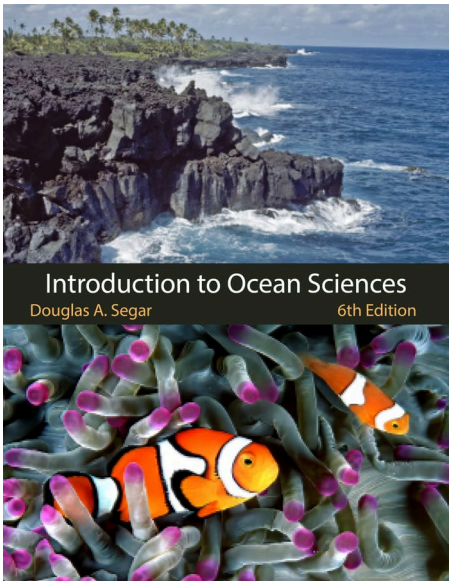
INTRODUCTION TO OCEAN SCIENCES 6E

Douglas A. Segar
American Meteorological Society

LibreTexts™


Introduction to Ocean Sciences
Douglas A. Segar
6th Edition

This text is disseminated via the Open Education Resource (OER) LibreTexts Project (https://LibreTexts.org) and like the thousands of other texts available within this powerful platform, it is freely available for reading, printing, and "consuming."

The LibreTexts mission is to bring together students, faculty, and scholars in a collaborative effort to provide an accessible, and comprehensive platform that empowers our community to develop, curate, adapt, and adopt openly licensed resources and technologies; through these efforts we can reduce the financial burden borne from traditional educational resource costs, ensuring education is more accessible for students and communities worldwide.

Most, but not all, pages in the library have licenses that may allow individuals to make changes, save, and print this book. Carefully consult the applicable license(s) before pursuing such effects. Instructors can adopt existing LibreTexts texts or Remix them to quickly build course-specific resources to meet the needs of their students. Unlike traditional textbooks, LibreTexts' web based origins allow powerful integration of advanced features and new technologies to support learning.

LibreTexts is the adaptable, user-friendly non-profit open education resource platform that educators trust for creating, customizing, and sharing accessible, interactive textbooks, adaptive homework, and ancillary materials. We collaborate with individuals and organizations to champion open education initiatives, support institutional publishing programs, drive curriculum development projects, and more.

The LibreTexts libraries are Powered by NICE CXone Expert and were supported by the Department of Education Open Textbook Pilot Project, the California Education Learning Lab, the UC Davis Office of the Provost, the UC Davis Library, the California State University Affordable Learning Solutions Program, and Merlot. This material is based upon work supported by the National Science Foundation under Grant No. 1246120, 1525057, and 1413739.

Any opinions, findings, and conclusions or recommendations expressed in this material are those of the author(s) and do not necessarily reflect the views of the National Science Foundation nor the US Department of Education.

Have questions or comments? For information about adoptions or adaptions contact info@LibreTexts.org or visit our main website at https://LibreTexts.org.

The LibreTexts name, logo, and book covers are trademarked by LibreTexts, Inc. (a 501(c)(3) not-for-profit organization) and protected against unauthorized use.

This text was compiled on 09/24/2026

Preface

Welcome to the American Meteorological Society's (AMS) *Introduction to Ocean Sciences*, 6th Edition. *Introduction to Ocean Sciences* explores the role of the ocean in the Earth system, with special emphasis on the flow and transformation of water and energy into and out of the ocean, the physical and chemical properties of ocean water, ocean circulation, marine life and habitats, interactions between the ocean and the other components of the Earth system (i.e., hydrosphere, atmosphere, lithosphere, and biosphere), and the human and societal impacts on and responses to those interactions. This book provides the physical, chemical, geological, and biological foundations for understanding Earth's ocean.

This sixth edition is a revision of *Introduction to Ocean Sciences*, 5th Edition, with updates to content and imagery. New, expanded, or significantly revised topics include metagenomics, the importance of microbes in the ocean, and recent climate science. It is accompanied by *AMS Ocean Sciences in Action*, a turnkey undergraduate-level curriculum package developed, licensed, and nationally implemented. These investigations engage students in observation, prediction, data analysis, inference, and critical thinking.

Introduction to Ocean Sciences covers the importance of ocean sciences (Chapter 1) and their history (Chapter 2). Ocean sciences are highly interdisciplinary, as demonstrated by the cross-chapter references throughout the text. Each chapter builds on the framework of Ocean Sciences subdiscipline by subdiscipline. Geological aspects of ocean sciences are covered in Chapters 4 and 6. Chemical aspects are covered in Chapter 5. Physical is covered in Chapters 8 through 11, and Marine biology is covered in Chapters 12 through 15. Chapter 16 brings these subdisciplines together to provide an overview of the direct anthropogenic influences on the ocean and a detailed discussion of how climate change, ocean acidification, and deoxygenation are altering the oceans that draws on the understanding of the interlinked ocean processes discussed in previous chapters.

Each chapter opens with a list of Critical Concepts. The Critical Concepts—including density, convection, and heat transfer—are essential to understanding the basic processes of the oceans. They are also concepts that students will need to understand before proceeding to more advanced science courses or making sense of many issues they will encounter in the future. Key terms are bolded in each chapter when they are first used and are defined in the glossary. Appendices cover unit conversions and milestones in the history of ocean science. The 6th Edition features many full-color photographs and line drawings that clarify the scientific concepts.

A note on the use of units in *Introduction to Ocean Sciences*: Generally, the International System of Units (abbreviated SI, for *Système International d'Unités*) is employed, with equivalent English or other units following in parentheses. Exceptions are units used by convention or convenience in oceanography. The equivalence between units is given in context, where general estimates are provided and approximate values are shown for all units. Conversion factors are in Appendix I.

Acknowledgments

Introduction to Ocean Science, 6th Edition, is the product of collaboration among many individuals with extensive scientific backgrounds and teaching experience. The lead author is Douglas A. Segar (Ph.D., Oceanography), an oceanographer with an extensive background in marine science, natural resource management, and environmental pollution research, marine conservation policy, and university instruction. Dr. Segar is joined by the managing editor Erin Smoak (M.S., Geology and M.Ed., Educational Technology Leadership), technical editor Elizabeth W. Mills (M.S., Meteorology), and copy editor Katie O'Neill (M.S., Environmental Science and Policy) from the AMS Education Program.

We are grateful to contributing authors Emily Brownlee, Ph.D. Biological Oceanography from Massachusetts Institute of Technology/Woods Hole Oceanographic Institution, currently an Associate Professor of Biology at St. Mary's College of Maryland; Michelle Valkanas, Ph.D., Biological Sciences from Duquesne University, currently an Assistant Professor of Biology at Pennsylvania Western University; and Elaine Stamman, M.S., Marine Biology from Texas A&M University, currently retired, who authored and edited sections of the text.

This page titled Preface is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

TABLE OF CONTENTS

Preface

Licensing

1: The Ocean Planet

1.1: The Ocean's and Earth's Environment

1.2: Oceans and the Origins of Life

1.3: How to Study Ocean Data

1.4: Chapter Summary

1.4.1: Key Terms

1.4.2: Chapter Questions

1.4.3: References

2: History and Importance of Ocean Studies

2.1: What is Oceanography

2.2: Exploration and Mapping

2.3: National and International Law Applied to Oceans

2.4: Ocean Resources

2.5: Chapter Summary

2.5.1: Key Terms

2.5.2: Chapter Questions

2.5.3: References

3: Studying the Oceans

3.1: Difficulties of Studying the Ocean Environment

3.2: The Unseen Domain

3.3: Bathymetry

3.4: Seafloor Sediments

3.5: Chemical and Physical Oceanography

3.6: Sampling Living Organisms in the Sea

3.7: Scuba, Manned, and Unmanned Submersibles

3.8: Satellites

3.9: Computers and Modeling

3.10: Chapter Summary

3.10.1: Key Terms

3.10.2: Chapter Questions

3.10.3: References

4: Plate Tectonics-Evolution of the Ocean Floor

4.1: Structure of the Earth

4.2: Plate Tectonics

4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.5: Divergent Plate Boundaries
4.6: Transform Faults and Fracture Zones
4.7: Hot Spots
4.8: Plate Interiors
4.9: Sea-Level Change and Climate
4.10: Present-Day Oceans
4.11: Chapter Summary
4.11.1: Key Terms
4.11.2: Chapter Questions
4.11.3: References

5: Water and Seawater
5.1: Origins and Distribution of the Earth's Water
5.2: The Water Molecule
5.3: The Dissolving Power of Water
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
5.5: Salinity
5.6: Dissolved Chemicals in Seawater
5.7: Heat Properties of Water
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
5.9: Ice Formation
5.10: Surface Tension and Viscosity
5.11: Transmission of Light and Other Electromagnetic Radiation
5.12: Transmission of Sound
5.13: Chapter Summary
5.13.1: Key Terms
5.13.2: Chapter Questions
5.13.3: References

6: Ocean Sediments
6.1: Sediments and Biogeochemical Cycles
6.2: Classification of Sediments
6.3: Lithogenous Sediments
6.4: Biogenous Sediments
6.5: Hydrogenous Sediments
6.6: Cosmogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments
6.9: Distribution of Surface Sediments

6.10: The Sediment Historical Record
6.11: Chapter Summary
6.11.1: Key Terms
6.11.2: Chapter Questions
6.11.3: References

7: Ocean-Atmosphere Interactions
7.1: The Atmosphere
7.2: Water Vapor in the Atmosphere
7.3: Water and Heat Budgets
7.4: Climatic Winds
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
7.7: Land-Ocean-Atmosphere Interactions
7.8: Global Climate Zones
7.9: Weather Systems
7.10: Local Weather Effects
7.11: Chapter Summary
7.11.1: Key Terms
7.11.2: Chapter Questions
7.11.3: References

8: Ocean Circulation
8.1: Energy Sources
8.2: Wind-Driven Currents
8.3: Ekman Motion
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
8.7: High-Latitude Surface Currents
8.8: Upwelling and Downwelling
8.9: Coastal Currents
8.10: Eddies
8.11: Inertial Currents
8.12: Langmuir Circulation
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
8.16: Chapter Summary
8.16.1: Key Terms
8.16.2: Chapter Questions

8.16.3: References
8.17: Box 1- Westward Intensification
8.18: Box 2 - Equatorial Currents

9: Waves
9.1: Complexity of Ocean Waves
9.2: Progressive Waves
9.3: What is Wave Motion?
9.4: Making Waves
9.5: Sea Development and Wave Height
9.6: Wave Dissipation
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
9.9: Tsunamis
9.10: Internal Waves
9.11: Rossby and Kelvin Waves
9.12: Standing Waves
9.13: Chapter Summary
9.13.1: Key Terms
9.13.2: Chapter Questions
9.13.3: References

10: Tides
10.1: Tide-Generating Forces
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
10.7: Tide Prediction
10.8: Marine Species and the Rhythm of Tides
10.9: Energy from Tides
10.10: Chapter Summary
10.10.1: Key Terms
10.10.2: Chapter Questions
10.10.3: References

11: Coasts
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures

11.5: Coral Reefs and Atolls
11.6: Wetlands
11.7: Deltas
11.8: Chapter Summary
11.8.1: Key Terms
11.8.2: Chapter Questions
11.8.3: References

12: Foundations of Life in the Oceans
12.1: How do we describe life?
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
12.4: Studying Microbes in the Ocean
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.7: Food Webs
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
12.10: Organic Carbon
12.11: Biological Provinces and Zones
12.12: Plankton
12.13: Nekton
12.14: Benthos
12.15: Chapter Summary
12.15.1: Key Terms
12.15.2: Chapter Questions
12.15.3: References

13: Coastal Oceans and Estuaries
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.5: Fisheries
13.6: Estuaries
13.7: Chapter Summary
13.7.1: Key Terms
13.7.2: Chapter Questions
13.7.3: References

14: Marine Ecology
14.1: Ecological Requirements

14.2: Habitats
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.7: Associations
14.8: Communication and Navigation
14.9: Chapter Summary
14.9.1: Key Terms
14.9.2: Chapter Questions
14.9.3: References

15: Ocean Ecosystems
15.1: Communities and Niches
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
15.6: Polar Regions
15.7: Beyond the Sun's Light
15.8: Hydrothermal Vents
15.9: Chapter Summary
15.9.1: Key Terms
15.9.2: Chapter Questions
15.9.3: References

16: Impacts of Humans on the Oceans
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.4: Waste Disposal
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
16.8: Chapter Summary
16.8.1: Key Terms
16.8.2: Chapter Questions
16.8.3: References
16.9: Box 1 - Petroleum in the Marine Environment
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.11: Box 3 - Radioactive Wastes in the Marine Environment

17: Critical Concepts

17.1: CC1 Density and Layering in Fluids

17.2: CC2 Isostasy, Eustasy, and Sea Level

17.3: CC3 Convection and Convection Cells

17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension

17.5: CC5 Transfer and Storage of Heat by Water

17.6: CC6 Salinity, Temperature, Pressure, and Water Density

17.7: CC7 Radioactivity and Age Dating

17.8: CC8 Residence Time

17.9: CC9 The Global Greenhouse Effect

17.10: CC10 Modeling

17.11: CC11 Chaos

17.12: CC12 The Coriolis Effect

17.13: CC13 Geostrophic Flow

17.14: CC14 Phototrophy, Light, and Nutrients

17.15: CC15 Food Chain Efficiency

17.16: CC16 Maximum Sustainable Yield

17.17: CC17 Species Diversity and Biodiversity

17.18: CC18 Toxicity

17.19: CC19 Metagenomics

17.20: References

Index

Glossary

Appendix 1 - Units and Converstion Factors

Appendix 2 - Dimensions of Earth and Ocean

Appendix 3 - Classification of Marine Organisms

Appendix 4 - Milestones in the History of Ocean Study

Detailed Licensing

Detailed Licensing

This page titled Introduction to Ocean Sciences 6e (Segar) is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

Licensing

Introduction to Ocean Sciences 6th Edition is under a CC BY-NC-ND 4.0 license

A detailed breakdown of this resource's licensing can be found in ***Back Matter/Detailed Licensing***.

This page titled Licensing is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1: THE OCEAN PLANET

Critical Concepts

- **CC5 Transfer and Storage of Heat by Water**: Water's heat properties are critical to maintaining a climate suitable for life as we know it.
- **CC9 The Global Greenhouse Effect**: Perhaps the greatest environmental challenge faced by humans is the prospect that major climate change may be an inevitable result of our burning of fossil fuels. The burning of fossil fuels releases carbon dioxide and other gases into the atmosphere, where they accumulate and act like the glass of a greenhouse retaining more of the sun's heat.
- **CC10 Modeling**: Complex environmental systems, including the oceans and atmosphere, can best be studied by using conceptual and mathematical models.
- **CC11 Chaos**: The nonlinear nature of many environmental interactions makes complex environmental systems behave in sometimes unpredictable ways. It also makes it possible for these changes to occur in rapid, unpredictable jumps from one set of conditions to a completely different set of conditions.

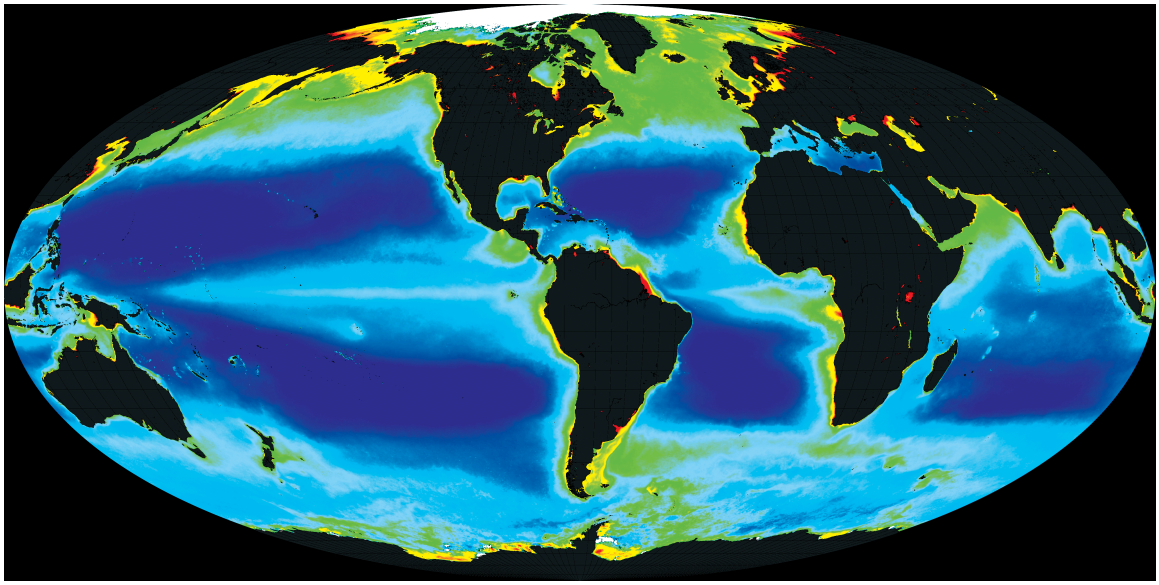


Figure 1-1. Why is this map of the world not drawn on a sphere or represented in a rectangular map shape such as we are used to? Why are the continents oddly misshapen? Why are the oceans those odd colors? Maps of the Earth are made on two-dimensional surfaces. Because they must represent the Earth's continents, which lie on a sphere, they distort the shapes and sizes of the Earth's features. This map is drawn to distort the Earth's surface features in a different way than is done in most of the maps that are familiar to us. Second, the strange colors that shade various areas of the oceans do not represent the color of the ocean surface itself, but instead represent biological productivity. The colors reveal details of the distribution of this property in the oceans.

About half of the world's human population lives within a few tens of kilometers of the oceans. A large proportion of the remainder live within a few tens of kilometers of a river or estuary that is connected to the oceans. These facts hint at, but do not fully reveal, the importance of the oceans to human civilizations.

In this text, we will explore the fundamental physical, chemical, geological, and biological features and processes of the oceans, why the oceans are difficult to study, and review the difficulties of studying the oceans and the methods ocean scientists have used. We will discuss the range of resources that the oceans provide us, and the impacts from human use of the oceans and the exploitation of their resources. We will also examine the direct and indirect effects of climate change, not just on human civilization and natural ecosystems but also on the fundamental chemistry of the global ocean. Some of these changes pose a long-term threat of extinction for many ocean species. After reading this text, you will have a new appreciation for the intimate and intricate linkages between our lives and the oceans. More importantly, you will be better prepared to analyze, understand, and assess the growing influx of scientific and popular reports on ocean-related findings, theories, and the impact of human activities.

In recent decades, human interest in the oceans has grown in response to our rapidly deepening knowledge and understanding of the creatures that inhabit the underwater world. Scuba diving has become a popular pastime, avocation, or sport, and all forms of ocean recreation have grown explosively. New and higher-quality videos are becoming available, allowing us to experience the ocean from the comfort of our homes. Many cities now have ocean aquariums that allow the public to see ocean life up close. A large portion of the planet's population experiences the ocean at beaches, when boating or enjoying other water sports, or by sailing the oceans on cruise ships. On a more somber note, humans are also fascinated and terrified by the death and destruction caused by tsunamis, hurricanes, severe storms, ships running aground or sinking, and sharks. These interactions with the ocean lead to such questions as:

- Why do some say that we know less about the oceans than we do about the moon or the planet Mars?
- Why are the world's active volcanoes located where they are?
- Why does the east coast of the United States have many barrier islands and tidal wetlands (marshes and mangrove swamps), whereas the west coast does not?
- What is a tsunami, and what should I do if I am ever on a beach when one is approaching?
- What is a rip current (rip tide), and how can I survive one?
- What is unique about water, and how do these unique properties affect daily life and our environment?
- What do we need to know about the oceans to understand and predict climate changes?
- Why are fisheries concentrated mostly in coastal zones, and why do some areas of the coastal ocean produce more fish than others?
- Almost all hurricanes in the Atlantic Ocean, especially large hurricanes causing severe damage, make landfall on the coast of the Caribbean, Gulf of America, or the Southeast United States. Only a very few weak hurricanes make landfall in the Northeast United States and hurricanes virtually never make landfall on the Atlantic Africa or Southern European coasts. Why is this?
- Is ocean pollution from oil spills and industrial discharges as bad as many reports often portray it?
- What are the main causes of the widely reported destruction of coral reefs and damage to other ocean ecosystems?
- What is ocean acidification, and why is it important?
- What is ocean deoxygenation? What are its causes and consequences?
- There have been at least 5 mass extinctions in Earth's 4.6-billion-year history. Why do many scientists now believe that the planet may already have entered a new mass extinction of species? What long-term changes in ocean physical characteristics and chemistry are expected to contribute to this mass extinction?

Answers to some of the questions listed above may be found in a single section or chapter of this text. However, the oceans are a complex environment in which geology, physics, chemistry, and biology are all linked in intricate ways. As a result, finding answers to many of these questions requires fitting together information from several different chapters. You will be reminded numerous times of the interdisciplinary nature of ocean sciences as you study this text.

One particular theme that exemplifies the interdisciplinary nature of ocean sciences is revisited many times throughout this book. This theme concerns what may be the most important scientific questions facing contemporary human society: Is human activity permanently altering the environment in a manner that will ultimately damage or diminish civilization itself? Specifically, will the release of gases from the combustion of **fossil fuels** alter the oceans and atmosphere and cause major changes in the global climate and ecosystems.

1.1: The Ocean's and Earth's Environment
1.2: Oceans and the Origins of Life
1.3: How to Study Ocean Data
1.4: Chapter Summary
1.4.1: Key Terms
1.4.2: Chapter Questions
1.4.3: References

This page titled 1: The Ocean Planet is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.1: The Ocean's and Earth's Environment

Only very recently have we come to realize that humans have already caused profound changes, not just in local environments but in the global ocean, atmosphere, and terrestrial environments as a whole. Too little is known about the global or regional consequences of environmental changes already caused by human activities. Even less is known about the future consequences if our civilization maintains its current exponential growth and development. The urgent need to assess the unknowns has been felt throughout the environmental science community. The oceanographic community has been particularly affected because most global environmental problems involve the oceans and ocean ecosystems, and our knowledge of the oceans is much poorer than our knowledge of the terrestrial realm. Changes in the marine environment caused by human activities are many and varied. They include many forms of pollution (**Chap. 16**) and physical changes in the coastal environment (**Chaps. 11, 13**). Many pieces of information must be obtained by different oceanographic disciplines before human impacts on the ocean can be identified, assessed, and effectively managed. Throughout this text, we refer to how oceanographic findings, principles, or studies can be applied to the practical problems of ocean management. The intent is to facilitate understanding of similar problems reported almost daily in the media.

Among contemporary environmental problems associated with human activities, three related impacts stand out as the most important and complex: global climate change due to the enhancement of the **greenhouse effect** (**CC9**), **acidification** of the oceans (**Chaps. 5, 13, 16**), and **deoxygenation** of the oceans (**Chaps. 12, 13, 16**). Each of these problems are caused largely by the release of large quantities of carbon dioxide to the atmosphere, primarily as a result of fossil fuel burning.

A greenhouse maintains a higher internal temperature compared to the outside temperature because the windows allow more solar energy into the greenhouse than they allow radiated out, warming the interior. The Earth's atmosphere acts in a similar way to control temperatures at the Earth's surface, with several gases, especially carbon dioxide, functioning like the glass in a greenhouse. The greenhouse effect is crucial for maintaining Earth's climate, but an increase in greenhouse gas concentrations increases its efficiency, leading to rising global temperatures. Since the Industrial Revolution, the burning of fossil fuels has steadily increased the concentration of carbon dioxide in the atmosphere, as shown in **Figure 1-2**. The concentrations of other greenhouse gases, such as methane and chlorofluorocarbons, are also increasing as a result of human activities. If there are no other changes to compensate for the consequent increase in greenhouse efficiency, the Earth's temperature will rise as concentrations of greenhouse gases rise.

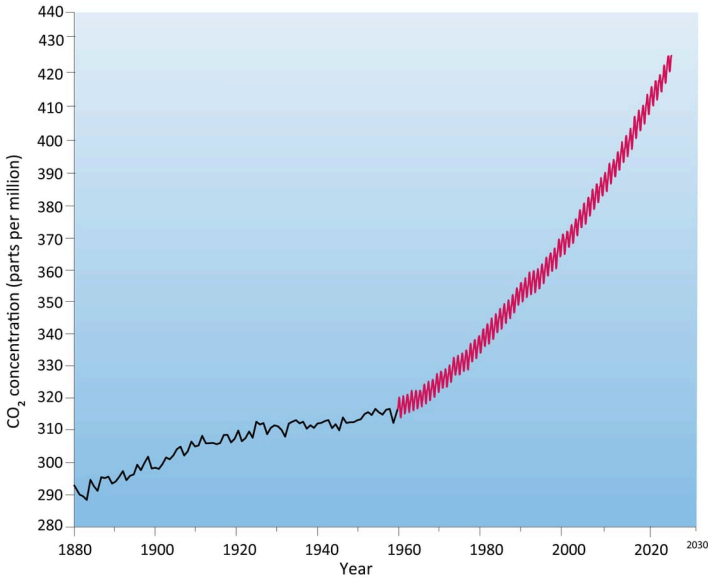


Figure 1-2. Change in the concentration of carbon dioxide in the atmosphere since 1880. Data for the smooth part of the curve were obtained from ice cores. Data from 1960 were obtained from direct measurements at Mauna Loa, Hawaii, and show both seasonal oscillations and a long-term upward trend. The upward trend continues. The seasonal peak in concentration exceeded 430 ppm for the first time in 2025, and the increase in concentration from May 2024 to May 2025 was 3.5 ppm.

The trend in atmospheric carbon dioxide concentration has accelerated from about 0.7 ppm per year in the 1960s to 2.6 ppm per year in the 2020s and is expected to continue increasing. The concentration is now more than 50% above the pre-industrial

concentration of about 280 ppm and has been consistently above 420 ppm since 2023. The data has not shown any significant slow down of the rate of increase due to the COVID-19 pandemic or the emissions reductions of Western nations, including the U.S.

Since 1950, Earth's temperature has risen 1–2 degrees Celsius and is projected to rise by an additional several degrees Celsius in the next two to three decades.

If the predicted increase in the global temperature does indeed occur, it will cause dramatic climate changes throughout the world. These changes are likely to be devastating to agriculture, the environment, and our entire civilization. In addition, if the predicted warming occurs, sea level will rise as a result of thermal expansion of the warmed ocean water and melting of ice from glaciers and polar ice sheets in Greenland and Antarctica. Some experts predict that sea level will rise a meter or more during the next several decades, inundating large areas of low-lying coastal land and entire low-lying island chains.

More than 30% of the carbon dioxide released by fossil fuel burning has been absorbed by the oceans. However, the added carbon dioxide reacts to form a weak acid. Consequently, although ocean waters are still weakly alkaline, with an average pH of approximately 8.1, the acidity of the oceans is rising. The acidity of a substance is measured by its **pH**, with lower pH values indicating more acidity. The oceans have been more acidic in the ancient past than they are today, but the rate at which acidification is currently taking place is believed to be many times faster than has ever occurred before. Because the rate of acidification is far too fast to allow many species to evolve and adapt, it is anticipated that ocean acidification will result in the extinction of many marine species and a drastic alteration of marine ecosystems. Species that are especially at risk include those that create hard parts (skeletons, shells, and other structural materials) of calcium carbonate. These include many of the most important groups of marine organisms, such as shellfish and, perhaps more critically, the pteropods that are the base of many marine food chains that provide food for fishes and marine mammals.

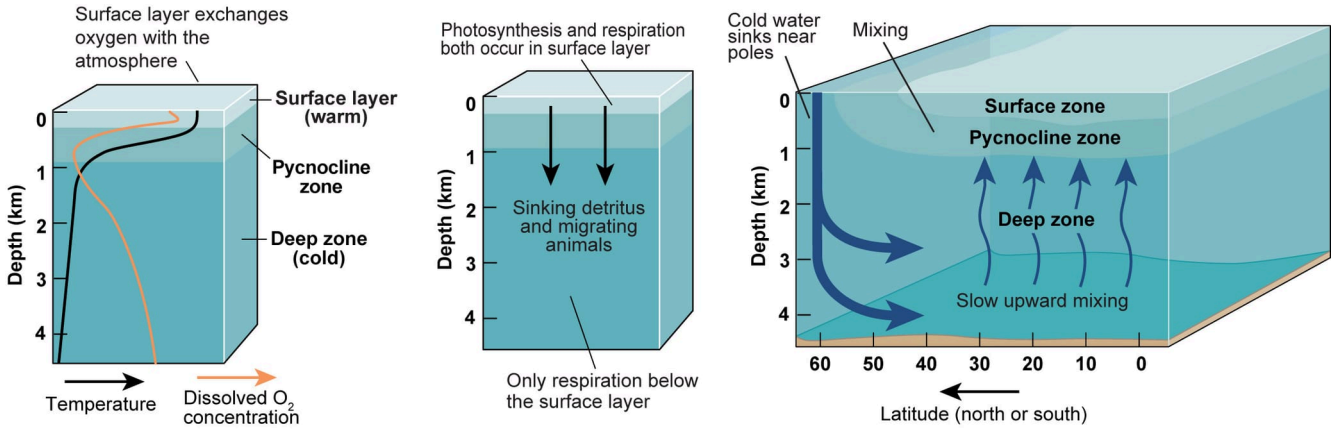
The effects of ocean acidification on marine ecosystems are already significant. For example, the acidity of coastal waters of the Pacific Northwest has risen in recent decades to a level that prevents oyster larvae from growing to maturity. This has significantly affected the large oyster industry in the region. Oysters must now be hatched and grown in remote locations such as Hawaii, where the water is less acidic, or in tanks with controlled pH water. Due to its cold water and high biological productivity, the Bering Sea off Alaska is the most acidic part of the world ocean, with pH values of 7.7 already observed there and continuing to decline. At these low pH values, there is strong evidence that the Alaska King crab population will decline and that populations of the small animals called pteropods, a principal food source for Bering Sea fishes, including salmon, will also decline since the water is too acidic for the pteropods to make their calcium carbonate shells.

Research suggests that future ocean acidification will cause major changes in the species composition of diverse marine food chains and is likely to severely damage coral reefs before the end of the century. However, some species that rely on calcium carbonate skeletal material do appear to be capable of adapting to this predicted increase in acidity levels. If we are going to mitigate or minimize the ecosystem effects of ocean acidification, we must not only reduce the amount of carbon dioxide we release to the atmosphere but actually remove much of the **anthropogenic** carbon dioxide already in the atmosphere.

Another consequence of anthropogenic releases of carbon dioxide is expected to be widespread deoxygenation of the oceans. This threat is caused by a combination of anthropogenic releases of carbon dioxide and civilization's releases to the oceans of large amounts of nitrogen and phosphorus, primarily from agriculture and in treated sewage discharges. This threat is a little complicated as several of the physical, chemical, and biological processes that are explained in later chapters of this book are involved.

For now, it is important to understand the following points that are illustrated in **Figure 1-3**. The oceans are essentially separated into a warm surface layer and colder deep water, and oxygen concentration in ocean water is controlled by both physical and biological processes. The surface layer is constantly mixed and in contact with the atmosphere, allowing it to exchange gases; therefore, its oxygen concentration is primarily determined by the atmospheric oxygen concentration. The much thicker deep-water layers below the warm surface layer are colder and mix with the surface layer only slowly (**Fig. 1-3a, CC1, Chaps. 5, 8**). In the deep-water layer, **respiration** (including microbial decomposition of **detritus**) removes oxygen from the water and replaces it with carbon dioxide (**Fig. 1-3b, Chaps. 5, 12**). If water in the deep layers of the oceans was not continuously replaced by oxygenated surface waters, the deep layers would lose all oxygen and would become void of all life, except for certain microbial species (**Chap. 12**). However, surface layer oxygen-rich water does sink at certain locations in high latitudes where it is cooled and becomes dense enough to sink. This water moves slowly through the deep oceans, continuously losing oxygen and gaining carbon dioxide due to respiration, then eventually mixes back into the surface layers to be re-oxygenated (**Fig. 1-3c, Chap. 8**). This circulation results in a vertical profile of dissolved oxygen concentration with high oxygen concentrations in the surface layer and in the deepest water layers with an oxygen minimum at depths in or just below the pycnocline zone (**Fig. 1-3, Chap. 8**). The concentration of dissolved oxygen at this oxygen minimum depth varies from location to location but, in some areas, is too low for

animal life to survive. Areas that exhibit such low oxygen concentrations are referred to as Oxygen Minimum Zones (OMZ). OMZs are usually found at depths of between about 200 and 1000 m (**Fig. 1-3**, **Chap. 8**), and at least some such oxygen-deficient areas are currently expanding.



(a) Layered structure of the oceans. Pycnocline zone is a transition zone where water temperature decreases and density increases rapidly with depth and vertical mixing is slow.

(b) Oxygen is used up by respiration in the pycnocline and deep zones. Rate of respiration is higher just below the pycnocline zone than at deeper depths.

(c) Ocean circulation carries oxygen rich surface water into deep layers. This oxygen rich water is slowly mixed back toward the surface layer while its oxygen concentration is progressively decreased by respiration.

Figure 1-3. Oxygen concentration in ocean water is controlled by both physical and biological processes. The oceans consist of two main layers, a warm surface layer and a cold deep layer, with a transition zone, called the pycnocline zone, between them. (a) Surface layer water is well mixed and in contact with the atmosphere, and its oxygen concentration is primarily controlled by gas exchange with the atmosphere. (b) Photosynthesis by marine life in this layer converts carbon dioxide to organic matter and releases oxygen, but this is offset by respiration, which converts oxygen to carbon dioxide. Respiration also occurs in the deep layer, fueled by animals that migrate vertically or live in the deep and decomposition of organic material (detritus), but there is no photosynthesis and no contact with the atmosphere, so respiration continuously consumes oxygen in this deep layer, and, therefore the longer the water remains in the deep layers, the less oxygen it contains. (c) The deep layer water is continuously replenished by sinking of cold water in some near-polar regions and then circulates through the oceans while being slowly mixed upward, eventually into the surface layer again. The pycnocline provides a barrier that inhibits mixing, so the upward mixing process is slow.

For many centuries, the balance between the supply of oxygen-rich water to the deep layers and the rate of respiration by organisms in the deep layer of the ocean has maintained enough dissolved oxygen for the organisms that live there. However, that has not always been true in several episodes of Earth's past when the deep ocean layers were devoid of oxygen. No animals or any other marine life, except for some species of microbes, can exist in water devoid of oxygen.

There are two ways that the deep oceans can lose all their oxygen and become **anoxic**. This can occur if the rate at which water circulates through the deep oceans is reduced, allowing a longer time for respiration to consume the oxygen, or if the rate at which respiration takes place in the deep oceans increases, or a combination of these. Respiration of living organisms in the deep layers of the oceans is fueled by the decomposition of organic matter (food) just as it is elsewhere. However, since no light penetrates down to the deep layer to support **phototrophy,** most of the food supply in the deep layers comes from the surface layer, as falling detritus or vertically migrating organisms (**Fig. 1-3b**). This food supply is limited, which limits the total respiration and, therefore, oxygen loss as water flows through the deep oceans. If the supply of food from the surface layer to the deep ocean water is increased, the additional food supports more deep-layer marine life, which means more respiration, and the oxygen concentrations in deep water decline.

The Earth has experienced at least 5 known mass extinctions of species, including the one that occurred when the dinosaurs became extinct (**Fig. 1-4**). While deoxygenation of the oceans is thought not to have been the sole cause of these extinctions, both acidification and deoxygenation did occur during most of these extinctions and are thought to have contributed to the extinction.

Now, we can consider why scientists believe that deoxygenation of the oceans is a major long-term threat. First, less oxygen can dissolve in seawater at higher temperatures, so climate warming will reduce oxygen concentrations in the surface layers of the oceans that are in contact with the atmosphere. Second, warming of the oceans is likely to enhance the temperature differences between deep ocean layers and surface layers, which will inhibit vertical mixing of the two. Warming of ocean surface waters is also expected to reduce the rate at which cold surface waters near the poles sink into the depths. Each of these effects will lengthen the time that deep ocean water spends out of contact with the atmosphere before returning to the surface layers to exchange gases and re-equilibrate with the atmosphere. Third, warmer ocean water is expected to increase the rate of production

of food by phototrophy in the ocean's surface layer, which will increase the supply of detritus to the deep layer, accelerating oxygen consumption. Current research indicates that each of these three changes have already begun to occur. Moreover, **photosynthetic** food production in most of the oceans is limited by the low concentrations of nutrients, especially nitrogen and phosphorus. Humans release large amounts of these nutrients, especially in agricultural runoff and treated sewage wastes. This is expected to enhance the rate of photosynthesis and production of organic matter, which will increase the production of detritus and the amount of detritus transported below the surface layer. Decomposition of this detritus is expected to increase, which would increase the rate of oxygen loss in the deep ocean layers.

Our understanding does not yet allow us to assess the future of ocean acidification and deoxygenation in the global ocean with any certainty. However, we do have observations of the ecological damage that they can cause in limited regions. For example, as mentioned earlier, acidification of the coastal waters off the west coast of North America has already been observed to have exceeded the ability of oysters to reproduce successfully (**Chaps. 5, 16**). Also, there are numerous coastal and estuarine areas, including the Chesapeake Bay and the coastal waters west of the Mississippi Delta, that experience low or totally depleted oxygen in their deeper water layers (below a shallow **pycnocline**), either periodically or permanently, due primarily to excess anthropogenic nutrients. These areas have been called "dead zones" because living organisms that are unable to move out of these areas as they develop die due to a lack of oxygen (**Chaps. 13, 16**). Also, water from an OMZ in the North Pacific Ocean has periodically encroached on the Oregon continental shelf, causing a periodic "dead zone" to appear.

The oceans and atmosphere act together as a complex system that regulates our climate and the concentrations of carbon dioxide and other gases in both the oceans and atmosphere. An essential role of the oceans is to capture, store, and redistribute the sun's heat energy to the atmosphere (**CC5**), so the oceans are integral to weather and climate. Within the ocean-atmosphere system, numerous complicated changes and feedbacks occur as a result of increases in atmospheric concentrations of carbon dioxide and other gases. Some changes add to the predicted global warming, whereas others reduce or even negate it. This is also true for ocean acidification and deoxygenation. However, our current understanding of the ocean-atmosphere system is poor, especially those segments of the system that are associated with ocean processes. To predict Earth's future with less uncertainty so that we can take appropriate actions, we must improve our understanding of the oceans.

In addition to studying contemporary ocean processes, oceanographers study the oceans to uncover important information about past changes in the world's climate. Such historical information, found primarily in ocean seafloor **sediment** and sedimentary rocks, can reveal how the Earth's climate has changed over tens of millions of years and help us assess how it might change in the future, either as a result of natural changes or as a result of the enhanced greenhouse effect.

Climate change, ocean acidification, and deoxygenation are complex processes involving many concepts referenced in the preceding paragraphs. You might ask why they are summarized in this introductory chapter. The reason is simple. Each chapter contains information needed to fully understand how these processes work. If climate change, acidification, and deoxygenation continue unchecked, the ultimate likely consequences include global mass species extinction, a far greater threat than the damage that may be caused by any other environmental issue. Keep this in mind as you learn about the oceans, and you will realize that what you are learning is not just science but information about things that are vital to your life and the lives of all other humans and living creatures. We will return to the subjects of ocean acidification and deoxygenation in **Chapter 16**, when we discuss human impacts on the oceans. By then, you will have learned about the fundamental physical, chemical, and biological processes operating in the oceans and will be able to much better understand how acidification and deoxygenation are developing and that they represent a major long-term threat to both ocean ecosystems and human civilization.

Many important decisions must be made by current and future generations about human use and the protection of our environment. It is important for all of us to participate in these decisions with an understanding of the complexity of the ocean-atmosphere system and the uncertainties inherent in all scientific studies and predictions of such complex environmental systems (**CC10, CC11**). We must also recognize that science cannot provide definitive answers to even the most intensively studied environmental problems, as all environmental systems include non-linear relationships, making them chaotic. This means that predictions based on even the best science and modeling will always have high uncertainty.

This page titled 1.1: The Ocean's and Earth's Environment is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.2: Oceans and the Origins of Life

To investigate the origins of life, a number of laboratory studies have attempted to reproduce the conditions thought to exist in the early oceans and atmosphere, including high ultraviolet radiation levels and frequent lightning discharges. Under these conditions, a wide range of organic compounds are created from mixtures of carbon dioxide, methane, ammonia, hydrogen, and water. The organic compounds formed in these experiments included amino acids, which are considered the most important building blocks of the complex molecules needed for life to exist. However, we do not understand how or know where these building blocks were assembled into the much more complex molecules that were needed to form the first living matter. Several hypotheses have been proposed. The chemical reactions necessary to construct complex molecules may have taken place on the surface of solid particles, deep within the Earth, or at hydrothermal vents (**Chap. 15**). Another possible explanation for the origin of these complex compounds is that they first reached the Earth in meteorites.

All life as we know it depends on the transport of chemical elements and compounds within cells and between cells and the surrounding environment. In all known living organisms, from **archaea** to mammals, these chemical substances are transported dissolved in water. Therefore, water is essential to all life on Earth. As a result, it is not surprising that life appears to have been nurtured and developed in the oceans for most of the billions of years of Earth's history. **Figure 1-4** shows the 4.6 billion years of Earth's history compared with a year with significant dates including the formation of the oceans, the first primitive animals, evolution of humans, and the 5 well-known mass extinctions.

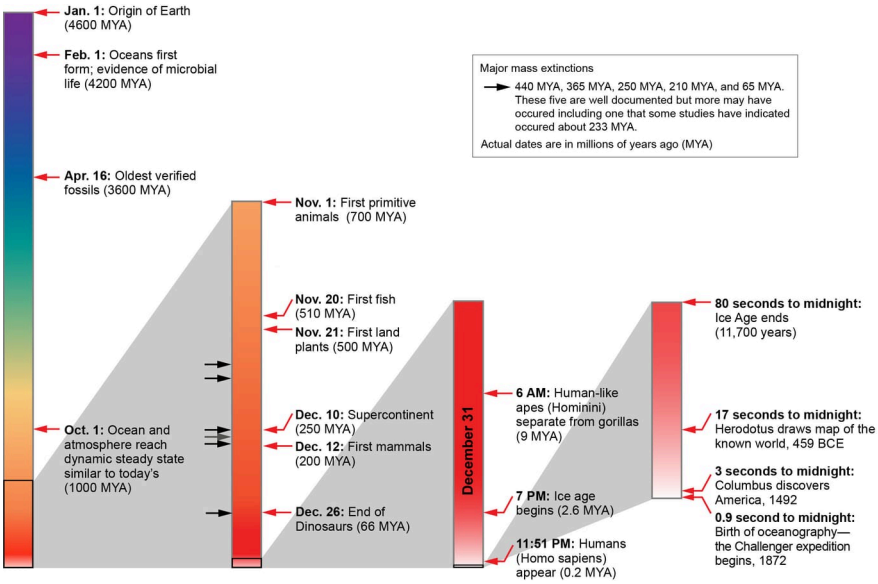


Figure 1-4. The history of the Earth, shown as if it were 1 year. The Earth formed about 4.6 billion years ago. Animals have existed on the Earth for only an extremely brief period of its history. If the age of the Earth were only 1 year, animals would have existed for only 2 months, our species Homo sapiens for about 9 minutes, and the science of oceanography for less than a single second. Most actual dates on the figure are expressed as millions of years ago (MYA) and are approximate.

Evidence from the fossil record indicates that life has existed on the Earth for a very long time. Although there is evidence that life began much earlier, the first life forms for which we have direct evidence were bacteria-like microorganisms that existed about 3600 million years ago (MYA), approximately 1 billion years after the Earth was formed (**Fig. 1-4**). When the oceans were first formed about 4200 MYA, the atmosphere consisted mostly of nitrogen, with smaller amounts of carbon dioxide and methane. Because there was no free oxygen in the atmosphere or oceans, it is thought that some of the earliest life forms were similar to the **chemosynthetic** microbes that are currently found in many isolated environments such as hydrothermal vents and deep within the Earth's **crust** (**CC14, Chap. 15**). These environments contain no free oxygen and are often characterized by conditions of extreme temperature and pressure, similar to the conditions that are thought to have existed in the oceans and within Earth's crust when the oceans were first formed.

If the first organisms did utilize chemical energy to support their energy needs, then at some unknown time the first species of microorganisms that utilized the sun's light energy developed. These organisms were **phototrophic** (**CC14**). Early phototrophic microorganisms probably used hydrogen sulfide as their source of hydrogen, which is needed for building chemical compounds.

Later, photosynthetic microorganisms, likely cyanobacteria, developed the ability to split water molecules, using the hydrogen to build their chemical compounds, releasing oxygen in the process. Over a period of time, between about 2000 and 1000 MYA, these photosynthetic organisms changed the composition of the Earth's atmosphere. Oxygen concentrations steadily increased until they reached approximately the 20% present in the current atmosphere (**Fig. 1-4**). This change was fundamental to the development of the majority of living species now on the Earth. As the free oxygen accumulated in the atmosphere, it reacted with sulfides and other chemical compounds in the oceans that supported the life cycles of chemosynthetic species. As a result, chemosynthetic species largely disappeared or became restricted to extreme environments, such as those where they are found today. Furthermore, the free oxygen permitted the development of the animal kingdom. All species of animals depend on respiration using oxygen obtained from their environment.

The oceans and atmosphere reached a **steady state** (maintaining approximately the same chemical compositions that they have today) approximately 1000 MYA, and this relative stability permitted the development of more complex life forms. The first primitive higher animals were marine **invertebrates**, perhaps similar to sponges and **jellies** (**jellyfish**), that developed about 700 MYA, followed by the first fishes about 500 MYA (**Fig. 1-4**). The first plants appeared on land about 500 MYA, and the first mammals only about 220 MYA. Humans are latecomers. Hominids (humanlike species) have existed for only about 4 million years—less than one-tenth of 1% of the history of the Earth. In the 1-year model, this is less than 9 minutes (**Fig. 1-4**).

Although the specific environment in which life first developed is not known, the oceans clearly played a major role in nurturing the development of life on the Earth.

This page titled 1.2: Oceans and the Origins of Life is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.3: How to Study Ocean Data

Before we begin to explore the intricacies of the relationships among humans, the oceans, and the environment, it is important to review some of the basic tools used to describe the Earth and its oceans. These include various forms of maps, graphical representations of data that vary geographically or with time, graphs that can show the relationship between two variables, and standard scientific notation for numbers and units of measurement.

As is true for other sciences, describing what we know about the oceans and ocean processes requires the use of graphs and similar diagrams, and sometimes mathematics. For ocean sciences, we must also be able to represent data in a geographic context, which requires the use of maps or **charts**. Graphs, diagrams, and maps make it easy to understand certain features of even complex data sets without using mathematics (a goal we have set for this text). However, these visual representations can be properly interpreted only if we understand how a particular graph or map presents, and often distorts, the data. To make it possible for you to properly study and understand the rest of this text, the remainder of this chapter discusses some of the characteristics of graphs and maps that are used extensively in this text. Even if you are familiar with scientific diagrams, you are strongly urged to review this material.

Graphs

Graphs are probably the most widely used way to present data in science and elsewhere. They provide a means of visualizing relationships between two or more variables or properties, but they can be extremely misleading unless they are read properly. To understand a graph, you must not only look at the general shape of the line or curve connecting points or the apparent difference in sizes of bars in a bar graph, for example, but you must also carefully examine the axes. Two simple examples in **Figure 1-5** illustrate why.

In **Figure 1-5**, three simple values that could be, for example, the prices of an item at different stores or the depths of the thermocline at three locations in the ocean are plotted in a simple bar graph. The same data are plotted in both parts of the figure. Why do the plots look so different? In **Figure 1-5a**, the y-axis (vertical axis) extends from 0 to 250. In **Figure 1-5b**, the data are “expanded” by showing only 228 to 240 on the y-axis. Items 1, 2, and 3 have nearly the same values, but one plot seems to show that item 3 has a much higher value than items 1 and 2. Have you ever seen this technique used in advertisements or news reports to present data differences in a misleading way?

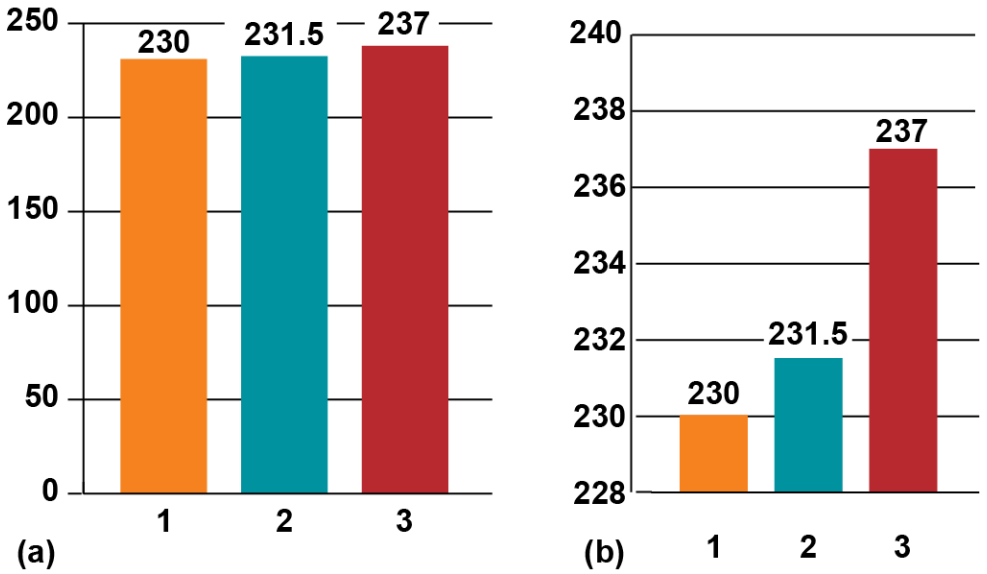


Figure 1-5. Two bar graphs that show the same data but use different y-axis ranges. They give very different impressions about the differences among the three values. In part (b), the true nature of the differences can be found only by carefully examining the scale on which the data is plotted.

The second example is a simple nonlinear (**CC10**) relationship between two variables (x^2= y). These variables could be, for

example, light intensity and depth, or primary production rate and nutrient concentrations (although the relationships between these pairs of parameters are actually more complicated). Notice that the first plot (**Fig. 1-6a**) shows that y increases more rapidly as x increases, which is exactly what we would expect from the equation. However, the same data in **Figure 1-6b** appear to show the exact opposite (y increases more slowly as x increases). The reason for this difference is that the y-values are plotted as the log of the value, not the value itself. Log plots are commonly used for scientific data but may be confusing unless you examine the axes carefully. One of the reasons why log plots are used is illustrated in **Figure 1-6c**. This figure is a plot of the same data as in **Figure 1-6a** and **Figure 1-6b**, but it plots the log value for both variables. The plot is a straight line, revealing a feature of this nonlinear relationship that is often important to scientists.

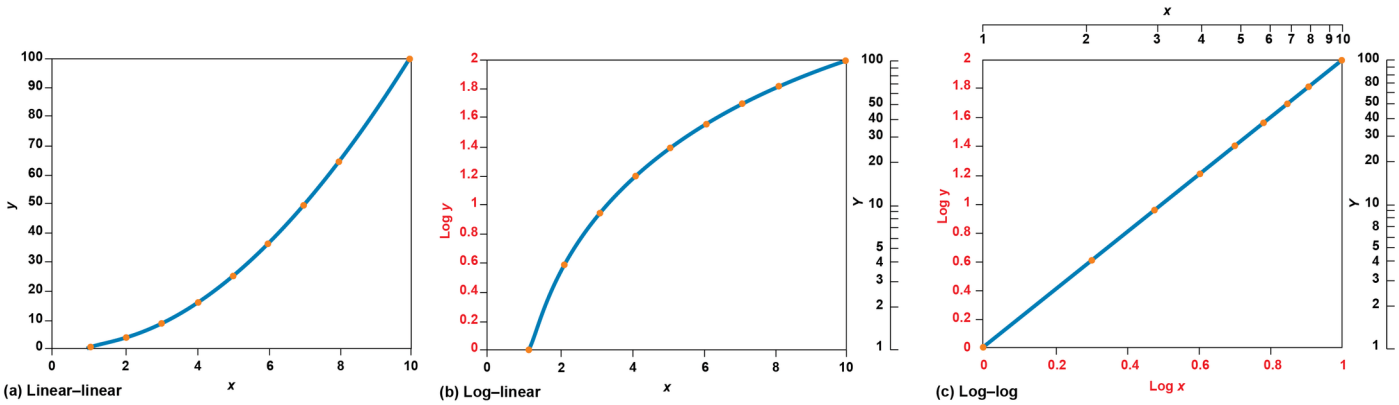


Figure 1-6. A simple nonlinear relationship is plotted in three different ways to show how logarithmic scales can change the apparent nature of the relationship between the two variables: (a) a linear–linear plot, (b) a log–linear plot, (c) a log–log plot. Notice that in parts (b) and (c), the scale on the log axis can be expressed as the log of the number (left side) or the number on a log scale (right side). This shows how important it is to analyze every part of a figure to distinguish the details like those seen in **Figure 1-5**.

Contour Plots and Profiles

Contour plots are used to display the two-dimensional spatial distribution of a variable, such as atmospheric pressure on weather maps. The variable plotted can be any parameter, such as pressure, temperature, soil moisture, vegetation type, or concentration of a substance.

Contour plots show how variables are distributed across flat surfaces. In some cases, the variable being plotted represents a measurement in the third dimension—perpendicular to the surface—such as depth or height. The most common examples are **topographic** maps that depict the height of the land surface above sea level and **bathymetric** maps that depict the depth of the ocean below sea level. Sea level is used as the surface on which the third dimension variable, the height or depth, is plotted. Contours of pressure are usually called **isobars**, temperature contours are **isotherms**, density contours are **isopycnals**, and depth contours are called isobaths. The prefix “iso” means “equal” or “the same,” so these lines connect locations with the same value of the variable, with each line representing a different value. Often, but not always, the values contoured are spaced at equal intervals of the variable, such as 50 m, 100 m, 150 m, and 200 m above sea level for a topographic map. One very important rule of contouring is that contours cannot cross or merge with each other. This is because each contour line always follows its own value. If the plotted parameter is elevated or depressed (forming hills or depressions) in two or more areas, contours around any two such features do not connect with each other. Instead, they form closed loops unless they are at the edge of the plot (**Fig. 1-7**). Contour lines show the distribution and magnitude of highs and lows of the plotted parameter. In any particular contour plot, high and low features (hills and valleys in a topographic map) that have more contours surrounding them are of greater magnitude than those with fewer surrounding contours. For example, higher hills are shown with more contour lines between sea level and the summit than lower hills. Contours also reveal the strength of gradients at different points on the plotted surface. Where the contours are closely bunched together, the value of the parameter changes quickly with distance across the surface, and the gradient is relatively strong (steep terrain on a topographic map). Conversely, where the contours are further apart, the gradient is relatively weak. The term relatively is important because the number of contours and the intervals between the contours can be different in different plots; even plots of the same data, which can be seen by comparing **Figure 1-7a** and **Figure 1-7b**.

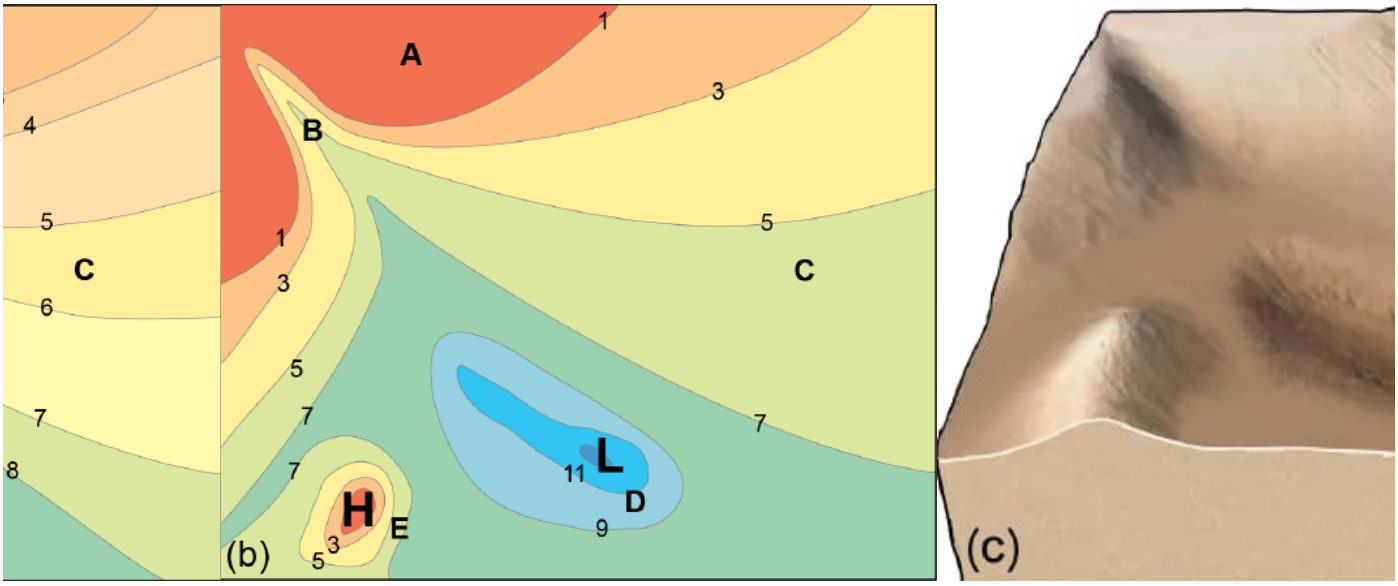


Figure 1-7. Contour plots reveal areas of higher and lower values of the plotted parameter, identified as H and L. They also indicate the strength of the gradient, which is stronger where adjacent contours are closer to each other. (a) In this example, which is a topographic map of the seafloor, there is a flat plateau at point A, a steep-sided valley at point B, an area of gentle slope at point C, a depression at point D, and a hill or mountain at point E, which has a very steep slope, especially on its right side. (b) Here, the same data are plotted with a greater interval between contours. Unless you examine it carefully, this plot can give the appearance that gradients are not as steep as they appear in part (a). (c) A three-dimensional representation of the topography plotted in parts (a) and (b).

When you study contour maps, make sure you look at the values on each of the contour lines to determine the value assigned to each contour. This will enable you to see which features are highs and lows and to assess the relative magnitude of gradients at different points on the plot accurately. Many contour plots are now produced by computers, and in some of these, contour lines are not used. Instead, the entire plot is filled with color that varies according to the value of the parameter plotted. Usually, the order of colors in the spectrum of visible light is used. Red represents the highest value, grading progressively through orange, yellow, green, and blue to violet, which represents the lowest value (**Fig. 1-7**). These plots are essentially contour plots that have an infinitely large number of contour lines and in which the range of values between each pair of adjacent contours is represented by a slightly different shade of color. **Figure 3-4**, **Figure 7-12**, and **Figure 8-24** are additional examples of this convention. This color convention, from red to violet, is now universally accepted and makes it easy to visually identify the areas of high and low values. Consequently, with a few exceptions made for specific reasons, we have used this convention to color-code the regions between contours in all of the contour plots in this text.

Cross-sectional profiles show the distributions of properties on slices (cross sections) through, for example, the Earth or an ocean. Oceanographers commonly use cross-sectional profiles with contour plots to show the distribution of properties with depth and distance across the vertical slice. Because the oceans and the Earth's crust are extremely thin in comparison with the widths of the oceans and continents, these profiles almost always have a large vertical exaggeration. For example, the maximum depth of the oceans is approximately 11 km, whereas their widths are several thousand kilometers. If we were to draw a vertical profile across the ocean with no vertical exaggeration on a page in this book and use an ocean depth of 11 km scaled to a height of 11 cm on the vertical axis, then the profile would have a width of 1000 cm (more than 50 page widths) for each 1000 km of ocean width plotted. Because this is obviously not possible, the scale on which distance across the ocean is plotted is reduced. This decreases the width of the plot but also introduces vertical exaggeration, which distorts the data.

Figure 1-8 shows the effects of 10 and 100 times vertical exaggerations of a topographic profile. Most profiles used in ocean sciences have exaggerations greater than 100 times. Vertically exaggerated profiles make the gradients of topographic features appear much greater than they really are. Two points are important to remember as you examine vertically exaggerated profiles in this textbook. First, the seafloor topography is much smoother, and the slopes are much shallower than depicted in the profiles. Second, ocean water masses are arranged in a series of layers that are very thin in comparison with their geographic extent, and this is not adequately conveyed by the vertically exaggerated cross-sectional profiles.

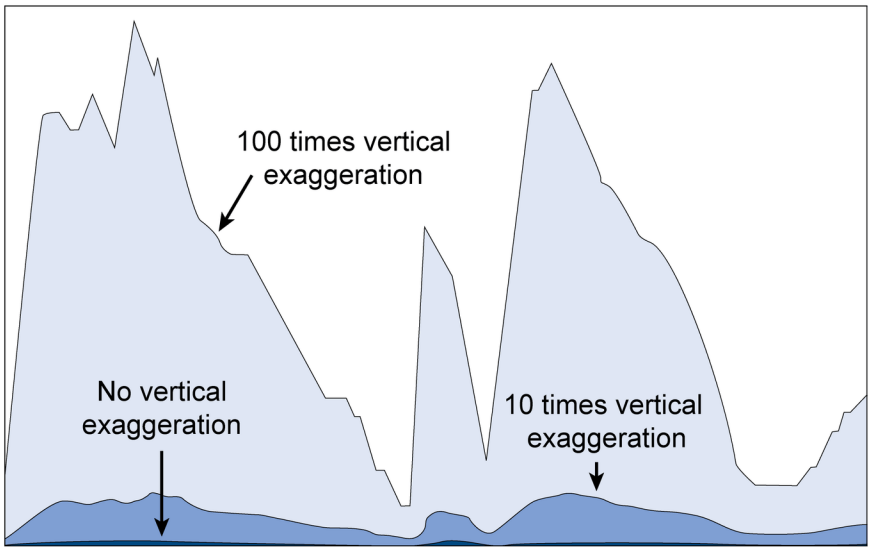


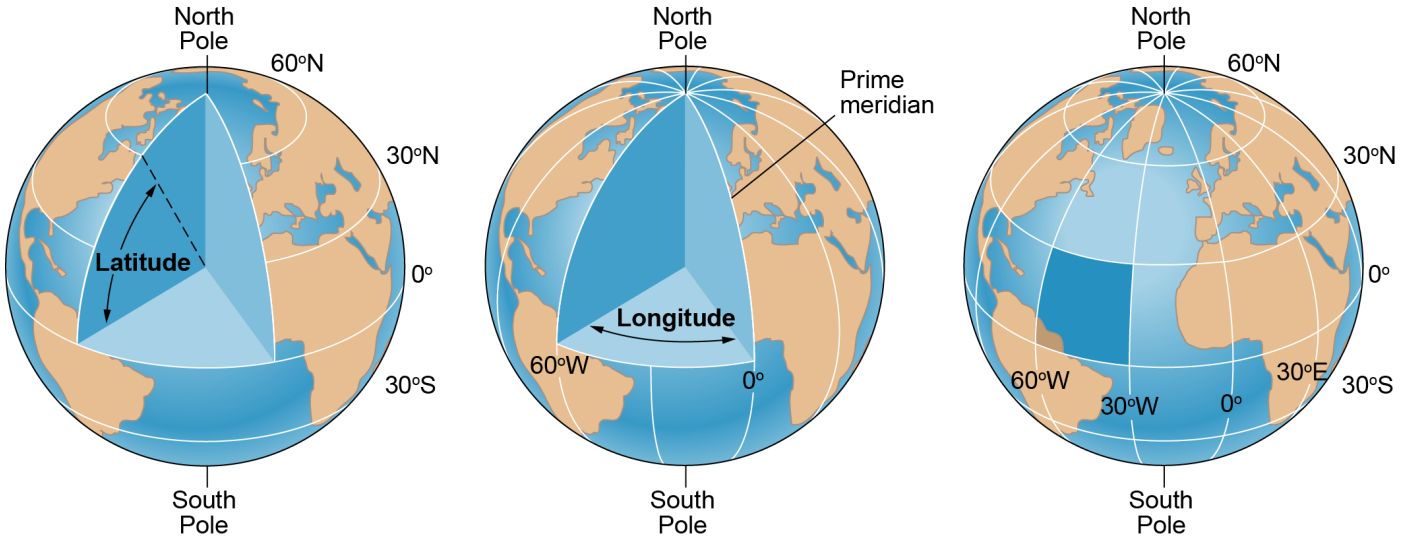
Figure 1-8. Cross-sectional vertical profiles used to depict ocean or atmospheric data are almost always vertically exaggerated. The effect of vertical exaggeration is shown in this figure. If the dark blue area below the lowest line in this graph were a plot of seafloor depth or land surface elevation, we would conclude that the topography was relatively flat. However, when vertically exaggerated 10 times (the mid blue plot area), it appears to be much more rough. When exaggerated 100 times (the light blue plot area), the topography looks like extremely rugged mountain chains with very steep slopes. Cross-sectional profiles in this book are plotted with vertical exaggerations greater than 100 times. This distortion cannot be avoided and must be considered when interpreting the data these cross sections depict.

Maps and Charts

The Earth is spherical, and the ideal way to represent its geographic features would be to depict them on the surface of a globe. Although virtual-reality computer displays make such representation possible, it is impractical for most purposes. Consequently, geographic features of the Earth's surface are almost always represented on two-dimensional maps or charts. To represent the spherical surface of the Earth on a two-dimensional map, a projection must be used. A projection is a set of rules for drawing maps of Earth's features on a 2-dimensional screen or a piece of paper. All projections distort geographic information, each in a different way. The projection that most people are familiar with is the Mercator projection, which was designed to be used for navigation. However, several different projections are used in this textbook, and you may see other projections elsewhere.

To draw accurate maps of the Earth, we must relate each location on the Earth's surface to a particular location on the flat map's surface. Therefore, we must be able to identify each point on the Earth's surface by its own unique "address." **Latitude** and **longitude** are used for this purpose.

In a city, the starting "address" is usually a specific point downtown, and the street numbers and house numbers on each street increase with distance from that point. On the Earth's surface, there is no center from which to start. However, two points, the North and South Poles, are fixed (or almost so), and the equator can be easily defined as the circle around the Earth equidistant from the two poles. This is the basis for latitude. The equator is at latitude 0°, the North Pole is 90°N, and the South Pole is 90°S. Every location on the Earth other than at the poles or on the equator has a latitude between either 0° and 90°N or 0° and 90°S. Why do we use degrees (°)? **Figure 1-9a** shows that if we draw a circle around the Earth parallel to the equator, the angle between a line from any point on this circle to the Earth's center (dashed line) and a line from the Earth's center to the equator is always the same. If the circle is at the equator, this angle is 0°; if it is at the pole, the circle becomes a point, and this angle is 90°. If the circle is not at the equator or one of the poles, this angle is between 0° and 90° and is either south or north of the equator.



(a) Latitude **(b)** Longitude **(c)** Latitude and longitude

Figure 1-9. Latitude and longitude are measured as angles between lines drawn from the center of the Earth to the surface. (a) Latitude is measured as the angle between a line from the Earth's center to the equator and a line from the Earth's center to the measurement point. (b) Longitude is measured as the angle between a line from the Earth's center to the measurement point and a line from the Earth's center to the prime (or Greenwich) meridian, which is a line drawn from the North Pole to the South Pole passing through Greenwich, England. (c) Lines of latitude are always the same distance apart, whereas the distance between two lines of longitude varies with latitude.

Latitude can be measured without modern instruments. Polaris, the Pole Star or North Star, is located exactly over the North Pole, so at 90°N, it is directly overhead. At lower latitudes, Polaris appears lower in the sky until, at the equator, it is on the horizon (at an angle of 90° to a vertical line). In the Northern Hemisphere, we can determine latitude by measuring the angle of the Pole Star to the horizon. In the Southern Hemisphere, no star is directly overhead at the South Pole, but other nearby stars can be used and a correction made to determine latitude. Star angles can be measured accurately with very simple equipment, and the best ancient navigators were able to measure these angles with reasonable accuracy without instruments.

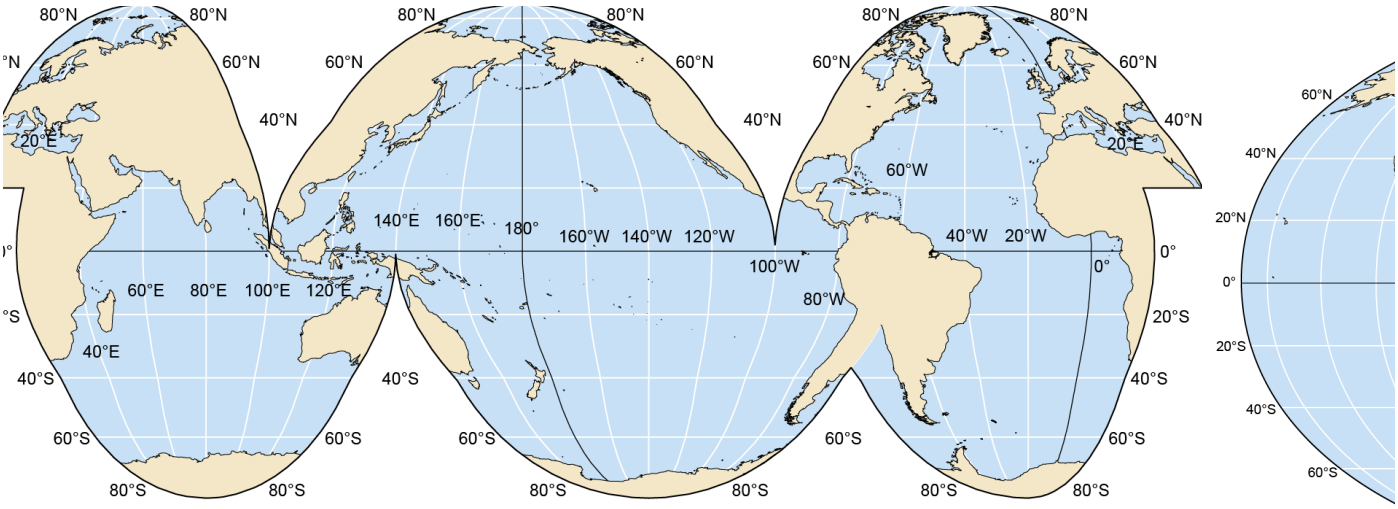
Latitude is a partial address of a location on the Earth. It specifies only the hemisphere, northern or southern, in which the location lies and that the location is somewhere on a circle (line of latitude) drawn around the Earth at a specific distance from the pole (equivalent to "somewhere on 32nd Street"). The other part of the address is the longitude. **Figure 1-9b** shows that the relative locations of two points on a line of latitude can be defined by measurement of the angle between lines drawn from the Earth's center to each of the locations. This angle is longitude, but it has no obvious starting location. Consequently, a somewhat arbitrarily chosen starting location, the line of longitude (north-south line running between the North and South Poles) through Greenwich, England, has been agreed upon as 0° longitude and is known as the "prime meridian."

Longitude is measured in degrees east or west of the prime meridian (**Fig. 1-9b**). Locations on the side of the Earth exactly opposite the prime meridian can be designated as either 180°E or 180°W. All other locations are between 0° and 180°E or between 0° and 180°W. Longitude is not as easy to measure as latitude because there is no fixed reference starting point, and no star remains overhead at any longitude as the Earth spins. The only way to measure longitude is to accurately fix the time difference between noon, when the sun is directly overhead, at the measurement location and noon at Greenwich. The Earth rotates through 360° in 24 hours, so a 1-hour time difference indicates a 15° difference in longitude. To determine longitude, the exact time (not just the time relative to the sun) must be known both at Greenwich and at the measurement location. Before radio was invented, the only way to determine the time difference was to set the time on an accurate clock at Greenwich and then carry this clock to the measurement location. Consequently, longitude could not be measured accurately until the invention of the **chronometer** in the 1760s.

Latitude and longitude lines provide a grid system that specifies any location on the Earth with its own address and enables us to draw maps. Before we look at these maps, notice that 1° of latitude is always the same length (distance) wherever we are on the Earth, but the distance between lines of longitude is at a maximum at the equator and decreases to zero at the poles (**Fig. 1-9c**).

Figure 1-10a is a familiar map of the world. This representation of the Earth's features is a Mercator projection, which is used for most maps. In the Mercator projection, the lines of latitude and longitude are drawn as a rectangular grid. This grid distorts relative distances and areas on the Earth's surface. The reason is that, on the Earth, the distance between two lines of longitude varies with latitude as shown in **Figure 1-10c**, but the Mercator projection shows this distance as the same at all latitudes.

Furthermore, the Mercator projection shows the distance between lines of latitude to be greater at high latitudes than near the equator, even though on the Earth's surface, they are the same.



oode's Interrupted projection **(c)** Robinson projec

Figure 1-10. Typical map projections. (a) Mercator projection. Higher latitudes near 90°N and 90°S are not shown because the shape and area distortions introduced by this projection increase rapidly with latitude near the poles. (b) Goode's interrupted projection preserves relative areas and shows each of the oceans without interruption. However, it distorts the shapes of the continents. (c) Robinson projection. This projection preserves none of the four desirable characteristics perfectly, but it is a good approximation for many purposes.

Why, then, has the Mercator projection been used for so long? The answer is that it preserves one characteristic that is important to travelers: on this projection, the angle between any two points and a north-south line can be used as a constant compass heading to travel between the points. However, the constant-compass-heading path is not the shortest distance between the two points. The shortest distance is a great-circle route. A great circle is any circle around the full circumference of the Earth. For navigation, this great circle passes through both the starting and end point of the journey. The Mercator projection suggests that the "direct" route between San Francisco and Tokyo is almost directly east to west. However, the normal flight path for this route is almost a great-circle route that passes very close to Alaska and the Aleutian Islands. You can see the great-circle route and why it is the shortest distance by stretching a piece of string between these two cities on a globe. Over short distances, the compass-direction route and the great-circle route are not substantially different in terms of distance, and the simplicity of using a single, unchanging compass heading makes navigation easy. Ships and planes now use computers to navigate over great-circle routes that require them to fly or sail on a continuously changing compass heading.

Although the Mercator projection is very widely used, other projections preserve different characteristics of the real spherical world. The following four characteristics are desirable to preserve in a map projection:

- Relative distances. Measured distance between any two points on the map can be calculated by multiplying by a single scale factor. Most projections do not preserve this characteristic.
- Direction. Compass directions derived from the straight lines between points on the map could be used for navigation. The Mercator projection preserves this characteristic.
- Area. Two areas of the same size on the Earth would be equal in area on the map. The Mercator projection is one of many projections that do not preserve this characteristic.
- Shape. The general shape of the oceans and land masses on a map would be similar to the shapes of these features on the globe. Most projections do not preserve this characteristic.

Because no projection (only a globe) preserves all four of these characteristics, different projections are used for different purposes. In this textbook, for example, Goode's interrupted projection (**Fig. 1-10b**) is used often because it preserves relative areas correctly and shows each of the three major oceans uninterrupted by the edge of the map. Other projections, such as the Robinson projection (**Fig. 1-10c**), are also used in this textbook. The projection used for each map is identified in each figure. **Table 1-1** lists the relative ability of each of these projections to satisfy the four desirable map characteristics described here.

Table 1-1. Characteristics of Selected Map Projections

Map Projection	Preserves Relative Distances	Preserves Directions	Preserves Relative Areas	Preserves Shape	Areas Most Distorted	Common Uses
Mercator	Poor	Excellent	Poor	Fair	Mid to high latitudes	Navigation charts, world maps, maps of limited areas
Miller's cylindrical	Good	Good	Excellent	Poor	High latitudes	World maps
Robinson	Good	Fair	Good	Fair	High latitudes, 4 "corners"	World maps
Mollweide	Good	Fair	Good	Poor	High latitudes, 4 "corners"	World maps
Goode's interrupted	Good	Poor	Very good	Poor in most areas	Relative positions of continents	World maps, global oceans
Conic	Good	Excellent	Very good	Excellent	Distortion equally distributed	Maps of the U.S., individual continents
Polar azimuthal	Fair	Excellent	Good/poor	Good	Outer edges	Maps of polar regions

Scientific Notation and Units

Some of the numbers mentioned in this chapter are extremely large. For example, the history of the Earth presented in **Figure 1-4** spans nearly 5 billion (5,000,000,000) years. On that figure, the abbreviation MYA was used to represent millions of years ago. In ocean sciences, such large numbers are common, as are very small numbers. For example, the concentration of lead in seawater is about 0.000,000,000,5 gram per kilogram of water. In order to avoid using long strings of zeros, scientists use a type of shorthand for numbers like these. For example, the age of the universe is $4 \bullet 10^{10}$ years, and the concentration of lead in seawater is $5 \bullet 10^{-10}$ grams per kilogram.

These numbers might look odd, but they are really quite simple. The numbers are written using powers of 10. The • symbol can be stated as "times" or "mutiplied by" and is always followed by 10, so together they make "times ten" or "multiplied by ten". The superscript number that always follows the 10, represents the number of orders of magnitude (factors of 10) by which the digit before the •10 must be multiplied. Thus, $5 \bullet 10^{9}$ is 5 multiplied by ten, nine times ($5 \times 10 \times 10 \times 10 \times 10 \times 10 \times 10 \times 10 \times 10 \times 10$) which is usually shortened to 5 times ten to the 9th. A quicker way to use this notation is to take the number before the •10 and add nine zeros. The number is said to be raised by 9 orders of magnitude (powers of 10). Similarly, $5 \bullet 10^{-10}$ is 5 times ten to the power of negative 10 (the number is multiplied by 0.000,000,000,1, which gives 0.000,000,000,5. Notice that the negative sign in the superscript part of the scientific notation denotes the number of powers of 10 by which the number is reduced. When a number has more than one nonzero digit, such as 5,230,000,000, the scientific notation becomes $5.23 \bullet 10^{9}$. The scientific notation places one nonzero digit to the left of the decimal so, in this example (5.23) the simple way to use the notation is to move the decimal point 9 times, adding zeros only after the decimal point has passed the numbers to the right of the decimal point in the scientific notation (23 in our example). Notice that this shorthand system makes it easy to compare two very large or two very small numbers. In our example, it is easy to see that $5 \bullet 10^{9}$ and $5.23 \bullet 10^{9}$ are not very different without having to count all the zeros. Similarly, $5 \bullet 10^{9}$ is clearly much larger than $5.23 \bullet 10^{7}$. This text makes extensive use of scientific notation. Appendix 1 provides a simple conversion table if you need it.

Like other scientists, oceanographers use a variety of scientific units to identify such parameters as length, speed, and time. To avoid problems with unit conversions, such as miles to kilometers, and to make the comparison of data easier, scientists have developed an International System of Units (SI) that is steadily progressing toward being used universally, although some other units are still widely used. The seven SI base units are shown in **Table 1-2**.

Table 1-2. Base SI units

Parameter	Unit	Symbol
Length	meter	m
Mass	gram	g
Time	second	s
Electrical current	ampere	A
Thermodynamic temperature	kelvin	K
Amount of substance	mole	mol
Luminous intensity	candela	cd

All other units are “derived” units. For example, volume is measured in cubic meters (m^3), and speed is measured in meters per second (m/s or $m \bullet s^{-1}$). SI units are used in much of this text. However, as is still common in science, they are not used exclusively because some of them remain unfamiliar even to many members of the scientific community. **Appendix 1** includes more information about SI units and a table of all scientific unit abbreviations used in this text.

In this text, we use simple SI units to express concentrations of chemicals dissolved in water, for example, grams per kilogram ($g \bullet kg^{-1}$) or milligrams per kilogram ($mg \bullet kg^{-1}$). However, you may see on the web or in other publications concentrations expressed in **molality** or **molarity**. Molality is simply the concentration of the dissolved substance in grams of dissolved substance per kilogram ($g \bullet kg^{-1}$) divided by the molecular weight of the dissolved substance. Molarity is calculated in the same way but uses the concentration of dissolved substance per liter and is expressed in grams of dissolved substances per liter ($g \bullet l^{-1}$). Dividing the mass of a substance by its molecular weight gives you the amount of substance in “moles,” a unit that is related to the number of molecules of the substance present. So 1 mole of a substance has exactly the same number of molecules as 1 mole of any other substance. Chemists use molality or molarity because it can simplify some calculations.

For example, seawater contains about 19.8 $g \bullet kg^{-1}$ of chlorine, present as the chloride ion (Cl^-)ion, and about 10.78$g \bullet kg^{-1}$ of sodium as the sodium ion (Na+). This looks like it is a lot more chloride ion than sodium ion but, if we convert these to molality (by dividing each number by the molecular weight of the ion - about 35 for the chloride ion and 23 for the sodium ion, we find that chloride concentration in seawater is 0.55 molal and sodium is 0.47 molal - much closer. In fact, if seawater contained only dissolved sodium chloride, these molarity numbers would be the same, so we know that 0.08 moles (0.55 molal minus 0.47 molal) of other positive ions, such as potassium (K^+) and calcium (Ca^{2+}), must be present in seawater. In this example, molality (or molarity) is an expression of the number of atoms of sodium or chlorine. If we wanted to make sodium chloride from chlorine and sodium, we would need to add 1 mole of each to get 1 mole of sodium chloride, which is simpler to calculate than adding 35.5 g of chlorine and 22.99 g of sodium as we would have to do otherwise. With more complex molecules than sodium chloride, this simplification becomes even more valuable.

This page titled 1.3: How to Study Ocean Data is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.4: CHAPTER SUMMARY

Introduction.

The oceans influence our daily lives in many different ways. Understanding how they affect us requires an interdisciplinary approach that includes knowledge of the geology, physics, chemistry, and biology of the oceans and how these aspects interact.

Most of the world's population lives near the oceans or a river that connects to the oceans. The oceans provide many resources, especially food and transportation corridors, but they are also susceptible to pollution, habitat damage, and other impacts of human activities. Recreational uses of the oceans and awareness of the unique nature of marine ecosystems and species have grown explosively in the past several decades.

The Oceans and Earth's Environment.

Humans have altered Earth's environment in a number of ways. The most important of these has almost certainly been the use of fossil fuels that has raised the atmospheric concentration of carbon dioxide from about 280 ppm prior to the industrial revolution (ca. 1750) to nearly 430 ppm in 2025. The rate at which this concentration has risen has accelerated and is now about 2.6 ppm per year. The rise in atmospheric carbon dioxide, climate change, ocean acidification, and ocean deoxygenation, which together are the most serious long-term threats to Earth's environment resulting from human activities. Each of these threatens to disrupt global ecosystems, and, when combined, they pose the risk of a mass extinction of species. Climate change also threatens to adversely impact humans directly in a number of ways, including increasing the severity and probably the rate of occurrence of extreme weather events, rising sea level, and increased coastal erosion. Climate is controlled by the atmosphere and oceans working together in complex ways, which will be explored in this text. Understanding these complex interactions is the key to understanding the oceans and the future of climate change, acidification, and deoxygenation.

Oceans and the Origins of Life.

Many organic compounds probably were formed in the early oceans, but it is not known how these simple compounds were combined into more complex molecules that formed the basis for the creation of the first life form. However, it is thought that early life was found predominantly in the oceans. The first life for which we have evidence was bacteria-like, existed about 4.2 billion years ago, and must have been chemosynthetic, as there was no oxygen in the early atmosphere. Eventually, microorganisms developed that used photosynthesis, which splits water into hydrogen (which is utilized by the organism in its metabolism) and oxygen (which is released). Between about 1 and 2 billion years ago, photosynthetic organisms added oxygen to the atmosphere until it reached its current concentration of about 20%. Chemosynthetic organisms either died out or were restricted to limited oxygen-free environments, and species developed that depended on respiration using oxygen from their environment. Today, all species of animals and most other species respire and need oxygen in their environment. The first primitive higher animals, invertebrates such as sponges, appeared about 700 million years ago. Fishes developed about 510 million years ago. The first land plants and mammals developed about 500 and 200 million years ago, respectively.

How to Study Ocean Data.

Maps and charts are used extensively to display ocean science data. Each graph, map, or chart may represent data in a different way using different distortions of the real world. Graphs may have axes that do not start at zero or are nonlinear. The choice of axis can substantially change how the same set of data is perceived at first glance.

Contour plots are used extensively to represent the distribution of properties on a two-dimensional surface, such as the Earth's surface. If the intervals between contours on the contour plot are all the same, the relative distance between two contours is a measure of the gradient of the plotted property. However, on some maps, the contour interval changes from one area of the plot to another (usually to avoid crowding of contours), so the values of each contour must be examined to interpret gradients on the plot properly. Most contour plots in this text are color-shaded between the contours. Red always represents the highest value of the parameter plotted, grading through the spectrum to blue as the lowest value.

Maps and charts are used to represent horizontal distributions. Profiles are vertical cross sections through the Earth or the

oceans and generally display contours to show the vertical distribution of properties. Most profile plots in this text are greatly vertically exaggerated. This exaggeration is necessary because the depth of the oceans, and the thickness of the Earth's crust, are extremely small compared to the width of the oceans or land masses.

The Earth is spherical, and a system of latitude and longitude has been developed to identify specific locations on this sphere. Latitude is referenced to the equator, a circle around the Earth halfway between the poles, which is designated as 0° latitude. Other latitudes are expressed by the angle between a line from the Earth's center to the location in question and a line from the Earth's center to the equator. There are 90° of latitude; the North Pole is at 90°N and the South Pole at 90°S. Longitude is referenced to the prime meridian, a line designated as 0° that is drawn through the North and South Poles and that passes through Greenwich, England. Locations on the continuation of that same circle connecting the poles on the side of the Earth away from Greenwich are designated as both longitude 180°W and 180°E. Other longitudes lie between 0° and either 180°W or 180°E, and they are determined by the angle and direction between a line drawn between the Earth's center and the prime meridian and a line between the Earth's center and the specific location. One degree of latitude is the same distance regardless of its location on Earth. The distance represented by 1° of longitude varies from a maximum at the equator to zero at the poles.

Maps and charts must represent the spherical surface of the Earth or oceans in only two dimensions. No two-dimensional projection can correctly maintain relative distances, compass directions, relative areas, and the proper shape of features on a sphere. Therefore, all maps and charts distort one or more of these characteristics. The Mercator projection, which is the most widely used projection in atlases, depicts the distance between degrees of longitude as the same regardless of latitude. Thus, this projection preserves the relative directions between locations, but it distorts all three of the other relationships. The distortion is not important in regional maps of low latitudes, but it becomes greater at high latitudes and for global maps. Ocean scientists and this text use a variety of other projections for specific purposes. For example, Goode's interrupted projection is often used because it generally preserves relative areas and distances, and it can be drawn to represent all the major oceans without having to split them at the edge of the map.

To represent very large or very small numbers, scientists use scientific notation based on powers of 10. A standard system of scientific units (SI) is now in place and is becoming more widely used, but it is not yet universal. Both scientific notation and SI units are used extensively in this text.

1.4.1: Key Terms

1.4.2: Chapter Questions

1.4.3: References

This page titled 1.4: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.4.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- acidification
- bathymetric
- chart
- chronometer
- chemosynthetic
- contour
- crust
- deoxygenation
- detritus
- fossil fuel
- greenhouse effect
- invertebrates

- isobar
- isopycnal
- isotherm
- latitude
- longitude
- photosynthetic
- phototroph
- phototrophic
- pycnocline
- respiration
- steady state
- topographic

This page titled 1.4.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.4.2: Chapter Questions

Study Questions

1. What causes the greenhouse effect?
2. Why weren't the first living organisms capable of photosynthesis?
3. What are the essential things to look at the first time you study a graph?
4. What information can we get from studying a contour plot?
5. What is vertical exaggeration in a profile?
6. Are lines of longitude parallel to each other?
7. What are the four characteristics we would like a map projection to preserve?
8. On a Mercator projection map of the world, which countries or areas appear larger than they really are compared to other areas on the map?

Critical Thinking Questions

1. We are concerned that the Earth's climate may change because of the increase in atmospheric carbon dioxide concentration that has resulted from human activities over the past century or more. It has been suggested that, since trees absorb carbon dioxide and release oxygen, we can solve the problem by replanting forests and planting many more trees in our urban areas. If we were to plant trees wherever they would grow on the entire planet, would this be enough to reverse the trend of increasing carbon dioxide concentration in the atmosphere? Explain the reasons for your answer. If you are not able to answer this question, what information would you need to do so?
2. A number of developing nations, especially those in Africa and South America, have suggested that world maps used in all textbooks and atlases should use a projection other than the Mercator projection. What do you think is their reason for this suggestion?

This page titled 1.4.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

1.4.3: References

Figure 1-1 – NASA Ocean Color, 2025: Level 3 & 4 browser. Accessed 21 May 2025, Source.

Figure 1-2 – Data from NOAA Global Monitoring Laboratory, 2025: Trends in CO_2, CH_4, N_2O, SF_6. Accessed 21 May 2025, Source.

This page titled 1.4.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2: HISTORY AND IMPORTANCE OF OCEAN STUDIES

Critical Concepts

- **CC5 Transfer and Storage of Heat by Water**: Water's high heat capacity allows large amounts of heat to be stored in the oceans and released to the atmosphere without much change of ocean water's temperature. Water's high latent heat of vaporization allows large amounts of heat to be transferred to the atmosphere in water vapor and then transported elsewhere. Water's high latent heat of fusion allows ice to act as a heat buffer, reducing climate extremes in high latitude regions.
- **CC9 The Global Greenhouse Effect**: Perhaps the greatest environmental challenge faced by humans is the prospect that major climate changes may be an inevitable result of our burning fossil fuels. The burning of fossil fuels releases carbon dioxide and other gases into the atmosphere where they accumulate and act like the glass of a greenhouse, retaining more of the sun's heat.
- **CC10 Modeling**: Complex environmental systems including the oceans and atmosphere can best be studied by using conceptual and mathematical models. Many oceanographic and climate models are extremely complex and require the use of the fastest supercomputers.
- **CC11 Chaos**: The nonlinear nature of many environmental interactions, including some of those that control annual fluctuations in fish stocks, mean that fish stocks change in sometimes unpredictable ways.
- **CC16 Maximum Sustainable Yield**: The maximum sustainable yield is the maximum biomass of a fish species that can be depleted annually by fishing but that can still be replaced by reproduction. This yield changes unpredictably from year to year in response to the climate and other factors. The populations of many fish species worldwide have declined drastically when they have been overfished (beyond their maximum sustainable yield) in one or more years when the yield was lower than the average annual yield on which most fisheries management are based.
- **CC17 Species Diversity and Biodiversity**: Biodiversity is an expression of the range of genetic diversity; species diversity; diversity in ecological niches and types of communities of organisms (ecosystem diversity); diversity of feeding, reproduction, and predator avoidance strategies (physiological diversity), within the ecosystem of the specified region. Species diversity is a more precisely defined term and is a measure of the species richness (number of species) and species evenness (extent to which the community has balanced populations with no dominant species). High diversity and biodiversity are generally associated with ecosystems that are resistant to change.

Figure 2-1. The dawn of ocean sciences. HMS Challenger is made fast to St. Paul's Rocks on the Mid-Atlantic Ridge—a hazardous task—during the first part of its 1872–1876 expedition. The Challenger expedition was the first expedition to perform scientific studies of the oceans. The Challenger, which was a converted Navy Corvette, was a wooden-hulled sailing vessel with auxiliary steam power and no electricity. The scientists and crew, numbering about 236 at the start of the voyage, spent 1250 days on board, with 713 of those days at sea. Living conditions were arduous on the vessel that was less than two-thirds of a football field long and less than one quarter as wide. The vessel was packed with supplies as well as sampling equipment and sample containers, leaving little space for the crew to live and work.

The oceans have played a key role throughout human history. They have provided sustenance for humans for millennia. They have facilitated long-distance trade, transported armies for foreign wars, and provided the arena for countless naval battles. They currently offer a source of recreation for hundreds of millions of people. However, the oceans' key role in mediating global climate and local weather is likely more important to humans than any of these other functions.

Given the enormous importance of the oceans to human existence, one might expect our knowledge and understanding of them to be extensive and thorough. However, the oceans, which cover approximately 70% of the Earth's surface to an average depth of almost 4 km, are a vast domain that remains largely unvisited and unexplored. Amazingly, until the 1930s, no human had ever descended below 150 m.

In this chapter, we will review some of what has been learned about the oceans and about why they are so important to humans. Compared to other sciences, oceanography is young, and oceanographers continue to make new, sometimes startling, discoveries.

2.1: What is Oceanography

2.2: Exploration and Mapping

2.3: National and International Law Applied to Oceans

2.4: Ocean Resources

2.5: Chapter Summary

2.5.1: Key Terms

2.5.2: Chapter Questions

2.5.3: References

This page titled 2: History and Importance of Ocean Studies is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.1: What is Oceanography

Until recent decades, oceanography was primarily an observational discipline with greater similarity to geography than to other sciences. For many thousands of years, study of the oceans was limited almost entirely to exploring and mapping the oceans, especially shallow underwater obstructions such as **reefs**. These studies were important to facilitate trade and travel between known landmasses and islands, but were very limited. Considerable information was also gathered about the best places to find fish and **shellfish**, especially in the coastal oceans. However, for many centuries this biological information was documented only in the oral histories of local fishers.

Aside from limited studies by scholars of a few ancient civilizations, almost all systematic studies of the oceans have been carried out within the past 250 years. During these two centuries, mapping of the oceans has continued with ever-increasing sophistication (**Appendix 4**). Geological oceanographers have mapped seafloor **topography** and **sediments**. Physical oceanographers have mapped distributions of salt content, temperature, and currents. Chemical oceanographers have mapped distributions of chemicals in seawater and in seafloor sediments. Biological oceanographers have mapped distributions of algae, bacteria, and animal **species**. Mapping the oceans has been an enormous task that is far from complete, for reasons we examine later.

During the past several decades, oceanographers have moved beyond simply mapping the ocean to studying ocean processes. As revealed in this text, the geological, physical, chemical, and biological processes in the oceans and overlying atmosphere are interdependent and complex. Consequently, oceanography must be interdisciplinary. Oceanography is now often called "ocean sciences" to acknowledge the discipline's transition from primarily mapping oceans to interdisciplinary studies of ocean processes.

To perform their work properly, oceanographers must understand all basic ocean processes, although they often specialize in one aspect. In oceanography, much information must be integrated to describe complicated processes. Therefore, the scientist must take a step beyond many of the traditional sciences that rely on reductive methodology, in which investigators reduce the complexity of a problem by studying only part of it (e.g., studying the effect of varying temperature on a single species rather than the combined effects of all environmental variables and competition from other species). Oceanographers do perform reductive methodology experiments in controlled laboratory or field conditions but they must always integrate and compare them with detailed observations of the ever-changing marine **environment**. This integration and comparison often requires the use of computer models (**CC10**).

This page titled 2.1: What is Oceanography is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.2: Exploration and Mapping

Through most of human history, the oceans have been studied primarily to facilitate seaborne travel and transport. Hence, early studies were directed primarily toward exploring and mapping the oceans.

Prehistory

The oceans provide a food source and an important mode of transportation, but we do not know when humans first began to use the oceans for these purposes. However, humankind has used these resources for tens of thousands of years and they remain the most valuable uses of the oceans today. The bones of marine fishes discarded by Stone Age people have been found in caves in South Africa that date back to between 100,000 and 70,000 years ago. In East Timor, near Indonesia, fish hooks that date back more than 38,000 years and bones of fishes that date back as far as 42,000 years ago have been found in a limestone cave. The fish bones came from species that live both in coastal waters and the deep ocean far from shore, providing evidence that humans were using boats or rafts to fish in the open ocean. Also, humans are thought to have arrived on the island of Australia about 50,000 years ago. The first boats were probably built even earlier, perhaps for fishing on lakes or in the shallow coastal ocean, then later for transportation and colonization. The first boats or rafts may even have been made and used by *Homo erectus*, our immediate predecessor in the evolutionary tree. Archaeological evidence shows that some *Homo erectus* lived on the island of Flores, Indonesia, at least 750,000 years ago. To reach Flores, *Homo erectus* must have crossed a wide, deep water strait that acted as a barrier to the migration of most other species.

Early human boats were made of wood, reeds, animal skins, or tree bark. These materials do not generally survive the centuries for archaeologists to study. The oldest boat found by archaeologists is thought to be a dugout canoe found in the Netherlands that was made between 10,200 and 9600 years ago.

Ocean Exploration in Early Civilizations

Recorded history tells us that early development of ocean transportation and trade centered in the Mediterranean Sea, although the Polynesian and Micronesian cultures in the Pacific Ocean may have pursued these activities during the same period.

The Mediterranean

The Minoan civilization, which prospered on the island of Crete in the Aegean Sea from about 3000 BCE, is considered the first recorded civilization to have used boats extensively for transport, trade, defense, and conquest. The Minoans' influence extended throughout the many islands of the Aegean, and oral and written history records that a Minoan navy fought and controlled pirates in the region. Although seafaring capability continued to develop in the Mediterranean in both the ancient Greek and Egyptian civilizations, the Phoenicians, who inhabited areas that are now parts of Israel, Lebanon, and Syria, were the greatest of all the early Mediterranean seafarers. From about 1100 to 850 BCE, the Phoenicians were a great sea power, voyaging throughout the Mediterranean to Spain, Italy, North Africa, and even the British Isles, where they traded for tin. The Phoenicians also claimed to have made a 3-year voyage around the entire continent of Africa, but that claim has not been confirmed.

Much of the information needed to navigate from one port to another was a closely held secret of the navigator's craft. However, maps that showed the shapes and sizes of **coastlines** and seas were made several thousand years ago. In about 450 BCE, the Greek historian Herodotus drew a map that is a surprisingly recognizable map of the Mediterranean and Red Sea region (**Fig. 2-2**). The most famous of ancient mapmakers was the Greek geographer Ptolemy, who lived around 150 CE. Ptolemy's maps, although little known until the end of the Dark Ages, about 1400 CE, were the basis of most maps until the 1500s.

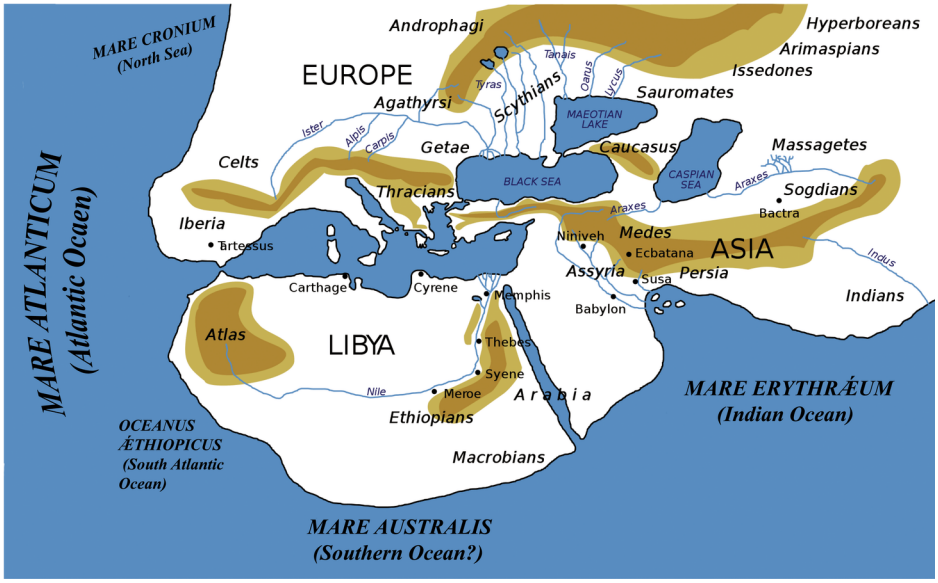


Figure 2-2. Map of the known world drawn by Herodotus in approximately 450 BCE. The oceans surrounding the known landmasses were thought to extend to the edge of the world. Compare this map carefully to a current Mercator projection, and you will see that the map drawn by Herodotus somewhat distorts longitude but quite accurately reproduces latitude.

Ptolemy's maps are remarkably detailed and accurate in their reproduction of north-to-south positions (**latitude**). However, they are distorted by substantial errors in east-to-west positions (**longitude**). For example, Ptolemy shows the east-to-west length of the Mediterranean to be about 50% too long in relation to its north-to-south dimension. These errors are a consequence of the mapmaker's inability to measure time accurately. Without precise and accurate time measurements, it was not possible to establish longitude correctly, as is discussed in **Chapter 1**. The east–west distortion of maps was not corrected until about the 1760s, when the first practical **chronometers** were developed that could keep accurate time on a ship.

In the 2000 years following the Phoenician era, ocean exploration slowed, and much of what may have been learned was lost in the turmoil of the Dark Ages. We know that philosophers did make several significant observations, especially during the early part of this period. In the sixth century BCE, Pythagoras declared the Earth to be spherical. In the fourth century BCE, Aristotle concluded that total rainfall over the Earth's surface must be equal to total evaporation because the oceans did not fill up or dry out. In the same century, another Greek, Pytheas, sailed out of the Mediterranean to Britain, Norway, Germany, and Iceland. He developed a simple method of determining latitude by measuring the angle between the horizon and the North Star—a method still used today. Pytheas also proposed the concept that **tides** were caused by the moon.

In the third century BCE, Eratosthenes, a Greek studying at the library in Alexandria, Egypt, calculated the circumference of the Earth along a circle through the North and South Poles. His value of 40,000 km was very close to the 40,032 km that has been determined by extremely precise and sophisticated modern methods. Approximately 200 years later, Poseidonius incorrectly recalculated the circumference of the Earth to be about 29,000 km. Ptolemy accepted Poseidonius's incorrect value for his maps, and the error was not corrected for centuries. In fact, this error led Christopher Columbus to believe he had reached Asia when he arrived in the Americas in 1492.

Micronesia and Polynesia

About 4500 years ago, at about the same time that ocean voyaging and trade were beginning to expand in the Mediterranean, the large islands of the far western Pacific and Micronesia were colonized. Humans traveled from mainland Asia to these islands, likely through Taiwan. The expansion may have been made possible by the development of the outrigger canoe (**Fig. 2-3**). After an apparent pause of about 1000 years, Polynesians colonized the islands of the western and central tropical Pacific, reaching Hawaii and Easter Island about 900 AD. Unlike Mediterranean sailors, who could follow the coastline, the Polynesians had to navigate across broad expanses of open ocean to colonize the Pacific islands. The Polynesians crossed the ocean in double-hulled sailing canoes made of wood and reeds. A larger hull provided living space for up to 80 sailors plus plants and animals; a smaller hull or outrigger stabilized the vessel. The double-hulled design is still used in much of Polynesia (**Fig. 2-3**).

Figure 2-3. A typical Polynesian outrigger canoe. This design has been used for centuries throughout much of Polynesia and Micronesia. This canoe was photographed in Papua New Guinea.

The Polynesians were arguably the greatest navigators in history. They successfully navigated across the open ocean using only their memorized knowledge of the stars, winds, wave patterns, clouds, and seabirds. This knowledge was passed down through generations of the families of navigators, who were justifiably venerated in Polynesian culture. Both the Polynesians and Micronesians created maps from wood sticks or rattan, but most Polynesian navigators did not use or need such maps. Some of the ancient Polynesian navigational knowledge persists today in the oral histories and experiences of just a few remaining descendants of the early navigators. Efforts have been made recently to sustain and preserve this knowledge through a series of ocean voyages across the Pacific in reconstructed replicas of ancient Polynesian boats, such as the *Hokule'a* (**Fig. 2-4**).

Figure 2-4. The *Hokule'a*. This vessel is a reproduction of the double-hulled voyaging canoes that the Polynesians used to explore and colonize the Pacific.

The recent journeys have proved that the ancient Polynesian navigation techniques were, and still are, remarkably accurate and reliable. There is great interest in discovering and documenting all the observational clues the navigators used. Much of the Polynesian art of navigation is mysticized and undocumented, but the navigators must have known principles of wave shapes and directions, cloud formations, and other ocean phenomena that even now are not fully understood by modern scientists. For example, Polynesian navigators can deduce the location of an island hundreds of miles away by observing wave patterns created in the wake of the island. Modern science has been able to document such large-scale wave patterns around islands only since satellite observations of the sea surface became possible.

The Dark Ages and the New Era of Discovery

During the Dark Ages, when ignorance and anarchy engulfed Europe, ocean exploration by Europeans was limited. At that time, however, the Arabs were developing extensive seaborne trade with East Africa, India, and Southeast Asia. They were the first to use the **monsoons** to their advantage: voyaging from East Africa to Asia during the northern summer, when monsoon winds blow from the southwest, and returning in the winter, when the winds reverse and blow from the northeast. The Vikings also made great ocean voyages during the Dark Ages, reaching Iceland in the ninth century and Greenland and Newfoundland late in the tenth century.

The middle and late 1400s brought the dawn of a new era as Europeans set out on many ambitious voyages of discovery. The Canary Islands off Northwest Africa were explored in 1416, and the Azores, in the middle of the Atlantic Ocean, were discovered in about 1430. The southern tip of Africa was rounded by Vasco da Gama in 1498; Columbus rediscovered the Americas in 1492; and the world was first circumnavigated by Ferdinand Magellan's expedition of 1519–1522, although Magellan himself was killed in the Philippines during the voyage.

Systematic mapping of the oceans began during the middle of the second millennium. After the mid-1400s, exploration of the oceans was incessant, and maps improved rapidly. However, most exploration was undertaken only to discover and colonize lands that could be reached by crossing the oceans. Systematic study of the oceans themselves is generally not considered to have begun until 1768, when Captain James Cook began the first of three voyages to explore and map the Pacific Ocean (**Fig. 2-5**). Captain Cook was the first navigator to carry an accurate chronometer to sea, so he was the first navigator able to measure longitude precisely, enabling dramatic improvement of existing maps. On his voyages, Cook made many measurements of water depth, or **soundings**, which he accomplished with a lead weight attached to a rope. He measured depths to 200 fathoms (366 m). He also made many accurate measurements of water temperature, currents, and wind speed and direction, and he documented the occurrence and general characteristics of **coral reefs**. On his final voyage, Captain Cook died at the hands of Polynesian natives at Kealakekua Bay, Hawaii.

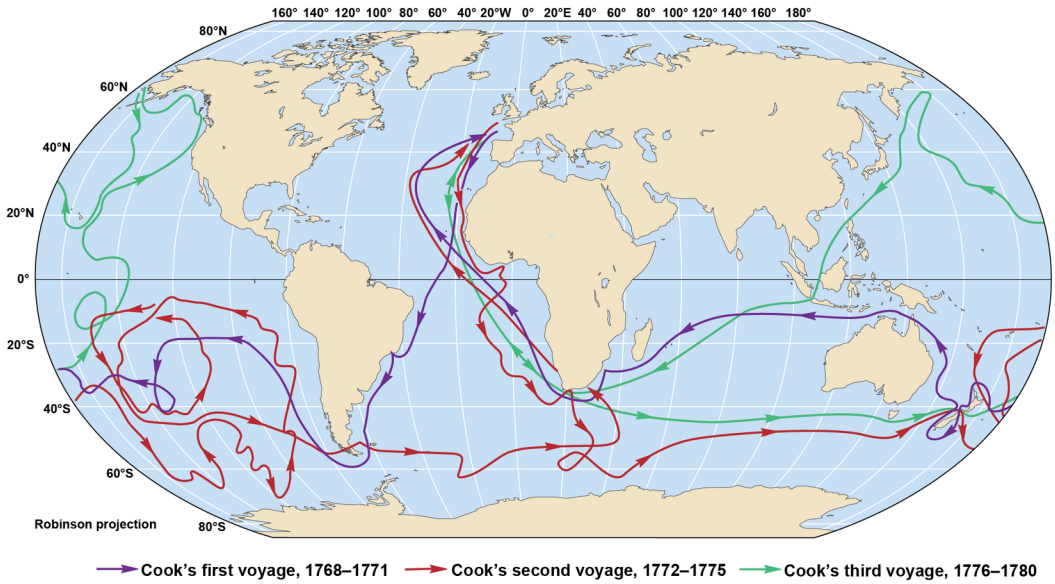


Figure 2-5. Routes of the three voyages of Captain Cook: 1768–1771, 1772–1775, and 1776–1780. Cook visited all the major oceans, traveling south into the waters near Antarctica and as far north as the Bering Sea. Places named after Captain Cook include Cook Inlet, Alaska; Cook Strait between the North and South Islands of New Zealand; and a group of Polynesian islands in the tropical Pacific near Tahiti.

At the same time that Cook was exploring the Pacific, Benjamin Franklin, then deputy postmaster for the American colonies, learned that it took 2 weeks longer for mail ships to sail the northerly route from Europe to the colonies than to sail the longer southerly route. Franklin's interest in that phenomenon led him to have many discussions with ships' navigators, from which he concluded that ships coming from the north were sailing against a great ocean current. From his observations, he and his cousin Timothy Folger were able to develop a remarkably accurate map of the Gulf Stream current in 1770 (**Fig. 2-6**).

Figure 2-6. Map of the Gulf Stream originally drawn by Benjamin Franklin and Timothy Folger in 1770.

The Birth of Oceanography

By 1800, several seafaring nations had established government offices with primary responsibility for producing charts that could be used by mariners to navigate safely and avoid reefs and **shoals**. Matthew Fontaine Maury, a U.S. Navy officer in charge of the Depot of Naval Charts, made a particularly significant contribution to the intensive ocean mapping efforts of the nineteenth century. Maury gathered data on wind and current patterns from numerous ships' logbooks, and he published his detailed findings in 1855 in a volume entitled *The Physical Geography of the Sea*. Maury also initiated cooperative efforts among seafaring nations to standardize the means by which meteorological and ocean current observations were made. Because of his many contributions, Maury has often been called the "Father of Oceanography."

The Beagle

In 1831, the HMS *Beagle* started a 5-year epic voyage of discovery that forever changed the way humans view their world (**Fig. 2-7**). The major objective of this voyage was to complete a **hydrographic** survey of the Patagonia and Tierra del Fuego coastal regions to improve maps used by English ships sailing between the Pacific and Atlantic Oceans around the tip of South America. The *Beagle* also visited the Galápagos Islands off the **coast** of Peru, crossed the Pacific to New Zealand and Australia, and returned to England across the Indian Ocean and around the southern tip of Africa.

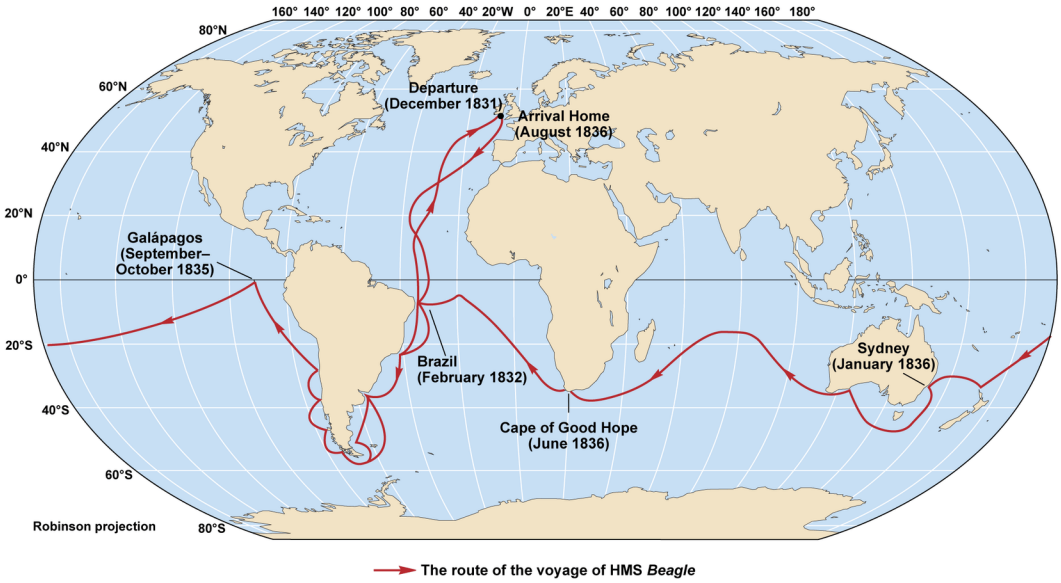


Figure 2-7. The voyage of HMS *Beagle* and Charles Darwin, 1831–1836. More than half the voyage was spent near South America.

The *Beagle* expedition has become famous primarily because of the young naturalist who traveled on the ship, observing the plant and animal life of the many places where the *Beagle* touched land. This naturalist was, of course, Charles Darwin. Darwin's observations on the *Beagle* expedition were part of the basis for his book *Origin of Species*, in which he proposed the revolutionary theory of natural selection.

Although the observations that led to the theory of evolution were the most famous of his findings on the *Beagle* expedition, Darwin also made several other major discoveries. For example, he proposed a theory to explain the formation of coral **atolls** that is still accepted (**Chap. 4**). In addition, he made a startling observation on one of his land excursions during the voyage: Darwin climbed high into the Andes Mountains, which run along the west coast of South America. At the top of these mountains, he found **fossils** in the rocks that were undeniably the remains of marine creatures. Darwin concluded correctly that the rocks had originated beneath the sea. Thus, he deduced that continents were not permanent and unchanging, as was widely accepted at the time, but that they must move, at least vertically. That observation remained almost unnoticed among Darwin's works until the twentieth century, when the theories of continental drift and **plate tectonics** were developed (**Chaps. 4**).

The Challenger

During the 1860s, British vessels investigating the seafloor prior to laying a transatlantic telegraph cable brought up living creatures in mud samples from the bottom of the deep sea. At the time, the prevailing scientific opinion was that the deep ocean was devoid of life, because of the high pressures and low temperatures. Thus, the discovery of life on the deep-ocean floor was perhaps as dramatic a finding as the discovery of life on Mars would be if it were to occur today. The discovery of life in the deep sea led to the true birth of oceanography as a modern science, an event that can be precisely dated to 1872 through 1876. It was during these years that HMS *Challenger* sailed the world's oceans as the first vessel outfitted specifically so that its crew could study the physics, chemistry, geology, and biology of the oceans (**Fig. 2-8**). The *Challenger* was a sail-powered navy corvette with an auxiliary steam engine. For its scientific expedition, sponsored by the Royal Society of London, the corvette's guns were removed and replaced with laboratories and scientific gear. Included was equipment for measuring the ocean depths, collecting rocks and sediment from the ocean floor, and collecting seawater and organisms from depths between the ocean surface and the seafloor. The scientific additions crowded the vessel and left only spartan living quarters for the crew and the six scientists.

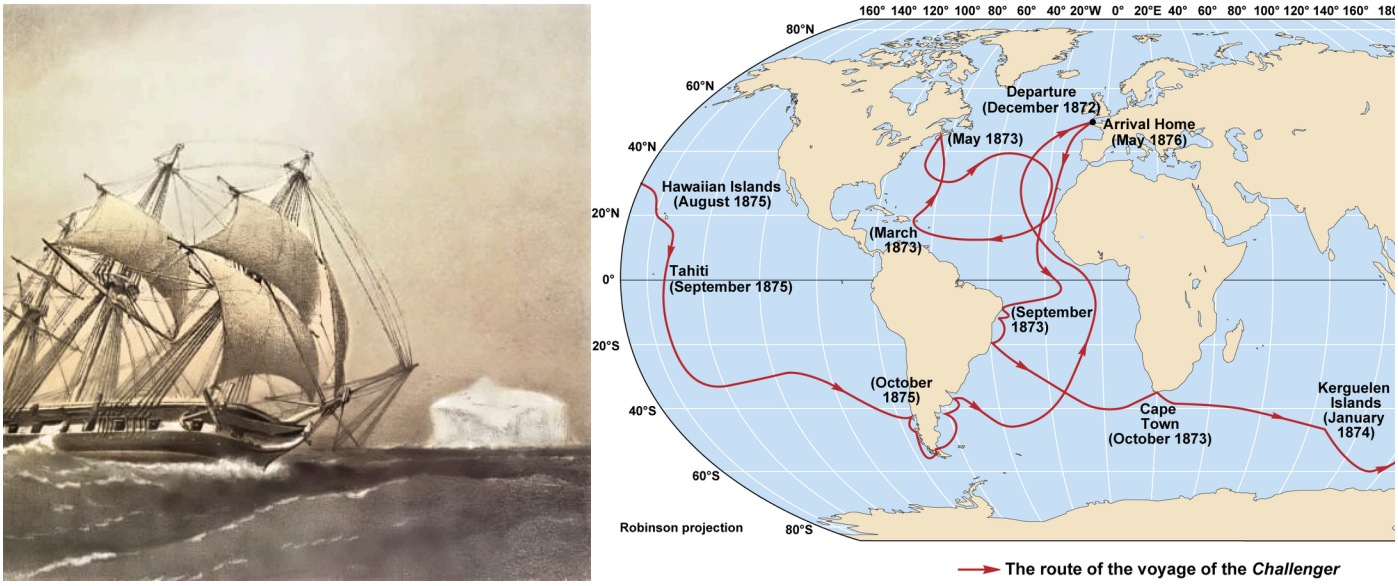


Figure 2-8. The Challenger expedition. (a) A painting of HMS Challenger. (b) Route of the voyage of the Challenger, 1872–1876, which sailed through all the major oceans, but not into the high-latitude regions.

The *Challenger* sailed 127,500 km across the oceans during its 5-year expedition to study the North and South Atlantic Ocean, the North and South Pacific Ocean, and the southern part of the Indian Ocean. The expedition made hundreds of depth soundings using a lead weight on a hemp line that was hauled in by hand over a steam capstan. With this equipment, a single deep sounding required an entire day. Despite the extreme difficulty and tedium of obtaining deep soundings, the *Challenger* expedition was able to measure a depth of 8185 m in the Mariana Trench east of the Philippines.

The *Challenger* also conducted:

- Hundreds of observations of ocean water temperature, both near the surface and at depth
- More than 100 dredge samples of rocks and sediment from the seafloor
- One hundred and fifty open-water net trawls for fishes and other organisms
- Numerous samples of seawater
- Many readings of ocean current velocity and meteorological conditions
- Countless visual observations of fishes, **marine mammals**, and birds

The expedition brought back a wealth of samples and new scientific information about the oceans, including the identification and classification of almost 5000 previously unknown species of marine organisms. The quantity of data and samples obtained was so great that a special government commission was established to analyze the information. Indeed, the 50 volumes of research reports generated by the expedition were not completed until decades after the ship returned to England. The volumes contained so much information that they provided the foundation on which almost all major disciplines of oceanography were later built.

The Modern Era

In the 150 years since the *Challenger* expedition, oceanographers have traveled the seas in research ships with increasing frequency, and with observation and sampling equipment of ever-increasing sophistication. A significant proportion of ocean research is still performed from research vessels in much the same way that it was in 1872. However, during the twentieth century, oceanographic research expanded to include exploration and study using **scuba** and manned **submersibles**, as well as unmanned observation using robotic vehicles and instrument packages either free-floating or attached to cables moored to the seafloor. Many of these instruments now report their data by **acoustic** links to the surface, and by radio signals sent through satellites to land-based facilities. In addition, oceanographers make many observations from aircraft and satellites using remote sensing techniques. The introduction in 1978 of satellites specifically designed to look at ocean processes was an important milestone because it allowed almost simultaneous observations to be made across an entire ocean basin for the first time. Satellites are now among the most important observing platforms used by oceanographers (**Chap. 3**).

The detailed history of oceanography since the *Challenger* expedition is too voluminous to include in this text. However, some of the important events are summarized in Appendix 4, and subsequent chapters review many of the more important findings

of ocean study and exploration in the modern era. Among the most important events or discoveries during this author's lifetime have been:

- The 1959 publication by Heezen and Tharp of the first comprehensive map of the ocean floor topography (**Chap. 3**).
- The discovery in 1964 and 1965 of hot, high-salinity **brines** and unusual black ooze-like sediments at the bottom of the Red Sea. This was the first observation of **hydrothermal vents**, although that was not recognized at the time.
- The first visit to a hydrothermal vent by the submarine *Alvin* in 1977, where a previously unknown type of **ecosystem**, based on **chemosynthesis** and sustaining hundreds of previously unknown species was discovered.
- The steady accumulation of evidence that led to the acceptance of Wegener's 1912 theory of continental drift. Much of the crucial evidence supporting this theory was gathered by the Deep Sea Drilling Program, which obtained samples from deep within ocean sediments for the first time.
- Satellite observations of the world ocean provided the first comprehensive snapshot view of the oceans. This view did not have to rely on many individual observations from ships and enabled the observation of large-scale dynamic processes as never before.
- The development of autonomous robotic measuring devices, including *Argo* floats, allow the gathering of data of unprecedented detail about processes that occur below the surface layer and that cannot be observed by satellite sensors.
- The discovery of the horizontal gene transfer mechanism in the mid-1950s.
- The finding, at the end of the 20th century (based on DNA analysis) that ocean life is dominated by microbial organisms.
- The finding that microbial life exists deep within ocean sediments and even deep within the Earth's crust below these sediments.

Each of these can arguably be claimed to be as important a breakthrough or advancement for ocean sciences as, for example, Darwin's observations on the voyage of the *Beagle* were to biology. We live in an age of discovery for ocean sciences and many more surprises, advancements, and revelations are to come.

This page titled 2.2: Exploration and Mapping is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.3: National and International Law Applied to Oceans

Before we look at the various categories of ocean resources, it is important to realize that the oceans and their resources are not owned by individuals or corporations, as is much of Earth's land area and its corresponding resources. Instead, ocean areas adjacent to coasts are owned by governments, and ocean areas remote from land are not owned at all. To understand why this is true and what it means, we must review a little about the history of national and international laws as they relate to the sea.

Throughout most of history, the oceans have not been considered territory that could be claimed or owned. Indeed, until the 1600s, there was little thought or deliberation on the legal status of the oceans or the ownership of its resources. Individuals and nations were generally free to travel and fish anywhere in the oceans. In 1672, however, the British declared that they would exercise control over a **territorial sea** that extended 5.6 km (in British units, 1 league or 3 nautical miles) from the coastline. This distance corresponded roughly to the range of **shore**-based cannons. The 5.6-km territorial sea became commonly accepted as the standard for most nations and was formally accepted by the League of Nations in 1930. The remainder of the oceans outside the territorial seas, called the **high seas**, was considered to belong to no one, and in general, all nations were free to use and exploit high-seas resources.

The Truman Proclamation

The 5.6-km territorial sea remained almost universally accepted until after World War II, when President Harry Truman proclaimed in 1945 that "the exercise of jurisdiction over the natural resources of the subsoil and seabed of the continental shelf by the contiguous nation is reasonable and just" and that "the U.S. regards the natural resources of the subsoil and seabed of the continental shelf beneath the high seas, but contiguous to the coasts of the U.S., subject to its jurisdiction and control." This proclamation, known as the Truman Proclamation on the Continental Shelf, was motivated by the "long-range worldwide need for new sources of petroleum and other minerals." The **continental shelf** was not defined in the proclamation. However, in a statement accompanying the proclamation, it was described as the land that is contiguous to the continent and covered by no more than 183 m (100 fathoms) of water. The 183-m depth was arbitrary because the continental shelf cannot be defined by a single depth (**Chap. 4**).

The Truman Proclamation caused a controversy and motivated a series of unilateral decisions by various coastal nations to claim the resources under their own coastal oceans. However, there was no uniformity in the claims. Individual nations claimed jurisdiction over different widths of ocean off their coasts. For example, nations along the west coast of South America, where the continental shelf is very narrow (**Chap. 4**), claimed the ocean resources to 370 km (200 nautical miles) from shore, approximately 66 times the distance of previously defined territorial seas. These nations went beyond the U.S. proclamation in another important way too: they extended their sovereignty over the entire 370-km zone, whereas the Truman Proclamation claimed jurisdiction only over the resources on or under the seafloor. Thus, these other nations were claiming ownership of the waters and fisheries in this zone, and the right to control any access to the zone by other nations' vessels.

"Law of the Sea" Conferences

In 1958 and again in 1960, the United Nations convened a conference on the "Law of the Sea." The conferences were intended to establish international uniformity in the ownership and access rights of nations to the resources of the oceans. The two conferences, attended by more than 80 nations, led to the adoption of several conventions. The conventions established a zone 22 km (12 nautical miles) wide within which nations had jurisdiction over fishery resources, and they affirmed that the "high seas" area beyond that zone was free for all nations to use for navigation, fishing, and overflight.

The conventions also attempted to define the right of coastal nations to own the seabed and sub-seabed minerals on the continental shelf. The continental shelf was defined as the area beyond the territorial sea out to a depth of 200 m or "beyond that limit, to where the depth of the superadjacent waters admits of the exploitation of natural resources." This definition was ambiguous. Coastal nations could claim the minerals out to a depth at which they could be exploited, and the size of this zone would expand as exploitation technologies improved. At the time, exploitation of seabed resources at depths greater than 200 m was considered unlikely. However, oil-drilling technology improved rapidly after 1960, and the U.S. was soon drilling in waters more than 200 m deep off California and more than 185 km (100 nautical miles) offshore in the Gulf of America (Golfo de México).

The potential mineral resources of the deep oceans, particularly **manganese nodules** (**Chap. 6**), were first recognized during

the 1960s. By the late 1960s, it was clear that more needed to be done to define and determine the rights of nations to use the oceans, particularly the rights to own and to exploit ocean mineral resources.

In 1967, the Maltese ambassador to the United Nations, Dr. Arvid Pardo, proposed that another Law of the Sea Conference be held. Pardo suggested a concept for a treaty that history may recognize as a critical turning point in the development of human civilization. His suggestion was that the ocean floor outside the zones of national jurisdiction should be reserved for peaceful uses, and that its resources should become the "common heritage of mankind." Pardo suggested that money generated from the exploitation of these resources be used for the benefit of less developed nations.

The common heritage principle has subsequently been applied to other global resources, including the atmosphere, Antarctica, tropical rain forests, and ocean biodiversity. It is now an important basis for wide-ranging and growing international cooperation among nations. The common heritage principle was a major catalyst for the growing global understanding that the Earth's natural resources belong to all its peoples. The essence of the principle is that all nations have rights and responsibilities to protect the global environment and to use its resources wisely.

The Law of the Sea Treaty

In response to Pardo's call, the United Nations convened the Third United Nations Law of the Sea Conference in 1973. The conference met many times over the next 9 years. Finally, in 1982, the 151 participating nations adopted a new Convention on the Law of the Sea (commonly referred to as the "Law of the Sea Treaty") by a vote of 130 to 4, with 17 abstentions. The U.S. was among the four nations that voted against adoption, even though many of the Treaty provisions were originally written or supported by the U.S. delegation. The U.S. declared that it would not sign or ratify the Treaty, because it did not agree with the ocean mining provisions, although it would follow many other individual provisions. Abstaining nations included the Soviet Union, Great Britain, and West Germany. As of 2023, there are 169 nations that have signed and 151 nations plus the European Union have both signed and ratified the Treaty. The U.S. is the only major nation who has not ratified the Treaty. The U.S. signed the Treaty in 1994 but the U.S. Senate has consistently refused to ratify it despite being urged to do so by several Presidents. The ratification would be necessary for the U.S. to formally join the Treaty organization. However, the U.S. does recognize the Treaty as general international law and abides by most of its provisions.

Since 1982, the Law of the Sea Treaty has been the basis for numerous laws written by many nations, including the U.S., about their individual rights to the oceans and ocean resources. Many provisions of the Treaty are now widely accepted, and the Treaty itself was ratified by the required number of 60 nations and became fully effective in November 1994. The principal provisions of the Law of the Sea Treaty can be summarized as follows.

A territorial sea of 12 nautical miles (22.25 km) was established, within which each individual coastal nation has full sovereign rights to all resources and to controlling access by foreign nationals. (Some nations, including Peru, Ecuador, Somalia, and the Philippines still claim territorial seas that extend to 200 nautical miles, despite the Treaty.)

1. An **exclusive economic zone** (EEZ) was established outside the territorial sea. The EEZ normally extends out to 200 nautical miles (370 km), but exclusive rights to the sea bottom and resources below the bottom (but not the water column) can extend as far as 350 nautical miles (649 km) to the edge of the continental shelf where the shelf extends beyond 200 nautical miles. The edge of the continental shelf is defined by the Treaty in geological terms, but in a way that is complicated and difficult to interpret in some areas. Within its EEZ, the coastal nation has jurisdiction over mineral resources, fishing, and pollution, and it may exercise control over access to the zone for scientific research.
2. Complicated procedures were established for drawing ocean boundaries of EEZs between nations that are closer to each other than 400 nautical miles (740 km), or whose coastlines meet in complex ways.
3. The right of free and "innocent" passage was guaranteed for all vessels outside the territorial seas and through straits used for international navigation that are within a territorial sea (straits narrower than 24 nautical miles, or 44.5 km).
4. Complicated rules were established for exploiting mineral resources from high-seas areas outside the EEZs. All such exploitation would be regulated by a new International Seabed Authority (ISA). Any nation or private concern wanting to extract minerals from the high-seas areas would have to obtain a permit from the ISA to mine a given site. In return for the permit, the nation or private concern would provide its mining technology to the "Enterprise," a mining organization set up within the ISA. The Enterprise would mine a separate site that would be paired with the permit site. Revenue from the permit site would go to the permittee, whereas revenue from the Enterprise site would be divided among the developing nations.

The Treaty did not include precisely defined procedures and rules for technology transfer to the Enterprise or for the disbursement of Enterprise profits among the developing nations. Those and all other decisions of the ISA would require unanimous approval of all nations party to the Treaty. Unanimous consent of all the nations is virtually impossible to achieve, especially when resources

and profits are at stake. Therefore, this provision was felt by many to destroy any reasonable chance that the ISA could succeed. Nations that voted no or abstained when the Law of the Sea Treaty was approved included most of the nations that were interested in deep-ocean mining of manganese nodules. No mining has yet been done under ISA auspices, partly because deep-ocean mining may not yet be economical. The ISA has granted numerous contracts for exploration (but not production) at sites including sites in the eastern equatorial Pacific Ocean and established protocols for monitoring the environmental effects of mining and test mining has been initiated. However, ISA has stated it will permit no production mining until it has finalized its mining regulations and has set a target date of 2025 for completing these regulations.

Exclusive Economic Zones and Extended Continental Shelf Rights

The Law of the Sea Treaty has exerted considerable influence over the actions of the world's nations. Most significantly, the EEZ concept is now almost universally accepted. Under the initial 200-nautical-mile EEZ definition, the total area placed under the control of the various nations was approximately $128 \cdot 10^6$ km^2, or about 35.8% of the world ocean (**Fig. 2-9**).

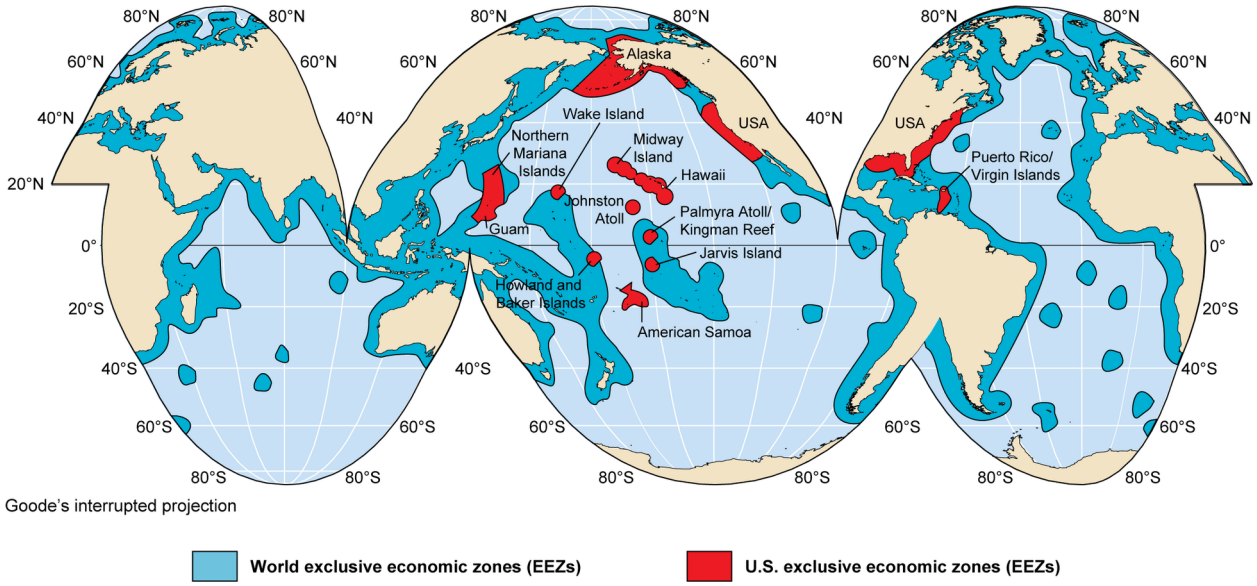


Figure 2-9. Exclusive economic zones (EEZs) of the U.S. and other countries. This map shows how large the U.S. EEZ is and how much of it is associated with Alaska and Hawaii, as well as various Pacific and Caribbean islands that are not among the current 50 states.

However, the Treaty allowed nations to claim continental shelf beyond 200 nm as part of their EEZ. The Treaty established a geologically based definition of continental shelf for these claims. The rules of how this definition should be applied were not adopted until 1999, and they were very complex and open to interpretation. Nations were required to provide extensive data and evidence to define where the seafloor descending from the landmass reaches the relatively flat ocean bottom. This location is almost impossible to define or measure precisely. In addition, nations could claim undersea mountains and ridges if they are "continental appendages." Again it is virtually impossible to precisely define and demonstrate this connection.

Since the Treaty entered into force in 1994 many nations have claimed EEZs beyond 200 nautical miles where their continental shelf extends beyond this line, as the Treaty allows. As a result about $138 \cdot 10^6$ km^2, (about 38% of the world ocean) are currently designated as EEZ, although some claims are still unsettled. The U.S. has a large EEZ, the second largest of any country (behind France). The U.S. EEZ is large not only because of the lengthy Atlantic, Pacific, and Gulf of America (Golfo de México) coastlines of the contiguous states, but because of the vast additional EEZ areas contributed by Alaska, Hawaii, Puerto Rico, the U.S. Virgin Islands, and the various U.S. Pacific protectorates and islands, including American Samoa and Guam (**Fig. 2-9**).

The U.S. EEZ encompasses an area of approximately $11.3 \cdot 10^6$ km^2, which is larger than the nation's ($9.8 \cdot 10^6$ km^2) land area. However, the U.S. EEZ could have been larger if the U.S. had been able to claim its continental shelf beyond 200 nautical miles as other nations have done. A claim could not be filed by the U.S. as the U.S. still has not ratified the Treaty and therefore does not belong to the Treaty organization. However, the U.S. EEZ does contain many of the world's most productive fisheries and probably a large proportion of the ocean's mineral wealth.

The EEZ areas associated with islands are often very large. A single tiny atoll remote from any other island can command an

EEZ of more than $0.2 \cdot 10^6$ km^2, an area somewhat larger than the state of California. Ownership of such a large area is the real reason for many territorial disputes over the ownership of islands that formerly were considered inconsequential. For example, the war between Great Britain and Argentina over the Falkland Islands in 1982 was motivated by the resources in the islands' EEZ. A dispute still continues between Japan and Russia over the Kuril Islands north of Japan. Probably the best example of an EEZ-motivated sovereignty dispute is the case of the South China Sea's Spratly Islands, which various nations claim to own. Although these tiny islands are inconsequential aside from the ocean resources that they command, various islands of the group are claimed by three or four nations (**Fig. 2-10**). Each of the claimant nations has stationed troops on one or more of the Spratly Islands, China has constructed military facilities on several of the contested islands, and tension often runs very high.

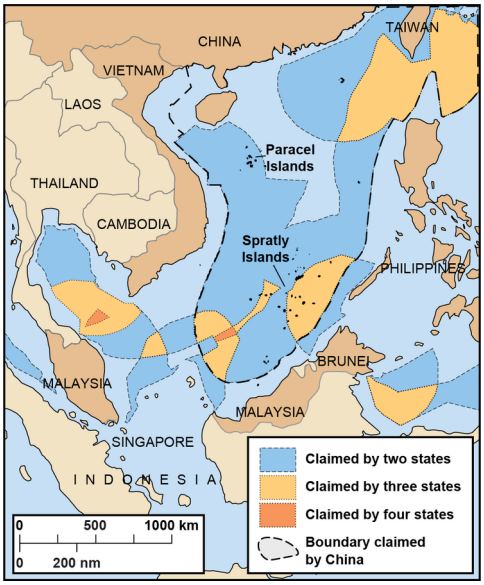


Figure 2-10. Boundary disputes in the South China Sea. Nations claiming ownership of one or more of the Spratly Islands include the Philippines, Vietnam, China, and Malaysia.

Sovereignty disputes over islands are not the only evidence of the importance of the EEZs. Many island nations, especially those in the Pacific, are extremely concerned about the possible reduction of their EEZs if sea level continues to rise, as predicted. In some locations, such as the Seychelle Islands in the Indian Ocean, a sea-level rise of only a few tens of centimeters would submerge low-lying islands, each of which commands an enormous area of EEZ. Even if sea level does not rise, the smallest islands may disappear through **erosion**. China claims sovereignty over a number of tiny islands in the South China Sea that are remote from any other island and the mainland. Many of these islands rise to only a few tens of centimeters above sea level, and they have essentially no economic value other than the up to 0.25 million km^2 of EEZ that each commands. China has committed large sums of money to dredge and fill and construct ports and airports and even military bases on a number of islands, yet wave erosion could soon eliminate some or all of these islands. The Chinese development on these disputed islands aims to further China's claim of sovereignty despite international court legal rulings that have already denied these sovereignty claims. While China's main motivation may be its claim of sovereignty of these islands and the accompanying EEZ, it also hopes to protect the island from disappearing due to sea level change. However, under the Law of the Sea Treaty, it is questionable whether such artificial means to preserve an island from sinking, and so retain its EEZ, would be accepted by international law.

The Treaty of the High Seas

While the Law of the Sea Treaty addressed the ownership of the mineral resources of the seabed of the high seas (all ocean outside of EEZs) it did not address the biological or other resources of the ocean waters of these high seas. In September of 2023, after several years of meetings, an addendum to the Law of the Sea Treaty formally titled the "Agreement under the Law of the Sea on the Conservation and Sustainable Use of Marine Biological Diversity of Areas beyond National Jurisdiction" generally shortened to the "Treaty of the High Seas" or BBNJ agreement, was adopted. By October of 2023 it had been signed by 82 countries, including the U.S. The Treaty entered into force in January 2026.

The Treaty of the High Seas establishes an authority, similar to the authority that controls mining in the high seas area, to ensure that the benefits of marine genetic resources (e.g. drugs produced from marine animals) are shared under the "common heritage of mankind" principle. The International Seabed Authority (ISA) has already established guidelines for conducting impact

assessments in the high seas, and a framework for establishing Regional Environmental Management Plans under the original Law of the Sea Treaty deep-sea mining provisions. While the addition of the Treaty of the High Seas represents the first attempt to provide protection and conservation for ocean ecosystems in high seas areas beyond those where seabed mining may occur, it does not address ownership of marine biological resources or fishing management in high seas areas. Rather it leaves this controversial subject to the existing patchwork of international fisheries management laws that largely apply to single species and not to ecosystem diversity and sustainability as a whole.

This page titled 2.3: National and International Law Applied to Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.4: Ocean Resources

Most people know that the oceans are fished for seafood and used for transport and recreation, and that oil and gas are extracted from the seafloor. However, it is not widely recognized how valuable these resources are. For example, according to the U.S. Department of Transportation, U.S. ports handled more than $2.1 trillion in goods in 2024 with 465,000 vessel calls to these ports. The recreational cruise industry is estimated to have contributed $65 billion to the U.S. economy in 2023. The National Oceanic and Atmospheric Administration reported that, in 2020, the seafood industry supported 1.1 million jobs and generated $154.7 billion in sales, while recreational fishing contributed 595,000 jobs and $98 billion. The Congressional Research Service reports royalties from offshore oil and gas resources paid to the US Treasury were $5.6 billion in 2019 but fell to $3.7 billion in 2020. The National Economics Program also reports that in 2018, tourism and recreation based on the coasts and oceans supported more than 2.5 million jobs and generated $143 trillion in Gross Domestic Product in the U.S. There are also a number of other ocean resources or potential ocean resources that are not included in these figures. In the following sections, we will briefly survey the wide range of ocean resources. For this purpose, we grouped these resources into eight categories:

- Biological resources
- Transportation, trade, and military use
- Offshore oil and gas
- Methane hydrates
- Minerals and freshwater
- Recreation, aesthetics, and endangered species
- Energy
- Waste disposal

Biological Resources

Fisheries

Fish and shellfish are probably the most valuable ocean resources. Seafood has a very high protein content and therefore contributes significantly to global dietary needs. In many coastal areas, seafood is the basic subsistence food because no other significant source of food protein is available. Iceland and many Pacific Ocean island nations, including Japan, are good examples of such seafood-dependent areas. Fisheries are also a major part of the U.S. economy.

Until the past century, seafood resources were, for all practical purposes, inexhaustible because they were replaced by reproduction faster than they were consumed. However, human populations have burgeoned in the last 100 years and the demand for seafood has increased accordingly. Consequently, the seafood resources of many parts of the oceans have been exploited so intensively that many species are **overfished** and can no longer reproduce fast enough to replace their populations. Overfishing, which has resulted from a variety of technical and socioeconomic factors, poses perhaps the greatest threat to the health of ocean ecosystems, other than climate change, ocean **acidification**, and **deoxygenation**. Most scientists feel that overfishing is a far greater threat than the oil spills, industrial waste, and domestic sewage discharges that often dominate media coverage of the oceans. Indeed, the increasing need for the greatest possible utilization of ocean fishery resources to feed a hungry and growing population, and the damage done to these resources by unwise exploitation and management, have been important factors in the development of the field of oceanography.

Historically, fisheries were, and many still are, resources not owned by any person, organization, or government. The fact that they were free for anyone to exploit led to a vicious cycle repeated in fishery after fishery. When a new fishery opens, a few fishing boat operators are able to catch large quantities of the resource species with little effort and make substantial profits. Other operators, aware of these profits, quickly enter that fishery. As the number of boats increases and the population of the target species is reduced, each operator must expend more effort to catch the same amount of fish. At the same time, the value of the catch may decline because the market is now flooded with this species. In response, each operator increases fishing efforts to catch more and recover profits lost in the price drop. Hunting and catching technologies are continuously improved, and the fishing pressure continues to rise. If unchecked, the continuous increase in fishing effort and effectiveness quickly leads to a collapse of the fish population as the **maximum sustainable yield** is exceeded (**CC16**). Many fishers then look for a new species to target, and the cycle begins anew.

In such instances, the fishery resource has declined, sometimes so precipitously that the commercial fishery has been essentially wiped out for decades. Perhaps the greatest known historical decline was the collapse of the anchovy catch off Peru from over 12 million metric tons in 1970 to less than 2 million metric tons in 1973. This decline constituted roughly 10% of the total world fishery catch at the time. A strong El Niño was certainly responsible for some of the decline, but many scientists believe that excessive fishing efforts contributed greatly to the problem. The California sardine fishery that supported the development of Cannery Row in Monterey, California during the 1940s (**Fig. 2-11a**) likely contributed significantly to the decline in sardine abundance that has not yet recovered. The number of king crab caught by fisheries in the Bering Sea (**Fig. 2-11b**) peaked and then declined rapidly in the 1980s, which was mirrored closely by abundance studies in the area. This decline occurred at a time of sustained increase of fishing pressure.The menhaden fisheries off the middle Atlantic coast of the U.S. and the cod fisheries off the coast of New England and Canada are more prime examples of such decimated resources, although environmental factors may also have contributed to the declines. Many other fisheries have also collapsed or may be close to doing so.

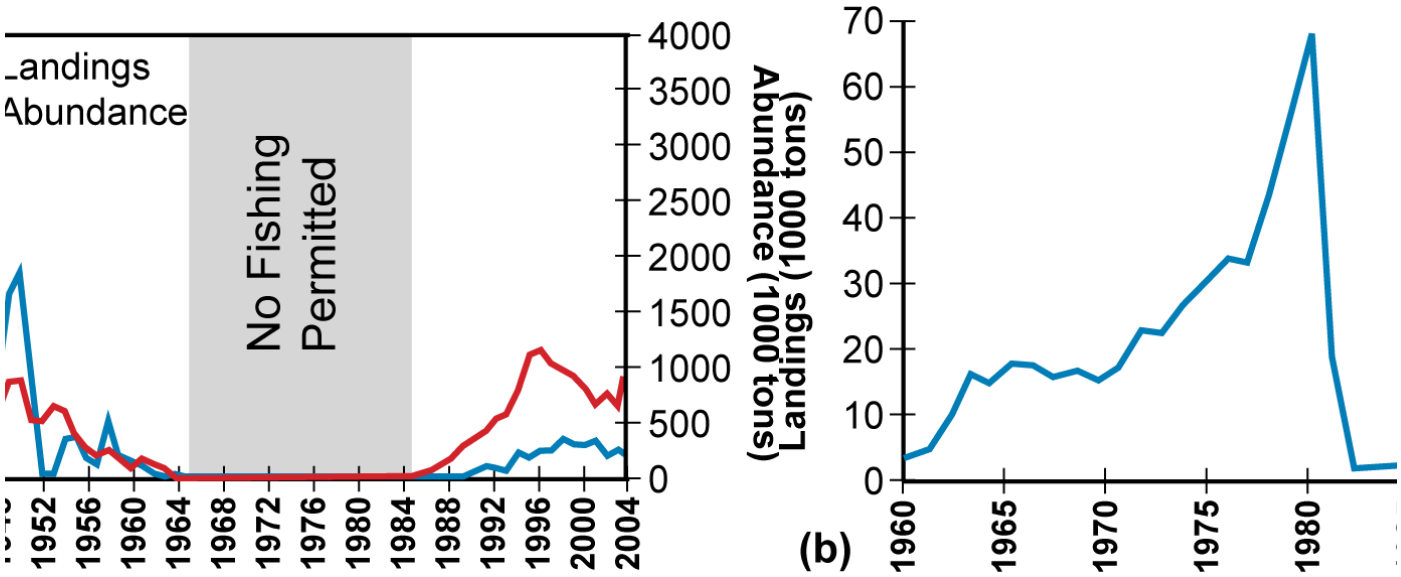


Figure 2-11. The effects of overfishing. (a) A history of changes in sardine stocks off California. Note that fishing pressure (landings data) remained intense despite the dramatic decline in abundance in the 1940s. California sardine stock declines have occurred periodically, so natural factors may have caused this decline. However, overfishing as the stock declined likely contributed and worsened the decline. (b) King crab abundance in Alaska declined dramatically in the early 1980s, when fishing pressure increased substantially. Neither sardine nor King crab populations have yet fully recovered.

Most major fisheries in the world are today being fished close to, or above, their maximum sustainable yield (**CC16**). When fishing is drastically reduced or halted for a species whose population has collapsed, the population of some species recovers. However, for other species, recovery never occurs because the species has been replaced in the food web by some other, often less commercially valuable, species. Evidence suggests that some fishes are steadily being replaced by **jellies** (jellyfish) in many parts of the world ocean. This replacement may be caused by global climate change, but it is very likely that overfishing is also largely or partly responsible for any such changes.

The United Nations Food and Agriculture Organization (FAO) monitors fisheries statistics worldwide. FAO reports that the share of fish stocks harvested unsustainably has grown significantly, climbing from 10% in 1974 to 37.7% by 2021. A further 50.5% were being fished at their maximum sustainable yield in 2021. FAO acknowledges that its data is less than perfect due to poor reporting of fishery statistics by a number of nations.

The rapidly growing recreational fishing industry also contributes to overfishing problems. For example, the recreational catch of striped bass on the east coast of the U.S. more than doubled between 2021 and 2022. The recreational catch in 2022 was estimated to be about 16 thousand metric tons, more than 8 times the amount caught by commercial fishers.

Most traditional fisheries are in coastal waters near the consuming population. However, as coastal fisheries have been depleted, fishers have exploited resources from the deeper parts of the oceans and from distant coastal regions. The movement to far-flung ocean fisheries was accelerated by the development of refrigeration, which allowed catches to be stored and transported long distances in fresh or frozen condition. In addition, deep-ocean fisheries were traditionally free and open for anyone to exploit because no nation owned or controlled the resources. Unfortunately, the species that were the easiest targets of deep-ocean fishers were also those most vulnerable to overexploitation because they reproduce and mature more slowly. The decimation of

populations of many species of whales and seals is among the better-known examples of the overexploitation of ocean resources.

Overfishing is particularly serious in many developing countries surrounded by coral reefs. As modern medicine enables human populations in these countries to expand, the limited fishery resources of reefs near each village can no longer provide sufficient seafood for the growing population. Fishers are forced to exploit reefs in an ever larger area and must often resort to technological "improvements" in fishing techniques, such as dynamiting and spearfishing. Some of the fishes killed by dynamiting float to the surface for easy harvesting, but many of them sink and are lost to the fishers. Furthermore, dynamiting destroys the reef and **habitat** for the fishes that escape the blast and for future generations of fishes. Intensive spearfishing, particularly with scuba, can quickly remove breeding adults of a population and thus hinder reproductive replacement of the population. In these instances, technological advances in fishing have proven to be very destructive to the resources.

Fishing threatens not only the targeted species, but also other species in the food web. For example, harvesting and drastic reductions in sea lion and elephant seal populations in California during the Gold Rush era (mid to late 1800s) led to a sharp reduction in the population of their natural predator, the great white shark. The marine mammal population became protected by law in 1972 and has recovered substantially since the early 1980s. Correspondingly, the great white shark population has also slowly increased. Another example of this pattern is the precipitous decline of the now rare Pribilof fur seal in the Bering Sea, which was almost certainly caused largely by increased fishing pressure on its principal food species, including pollack.

Beyond altering food web dynamics, fishing can negatively affect non-target species through unintentional capture known as **bycatch**. Many fishing techniques, including nets and trawls, are not efficient in selecting the target species. For example, turtles were often unintentionally caught in shrimp nets, and many dolphins were caught and killed in nets set to capture large **schools** of tuna. Considerable effort has been put into developing nets that, while they may be less efficient at catching the target species, do not catch and kill endangered turtles or marine mammals. Dolphins, sharks, turtles, and many nontarget fish species are also caught in kilometer-long **drift nets** that form a barrier across the ocean **photic zone** and that capture, and usually result in the death of, anything large enough not to pass through the mesh. As in other fisheries, this incidental bycatch of "nonvaluable" species is simply discarded overboard. Such drift nets are now outlawed by most nations but, as with other illegal fishing methods, may continue to be used in areas far from surveillance by law enforcement officials. Enforcement of fishing regulations on the high seas is extremely difficult because no nation can afford to patrol the high seas adequately.

Bottom trawling, a widely used fishing technique where a net is dragged across the sea floor to catch bottom dwelling species, has been identified as perhaps the most environmentally damaging fishing technique. These trawls destroy deep sea corals and other nontarget **benthos** in seafloor ecosystems. Many of the species in these seafloor ecosystems are extremely slow growing so trawling damage in some areas may not be repaired by re-colonization and re-growth for decades or centuries.

Many fisheries are now managed to avoid overfishing, but management is often ineffective because assessing the fishery stock size and its age composition is expensive and difficult. In addition, the maximum sustainable yield may vary dramatically from year to year because of the **chaotic** variations induced by natural factors (**CC16, CC11**). Consequently, managers have only a poor understanding of what maximum sustainable yield might be. Safe management requires that substantially less than the estimated maximum sustainable yield be caught each year to guard against errors inherent in the stock assessment data and against the inevitable years when stocks decline unexpectedly because of natural factors. However, managers are under pressure to allow the largest possible annual catch in order to provide adequate income to the owners and crews of fishing boats competing for the resource. Setting the catch "too low" would mean lost income and possibly jobs for fishers, as well as unnecessary "wasting" of some of the resource value.

A number of alternative management concepts are now applied to some fishery resources to address this problem. These include assigning catch quotas to individuals, individual boats, or communities, granting exclusive limited fishing access to defined geographic areas, or both. Each of these regulatory approaches embodies the principle that access to the fishery resources is a privilege and that access to these resources should no longer be open to any and all persons who choose to fish. Another conservation approach—establishment of fishing-free natural reserves where fish populations can reproduce freely—is also now widely applied.

To this day, humans are primarily hunters and gatherers in the oceans, much as Stone Age people were on land. Human development from ocean hunter-gatherer to ocean farmer is long overdue. Mariculture (ocean farming) historically has been used on only a small scale in very few locations. However, it has developed and continues to grow rapidly, particularly in China, where it has been practiced for thousands of years, and in several other developing nations of the Pacific Ocean basin.

Other Biological Resources

Apart from their aesthetic value, marine species are an important pool of genetic diversity. Many marine species have developed unique biochemical methods of defending themselves against predators, parasites, and diseases, and of detoxifying or destroying toxic chemicals. Therefore, marine species are a major potential resource for the development of pharmaceutical drugs and pollution control methods. The search for beneficial drugs and pollution-fighting organisms in the oceans is extremely tedious and has barely begun. However, a substantial number of potentially valuable pharmaceutical products have already been isolated from marine species and many are being tested for a variety of medical purposes. A number of pharmaceuticals derived from marine organisms have already become approved for human use. These include compounds isolated from marine sponges such as the antiviral acyclovir and the HIV/AIDS drug azidothymidine (AZT). Many of the compounds that show promise have come from rare ocean animals or algae found only in limited areas of certain coral reefs or other threatened ocean ecosystems.

Coral reefs are like the tropical rain forests of the oceans, in that they are the most promising sources of pharmaceuticals because of their extremely high species diversity. They sustain large numbers of candidate species, any one of which may contain numerous chemical compounds potentially valuable as drugs. Some such naturally occurring compounds have been used to design similar molecules that have similar drug properties but that can be industrially produced from widely available raw materials. Unfortunately, other pharmaceutically active compounds isolated from marine species may not be readily synthesized or redesigned. If this proves to be the case, conservation measures will need to be developed and enforced to prevent impoverished villagers who live near a reef from using destructive harvesting techniques to supply the pharmaceutical industry.

An example of destructive harvesting techniques is provided by the industry that provides fishes for tropical aquariums. A growing number of fishes bred in aquariums are entering the market. However, most tropical fishes sold in the aquarium trade in the U.S. are still captured from Philippine reefs. Some collectors in the Philippines and elsewhere capture fishes by releasing cyanide into cracks in the reef, even though this practice is illegal. When cyanide is released into the reef, many fishes and other species die, but some fishes are not killed immediately. They swim out of hiding to avoid the cyanide, but they are stunned by the chemical and, therefore, easily captured. As many as 90% of the fishes that survive the initial cyanide collection die of the cyanide's effects during transport to the U.S. or after the buyer has placed the fish in a home aquarium. The ornamental (aquarium) fish market in the U.S. was estimated to be worth about $6.3 billion in 2022. Only a very tiny fraction of this money ever reaches the harvesters or is used to promote sustainable collection practices.

Transportation, Trade, and Military Use

Despite the rapid growth in air transport, surface vessels remain the principal and cheapest means of transporting cargo and people across the oceans. Large numbers of commercial, recreational, and military vessels enter or leave U.S. ports every day. The importance of the oceans for transportation is evidenced by the estimated $2.1 trillion in shipped goods handled by U.S. ports, and a passenger cruise industry that contributed $36.0 to the U.S. GDP in 2023. In addition to this value for commercial ocean transportation, the oceans have important military uses. The oceans are plied by many surface naval vessels, but submarines have become particularly important, especially to the U.S. Navy, as platforms to transport and deploy ballistic missiles. As a result, during the past several decades, intensive efforts have been made to improve ways to hide and find submarines. Very extensive oceanographic studies have been conducted to support these efforts, particularly studies of ocean surface and subsurface **currents** and acoustic properties of the oceans.

The many benefits of our use of the oceans for transportation, trade, military, and recreation do come at a cost to the ocean environment. Ships have become progressively larger, particularly oil tankers, container ships, and naval vessels (especially aircraft carriers and submarines). Larger vessels require deeper ports and harbors, which has increased the need for dredging navigation channels in many bays, estuaries, and rivers. Dredging damages benthos at the dredging site, and the dredged material, which is generally dumped at a site in the estuary or coastal ocean not far from the dredging site, can also have serious impacts (**Chap. 16**).

Offshore Oil and Gas

Oil and gas are extracted from beneath the seafloor in many parts of the world. Most of the undiscovered oil and gas reserves are believed to lie beneath the continental shelves and **continental slopes**. The search for the sedimentary structures most likely to yield oil or gas under the oceans, and the development of technologies to drill for and produce oil and gas safely and efficiently, have been intense throughout the past several decades. Oil and gas have been produced from wells drilled in shallow water for many years. However, technological developments have steadily extended capabilities for drilling in deeper and deeper waters, and in areas where **weather**, waves, currents, and sometimes ice conditions are progressively more demanding. The search for

oil and gas, and the need to identify and control the environmental impacts of this search, are a consistent and important focus of recent oceanographic studies.

In the EEZ of the U.S., the most extensive known oil and gas deposits are in the Gulf of America (Golfo de México). However, substantial known reserves are present beneath the continental shelves of the northeastern and southeastern U.S., California, and Alaska (**Fig. 2-12**). There may also be many undiscovered deposits, particularly off Alaska. The offshore petroleum industry is among the largest natural resource development industries globally. Oil and gas are used primarily as fuel for vehicles, industry, and heating, but they are also the basic raw materials for plastics, pharmaceuticals, and other chemicals, cosmetics, and asphalt. Although fossil fuel burning may be reduced to prevent further buildup of atmospheric carbon dioxide (**CC9**), petrochemicals will still be needed in the future.

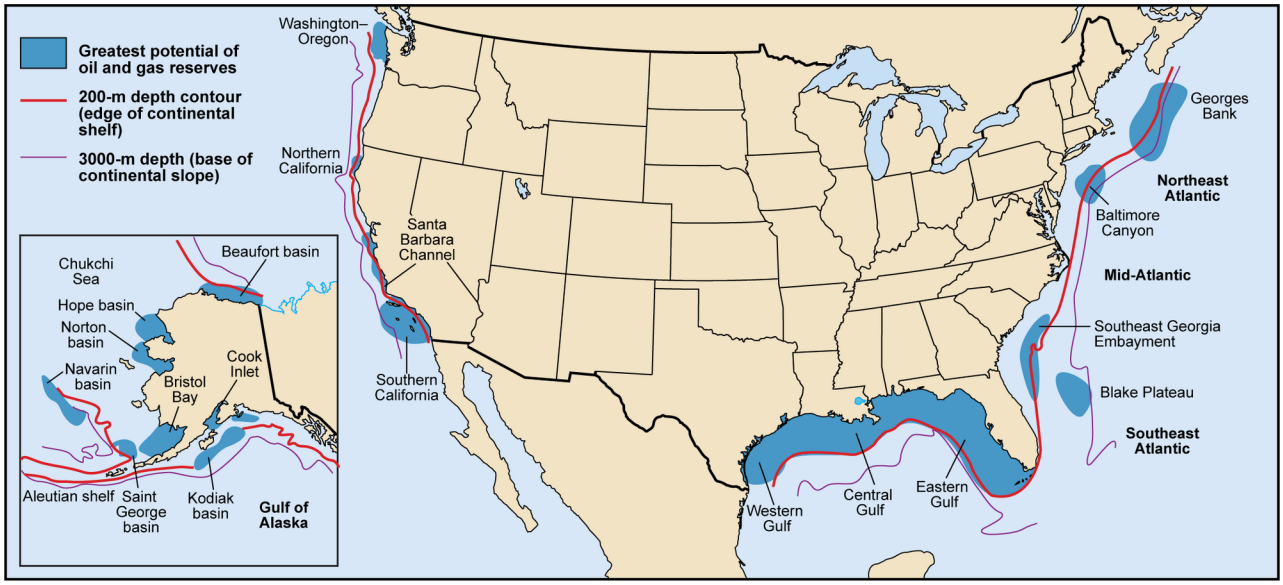


Figure 2-12. Offshore oil and gas deposits occur throughout much of the U.S. Atlantic, Pacific, and Arctic Oceans, the Gulf of America (Golfo de México), and the Bering Sea continental shelves. The deeper continental shelf and continental slope of the Gulf of America (Golfo de México) and the Arctic Ocean continental shelf are currently thought to hold the greatest potential reserves.

Drilling in relatively shallow water is usually done from offshore platforms supported by long legs anchored to the seafloor (**Fig. 2-13**). Floating platforms (**Fig. 2-13a**) are also used, especially in deep water, while drilling in the nearshore Arctic is done from artificial gravel islands. A production platform may tap into 100 or more wells drilled into the seafloor below. Directional drilling, a process in which the well pipe is drilled downward and then turned underground to drill at an angle or even horizontally away from the drill site, now allows a single well to drain oil from an area as much as 12 km in radius and to recover more than 20 times as much oil per drill site or platform. On an ocean production platform, the well pipes generally extend up through the platform's legs. On the platform, oil, gas, and water are separated, and oil and/or gas are usually transported ashore through a pipeline laid on, or buried in, the seabed.

Figure 2-13. Structures of the offshore oil and gas industry. (a) These offshore exploration drilling rigs are under maintenance in Galveston, Texas. On the left is a typical jack-up drilling rig; on the right is a typical floating rig that rides on pontoons below the depth of most wave action. (b) Production oil rigs permanently anchored to the seafloor of Cook Inlet, Alaska.

Methane Hydrates

The decomposition of organic matter releases methane. At low temperatures and high pressures, methane molecules can be trapped within the crystalline lattice of water ice crystals to form a combination called “methane hydrates” (on average, one molecule of methane for every five or six molecules of water ice). Methane hydrates were first observed several decades ago in samples of cores drilled on land and the ocean floor. These hydrates could be a potential source of methane, a clean-burning fossil fuel. (It is completely combusted to carbon dioxide and water with virtually no chemical waste by-products or products of incomplete combustion, such as are produced by the refining and burning of other fossil fuels). However, widespread exploitation of this resource is considered unlikely for several reasons. Most importantly, methane hydrates are widely dispersed, usually occurring in the pore spaces of sediments and rocks.

Methane is rapidly released from the hydrates at normal atmospheric pressures and temperatures, but the deposits are generally too deep to mine and bring to the surface. However, several methods are now being tested that release methane from the hydrate in the sediment or rock so that it could be collected in a drill hole and recovered in the same way that petroleum and natural gas are recovered from oil and gas deposits.

In 2000, interest in methane hydrates was revived when some fishers dragged their trawl net at a depth of 800 m in a canyon about 50 km east of the mouth of Puget Sound in the northeastern Pacific Ocean. The fishers were startled to see their net rise to the surface filled with 1000 kg of icy chunks that were fizzing and melting. They hauled the “catch” aboard their vessel but quickly shoveled it back overboard because they had no idea what it could be. They tried to save samples in a freezer, but the low temperature alone was not enough to stop the methane from escaping the hydrate. The gas, expanding as it was released from the hydrate, even broke the containers in which they tried to store the hydrate.

A year later, an ROV discovered what were described as “glaciers” of frozen methane hydrates forming outcrops on the seafloor about 50 km from where the fishers had found hydrates. Since that time, deposits of methane hydrates have been found on the seafloor or buried in shallow sediments in several other areas of the oceans, and they are now thought to occur in many parts of the world’s oceans, especially on the continental slope.

The mechanism that forms methane hydrate deposits is poorly understood. Some hypotheses suggest that the methane is a product of the decomposition of organic matter buried in the sediments in **marginal seas** as the continents were pulled apart—the same source thought to be responsible for most of the world’s oil and gas deposits (**Chap. 4**). Instead of being contained in nonporous rocks and converted to oil and natural gas, decomposing organic matter released methane that migrated through porous sediments and rocks until conditions were right for the formation of methane hydrates.

Methane hydrates are unstable except at high pressures and low temperatures, so they occur only below a depth of about 500

to 600 m in the oceans. However, these conditions are also present at relatively shallow depths beneath the frozen tundra of the Arctic regions. Not surprisingly, methane hydrate deposits have also been found in these environments. Estimates of the extent of methane hydrate deposits suggest that the total world resource may be more than 100 times the total volume of natural gas estimated to be recoverable from world oil and gas deposits. Thus, although most methane hydrate is widely dispersed and probably unrecoverable, the economic value would be huge if only 1% of the total were contained in concentrated deposits that could be recovered. The methane would also be a very efficient and clean-burning fuel that could be readily adapted to most commercial and industrial energy uses.

The first field test of the extraction of methane from methane hydrate deposits in the ocean was carried out in 2013 at a well drilled in the Nankai Trough, an ocean basin 80 km off the coast of Japan. Japanese researchers retrieved methane up to the surface over a period of one week from a water depth of 1000 m. Japan and China both successfully produced methane from test production wells in 2017, China drilling in the South China Sea and Japan drilling off Japan's central coast. Significant efforts to establish a methane hydrate extraction industry are concentrated in a few nations, with Japan and China leading the way. The U.S. has identified the presence of methane hydrate deposits in the Gulf of America (Golfo de México) and Alaskan permafrost, but interest in these deposits has, for now, focused on research and mapping.

Minerals and Freshwater

Ocean sediments contain vast quantities of mineral and material resources other than just oil and gas. They include sand and gravel (**Fig. 2-14b**), manganese nodules, **hydrothermal minerals**, **phosphorite nodules**, and heavy minerals such as gold that are often present in sediments of current or ancient river mouths (**Fig. 2-14a**). Sand and gravel are currently mined in large quantities from the shallow seafloor and used as construction materials in locations where no local land resources are available. At present, few efforts have been made to exploit other marine mineral resources. Cassiterite, a tin mineral, is dredged from shallow waters offshore from Thailand and Indonesia, and gold-bearing sands are dredged from shallow river mouth deposits offshore of Alaska, New Zealand, and the Philippines. Despite the limited scope of current ocean mining activities, there is substantial interest in future development. Mineral deposits thought to be most likely exploitable include phosphorite-rich sands as potential sources of phosphorus for fertilizer, manganese nodules, and seafloor massive sulfide (SMS) deposits laid down by hydrothermal vent activity as potential sources of metals including iron, zinc, copper, nickel, cobalt, manganese, molybdenum, silver, gold, and platinum.

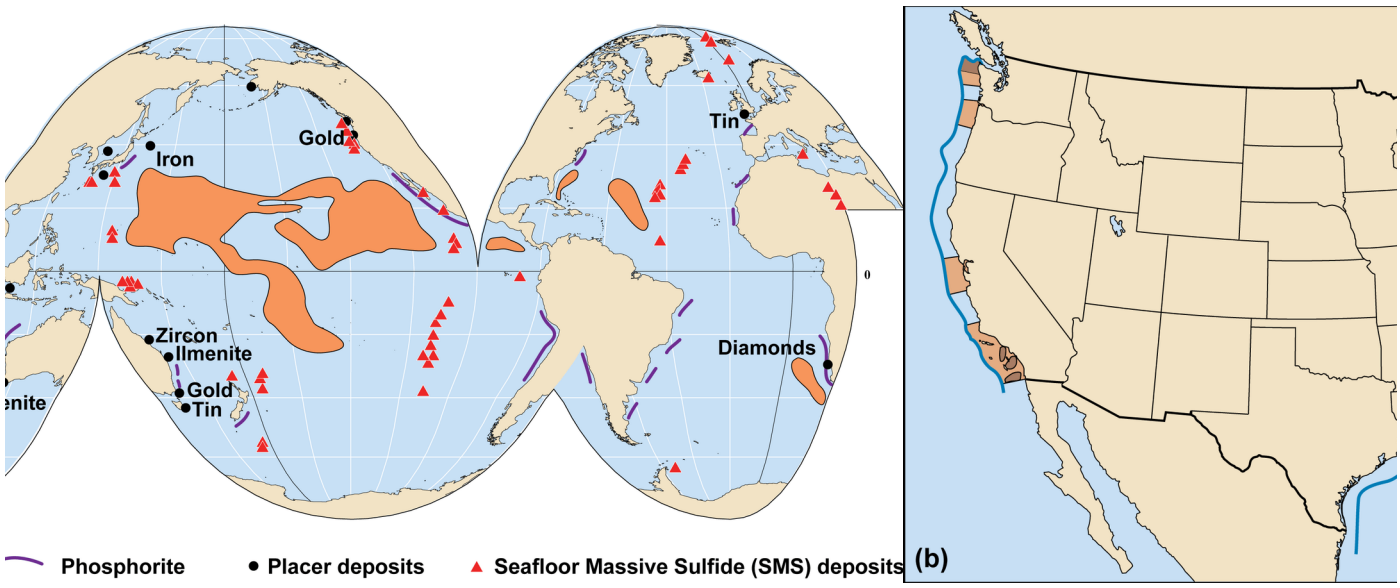


Figure 2-14. Some mineral resources of the oceans. (a) The principal mineral resources of the seafloor are manganese nodules that occur in areas of abyssal seafloor, phosphorite nodules and deposits that occur primarily on the continental shelf, and placer deposits (minerals containing gold and other heavy metals) that accumulate at river mouths. There are a number of commercial mining operations in different parts of the world, as well as other areas where potentially commercial deposits occur. (b) There are abundant sand and gravel resources on the continental shelves of the U.S. Atlantic and Pacific coasts in areas where fine-grained muddy sediments do not dominate.

Many mineral resources, particularly manganese nodules and hydrothermal minerals, are found primarily in the deep oceans far from land. The discovery of such mineral deposits has led to extensive oceanographic research to identify the processes that created them and to determine their distribution and abundance on the seafloor. Deep-ocean minerals are currently too expensive to mine compared to the dwindling, but still adequate, sources on land. However, the potential future value of those resources is

considerable. If deep-ocean mining is developed, it may have significant environmental impacts, especially if large quantities of fine-grained sediment are released into the naturally clear waters of the open-ocean photic zone.

Until 1982, the mineral resources of the deep-ocean floor were not owned by any nation or individual and could legally be mined by anyone. During the 1960s and 1970s, a widespread fear arose in the international community that deep-ocean minerals might be exploited and depleted to the benefit of only one or two nations that commanded the technology to mine them. This fear was a principal driving force behind the negotiations that led to the Law of the Sea Treaty, described earlier in this chapter.

Table salt and freshwater are both produced from ocean waters. Salt is produced by evaporation in coastal **lagoons** (**Fig. 2-15**). Freshwater is produced by evaporation or by reverse-**osmosis** extraction units. The process generates high-salinity brine wastes that are discharged to the oceans. The high salinity may have adverse effects on the **biota**, especially if water temperature is also high and causes additional stress. Such situations occur in locations, such as the Persian Gulf, especially off Saudi Arabia, where freshwater production from seawater is practiced most intensively.

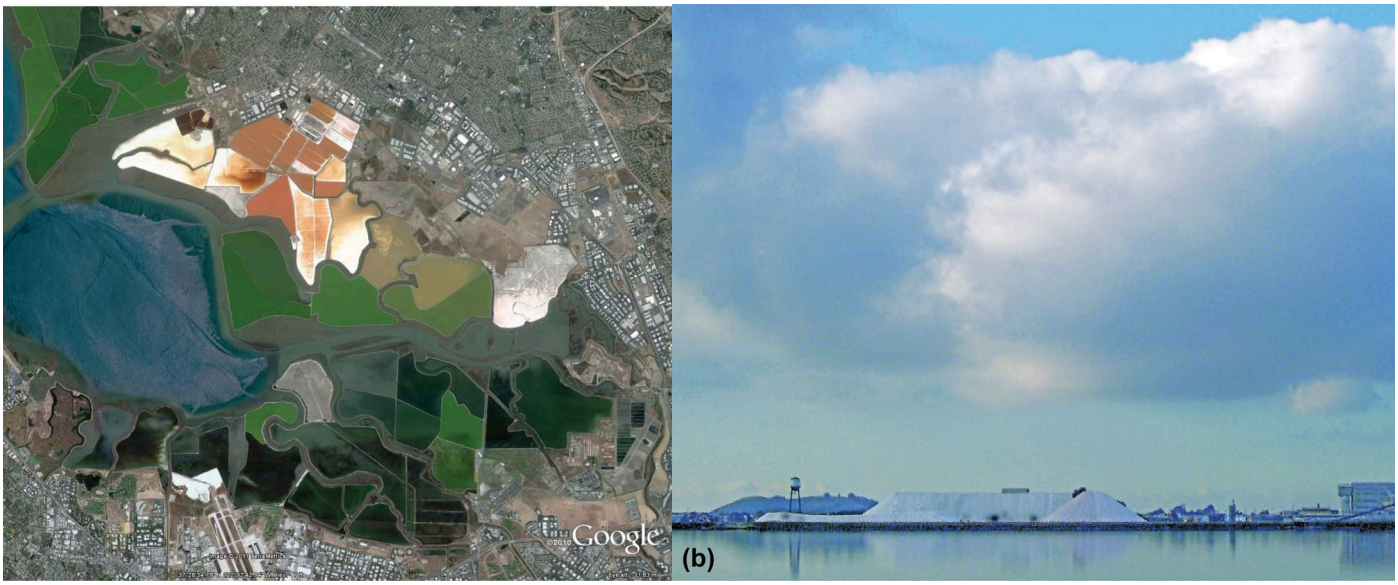


Figure 2-15. Solar evaporation ponds are used to produce common salt in South San Francisco Bay, California. (a) This satellite image shows the evaporation ponds in South San Francisco Bay. The bright red and green colors are due to blooms of different algae species that grow at different salinities as water evaporations. Levees separate the ponds. Many of these ponds have now been restored to wetlands. (b) The large white mounds are piles of salt collected from the ponds and awaiting processing, packaging, and sale.

Recreation, Aesthetics, and Endangered Species

Humans have probably always enjoyed living on the coast, not only for the ocean's food resources, but also for its aesthetic qualities and its moderating effects on climate (**Chap. 8, CC5**). It is the aesthetic qualities of oceans that have inspired poets and artists for millennia. However, only within the past generation or two have the oceans become popular for a variety of recreational activities, including sunbathing, swimming, surfing, sailing, luxury cruises, snorkeling, and scuba diving. As these pastimes have developed, a wider range of people now see the ocean environment as a vital part of nature that deserves protection, unlike much of the despoiled land. Underwater photography and video have introduced a large percentage of the human race to the beautiful, strange, and alien world of marine life.

As these recreational uses have become more popular, there has also been a growing recognition of the effects that such activities have on the marine environment. For example, anchor damage by boats carrying recreational scuba divers has become so severe in some locations that anchoring is now illegal, and boats must tie up to permanent mooring buoys installed at carefully selected locations on the most heavily dived reefs. Scuba divers also can damage coral reefs by breaking coral with their fins or hands. Although incidental contact with the reef and some coral damage is inevitable, most divers are now careful to protect the reefs on which they dive. Damage to reefs by divers is a significant problem only on the most popular reefs, where many divers are in the water every day. Where divers are only periodic visitors, the reefs are able to recover from any minor damage they may sustain. Indeed, damage occurs naturally as a result of the feeding and other activities of large reef animals, such as parrot fishes, and as a result of periodic strong storms. Sometimes such damage is even beneficial to the health and species diversity of coral reef communities (**CC17**).

Energy

As fossil fuels have been depleted, climate change concerns have grown, and nuclear energy has lost favor, a major search has begun for alternate sources of energy. The oceans offer several potential sources of energy, including tides, waves, ocean currents, and thermal gradients. Full development of these energy resources, if possible, would allow substantial reduction in the use of fossil fuels. This, in turn, would decrease the emissions of carbon dioxide and air contaminants, such as sulfur dioxide, nitrogen oxides, and particulates. However, at present very little energy is generated from any ocean resource. Substantial engineering problems must be solved before ocean energy resources can contribute significantly to global energy demands. In addition, the possible environmental impacts of ocean energy generation technologies must be assessed.

The only ocean energy source currently exploited on a large commercial scale is the tides (**Chap. 10**). The technology used is exemplified by the tidal power plant at La Rance, France (**Fig. 10-22**). Because this technology requires a dam across the mouth of a bay or other inlet in which the tidal range is very large, relatively few locations are suitable (**Fig. 10-21**). To generate power during most of the tidal cycle and adjust to fluctuating energy demands during the day, water must be stored behind the dam during part of the tidal cycle and restricted from moving into the inlet during the **flood**. Consequently, tidal currents may be considerably reduced, **residence time** increased, and sedimentation altered in a way that can be detrimental to benthic biota, including valuable shellfish and bottom-dwelling fishes. The dam also hinders the passage of organisms and vessels.

A number of technologies have been proposed for energy recovery from currents. The most developed technology involves placement of huge fanlike turbines in a current. However, many huge turbines would need to be placed side by side across the current to generate significant power. Existing facilities are generally few and small, serving primarily isolated communities.

The nature and scope of possible adverse environmental effects of the proposed technologies are unknown. The major concern may be that modifying currents could have other far-reaching effects. For example, if enough turbines were deployed across the Gulf Stream to generate the power equivalent of several nuclear power plants, energy withdrawn from the current and the **eddy** motion caused by the turbines could conceivably alter the direction and speed of the Gulf Stream to some small extent. Even the remote possibility of such adverse consequences will probably prevent investment of the large sums of money needed to develop this technology.

The energy potentially available from ocean waves and currents is truly enormous, but such energy is widely dispersed and will not be easy to harness. Coastlines where wave energy is greatest, and therefore the potential for generating energy from waves is highest, are shown in **Figure 2-16a**.

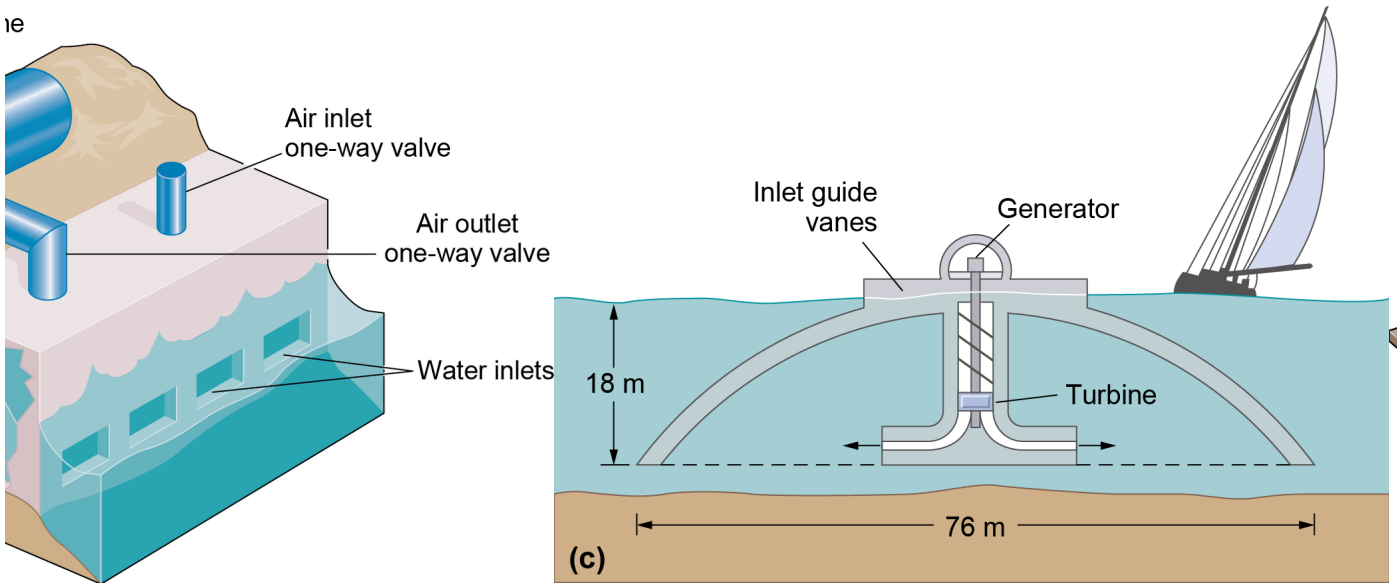


Figure 2-16. Harnessing energy from the ocean. (a) Waves could be harnessed for energy in areas where average wave energy is high. Such areas are primarily in high latitudes (where strong storms are more frequent), on west coasts of continents (because most storms move east to west with the westerlies), and on southern coasts of the Southern Hemisphere (because strong storms are frequent around Antarctica, and no continents limit the fetch or intercept the waves). (b) Various wave energy generators have been successfully used on some rocky coastlines. The design shown here has a chamber open to the sea just below mean sea level into which waves can force water. As water is forced in, the air in the chamber is compressed and drives a turbine. (c) One of several designs for wave energy generators that float on the ocean surface. This one is a "dam atoll." Most of its mass is below the depth of wave motion, such that it does not rise and fall as waves pass. Instead, water enters the generator at the surface, spirals downward, and drives a turbine. (d) A wave energy generator that uses a field of mechanical pumps on the shallow seafloor to pump seawater into a reservoir, where it can be released through a generator. (e) OTEC power plant in Kona, Hawaii. You can see the long vertical pipe bringing in cold deep water and the discharge pipe extending into the ocean at the top of the photograph.

Several design approaches have been developed to extract energy from waves, although none has yet been tested on a large power plant scale. One relatively simple design uses the rise and fall of water at the **shoreline** to compress air within a boxlike structure or shaft, which is open to the ocean only below the waterline (**Fig. 2-16b**). The compressed air drives a turbine. A variant of this design uses the upward and downward surge of water within an enclosed box or shaft to drive the turbine directly. Although these designs are simple and proven, they have limited value because they can be effective only on coasts with persistent strong wave action. In addition, multiple generators must be placed along a coast if large amounts of power are to be generated. Such structures would have adverse effects on the aesthetics of the shore and alter sediment transport along the coast. Generators of this type have been built to serve small coastal communities in Norway. Ironically, one of the first such installations was destroyed by severe storm waves soon after it was built.

Several wave power extraction devices have been designed to be deployed offshore. For example, the "dam atoll" is designed to focus waves toward a central generator shaft (**Fig. 2-16c**). A more recent device design appears to overcome some of the disadvantages of earlier systems (**Fig. 2-16d**). In this design, a matrix of simple mechanical pumps would be placed just beyond the surf zone, where they would use wave energy to pump seawater up into a reservoir on the adjacent land. Power could then be produced, even when the wave energy was low, by releasing the seawater from the reservoir through a turbine. These pumps would be submerged, minimizing the aesthetic impact. However, they would need to cover very large areas of seafloor and fill large reservoirs to produce as much power as a typical power plant, and the pumps would have to withstand wave impacts during even the largest storms.

Several wave energy generator designs have been tested with mixed success, but they would need to be deployed not far offshore in strings stretching along many kilometers of coast. They could be a navigation hazard and might interfere with movements of marine organisms, particularly marine mammals. Even more detrimental, a string of such wave generators along a coast would drastically reduce wave energy at the shore reducing **longshore drift**, changing sediment **grain size**, and altering both the sedimentation regime and the associated biology of the nearshore zone. On rocky coasts, the important **supralittoral zone** with its unique biota (**Chap. 12**) would be particularly affected.

The quantity of ocean thermal energy that potentially can be exploited is huge and may be easier to extract than wave or current energy is. Ocean thermal energy conversion (OTEC) systems exploit the temperature difference between deep and shallow waters in the oceans to drive a turbine and generate electricity. The process by which energy is extracted to run the turbines is analogous

to a refrigerator running backward, or a power plant operating at unusually low temperatures. Conventional power plants use heat from burning fossil fuel or a nuclear reaction to vaporize water in a closed container. The resulting high-pressure steam drives a turbine and is then condensed by cooling water to be recycled.

In OTEC, water is replaced by ammonia or another suitable liquid that is vaporized at much lower temperatures. The ammonia is heated by warm surface waters flowing over heat exchanger tubes through which the ammonia passes. Ammonia evaporates, and the resulting high-pressure gas drives a turbine. Once through the turbine, ammonia is condensed as it passes through another heat exchanger cooled by cold water pumped up from below the permanent **thermocline**. The ammonia is then recycled. An OTEC power plant is essentially built around a wide pipe that reaches to depths from which cold water can be pumped. Such systems have already been tested successfully.

OTEC is very promising and is probably the form of ocean energy generation most likely to contribute significantly to world energy needs. However, OTEC can be used efficiently only where surface waters are warm and cold deep waters are accessible—conditions found year-round primarily in tropical and subtropical latitudes. Locations where deep water is found close to land are ideal because long power transmission lines across the seabed would not be needed. In the U.S., only a few coastal locations, such as Hawaii (**Fig. 2-16e**), where an operating prototype and research OTEC facility has been operational since the 1980s, and the east coast of South Florida, are suitable. Pacific island nations are perfect locations, but the considerable investment needed to develop and build OTEC power plants is difficult for such nations to afford. Several operating OTEC plants are currently located in Japan, China and, on the French island of Reunion in the Indian Ocean. The largest operating OTEC plant is a 100 Megawatt (enough to supply about 50,000 homes) plant in Okinawa, Japan. Continued development in the U.S. is uncertain but negotiations began in 2016 to design and build an OTEC plant on the island of St. Croix in the U.S. Virgin Islands. OTEC may become a more attractive technology if proposed floating OTEC platforms are fully developed. Floating OTEC platform designs could use the electricity they generate to produce hydrogen by electrolysis of water. The hydrogen could be either transported ashore in tankers to be used as a fuel or used to generate ammonia (for use in fertilizer and chemical manufacturing) from atmospheric nitrogen.

Waste Disposal

The oceans have been used for waste disposal for thousands of years and, for most of that time, without significant harm to the marine environment (**Chap. 16**). Indeed, use of the oceans for sewage waste disposal has historically proven to be one of the most effective advancements ever made in human health protection.

Oceanographic research has documented a variety of problems caused by the disposal of certain wastes in parts of the oceans. Although this research has led to the mitigation of many of the worst impacts, much remains to be learned. The oceans will undoubtedly continue to be used for waste disposal, and this may be the most environmentally sound management approach for some wastes. However, a much better understanding of the oceans is necessary to determine which wastes can be disposed of in this way, in what quantities, where, and how, without adversely affecting the environment or human health.

The oceans have a very great capacity to assimilate, safely and completely, large quantities of natural wastes if these materials are widely dispersed or released slowly enough that they are thoroughly mixed into the huge volume of the open oceans. Hence, the oceans suffered little from waste disposal practices until the past century, when human populations and modern industry began to grow explosively. Problems arose when cities and industries grew and became concentrated in large urban areas. This concentration caused the rate of disposal in many coastal and estuarine areas to exceed the rate at which the wastes could be dispersed and assimilated. Additional problems arose with the development of synthetic chemicals and materials, because the ocean ecosystem had no mechanisms to destroy or neutralize some of these substances.

Most of the wastes currently disposed in the oceans are liquids or slurries that are discharged through pipelines, called **outfalls**, into rivers, estuaries, or the coastal ocean. At one time, quantities of a variety of solid wastes, including chemicals, low-level **radioactive** wastes, construction debris, and trash, were transported out to sea on vessels and dumped with the assumption that they would simply fall to the ocean floor, most of which was thought to be lifeless, and not cause any harm. Growing understanding of ocean ecosystems and the effects of ocean dumping have now led to the elimination of almost all dumping of solid wastes in the oceans. Dredged material is now the only waste material dumped at sea from vessels in large quantities.

Waste disposal and its effects in the oceans are discussed in more detail in **Chapter 16**.

This page titled 2.4: Ocean Resources is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.5: CHAPTER SUMMARY

What Is Oceanography?

Oceanography is an interdisciplinary science divided into subdisciplines of physical, chemical, geological, and biological oceanography.

Exploration and Mapping.

The oceans have been used for fishing for more than 100,000 years. Systematic ocean exploration began between about 6000 and 2000 years ago and was concentrated in the Mediterranean, Polynesia, and Micronesia. Subsequently, between about 2000 and 600 years ago, exploration apparently slowed for centuries, except in the Mediterranean.

After about 1400 CE, European and Mediterranean nations mounted many ocean expeditions, and the scientific study of the oceans burgeoned. The voyages of Captain Cook (1768–1780), Charles Darwin and the *Beagle* (1831–1836), and the *Challenger* (1872–1876) conducted some of the most important early systematic oceanographic studies.

Law of the Sea.

For most of recorded history, the resources of the oceans were not owned and could be exploited by anyone. In 1672, the British claimed ownership of a territorial sea that reached 5.6 km offshore. Such claims became common practice until 1945, when President Truman claimed U.S. ownership of the resources of the seafloor offshore to where the ocean reached a depth of 183 m. Many similar claims by other nations ensued. The resulting confusion was resolved by United Nations "Law of the Sea" conferences in 1958 and 1960, and passage in 1982 of a comprehensive Law of the Sea Treaty after more than a decade of negotiations.

The Law of the Sea Treaty grants ownership of all fisheries and mineral resources within an EEZ that is 200 nautical miles wide offshore from each coastal state. The most controversial treaty provisions apply the principle that minerals of the deep-ocean floor outside of EEZs are the "common heritage of mankind." The U.S. and several other nations did not sign the treaty because of these provisions. In 1994, enough nations ratified the treaty for it to become effective, and the U.S. signed it but has not yet ratified it.

The seabed and fishery resources of an EEZ can be very valuable. Consequently, sovereign nations now consider otherwise inconsequential islands valuable. The result has been a variety of territorial disputes, and even wars, over such islands.

In 2023, an addendum to the Law of the Sea Treaty, generally referred to as the "Treaty of the High Seas," was adopted. This Treaty ensures that the benefits of marine genetic resources (e.g. drugs produced from marine animals) are shared under the "common heritage of mankind" principle; establishes rules for conducting impact assessments in the high seas; and establishes a framework for the establishment of ocean sanctuaries within which area fishing will be prohibited.

Value of Ocean Resources.

The oceans provide abundant resources, including fisheries and other biological resources; transportation, trade, and military use; offshore oil and gas; methane hydrates; minerals and freshwater; recreation, aesthetics, and endangered species; energy; and waste disposal.

Biological Resources.

Most of the world's major fisheries are overfished. The principal reason is that most fisheries are an unowned resource open to anyone who wants to exploit them. Many fishing techniques collect and kill nontarget species that are often discarded. Bottom trawling also severely damages slow-growing seafloor ecosystems. Fishing line, nets, and other items are discarded or lost and cause beach pollution. Aside from fisheries and the aesthetic value of marine species, ocean biological resources include

ornamental species used in aquariums, and pharmaceuticals extracted or developed from marine species.

Transportation, Trade, and Military Use.

Most goods traded by humans are transported in surface vessels. The oceans are also extensively used by surface vessels and submarines of the world's navies. Research to support naval uses, especially hiding and hunting submarines, has been critical to the development of oceanography. Recreation on cruise liners and small craft is increasing rapidly. Occasional spills, especially from oil tankers, releases of oil from boat motors and ship engines, dumping of trash by some vessels on the high seas, construction of ports and portside facilities, discharge of bilge water, antifouling paints, anchors, and dredging to maintain navigation channels each have environmental effects, often deleterious.

Offshore Oil and Gas.

Most undiscovered oil is beneath the continental shelves and slopes. Offshore drilling and production platforms or islands are used throughout the world. These facilities have, on a few occasions, accidentally spilled large amounts of oil, and some of them discharge drilling muds and oily water.

Methane Hydrates.

Large accumulations of methane hydrates have been discovered in ocean sediments, especially on the continental slope. Because methane burns cleanly, methane hydrates could provide a large and desirable source of energy. Several nations are now testing technological means for extracting the methane economically. Environmental impacts are uncertain but could include accidental release of methane (a greenhouse gas) and disturbance of continental slope sediments that could cause turbidity currents and tsunamis.

Minerals and Freshwater.

Ocean mining for sand and minerals, although limited, is increasing. Mining alters habitat at the mine site and may lead to the discharge of tailings and other wastes. Coastal wetlands in some areas have been altered to construct evaporation ponds for salt production. Freshwater production from seawater generally discharges high-salinity brines.

Recreation, Aesthetics, and Endangered Species.

Human populations have historically been concentrated on the coast for its aesthetic values and moderation of climate. These values are becoming increasingly important as human use of the oceans for many forms of recreation increases. Humans and their recreational activities have many effects on the marine environment that are becoming better studied and understood. The protection of ocean ecosystems and species has emerged as an important goal, but difficult conflicts often occur between the need for protection and the need for resources.

Energy.

Winds, waves, tides, currents, and the temperature difference between surface and deep waters are all potential sources of energy. So far, only tidal energy is commercially developed. Facilities to generate energy from these sources may alter current, wave, and habitat characteristics and cause contamination from antifouling paints. OTEC may be the most promising technology, particularly for tropical island communities and is now in use in several locations outside the U.S.

Waste Disposal.

The oceans are capable of assimilating large quantities of some wastes without any significant negative impacts. Wastes have been disposed of in the oceans for thousands of years, but increasing amounts have caused significant negative impacts on the ocean environment, especially when dumped or discharged in locations or by means that allow them to accumulate locally. Except for dredged material, most wastes now disposed of are liquids or slurries, such as treated sewage.

2.5.1: Key Terms
2.5.2: Chapter Questions
2.5.3: References

This page titled 2.5: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.5.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- acidification
- acoustic
- atolls
- benthos
- biota
- bottom trawling
- brines
- bycatch
- chemosysthesis
- chronometer
- coastline
- continental shelf
- continental slope
- coral reefs
- currents
- deoxygenation
- drift net
- ecosystem
- eddy
- eutrophication
- environment
- erosion
- exclusive economic
- zone (EEZ)
- flood
- foul
- fossils
- grain size
- habitat
- high seas
- hydrographic
- hydrothermal minerals
- hydrothermal vent
- jellies
- lagoon
- latitude
- longitude
- longshore drift
- manganese nodules
- marginal seas
- marine mammals
- maximum sustainable yield
- monsoon
- osmosis
- overfished
- outfall
- phosphorite nodules
- photic zone
- plate tectonics

- reefs
- residence time
- schools
- scuba
- shellfish
- shoal
- shore
- sounding
- species
- submersible
- supralittoral zone
- territorial sea
- tides
- topography
- weather

This page titled 2.5.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.5.2: Chapter Questions

Study Questions

1. Why did oceanography develop as an interdisciplinary science?
2. Why was the art of navigation developed to a much greater degree by the early Polynesian and Micronesian civilizations than by the contemporary European civilizations?
3. Why were early mapmakers able to measure distances accurately in a north–south direction but not in an east–west direction?
4. Why was the *Challenger* expedition so important to the development of oceanography?
5. Why don't we generate a significant amount of energy from tides, waves, and currents?
6. List the human uses and resources of the oceans.
7. Why is an international treaty that establishes a special set of laws for the oceans important?
8. Discuss how you would apply the "common heritage of mankind" principle to the protection and hunting of terrestrial wild animals whose habitats cross national boundaries

Critical Thinking Questions

1. During the early part of the Dark Ages, observations about the oceans were made by people we now call "philosophers." Today those who study the oceans are known as "scientists," not philosophers. Do you agree with this distinction? Why? What are some differences between philosophy and science?
2. In 1492, Columbus used Poseidonius's highly inaccurate calculation for the circumference of the Earth to identify the New World as Asia. Recall that Eratosthenes had calculated the circumference of the Earth with great accuracy well before Poseidonius. For what reasons do you think Columbus used the later, incorrect circumference?
3. Darwin's observations about evolution are well known. His ideas of continent movement (mountains once having been under the sea, for example) are just now being noticed. Why do you suppose this is?
4. For what reasons do you think the Law of the Sea Treaty established a limit of 200 miles for exclusive economic zones (EEZs)? Why do you think the treaty allows some nations to establish limits that are farther offshore than 200 miles? Explain why you think such exceptions should or should not be allowed.
5. Who has the right to decide who can use the biological and mineral resources of the oceans in areas outside the EEZs of the nations? Who should benefit economically from this use? How should use of these resources be managed, and by whom? If a nation disagrees with a management decision and allows its citizens to exploit these resources without following the internationally agreed-upon rules, how should the situation be resolved?
6. Who do you think should finance the search for beneficial drugs and pollution-fighting organisms in the oceans? Who should benefit economically from the discovery of such commercially valuable compounds? Who should benefit from harvesting organisms to make the compounds? How should the harvesting be regulated, and by whom?
7. Nearly 71% of the Earth's crust lies under its oceans. Available supplies of valuable minerals and oil and gas in the deposits beneath land are steadily becoming scarcer as they are depleted. Should minerals and oil and gas be removed from under the ocean? What factors need to be considered to make this decision?

This page titled 2.5.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

2.5.3: References

Figure 2-1 – Cantillo, A. 1997: The dawn of ocean sciences. Oil on canvas. With permission D. Segar and E. Stamman Segar. Private collection.

Figure 2-2 – Adapted from Bibi Saint-Pol, 2006: The world according to Herodotus. Accessed 6 June 2025, Source, CC BY-SA 4.0.

Figure 2-3 – Segar, D., and E. Stamman Segar, 1997: A typical Polynesian outrigger canoe.

Figure 2-4 – HongKongHuey, 2009: Princess Taiping sails with the Hokule'a in Hawaii. Accessed 6 June 2025, Source, CC BY 2.0.

Figure 2-5 – Roberts, Henry, 1789: A general chart exhibiting the discoveries made by Captn. James Cook in this and his two preceeding voyages, with the tracks of the ships under his command. The National Library of Australia Trove, accessed 6 June 2025, Source.

Figure 2-6 – Franklin, B., 1770: Map of the Gulf Stream by Benjamin Franklin. NOAA Photo Library, accessed 6 June 2025, Source, Public Domain.

Figure 2-7 – Encyclopædia Britannica, Inc.: Charles Darwin: HMS *Beagle* voyage. Accessed 6 June 2025, Source.

Figure 2-8a – Mitchell, W. F., 1880: HMS *Challenger*. Lithograph, tinted. National Maritime Museum, accessed 6 June 2025, Source.

Figure 2-8b – Adapted from Wild, John James, 1877: Plate 2: Contour chart of the bottom of the ocean. *Thalassa; an essay on the depth, temperature, and currents of the ocean*. London, M. Ward & Co., Smithsonian Libraries, Internet Archive, accessed 6 June 2025, Source.

Figure 2-9 – Adapted from Marine Regions, 2025: Maritime boundaries. Accessed 6 June 2025, Source.

Figure 2-10 – Adapted from Asia Maritime Transparency Initiative, 2025: Maritime Claims of the Indo-Pacific. Accessed 6 June 2025, Source.

Figure 2-11a – Data from Takasuka, A., Y. Oozeki, H. Kubota, and S. Lluch-Cota, 2008: Contrasting spawning temperature optima: Why are anchovy and sardine regime shifts synchronous across the North Pacific? *Prog. Oceanogr.*, **77**, 225–232, Source.

Figure 2-11b – Data from Kruse, G. H., F. C. Funk, H. J. Geiger, K. R. Mabry, H. M. Savikko, and S. M. Siddeek, 2000: Overview of state-managed marine fisheries in the central and western Gulf of Alaska, Aleutian Islands and southeastern Bering Sea, with reference Steller Sea Lions. Regional Information Report 5J00-10, 185 pp, Source.

Figure 2-12 – Adapted from U.S. Energy Information Administration, 2024: Oil and petroleum products explained: Offshore oil and natural gas. Accessed 6 June 2025, Source.

Figure 2-13a – Segar, D., and E. Stamman Segar, 1997: OIl rigs in Galveston, Texas.

Figure 2-13b – Segar, D., and E. Stamman Segar, 1997: Oil rigs in Cooks Inlet, Alaska.

Figure 02-14a – Adapted from H.V. Thurman, 1994, Introductory Oceanography, 7th Edition, Macmillan Publishing Company, New York, NY;

Figure 2-14b – Adapted from U.S. Energy Information Administration, 2024: Oil and petroleum products explained: Offshore oil and natural gas. Accessed 6 June 2025, Source.

Figure 2-15a – Stevens, J., M. Carlowicz, and J. Bourgeois, 2015: From salt production to salt marsh. NASA Earth Observatory, accessed 6 June 2025, Source.

Figure 02-15b – Photo by Kevin on Flickr as “A Mountain of Salt”, Licensed under Creative Commons Attribution 2.0.

Figure 2-16a – Adapted from Changery, M. J., and R. G. Quayle, 1987: Global coastal wave energy resources. *Sea Frontiers*, **33**, No 4, 260–261. National Climatic Data Center, U.S. Department of Energy, accessed 6 June 2025, Source.

Figure 2-16b – Adapted from Duxbury, A. C., and A. B. Duxbury, 1994: An introduction to the world's oceans, 4th Ed. Wm. C. Brown Publishers, 472 pp, Source.

Figure 2-16c – Adapted from Wirt, S. L., 1996: Scattering and absorption of waves by Author's Island. Accessed 6 June 2025, Source.

Figure 2-16d – MIT Media Library, 2006: Wave and tidal energy. Accessed 6 June 2025, Source.

Figure 2-16e – United States Department of Energy, 2005: OTEC in Hawaii. Accessed 6 June 2025, Source.

AAPA: 2024: America's ports are responsible for $2.89 trillion in economic activity and 21.8 million American jobs per impact study. Accessed 6 June 2025, Source.

CLIA, 2025: North America. Accessed 6 June 2025, Source.

NOAA Fisheries, 2023: Atlantic striped bass populations. Accessed 6 June 2025, Source.

UN FAO, 2025: The status of fisheries resources. Accessed 6 June 2025, Source.

This page titled 2.5.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3: STUDYING THE OCEANS

Critical Concepts

- **CC1 Density and Layering in Fluids**: Water in the oceans is arranged in layers according to water density.
- **CC6 Salinity, Temperature, Pressure, and Water Density**: Sea water density is controlled by temperature, salinity, and to a lesser extent pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Movements of water below the ocean surface layer are driven primarily by density differences.
- **CC10 Modeling**: Complex environmental systems, including the oceans and atmosphere, can best be studied by using conceptual and mathematical models.
- **CC14 Phototrophy, Light, and Nutrients:** Chemosynthesis and phototrophy (which includes photosynthesis) are the processes by which simple chemical compounds are made into the organic compounds of living organisms. The oxygen in Earth's atmosphere is present entirely as a result of photosynthesis.

Figure 3-1. Developments in scuba diving technology are revolutionizing marine sciences. The strange-looking backpack on the diver shown in the image on the left is a scuba rebreather system that removes carbon dioxide that the diver breathes out so that the gases can be rebreathed many times. Rebreather systems use a computer to control the mixture of air, oxygen, and helium that the diver breathes allowing divers to make very deep dives and return to the surface safely without the many hours of decompression that are usually needed. The spectacular nudibranch shown in the photo to the right is one of many never-before-seen species found on a number of very deep dives made with the rebreather in Papua New Guinea. It was collected from a depth of about 130 m. Scuba is normally limited to a maximum depth of about 50 m.

Over 80% of the ocean remains unexplored. Far less of the deep oceans have been fully explored. For example, humans have observed the deep ocean floor in real time only from submersibles or by video from remotely operated vehicles (ROVs) and these observations are estimated to have covered less than 0.001% of the ocean floor. This chapter will introduce you to the tools and techniques used by oceanographers and explain why the oceans are so difficult to study. Some of the tools described in the chapter are now obsolete, but only by understanding how ingenious oceanographers need to be to overcome the difficulties of studying the oceans can we appreciate why we still do not know more about them. This understanding is also necessary for a full appreciation of the immense value of technologies, such as the Global Positioning System (GPS), advanced satellite sensors, autonomous underwater vehicles (AUVs) and computers that can rapidly analyze and graphically portray massive amounts of data.

3.1: Difficulties of Studying the Ocean Environment
3.2: The Unseen Domain

3.3: Bathymetry
3.4: Seafloor Sediments
3.5: Chemical and Physical Oceanography
3.6: Sampling Living Organisms in the Sea
3.7: Scuba, Manned, and Unmanned Submersibles
3.8: Satellites
3.9: Computers and Modeling
3.10: Chapter Summary
3.10.1: Key Terms
3.10.2: Chapter Questions
3.10.3: References

This page titled 3: Studying the Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.1: Difficulties of Studying the Ocean Environment

Why did it take so long to discover the most fundamental secrets of the oceans? Why did we not know of the existence of the immense mountain chains passing through all the oceans until we were already looking beyond the Earth and launching satellites and humans into space? The answer lies in the hostility of the oceans to oceanographers and to their instruments. In many ways, the ocean depths are more difficult to explore than the surface of the moon or Mars.

A principal focus of oceanographers in developing techniques and instruments to study the oceans has always been, and still is, to overcome the many problems unique to the oceans. The most significant of these problems are the following:

- Visiting the ocean depths is difficult because we cannot breathe in water
- Water absorbs light and other electromagnetic radiation, such as radar and radio waves, severely limiting their use for remote sensing in the oceans
- The oceans are extremely deep
- Pressure in the ocean depths is extremely high
- Seawater is corrosive
- The sea surface is dynamic

“Seeing” through Ocean Water

Compared to the atmosphere, water is a much more efficient absorber of electromagnetic radiation, including radio and radar waves and ultraviolet, infrared, and visible light. In all but the shallowest areas, the seafloor cannot be seen by the naked eye or with any type of optical telescope. Even in the clearest ocean water, we see at best a distorted image of the seafloor, and only where the maximum depth is a few tens of meters at most. Because we cannot see the seafloor, mapping the ocean floor was more difficult than mapping the surface of the moon. Only in the 1920s did oceanographers discover that sound waves could be used as their “eyes” to see the seafloor. Oceanographers also discovered that the magnetic and gravity fields of the seafloor could be sensed through the depths of ocean water. Even so, our ability to study the deep ocean is still limited by its lack of transparency to electromagnetic radiation. For example, radar and other instruments carried on satellites can produce extraordinarily detailed maps of the planet’s land surface in a matter of days, but they cannot map the seafloor directly because most electromagnetic radiation cannot penetrate the depths of the oceans. However, satellite sensors can map the seafloor topography indirectly by making very precise measurements of sea surface height. Satellite instruments can also be used to produce excellent maps of ocean surface features, including wave patterns, sea surface temperatures, and the abundance of photosynthetic life in the near-surface waters.

Inaccessibility

The average depth of the oceans is 3800 m, and the greatest depth is 11,040 m. These depths are farther below sea level than the average and greatest elevations of the land are above sea level. The average land elevation is 840 m, and the maximum elevation, at Mount Everest, is 8848 m. Most of the ocean floor is as remote from sea level as the highest mountain peaks are. Put another way, most commercial airplanes fly roughly as high above the land as the deepest parts of the ocean are below the sea surface.

Until the recent development of autonomous underwater vehicles, oceanographers had to lower instruments or samplers, usually on a wire, and then haul them back up to the ship to take a sample of the deep-ocean waters or **sediment**. Because the oceans are so deep, the process of lowering an instrument or sampler, probing or sampling the water column or seafloor, and retrieving the instrument is extremely time-consuming. Getting a single sample of mud or bottom water at one place on the ocean floor usually takes many hours. In contrast, a scientist studying the land can collect many samples of rock, soil, plants, and animals much more efficiently. AUVs can now perform some of these sampling missions, but they too take a lot of time to travel from surface to seafloor and back.

Research vessels, most of which travel at only about 20 $km \bullet h^{-1}$, consume large amounts of time and fuel going to and returning from sampling locations far from land. Until the development of the satellite-based Global Positioning System (GPS) for civilian use in the 1980s navigation far from land was difficult. Determining the location of a specific sampling site, or its relocation to resample, was much more difficult than on land. Because research vessels operating in the open ocean can cost tens of thousands

of dollars a day to operate, the large amount of time needed to sample the deep oceans means that few samples can be collected during any oceanographic cruise and that each sample is very expensive to obtain. Therefore, samples of the seafloor, oceanic waters, and organisms living in the oceans have been obtained only at intervals of tens of kilometers throughout most areas of the oceans, especially the deep oceans. The advent of AUVs lowers the cost and allows much greater sampling frequency, especially sampling of some water properties, and drastically lowers costs.

Pressure

The pressure of the atmosphere at sea level is about 1.03 $kg \bullet cm^{-2}$, or 1 atmosphere (atm). On a journey to space, a space capsule is subject to a 1-atm pressure change because the atmospheric pressure in outer space is effectively zero. Therefore, manned spacecraft must have hulls that can withstand a 1-atm pressure difference. Because most electronic equipment can operate without any problem at zero atmospheric pressure, unmanned satellites need no protection against pressure differences. In contrast, on a journey into the oceans, the pressure increases by 1.03 $kg \bullet cm^{-2}$ (or an additional 1 atm) for each 10 m of depth. Hence, the pressure at 100 m is 11 times as high as the pressure at sea level (1 atm of air pressure plus 10 atm of water pressure).

In the deepest part of the oceans, at 11,000 m, the pressure is a truly astounding 1101 times as high as atmospheric pressure, or over 1100 $kg \bullet cm^{-2}$, which is more than a tonne of pressure per square centimeter. Therefore, manned submersibles designed to dive to the greatest depths of the oceans need hulls capable of withstanding a greater than 1000-atm pressure difference. Most submersibles are not designed to dive that deep, but even shallow dives to 1000 m require hulls that can withstand a greater than 100-atm pressure difference. Submarines and submersibles must have hulls of thick metal, and viewing ports of thick, durable glass or plastic. Deep-diving manned submersibles must be massive, even when made of strong, light materials, such as titanium. In addition, submersible hulls must withstand the metal-fatiguing stresses of repetitive pressurization and depressurization. These requirements make deep-diving submersibles almost prohibitively expensive.

Conductivity, Corrosion, and Fouling

Seawater poses a problem for unmanned instrument packages because most of these rely on electrical components. Such components will not work if immersed in seawater, because seawater conducts electricity and causes short-circuiting. Oceanographic instruments must be placed inside watertight containers called “housings” that must be able to withstand oceanic pressures, because the interiors of the housings normally remain at atmospheric pressure.

Seawater is extremely corrosive, as divers and other water sports enthusiasts quickly discover when they forget to wash their equipment with freshwater after use. Therefore, all wires, cables, sampling devices, and instrument housings must be protected. Iron and most steels corrode quickly in seawater, so special marine-grade steel or other materials must be used to minimize corrosion. These materials were not available to early oceanographers, who used more expensive and heavier materials, such as brass and bronze. Steel is still the best material available for wires to lower and raise most instrument packages or samples. Even so, the most corrosion-resistant steel wires must usually be further protected by a coating of grease or plastic. Most measurements of trace metals and organic compounds dissolved in seawater were useless until the past several decades because of contaminants from corroding wire and metal sampler parts, and from the grease.

In addition to corrosion problems, a variety of marine organisms **foul** instruments that are left in the ocean to record data for days, weeks, or months, as is necessary for some studies. Some marine organisms, such as **barnacles**, quickly adhere to and colonize the surface of virtually any solid material. Instruments that rely on freely moving parts or on a clean surface-to-seawater contact can quickly be rendered inoperable by such biological fouling.

Wave Motion

Perhaps the most obvious difficulty faced by oceanographers is that the ocean surface is dynamic, and research vessels therefore cannot provide a stable platform on which to work, especially in bad **weather**. The perils of working on a rolling and pitching vessel are many. First and foremost, oceanographers must battle seasickness. In addition, they suffer mental and physical fatigue and disorientation caused by working long hours at odd times of day on an unstable platform. (Many shipboard research activities are continuous 24 h a day due to the cost of operating research vessels.) Oceanographers treasure the rare days of calm seas. Satellites now allow some scientists to remain on land and direct scientific studies remotely using fast data and video communications. However, except for those limited tasks that can now be performed by autonomous vehicles, vessels must still be manned by skilled technicians and crews. Besides the personal hardships, there are many dangers and difficulties associated

with the deployment and retrieval of often extremely heavy instrument packages over the side of a research vessel. The sight of heavy equipment swinging wildly on a wire from a crane over the deck of a ship when seas are rough is indeed frightening. Hanging over the side of a ship in a storm to clamp instruments that must be attached at certain intervals to a heaving wire is an experience few people would relish.

Less obvious than seasickness and the perils of equipment deployment and retrieval, but just as difficult, are the problems associated with using scientific instruments and performing scientific experiments in shipboard laboratories. Most scientific instruments are delicate and made to be used in a normal vibration-free and motion-free laboratory environment. The lurching, pounding, and vibrating to which such equipment is subjected at sea quickly expose any weaknesses. Equipment must be specially designed or modified to operate reliably at sea. In addition, all equipment must be clamped or tied down in bad weather.

Logistics

A profusion of other, lesser problems is associated with studying the oceans. For example, on a research vessel hundreds of miles, and therefore days and tens of thousands of dollars, away from port, broken equipment cannot be taken to a repair shop, a technician cannot be called in, and spare parts cannot be picked up at a store. Oceanographers and research vessel crews have become skilled and ingenious at using available materials to fix equipment at sea. Nevertheless, even the greatest ingenuity sometimes fails, and research efforts must be postponed until the next cruise to the appropriate location, which may be several years later. Such postponements can also be caused by bad weather that slows or prevents work at sea, although most ocean research cruises are planned to allow some leeway for bad-weather days.

In the rest of this chapter, just a few of the many and varied techniques, instruments, and samplers used by oceanographers of yesteryear and today are briefly reviewed. As you study this material, keep in mind the difficulties of working in and on the oceans.

This page titled 3.1: Difficulties of Studying the Ocean Environment is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.2: The Unseen Domain

By the middle of the nineteenth century, very few land areas remained unexplored by Western civilization. A wealth of knowledge had been obtained about the sizes, locations, and shapes of the landmasses, and their mountain ranges, plains, rivers, and lakes. However, there still was almost no knowledge of what lay below the surface of the oceans. The oceans were known to cover two-thirds of the Earth's surface and to be very deep, with the exception of limited areas around continents, islands, and **reefs**. It was also known that the near-surface waters of the seas abounded with fishes and other creatures, and that much of the seafloor was covered by various types of sediment. However, as late as the mid-nineteenth century, the deepest ocean waters were believed to lie stagnant and unmoving in the ocean basins, and the deep waters and deep-sea floor were thought to be devoid of life.

This state of ignorance began to change in the latter half of the nineteenth century, when the first telegraph cables were laid on the seabed and the HMS *Challenger* undertook its expedition. Despite subsequent intensive studies, the oceans remained a mystery for decades. For example, not until the 1950s did oceanographers begin to realize that an immense chain of undersea mountains runs through all the oceans. The first comprehensive and generally accurate bathymetric map was produced in 1959. Corrections and refinements continue today.

This page titled 3.2: The Unseen Domain is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.3: Bathymetry

People have long needed to map seafloor topography to navigate safely past obstructions such as submerged rocks and reefs. They have also long been curious about the depth of water over the seafloor.

Soundings

For centuries, the only way to explore any seafloor deeper than the few tens of meters that could be reached by pearl divers or in crude diving bells was by lowering a line into the water with a weight attached. The length of line payed out (on a ship, letting out line is referred to as "paying out") before the weight hit the bottom indicated the water depth—a measurement called a **sounding**. This method led to the unit of depth called the "fathom," which was used almost exclusively for nautical charts until it was supplanted in recent decades by the meter. The fathom was originally 5.5 ft., or the length of line between the outstretched arms of the man hauling the sounding line back aboard the ship. Originally, all depths were measured as a count of the number of such lengths of line that were hauled back aboard after the sounding weight hit the bottom. Later the fathom was changed to exactly 6 ft., and sounding lines, particularly those used for deeper soundings, had knots or ribbons tied at measured intervals to improve the sounding accuracy.

A modification of sounding with line and weight was used to collect samples of bottom sediment. The bottom of the weight, usually lead, was hollowed out and fitted with a lump of tallow. When the tallow hit the bottom, a small amount of bottom sediment adhered to it, unless the weight hit a rocky bottom. Early nautical charts included a description of the type of seafloor based on such samples. Seafloor composition was categorized as sand, silt, or mud, with the mud color sometimes noted. Many modern nautical charts, particularly those of shallow coastal waters, still include that information.

Sounding technology saw a major technological advance during the 1885 voyage of the *USS Tuscarora* to study possible routes for a telephone cable between America and Japan. On the ship's voyage, the sounding line was replaced by a single strand of piano wire with a weight attached at its end. The wire was deployed from a drum and hauled back by a winch. Because hauling with a winch was much faster than hand-hauling, the Tuscarora could make several deep soundings each day. Despite this improvement, fewer than 6000 soundings had been made in depths greater than 1000 fathoms (1800 m) by 1910. Hence, fewer than 6000 depth measurements had been made in an area that covers approximately 40% of the Earth's surface—an area nearly 30 times larger than the combined surface area of the 48 contiguous U.S. states. Knowledge of the topography of the deep-ocean floors was only rudimentary, even as late as the beginning of World War I in 1914. For comparison, consider a map of the mountains, plains, and river valleys of the U.S. if the country were covered in clouds and had been studied solely by lowering a wire through the clouds to the ground at only 200 locations.

Sounding Errors and Problems

Taking soundings using a line or wire is a tedious process that poses additional problems, many of which must still be overcome when wires are being used to lower instrument packages. For example, determining when the weight on the wire has reached the bottom is difficult in deep water. Several kilometers of line or wire is sufficiently heavy to continue to pull more wire from the drum, even when the weight at the end has hit the bottom. Watching for a reduction in how fast the wire pays out or for a slackening of the tension in the wire, each of which occurs when weight is reduced by bottom impact, can sometimes help to overcome this problem. However, these techniques are very difficult, especially in bad weather, even using electronic wire-tension measurement systems. As the ship rolls, the head of the crane or the A-frame over which the wire is payed out moves up and down in relation to the sea surface. When a considerable length of wire has been payed out, its weight and drag in the water prevent it from moving up and down with the ship's roll. As a result, it stretches and contracts, and the wire tension fluctuates. Therefore, wires must be several times stronger than would be necessary to carry only their own weight and the weight of any instruments attached to them.

When wires are used to lower instruments, **currents** and wind blowing against the research vessel can also cause problems. Because the ship is slowly blown along the sea surface by the wind, it tends to move sideways from the weight or instrument package on the wire that is far below. The drag of the wire and its weight tend to prevent the wire from following the ship's sideways motion. Therefore, the wire does not drop vertically to the bottom (**Fig. 3-2**). If subsurface currents flow in a different direction from the research vessel's wind drift, or from that of currents at other depths, the wire's vertical path through the water can have a complex S shape or other curve (**Fig. 3-2**). Because a wire never falls through a deep water column vertically, more wire

than the actual water depth beneath the research vessel must be let out if the end of the wire is to reach the seafloor. Therefore, all line and wire soundings in deep water were incorrect. The actual depth was always less than the measured depth.

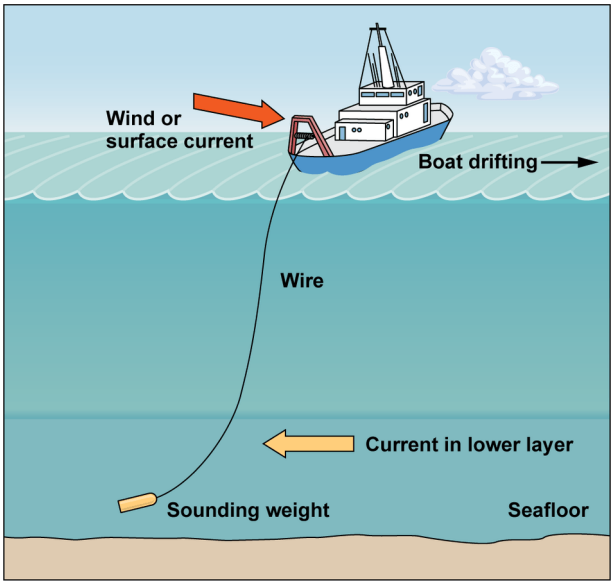


Figure 3-2. Winds and currents can deform the vertical path of a sounding weight or instrument package on the end of a wire as it falls to the seafloor. Currents in the surface layer or winds that blow the vessel across the ocean surface can move the vessel away from the vertical path of the weight. Therefore, a length of wire longer than the water's depth must be payed out for the instruments to reach the seafloor. Currents in lower layers can further complicate the path of the wire, particularly in very deep water where several kilometers of wire must be payed out for the instrument package to reach the seafloor.

A more practical problem associated with lowering instruments or wires over the side of a research vessel is that the wind or currents can blow the vessel over the top of the wire or bend the wire under the vessel. The wire can end up stretched tight against the curve of the ship's hull as it passes underneath. Continuing to lower or raise the wire in this situation could damage the ship's hull or break the wire. Therefore, most oceanographic research vessels are specifically designed to be capable of turning slowly around the wire without moving forward. This ability is usually provided by a bow thruster propeller located in a tunnel near the bow of the ship. The propeller is set at a right angle to the ship's normal direction of travel and can be used to push the bow of the ship to one side or the other.

Echo Sounders

In response to the sinking of the *RMS Titanic* in 1912, Reginald Fessender, a former assistant to Thomas Alva Edison, invented a device that could detect an iceberg almost 5 km away by sending a sound signal through the water and detecting the return echo. That sound navigation and ranging equipment, which became known as **sonar**, was quickly developed into a device for hunting submarines during World War I. After the invention of sonar, it was a simple matter to orient the sound source to point vertically downward (**Fig. 3-3**) and to detect the echo from the seafloor. The speed of sound in seawater is known, as are the relatively small changes in this speed with **salinity** and temperature. Hence, if the distribution of salinity and temperature with depth is known from other measurements, the depth of the water below a ship can be determined by measurement of the time taken for the sound to travel to the seafloor and back. The first truly successful echo sounder depth recordings were made in the North Sea in 1920 by the German scientist Alexander Behm. Subsequently, knowledge of bottom topography developed rapidly as echo sounding equipment was improved and installed in more vessels.

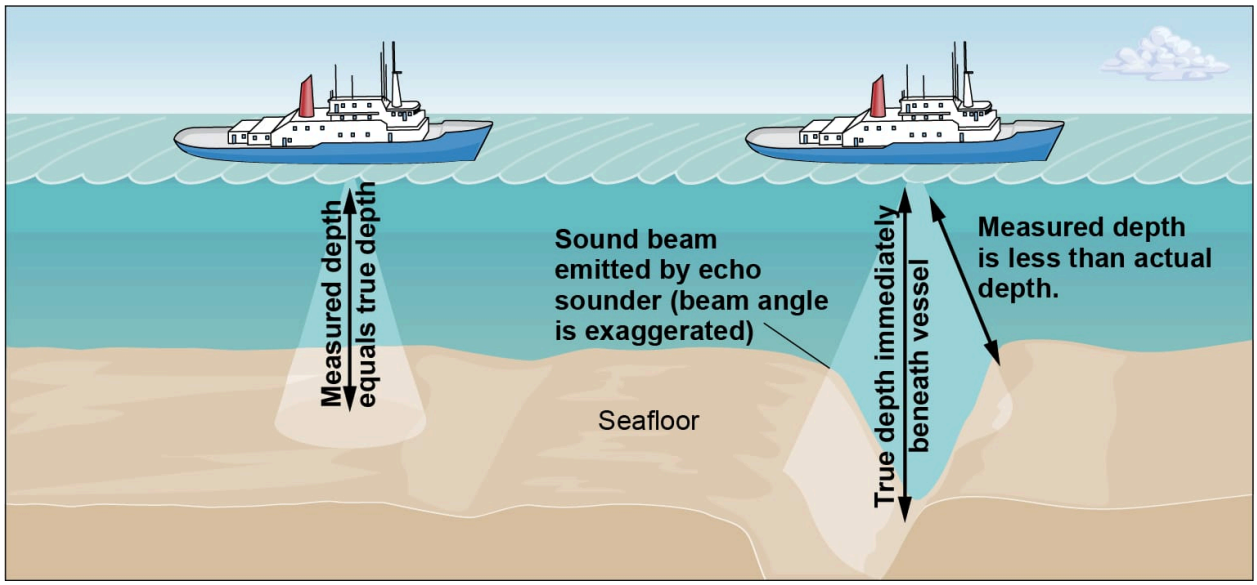


Figure 3-3. Echo sounders measure the depth of water beneath a vessel by measuring how much time a sound pulse takes to travel from the vessel to the seafloor and back. Because sound travels in seawater at about 1500 $m \bullet s^{-1}$, a sound pulse takes 2 s to return to the research vessel when the water depth is 1500 m. The sound pulses spread out over a narrow angle as they travel downward from the vessel. Thus, in deep water, they are reflected off a large area of seafloor.

The great advantage of echo sounders was that they could obtain essentially continuous records of the water depth below a moving ship. By the mid-twentieth century, every research vessel was equipped with an extremely precise echo sounder called a "precision depth recorder" (PDR). Standard operating procedures on most research vessels required that the PDR be operated continuously, and that depths and precise ship positions be recorded while the vessel was under way. PDR depth measurements, although much more accurate than soundings, also have limitations. Because the PDR records only the depth of water directly under the ship's track, depths between two ship tracks still must be inferred by interpolation. Unless depth recordings from many ship tracks cover a given area of ocean floor, major features such as undersea hills and mountains may still be overlooked. In addition, the PDR measures only the depth of the closest echo from under the vessel (**Fig. 3-3**). Even PDRs with very narrow beam widths receive echoes from a relatively large area of ocean floor. Hence, a nearby hill can cause the depth to be recorded as shallower than it really is, and narrow valleys and depressions can be missed completely (**Fig. 3-3**).

The depth information generated by research vessels after World War II had to be plotted by hand, which proved to be a daunting task. Of course, depths are now recorded electronically and processed by computers, but computers did not become practical for use at sea until the 1970s. The first truly comprehensive map of the ocean floor was completed by Bruce Heezen and Marie Tharp in 1959. Compiling it was an enormous and tedious undertaking that involved matching depth and position (navigation) data from thousands of hours of PDR recordings made by many vessels without the aid of computers. Navigation errors were corrected by comparison of the depths of each ship, recorded where ship track lines crossed. Depth data were then entered painstakingly on a blank map and carefully contoured, and then a three-dimensional representation was drawn exactingly by hand. The map generated by this massive project was truly revolutionary.

To understand this, look at a typical atlas map that shows all the oceans as a featureless uniform blue expanse, then look at **Figure 3-4**. This revelation of previously unseen seafloor topography that rivals the greatest mountain chains and other features of the continents was, for oceanographers and others, like being introduced to an entirely new planet. Heezen and Tharp, and the many other people who spent years gathering data or otherwise helping to create the map, made a contribution to human knowledge that today remains startling, profound, and beautiful.

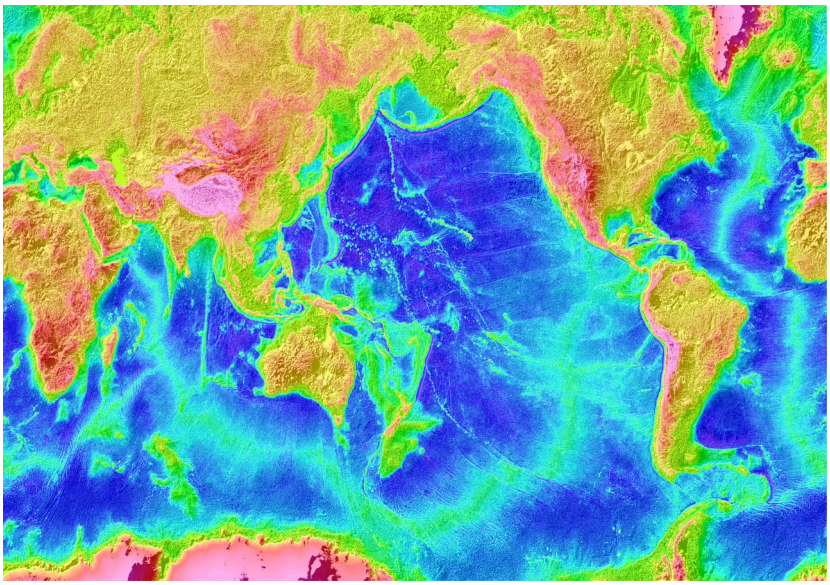


Figure 3-4. Mercator projection map of the seafloor and land surface elevation. Seafloor elevation data were obtained from ship generated precision depth recorder data and satellite measurements of sea surface height data. (Data were obtained from many sources.) The highest resolution sea surface height data used in this map was from satellites, permitting only features greater than about 20–25 km across to be shown in this map. Click for a larger version of the map.

Since 1959, PDR surveys of the oceans have continued on almost all research vessels during their entire time at sea. Nevertheless, most of the ocean floor is still very poorly mapped by this technique. In fact, vast areas of the deep oceans have been mapped by PDRs at a level of detail equivalent to mapping the U.S. only by measuring elevations along the interstate highways. Look at a road atlas and think about what topography we would have missed if we had mapped the U.S. from topographic data taken only along this highway system.

Wide-Area Echo Sounders

In the 1960s, echo sounding was revolutionized by the simultaneous development in the U.S. and England of somewhat different approaches to determining the seafloor topography within a wide swath under, and to either side of, a vessel track. In the American system, called “multibeam sonar” or “swath,” up to 200 or more narrow sound beams are broadcast in a fan pattern beneath the ship (**Fig. 3-5a**), and the depth (corrected for the angle) is recorded for each beam. In the British system, called “sidescan sonar,” two wider sound beams are broadcast at an angle, one to each side and downward from a streamlined instrument enclosure called a “fish” towed underwater behind the vessel (**Fig. 3-5b**). In this system, echoes from different broadcast angles within each beam return to the fish at different times, depending on the angle and therefore the distance from the fish. Within the prolonged returning sequence of echoes from each outgoing pulse, the intensity of the echo received varies according to distance and the bottom topography. Strong portions of the returning echo sequence indicate that the bottom is sloped up toward the fish. Weak echoes indicate an area sloped down away from the fish. These represent the front and back sides of the hill, respectively.

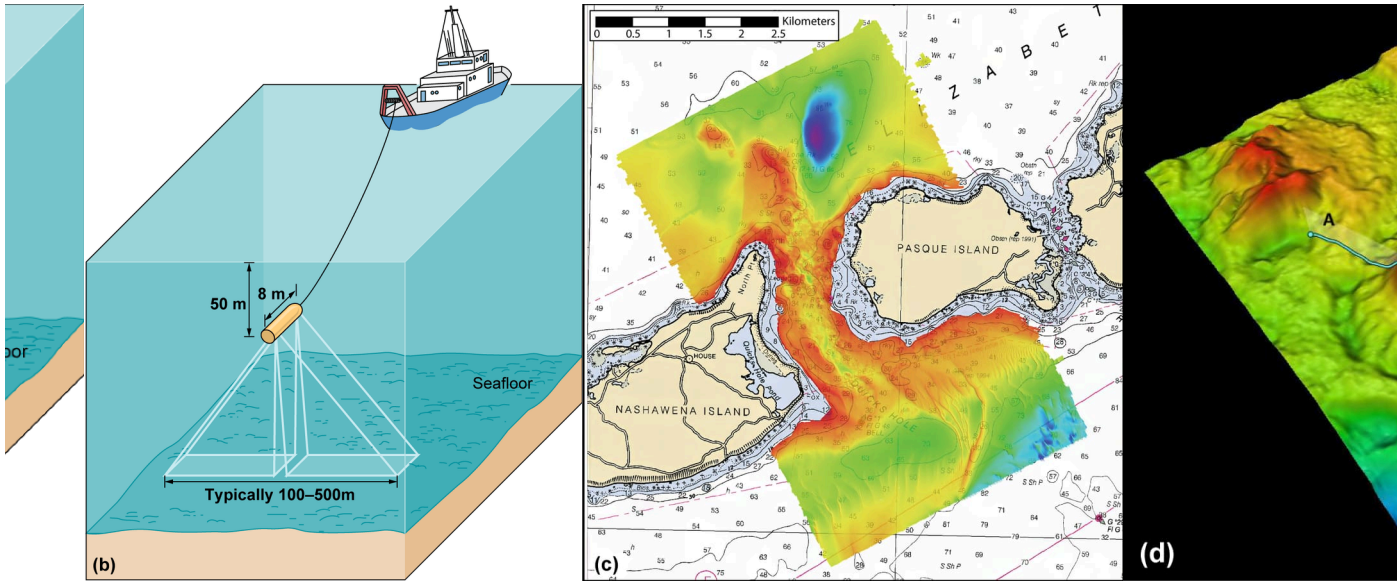


Figure 3-5. Modern seafloor topography measurements have been continuously improving and are now made with two different systems. (a) Multibeam sonar systems use up to 1000 or more separate narrow beams spread out under the ship or towed fish. Both systems require computer processing of return echo data to produce much more detailed seafloor maps than were previously possible. (b) Sidescan sonar uses a broad sound beam sent from a towed fish. (c) The colored area of this map shows the detailed topography revealed by a multibeam sonar survey of Quick's Hole, Massachusetts in 2004. The surrounding white nautical chart shows the much less detailed maps that were available before sidescan or multibeam sonar were developed. (d) This multibeam image of a deep-water ridge north of Sulawesi Island, Indonesia from a scan taken in 2010 shows the improvement in sonar systems.

The two wide-area echo sounding systems have somewhat different uses because each is better suited to mapping certain types of bottom terrain in certain depths of water depths. Each system requires powerful and sophisticated computer technology to process the signals received. Both methods provide dramatically improved maps and reveal previously unknown canyons, valleys, hills, and other features of the seafloor (**Fig. 3-5c**). Detailed charts of coastal waters developed with the new echo sounding systems are now being produced as each area is surveyed.

In the early part of this century, side scan sonar evolved into a newer system, synthetic aperture sonar (SAS). The main difference between side scan sonar and SAS is that SAS uses sophisticated signal processing to coherently recombine the overlapping areas of the pings to create ultra-high resolution images (**Fig. 3-6c**). The return echo data are then mathematically combined for up to several successive pings. This requires far more computing power than even side scan sonar requires but results in a far more detailed view of the seafloor (**Fig. 3-6**). SAS provides greater resolution than side-scan over a comparable area, more information about the sediment type, and details of small objects such as fish holes and burrows. While the resolution of side scan tends to degrade with distance between seafloor and the towed fish, SAS systems have a consistent resolution throughout the full operational range. The resolution of these systems can be degraded by random motions such as those that occur during bad weather when a ship and its towed array are affected by wave and wind action. SAS is now used extensively in marine archaeological surveys, and military applications such as mine detection.

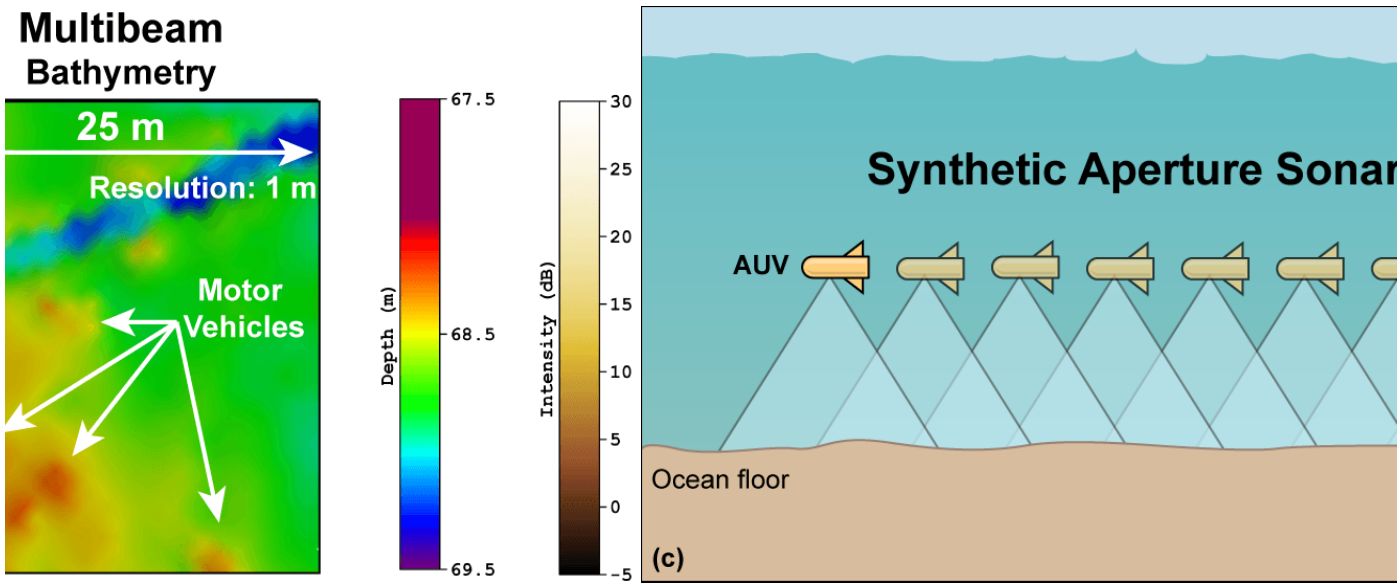


Figure 3-6. (a) Multibeam and sidescan sonar both rely on successive pings and return echos that are timed so there is minimal overlap, and the echo from each ping is computer processed separately. (b) Synthetic aperture sonar pings more often, causing the echo return areas from two or more successive pings to overlap. More complex computations then analyze the data by combining the echo returns from two or more pings. The following parts of this figure show the differences in resolution between the sonar types taken at the same distance from the seafloor (c) Multibeam and sidescan sonar can resolve objects about 1 m across. (d) Synthetic aperture sonar, when processed to extract bathymetric data, can resolve objects several times smaller, and (d) when processed as an image, can resolve objects just a few centimeters across. [Images courtesy of Kraken Robotics]

Multibeam, sidescan, and synthetic aperture sonar are now generally mounted in AUVs as each type of sonar is sensitive to the motions of a ship or towed fish. Also AUV's, which are discussed more later in the chapter, can fly at depth, closer to the seabed being surveyed, and they can survey an area without the need for constant attention from a surface vessel. Such systems can now operate at depths of up to 6,000 m but are expensive to build and operate. Although these systems are beginning to expand the area in which the ocean floor is surveyed in detail, it will still take decades or longer before the majority of the seafloor is surveyed at the level of detail possible with sidescan, multibeam, or synthetic aperture sonar.

Ocean Topography from Satellites

Unlike research vessels and autonomous floats, satellites have been able to survey the world's oceans with unprecedented comprehensiveness in just a few days, albeit at lower resolution than sonar can achieve.

The *Seasat* satellite launched in 1978 made the first satellite-based maps of the ocean surface, from which the topography of the seafloor can be deduced. The sea surface topography was measured by a radar altimeter carried on board the satellite. The height of the sea surface can be used to map the seafloor topography because it is affected by the depth of the ocean below it. Rock is denser than water, so rocks have a slightly higher gravitational attraction than water. Therefore, the sea surface is slightly higher over an undersea mountain than it is over a deeper area (the mountain "pulls" water toward it from the sides to create "mounds"). There are now several satellites in orbit that provide satellite altimetry data and made it possible to map the entire world's ocean floor. This data has led to the discovery of hundreds of previously unknown **seamounts**, **fracture zones**, and other topographic features in areas where depth recordings had not been made by ships. By 2023, high resolution radar satellites had identified more than 43,000 seamounts of which only about 16,000 have been mapped by sonar.

Data from satellites and from shipboard PDR surveys have all been combined to provide a detailed view of the ocean floor (**Fig. 3-4**). However, the resolution of these maps is still such that features as large as several kilometers across may be missed. Limitations inherent to satellite bathymetry make it unlikely that substantially better resolution will ever be obtained except where detailed wide-area sonar surveys are made.

In contrast to ocean floor mapping, scientists completed the mapping of other planets and their moons quickly and easily using radar, cameras, and modern computers aboard planetary probes such as the *Mariner*, *Voyager*, and *Mars Global Surveyor* spacecraft. In addition, the resolution and comprehensiveness of our maps of Mars, Venus, and the Earth's moon are much better than our seafloor maps. This comparison illustrates the difficulty of studying the oceans.

This page titled 3.3: Bathymetry is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.4: Seafloor Sediments

The seafloor is covered by sediment ranging in thickness from zero on a small fraction of the ocean floor to several kilometers. As **Chapter 6** discusses, many secrets of the Earth's history are to be found in these sediments. Because the sediments slowly accumulate layer upon layer, history is preserved in sequence; older sediment is found at progressively greater depths below the seafloor. The upper few to tens of centimeters of sediment are especially important because many living organisms inhabit these sediments, and because the processes that affect the fate of chemicals in and on sediment particles are also concentrated in this zone. Therefore, oceanographers are interested in obtaining two basic types of sediment samples:

- Samples that contain an undisturbed sequence of the layers of sediment from the sediment surface down as far as necessary to cover a long period of history
- Large samples of the top few tens of centimeters of sediment, within which most non-microscopic organisms live

Aside from samples taken in very shallow water, the earliest sediment samples retrieved from the oceans were those that adhered to the lump of tallow at the end of a sounding line. Such samples were useful only for a gross characterization of sediment color and the size and type of the sediment grains. When he explored Baffin Bay in search of a Northwest Passage in 1820, Sir John Ross had his blacksmith construct a "deep-sea clam." That device, the forerunner of grab samplers used today, collected several kilograms of greenish mud containing living worms and other animals from water depths of almost 2000 m.

Grab Samplers and Box Corers

In its basic design, a grab sampler consists of a sealed metal container, usually with two halves that open at a top hinge like a clamshell (**Fig. 3-7a**). The sampler is lowered with the clamshell jaws open. When it hits the bottom, it sinks into the sediments. As it is pulled back out of the sediment, a mechanism causes its two halves to close. It is then retrieved with its sediment sample within the clamshell, thus grabbing a sample of sediment.

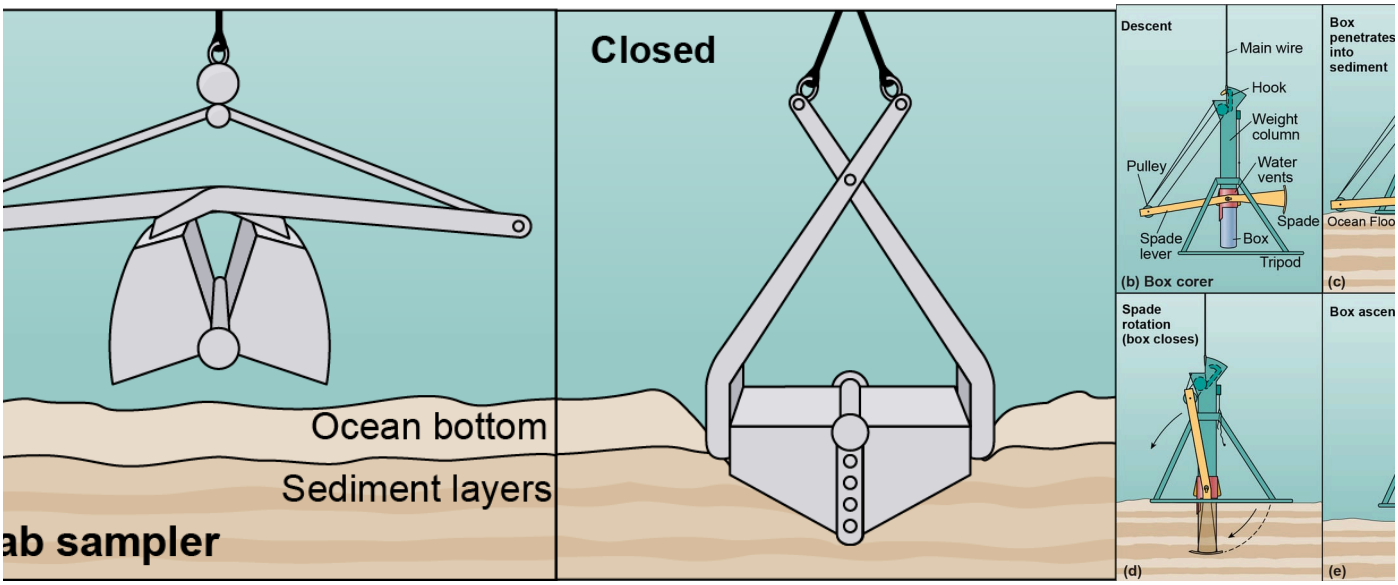


Figure 3-7. Methods of sampling near-surface seafloor sediments. (a) Grab samplers are used to obtain large samples of surface sediments. (b) Box corers are used to collect large-volume undisturbed samples of the upper few centimeters of sediment. (c) The box corer is forced into the sediments by its weight, (d) the box is and is then closed off by a spade arm that cuts through the sediment and under the box. (e) The box with the sample sealed inside is pulled up by its cable and retrieved.

Grab samplers are relatively light and simple to operate, but they can disturb and partially mix the sediment they retrieve. To minimize disturbance, box corers are often used. A box corer consists of a supporting framework that is lowered to sit on the seafloor, a heavily weighted box with an open bottom that sinks about 20 to 30 cm into the sediment, and a blade that slices under the box from the side to hold the sediment in the box when the frame is retrieved (**Fig. 3-7b**). Because they have a wide opening, box corers can collect large amounts of sediment, but they often weigh several hundred kilograms, making them difficult

to deploy from research vessels.

Gravity and Piston Corers

To take deeper sediment samples, corers are used. Corers consist of a tube that is open at the bottom end like an apple corer. The tube is forced vertically into the sediment. When the tube is pulled out of the sediment, the sediment core is usually held inside by a core catcher (**Fig. 3-8**). There are two basic types of corers: the gravity corer and the piston corer. Gravity corers are allowed to fall freely on the end of a cable; they strike the bottom with great force and are driven into the sediment by weights mounted at the top of the core tube (**Fig. 3-8a**). Piston corers, and sometimes gravity corers, are attached to a release mechanism at the end of a cable (**Fig. 3-8b**). In a piston corer, the action of the piston helps the core slide into the core barrel so that longer cores can be obtained, and it helps minimize vertical distortion and disturbance of the core's sediment layers.

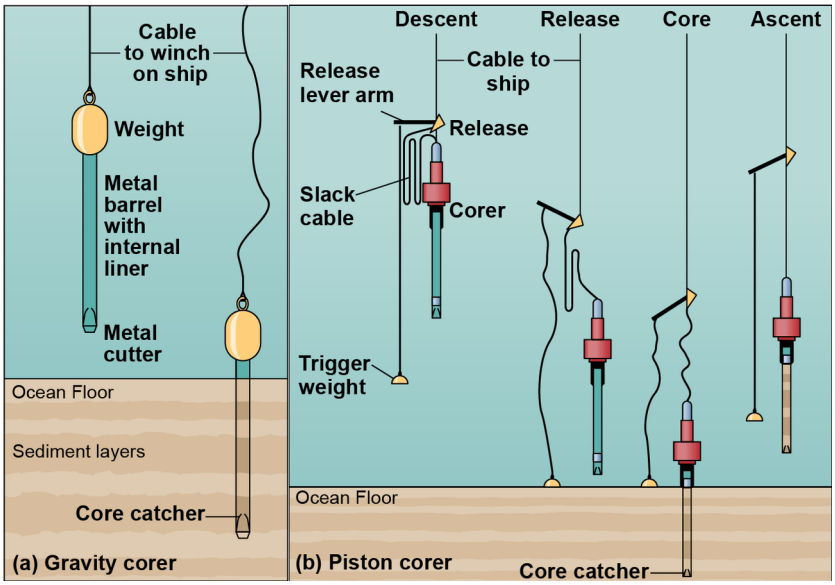


Figure 3-8. Sediment corers are used to obtain long, mostly undisturbed, vertical samples of soft seafloor sediments. Most corers have a core catcher that consists of a ring of flexible metal leaves around the inside of the core tip. The leaves are easily pushed aside as the corer enters the sediments, but they are bent backward to close off the tube as the corer is pulled out. (a) Gravity corers are driven into the sediment by their weight and can obtain cores up to about 4 m long. (b) Piston corers are usually suspended from a release arm, which is held in the closed position by a trigger weight that hangs from the end of the trigger arm to below the corer itself. When the trigger weight hits the bottom, the trigger arm is released, allowing the weighted corer to fall freely into the sediments. When the tip of the core barrel reaches the seafloor, a loop of cable between the trigger assembly and a piston inside the core barrel becomes taut holding the piston in position. The movement of the piston causes a small reduction of the water pressure in the barrel above the sediment, helping the sediment core slide into the barrel. The piston mechanism increases core length to longer than 50 m, depending on sediment type and minimizes vertical distortion of the sediment layers.

Most sediment cores are taken in muddy sediment, but gravity and piston corers cannot penetrate very sandy or some other types of sediment. In these situations, special corers are needed to force the core barrel into the sediment, either by vibrating it mechanically or by forcing air or water down the outside of the barrel to blow the sediment away.

The smallest corers weigh a few tens of kilograms and take short cores (up to about half a meter long); the largest weigh several tonnes and can take cores more than 50 m long. Deploying and retrieving one of the largest corers is an exacting task that is now performed by computer-driven mechanical systems on specially equipped research vessels.

Unfortunately, and to the frustration of researchers who may have waited hours for the sampler to be lowered and retrieved, especially in deep water, grab samplers and corers often fail to penetrate the sediment or close properly. There are several reasons that these failures occur. For example, if currents are strong and deflect the wire from vertical, the sampler might impact the seafloor at too great an angle. In other cases, the sampled material might jam in the closure mechanism. Even a tiny opening in the closure mechanism can cause the entire sample to be lost as the sampler is hauled back through the water column.

Drilling Ships

To study older layers, scientists must explore deeper in the ocean sediment than can be reached by corers. In addition, the bedrock beneath the ocean sediment holds valuable clues to the history of the Earth and its oceans. In 1968, the U.S. began using a unique drilling ship, the *Glomar Challenger*, to sample the deeper sediment layers and rocks. The ship was capable of drilling up to about 2000 m long sediment cores in water as deep as 6 km. The Deep Sea Drilling Program (DSDP) used the vessel to obtain more than a thousand cores from throughout the ocean. These core samples helped to confirm the theories of **seafloor spreading** and **continental drift** (Chap. 4).

In 1983, the DSDP was succeeded by the Ocean Drilling Program (ODP), an international cooperative program funded jointly by the U.S., Canada, West Germany, France, Japan, the United Kingdom, Australia, and, at one time, the Soviet Union. In 1985, the more sophisticated *JOIDES Resolution* replaced the *Glomar Challenger*. In 1990, in a water depth of 5700 m, 2500 km south of Japan, the *JOIDES Resolution* drilled through 200 m of recently formed volcanic rock and 460 m of sediment lying below it. A sample of sediment was retrieved that was estimated to be 170 million years old and is believed to be the oldest remaining ocean floor sediment, except for that found on small fragments of tectonic plates that have avoided subduction.

In 2003, the ODP was succeeded by the Integrated Ocean Drilling Program and then the Integrated Ocean Discovery Program (IODP), led jointly by the U.S. and Japan. This program operated two drilling ships, the *JOIDES Resolution* and a newer Japanese vessel called *Chikyu*. The *Chikyu* can drill holes as deep as 6 km beneath the seafloor, far surpassing the 2-km limit of the *JOIDES Resolution*. The *Chikyu* is also able to drill in shallower water, and it has the ability to prevent blowouts (uncontrolled releases of gas or oil) if it penetrates formations that contain oil and gas. This capability allows Chikyu to drill in many locations where the *JOIDES Resolution* could not. However, in 2023, *JOIDES Resolution* was able to recover the first-ever core sample of Earth's mantle material while drilling on the Mid-Atlantic Ridge. The IODP program ended, and the *JOIDES Resolution* was decommissioned in 2024. Ocean drilling is now performed by Japan's *Chikyu*, but a new drill ship, China's *Meng Xiang* is scheduled to begin operation in 2025 and is expected to be able to drill deeper into the mantle.

Seismic, Magnetic, and Gravity Studies

Because obtaining sediment core or drill-core samples is time-consuming and expensive, oceanographers can only sample a few locations directly. Fortunately, certain remote sensing techniques can provide information about the sediment in areas where actual seafloor samples are unavailable. Seismic profiling and measuring magnetic and gravitational fields are the principal techniques for obtaining such information.

Like sonar measurements, seismic profiling uses a sound wave or shock wave. The sound wave passes through the ocean water into the sediment and is partially reflected at each depth where the type of sediment changes or a volcanic rock layer begins (Fig. 3-9). Because ocean sediments are built up in layers, many echoes, corresponding to the top of each layer, are reflected from within most ocean sediments. Returning echoes are monitored with a string of **hydrophones**, devices that record sound waves, towed behind the research vessel. The sound waves received by each hydrophone have traveled different distances from the top of each sediment layer and are received at different times (Fig. 3-9a,b). Seismic profiles reveal the sediment's structural features hidden below the seafloor, including **faults**, tilting of the layers, reef structures, and buried mountain tops. Modern seismic profiling systems use multiple sound sources and multiple strings of hydrophones towed behind the research vessel. Using powerful computers to analyze the resulting data, these systems can produce detailed three-dimensional images of the sediments or rock beneath the seafloor (Fig. 3-9c,d).

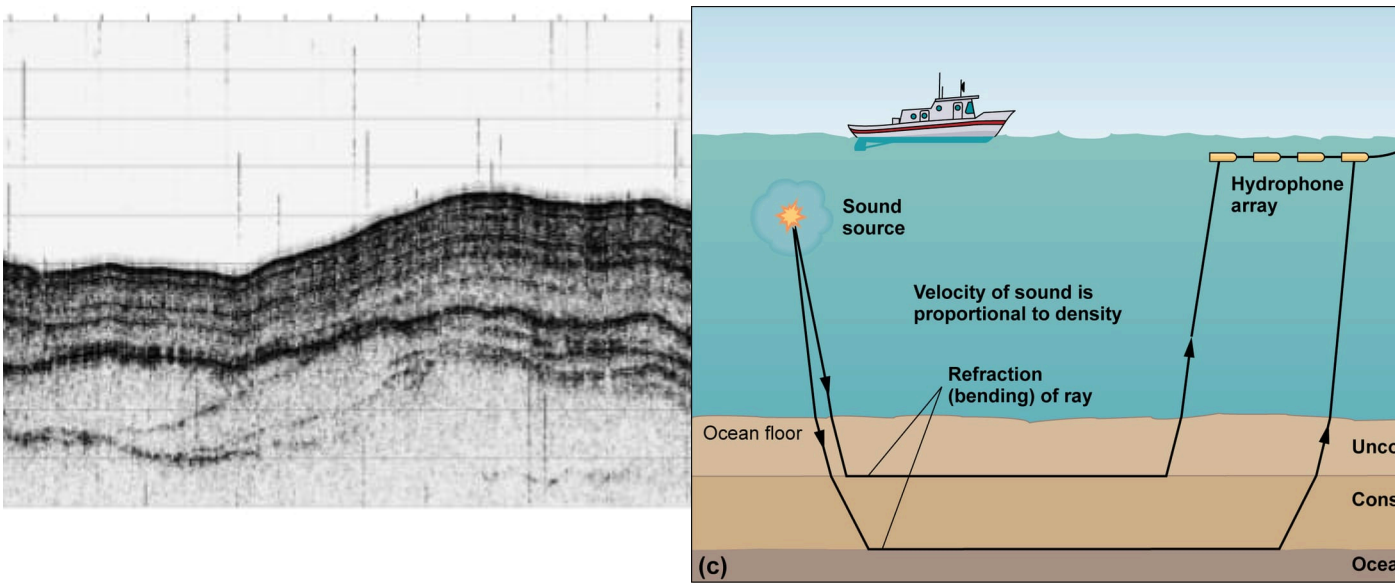


Figure 3-9. Two seismic profiling techniques are used to map subsurface sediment layers and structures. (a) Seismic reflection sends sound pulses, which penetrate the seafloor and are reflected off sedimentary layers. (b) A typical reflection seismic profile shows rugged seafloor topography overlaid by sediment layers. (c) Seismic refraction sends sound pulses, which are refracted (bent) at interfaces between layers of sediment and then travel back out to be received by a second ship. At one time, the sound source was provided by dynamite sticks thrown overboard to explode just under the water surface. However, the sound source typically used now is an air gun that periodically fires a burst of compressed air into the water at a depth of about 15 to 35 m. The sound waves are generated by the collapse of the air bubbles produced by each burst. (d) A 3-dimensional seismic refraction image of an ancient reef structure buried beneath the sediments off Nova Scotia, Canada. The light colored 2-dimensional sections across the center of the image and along the right side show how much more information the 3-dimensional image can convey.

Because sound travels at different speeds in different types of sediment and rock, seismic profiles can help determine the nature of each layer of sediment. Distinct sediment layers in many parts of the ocean can be traced for hundreds of kilometers in all directions. Hence, the layers found in seismic maps can often be correlated to the layers of sediment and rock retrieved by drilling or coring.

Sediment and rock on or below the seafloor can also be studied by precise measurement of changes in the strength of the gravitational-field or magnetic-field strength. Tiny changes in magnetic-field strength are detected by instruments towed behind a research vessel as it passes over seafloor sediment and rocks with variable magnetization. Gravitational-field strength also changes slightly at different locations because the Earth is not perfectly round, and because denser, heavier sediments and rocks exert a slightly greater gravitational pull than less dense or lighter sediments at the same depth. Therefore, extremely small changes in gravitational-field strength detected by instruments carried on research ships provide information about the sediment and rock below, especially the presence of mountains of volcanic rock overlain by less dense sediment.

Dredges

Although most ocean floor is covered by thick sediment, rocky outcrops are also present. Some of these rocky areas of the seafloor are partially covered with **manganese nodules** and **phosphorite nodules** (Chap. 6). Because corers and drills cannot sample such surface rocks, dredges are commonly used. A dredge consists of a metal chain or nylon mesh net with one end held open by a strong, rigid, metal frame (Fig. 3-10).

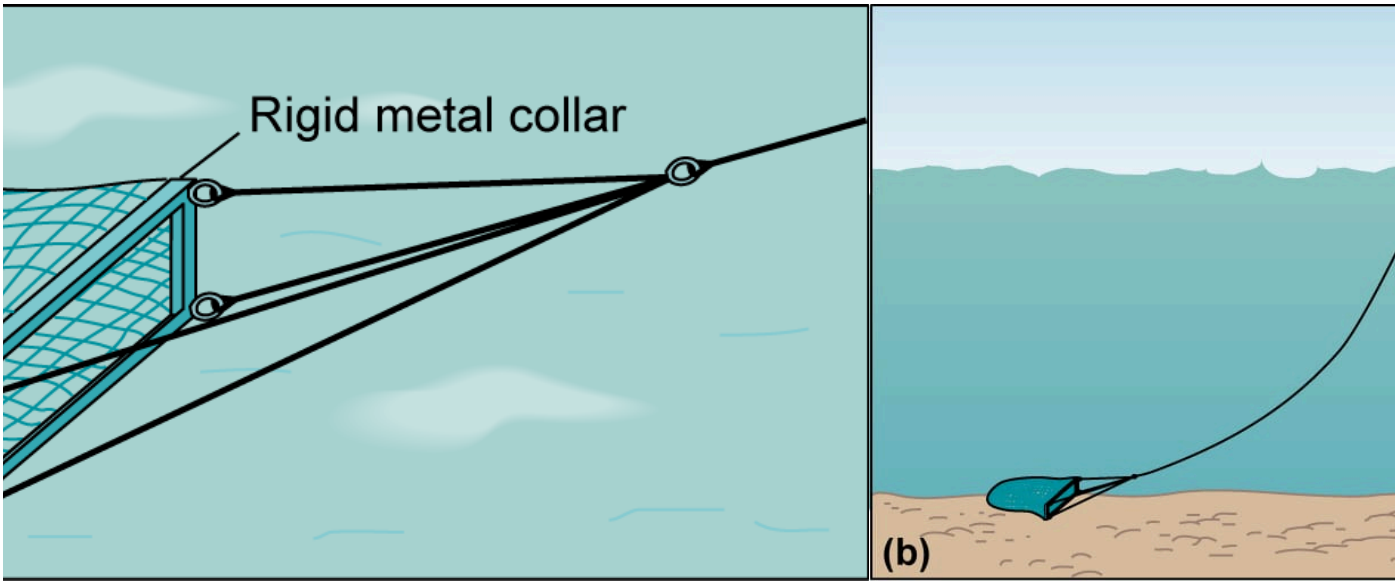


Figure 3-10. (a) Rock dredges are used to obtain rock samples, including nodules, from the seafloor with little or no sediment cover. (b) The dredge is lowered on a cable and towed across the seafloor. Sediment passes into and through the net, while nodules and loose rocks are retained. The metal collar breaks off rocks from a rocky seafloor.

Dredges are often massive because the rocks in many areas must be broken off solid **lava** flows. When a dredge snags on such rocks, considerable force is needed to break the rock and release the dredge. Therefore, the cable used for lowering and towing dredges must be extremely strong. Steel cables 1 cm or more in diameter are sometimes required. In very deep water, 10,000 m or more of cable weighing several tonnes is required. Because the drum on which the heavy dredge cable is stored and the winch needed to drive the drum are huge, only a few of the largest research vessels can deploy the largest dredges. Submersibles, ROVs, or AUVs discussed later in the chapter now perform some of this type of sampling.

This page titled 3.4: Seafloor Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.5: Chemical and Physical Oceanography

Oceanographers aim to understand how the concentrations of various chemicals in seawater, whether dissolved or associated with **suspended sediment**, change temporally or spatially (over time and space). Most of these chemicals are found in seawater at very low concentrations (**Chap. 5**) making measurement of their concentration very difficult. As a result, most chemical concentrations cannot yet be measured directly by instruments lowered into the ocean. Instead, scientists usually need to collect water samples from specific locations and depths, then analyze them on a research vessel or in a laboratory on land.

Sampling Bottles

In all but shallow water, where samples can be pumped up through a hose, water samples are collected in specially designed bottles. Usually, sampling bottles descend in an open configuration that allows seawater to flow through them, so they are not crushed by the increasing pressure. Several sampling bottles can be attached at intervals on a **hydrographic** wire to sample at different depths during one lowering. When the bottles are at the desired depths, a brass or stainless-steel messenger is attached to the wire. (**Fig. 3-11**). The messenger slides down the wire and hits a trigger mechanism on the shallowest bottle. This closes the bottle and releases another messenger attached under the bottle. Thus, each bottle releases a new messenger to slide down the wire and close the next-deeper bottle.

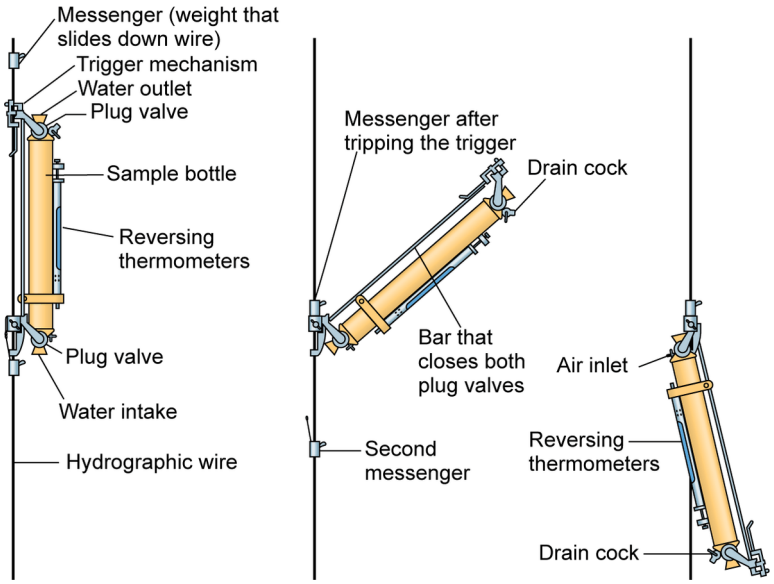


Figure 3-11. Nansen water-sampling bottle. When the messenger hits the trigger mechanism, the top of the bottle is released from the wire, the bottle falls into an inverted position, and a mechanical linkage closes valves at each end of the bottle. The trigger mechanism also releases a second messenger, which slides down the wire to the next sampling bottle.

For many years, most water sampling was done with Nansen bottles (**Fig. 3-11**). These have been replaced by newer designs developed primarily to collect larger samples or to avoid sample contamination. Because concentrations of some important trace metals and organic compounds in seawater are extremely low (**Chap. 5**), oceanographers must often collect large volumes (sometimes tens or hundreds of liters) of seawater per sample. They then use sophisticated chemical techniques to extract and concentrate the target chemicals before even the most advanced and sensitive analytical instruments can measure the concentration.

Avoiding Sample Contamination

Contamination of the sample must be avoided during its journey from the ocean depths to the laboratory. Contamination may come from many sources, including the metals, plastics, and other sample bottle materials, the metal of the hydrographic wire, and the grease that covers the wire to protect it from corrosion. Dust, oil, and vapors in the ship and laboratory atmosphere are other potential **contaminants**.

The thin **surface microlayer** (about 0.1 mm thick or less) that covers the entire ocean presents an especially difficult contamination problem. The microlayer always contains higher concentrations of many chemicals than the seawater below, and it can be further contaminated by discharges from the research vessel, such as oily cooling water and paint and corrosion chips from the vessel's hull. Sampling bottles that remain open as they are lowered through the sea surface retain a film on the inside of the sampling bottle deposited by the surface microlayer, which can significantly contaminate the sample. Several ingenious sampler designs, including the GoFlo bottle (Fig. 3-12), prevent surface microlayer contamination.

Figure 3-12. A GoFlo water-sampling bottle. This ingeniously designed bottle is lowered through the sea surface in a closed configuration to avoid contamination. At a depth of a few feet, a pressure-sensitive trigger opens both ends of the bottle. The bottle is closed again by means of a messenger or other triggering device when the bottle is at the required sampling depth.

The GoFlo bottle is carefully cleaned and sealed closed on the research vessel before being lowered through the surface. A pressure-sensitive mechanism opens the bottle automatically once it is a few meters below the surface. The bottle can then be closed at the required sampling depth. For chemical parameters not affected by surface microlayer contamination, one of many other sampler designs can be used.

Determining the Depth of Sampling

Ocean water forms a series of horizontal layers, each of which may be only a few meters thick (Chap. 8). Water moves great distances within these layers, but it mixes only slowly with the water in the next layer above or below. Chemical oceanographers usually want to sample within each of the layers, and sometimes at closer depth intervals across the interfaces between layers. Unless the precise depths of the layers are known, selecting the appropriate spacing of water-sampling bottles along the wire to sample all the layers is impossible. In addition, even if samplers are placed correctly along the wire, the depth at which each sampler is closed can be affected by curvature of the wire caused by currents or vessel drift (Fig. 3-2).

Density increases with depth in the oceans, so depth may be estimated by measuring density. The two primary parameters that determine the water density are temperature and salinity, so density may be calculated and depth estimated by measurement of these two parameters. Salinity, temperature, and density relationships are discussed in **CC6** and Chapter 5. The approximate depth from which a sample is obtained can often be determined if the temperature and salinity of the water at the depth at which a sampler is closed are measured. Salinity can be determined by measuring the **electrical conductivity** of the sample of seawater collected after it has been returned to the laboratory. However, the temperature of the sample changes as the sample bottle is retrieved so, before the development of sensitive electronic thermometers, the temperature had to be recorded at depth when the sampler is closed. This temperature measurement was achieved by a "reversing thermometer" mounted on the outside of the sampler (Fig. 3-13a). The thermometer on a Nansen bottle reverses with the bottle itself (Fig. 3-11), but other sampling bottles, which do not themselves reverse, have an externally mounted thermometer rack that rotates mechanically through 180° as the bottle is closed. To accurately determine the depth of sampling, two reversing thermometers can be used, one protected and one unprotected from pressure changes with depth (Fig. 3-13b). This method is so accurate it is still used today to calibrate the electronic temperature sensors.

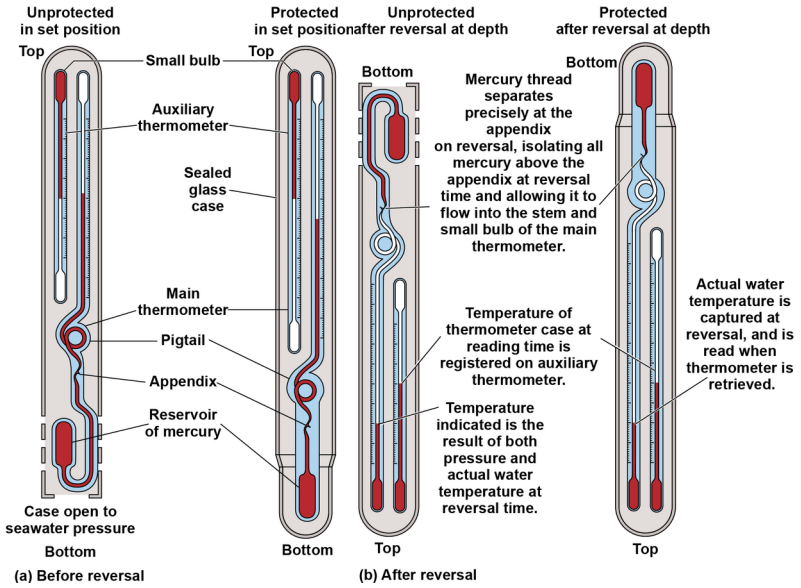


Figure 3-13. (a) Reversing thermometers like these are attached in their set position to a water-sampling bottle in a rack. (b) When the bottle is closed, the rack and thermometers are turned upside down (reversed). The protected thermometer records the true temperature, while the unprotected thermometer records a temperature that is slightly too high because the mercury reservoir is squeezed slightly by the increased pressure at depth. The temperature difference between the thermometers can be used to determine the depth at which the thermometers were reversed. Reversing thermometers are still used to calibrate CTDs, which are instrument packages that measure conductivity, temperature, and depth electronically.

Instrument Probes and Rosette Samplers

Using water sampling bottles and reversing thermometers to determine salinity, temperature, and depth is tedious and only provides data only at the sampled depths. This method was replaced in the 1970s by instrument packages containing electronic sensors attached to a wire that make measurements continuously as they are lowered and raised. These packages are called **CTDs**, for conductivity (electrical), temperature, and depth (or sometimes "STDs," for salinity, temperature, and depth), because electrical conductivity is measured to determine salinity (**Chap. 5**). CTDs have many advantages over sampling bottles, especially their ability to read salinity and temperature continuously as a function of depth. Such sensors, or probes, have enabled oceanographers to observe small-scale variations in the layered structure of ocean water that cannot be discerned from widely spaced bottle samples. The wire used to lower CTDs have electrical conductors running throughout their length (often tens of thousands of meters). These wires supply electrical power to the sensor package and also return the sensor signal to a processing unit in the ship's laboratory. This wire is more expensive than the simple steel wire, but the advantages justify the extra cost. Therefore, these CTDs enable scientists to see the variations of salinity and temperature with depth. The special CTD wires, which have electrical conductors running throughout their length (often tens of thousands of meters), are much more expensive than the simple steel wire used with sample bottles

Although oceanographers no longer need to use water-sampling bottles to determine salinity, temperature, and depth, samples of water must still be collected and returned to the laboratory for analysis of most dissolved chemicals. The dissolved constituents and the importance of variations in their concentrations are discussed in **Chapters 5**, **12**, and **16**. If the sampling bottles are mounted around a CTD (**Fig. 3-14**), samples can be taken at precise locations within the various water layers. Each bottle can be closed when the CTD readings in the shipboard laboratory show that the bottle is at the appropriate depth. For this purpose, a rosette sampler (**Fig. 3-14**) is used that consists of a rack for mounting 12 or more sample bottles around the CTD sensor package and an electronically operated trip mechanism. Signals sent down the support wire to the trip mechanism close each sampling bottle individually.

Figure 3-14. A rosette water sampler has sample bottles, some fitted with thermometer racks, arranged in a rosette pattern around an electronic device that enables researchers on the ship above to close individual sample bottles at selected depths through an electrical signal sent down the hydrographic cable. The round frame below the water-sampling bottle rack usually contains a CTD and possibly other sensors, such as turbidity-measuring devices.

There is great interest in developing sensors to add to the CTD that would continuously measure dissolved concentrations of important chemicals. Dissolved oxygen, **pH**, and **turbidity** are among the relatively few parameters for which reliable sensors exist. To date, none of the other sensors are as reliable and sensitive as the salinity, temperature, and depth sensors are. However, a wide range of chemical sensors are now beginning to reach the stage of development where they are providing useful data.

Measuring Currents

Oceanographers are interested in studying the movement of water in currents that are present throughout the ocean depths. The speed and direction of currents can be computed from salinity and temperature distributions in the oceans (**Chap. 8**), and water movements can also be studied using chemical or radioactive tracers dissolved in the water. These indirect methods, especially when combined with mathematical modeling techniques (**CC10**), are extremely valuable for studying the movements of water averaged over large distances (tens of kilometers or more) and long periods of time (months or longer).

Because tracers are much less useful in studying the small-scale details of current distributions and their variability with time, a variety of systems are used to measure currents directly. There are three basic types:

- *Passive devices* flow with a current wherever it goes and periodically or continuously report their position.
- *Current meters* are anchored and periodically or continuously measure the speed and direction of water flowing past them.
- *Remote sensing systems* can measure currents at various depths beneath a moving research vessel or at various depths from a fixed mooring.

Drifters, Drogues, and Floats

The simplest method of measuring currents in the ocean is to use drifters. Drifters are designed to float on the surface of the water or are weighted and designed to sink very slowly to the seafloor, where they are easily picked up and moved by even the gentlest bottom current (**Fig. 3-15a**). Drift cards (**Fig. 3-15b**) are thrown overboard in large numbers at a fixed location. Like a message in a bottle, they drift on the surface with the ocean currents until they wash up on a beach. Each card is numbered and bears a message asking whoever finds it to return it to the oceanographer with details of where (usually on a beach) and when it was found. The finder is often paid a small reward for returning the card. Drift cards are a very inexpensive means of gaining information about mean current directions, especially in coastal regions, where currents are often very complex and variable. Drift cards are particularly valuable in studies of the probable fate of wastes discharged or oil spilled at specific locations. Surface currents can also be studied with floats, whose movements can be followed remotely by radio signals received from a transmitter mounted on the float.

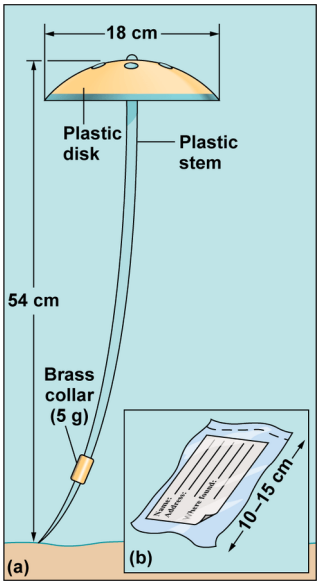


Figure 3-15. Surface and seabed drifters can be used to study ocean currents. (a) A seabed drifter is weighted such that its bottom tip drags along the seafloor as the plastic disk section "catches" the current and moves with it, while (b) a surface drift card is usually a simple postage-paid return postcard sealed in plastic that floats flat on the water surface and moves with surface currents.

Current speed and direction both vary with depth in the ocean. Currents flowing below the surface can be studied with two types of passive systems. In shallow coastal waters, parachute drogues are often useful. A drogue consists of a parachute attached to a weight and to a measured length of wire, which is attached to a float, usually with a radar reflector mounted on it. The parachute opens and is pulled down to a known depth by the weight (**Fig. 3-16**). Once at the assigned depth, the parachute drogue "sails" in the current, dragging the surface float behind it. Such devices can only be used at relatively shallow depths because of the excessive drag that is caused by a long wire.

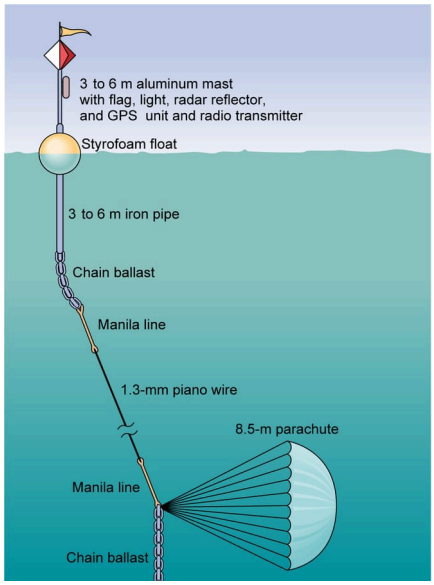


Figure 3-16. A parachute drogue consists of a parachute deployed at a selected depth (usually less than 100 m), where it sails along with the current. The parachute is attached to a surface float that drags the surface float after it. The researchers can follow the changing position of the float using visual observations, radar from a ship, or remotely using a GPS unit and radio transmitter mounted on the float.

Radio transmitters and global positioning system (GPS) sensors can be mounted on some surface drifters and parachute drogue floats. This enables them to be tracked from a remote location which reduces the need for, and cost of, surface vessels.

Currents in deep waters can be measured with neutrally buoyant floats. The simplest versions are self-contained instrument packages, mounted in one or more hollow tubes and weighted so that their density is precisely the same as the density of the seawater at the depth at which the float is to operate (**CC1**). These floats are deployed from a research vessel and sink through the

water column until they reach neutral buoyancy at a predetermined depth. The float contains a pinger, an electronic system that produces short sound pulses, or “pings” that can be used to triangulate the float’s location when heard by two or more listening stations aboard research vessels or onshore. This type of float may be followed for months as they move in the complex currents and eddies of the ocean depths (**Chap. 8**). Modern versions of these floats are called “autonomous floats,” and they are capable of moving back and forth vertically through the water column, recording conductivity, temperature and other data. These floats periodically revisit the surface to send their data back to a ship or shore station by radio. These floats do not have a means of recording their exact position during their time below the surface. GPS sensors are small enough that they can be mounted in the floats, but GPS and other electromagnetic wave-based positioning systems cannot work below the ocean surface because ocean water effectively absorbs this radiation. However, average current speed and direction, averaged over the depths visited by the floats and the time interval between surface visits, can be estimated from the precise locations of the floats measured each time the float surfaces to report its data.

Mechanical Current Meters

Many current meters remain in a fixed location, measuring the rate and direction of the water flowing past them. Three types of mechanical current meters were common before the development of acoustic current meters: (1) those having an impeller whose axis is oriented vertically (**Fig. 3-17a**); (2) those having a propeller oriented to face the current (**Fig. 3-17b**); and (3) those that rely on the current to tilt the meter body at an angle from its normal vertical position (**Fig. 3-17c**). In each case, the meter is oriented to align itself with the current direction by one or more vanes or fins, just like a weather vane. Current direction is measured using a magnetic compass inside the instrument housing. These current meters are usually suspended at intervals below the surface on wire moorings (**Fig. 3-17d**). The entire mooring, including the float at the top, is often deployed well below the water surface. The string of meters is left to record currents for days, weeks, or even months, and then recovered by releasing the weights from the bottom of the wire with an acoustic signal sent from the recovery ship to an acoustic release. This method of deployment below the surface is often necessary to avoid having the meters cut free or stolen by curious ship crews or by misguided fishers concerned about their nets catching on the mooring.

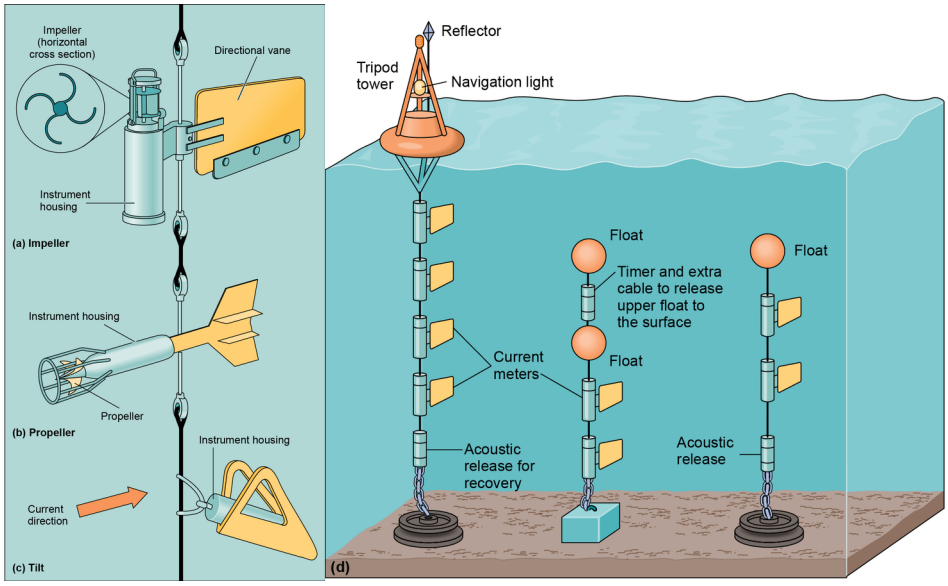


Figure 3-17. Typical designs for current meters measure current speed by recording the (a) rate of rotation of a rotor that behaves much like a waterwheel (this type of rotor is known as a “Savonius rotor” or an “impeller”), (b) rate of rotation of a propeller, and (c) angle to which the meter is pushed by the current. Vanes or fins are used to orient the meter in the current. Current speed and direction are usually recorded on tape in the meter and are read when the meter is retrieved. (d) Current meters are typically deployed in vertical strings, such as the three different configurations shown here, all moored to the seafloor. They are often left in place for weeks or months and then retrieved by activating a timer or an acoustic signal from the recovery ship to release the bottom weights. Often current meters are moored entirely below the surface to discourage theft and to reduce navigation hazards.

Current meters deployed on moorings normally contain internal devices that continuously record current speed and direction. Current direction is recorded by continuous readings of the position of a magnetic compass mounted inside the meter body. Depending on the meter design (**Fig. 3-17a–c**), current speed is measured by the speed of rotation of the impeller or rotor or by the angle of the meter’s tilt from the vertical.

Most current meters are sophisticated and expensive electronic instruments. However, currents have been measured with much simpler systems in some shallow coastal waters. The most ingenious and least expensive system consists of a sealed glass bottle partially filled with sealing wax and containing a magnetic needle attached to a piece of cork. The neck of the bottle is tied with string to a LifeSaver candy, and a weight, such as a rock, is attached to another string, which is also tied to the candy. The bottle is heated on board the ship to melt the wax, and the "current meter" is dropped into the water, where the weight pulls it to the bottom. Once on the seafloor, the bottle on its string is held at an angle by the current, the wax solidifies in the cold water, and soon the candy dissolves, releasing the buoyant bottle from the weight to float back to the surface. The wax surface, which was horizontal when the wax solidified, lies at an angle to the bottom of the bottle. That angle is a measure of the current speed. The bottle also records the current direction because the magnetic needle, which was facing north as it floated in the wax, is locked in place as the wax sets.

Acoustic Current Meters

Remote sensing Doppler acoustic current meters measure current speed and direction at multiple depths simultaneously. Such meters can be mounted looking upward from the seafloor or a seabed mooring (**Fig. 3-18a,b**), or mounted on a ship looking downward (**Fig. 3-18c**). They send several narrow-beam sound pulses into the water column that are angled away from the meter in different directions. The sound echoes off particles in the water column, and these echoes are recorded from each of the beams. The current speed and direction are calculated from the Doppler shift of the sound frequency in the returning echoes. The Doppler shift is the same phenomenon that makes the pitch of a train whistle change as it passes. When the train is approaching—or, in this context, when current is flowing toward the acoustic meter—the pitch is increased. When the train is moving away—or when the current is flowing away from the acoustic meter—the pitch is decreased, producing a deeper (more bass) tone.

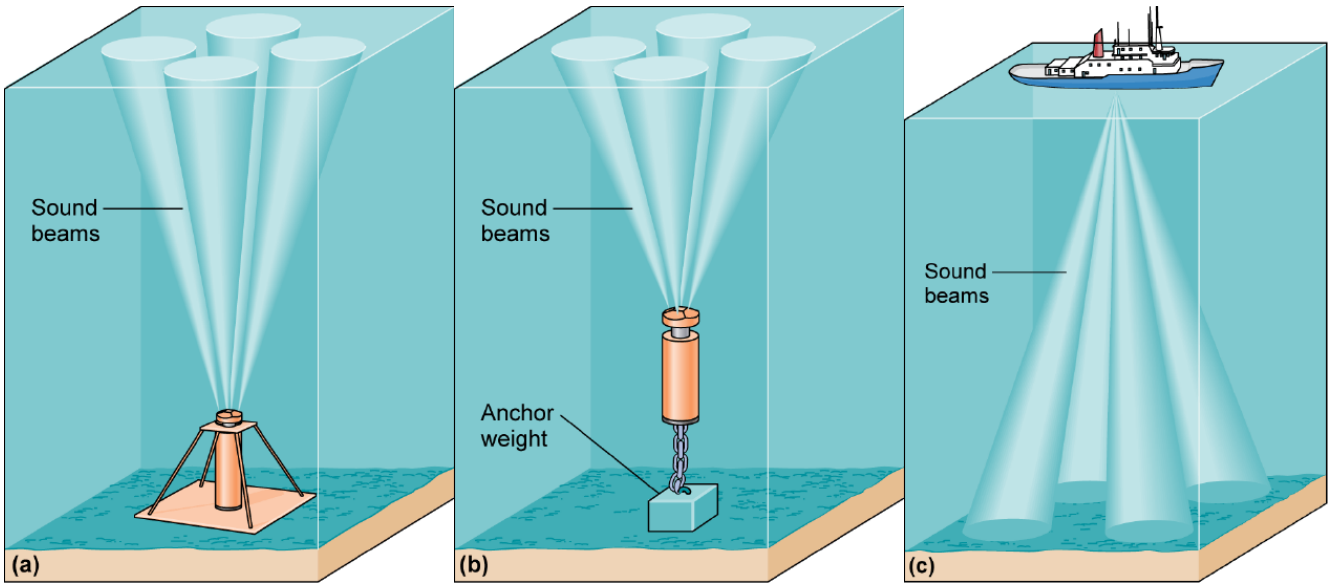


Figure 3-18. Doppler acoustic current meters can be (a) mounted on the seabed, (b) moored in mid water, or (c) mounted in a ship's hull facing downward. The sound beams sent out by the meter are reflected off particles in the water, and the frequency of the sound in the returning echo changes according to the direction and speed of the particles that move with the current.

Remote sensing Doppler current meters have the advantage of simultaneously measuring currents at all depths within the meter's maximum operating range of 100 m or more. They are particularly useful in locations where current-meter moorings are not possible, such as in busy shipping lanes and estuaries with very fast currents. Such meters are widely used but they are expensive and collect large amounts of data, including sound intensity, Doppler shift, and instrument orientation and tilt, which must be processed through complex computer analyses to yield data on current speed and direction.

Another technique, acoustic tomography, is capable of making simultaneous observations of water movements or currents within large areas of the ocean. The system is the acoustic equivalent of the computerized axial tomography (CAT) scan used to produce three-dimensional images of internal human body parts. The acoustic tomography system consists of several sound sources and receivers moored at different locations within a study area that can be hundreds of kilometers across (**Fig. 3-19**). Sound is emitted by each source and received by each receiver, so numerous pathways of sound traveling between sources and receivers are possible. The sound velocity along each of the pathways is affected by water characteristics such as depth, temperature, and salinity. When currents flow within the study area, they change the distribution of temperature and salinity within the study

area and therefore the average speed of sound along the source-receiver pathways. Powerful computer data analysis techniques convert the small variations in travel times between each source and receiver into a detailed picture of the water property distributions and movements within the study area.

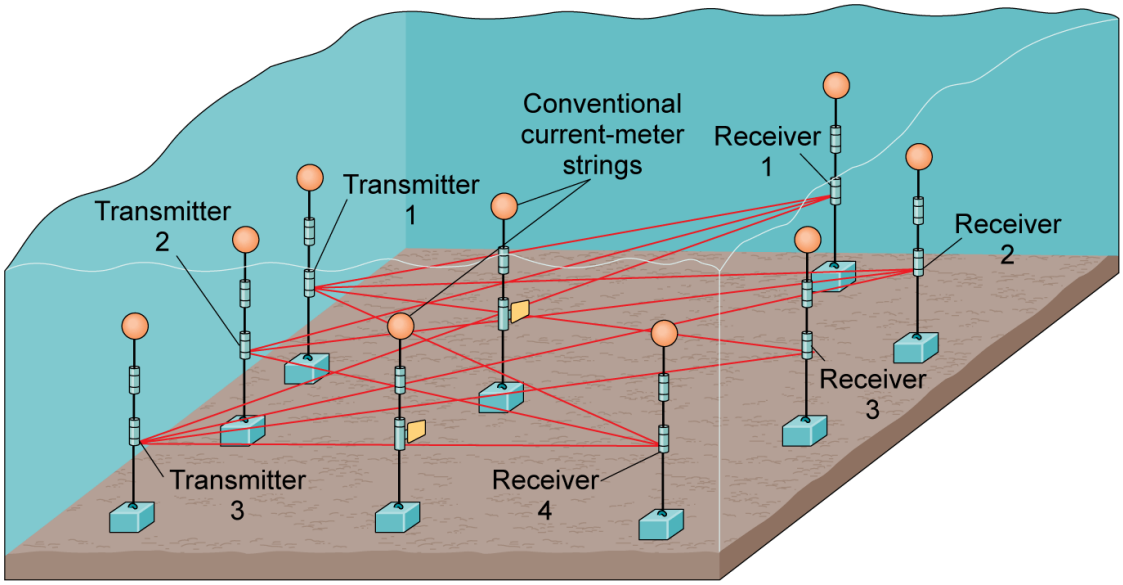


Figure 3-19. In a typical acoustic tomography array, sound pulses are sent from each transmitter and received by each of the receivers.

This page titled 3.5: Chemical and Physical Oceanography is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.6: Sampling Living Organisms in the Sea

Almost all terrestrial **species** spend their entire lives within a narrow zone that extends from a few tens of meters above the tops of the trees to a few meters below the Earth's surface. Therefore, terrestrial organisms are relatively easy to study, as they are readily accessible and visible to the scientist. In contrast, ocean life is present throughout the thousands of meters of ocean depth (**pelagic** species) and for several meters, or more, below the seafloor (benthos or **benthic** species).

Special Challenges of Biological Oceanography

Although pelagic marine life is concentrated in the ocean's upper few hundred meters, scientists cannot readily enter the oceans to observe it. Therefore, most marine biological studies require capturing the undersea creatures and bringing them back to a research vessel or onshore laboratory. Collecting biological samples from a specific ocean location and depth is a daunting task that poses five major problems, which add to the general problems of working in the oceans discussed previously:

- Pelagic life is widely dispersed in three dimensions. Therefore, large volumes of water must be sampled or searched before representatives of all, or even most, species in any part of the ocean are captured.
- The species that make up ocean life range in size from microscopically small **bacteria**, **archaea**, and **viruses** to giant whales. Therefore, no single method of sampling can capture all important species in a particular study location.
- Many species of ocean life actively swim and avoid any sampling device lowered into their **environment**.
- Many species are extremely delicate and are literally torn apart by contact with sampling devices that are dragged through the water.
- Many species that live in the cold, dark, high-pressure, deep-ocean environment cannot survive the changes in temperature, pressure, or both as they are brought to the surface.

Nets, Water Samples, and Traps

Nets towed through the water are the sampling devices most often used to collect pelagic species. Nets of different overall sizes and mesh sizes are required to capture different types of organisms (**Fig. 3-20**). Small nets with very fine mesh are used to sample **phytoplankton** and **zooplankton**, and progressively larger nets and meshes are used to capture fishes and **invertebrates**. Net samples may be collected by simply lowering a net from a stationary ship and hauling it back on board. However, biologists usually want to collect samples from a specific depth, so most sampling with nets is performed by towing a weighted net slowly behind the research vessel.

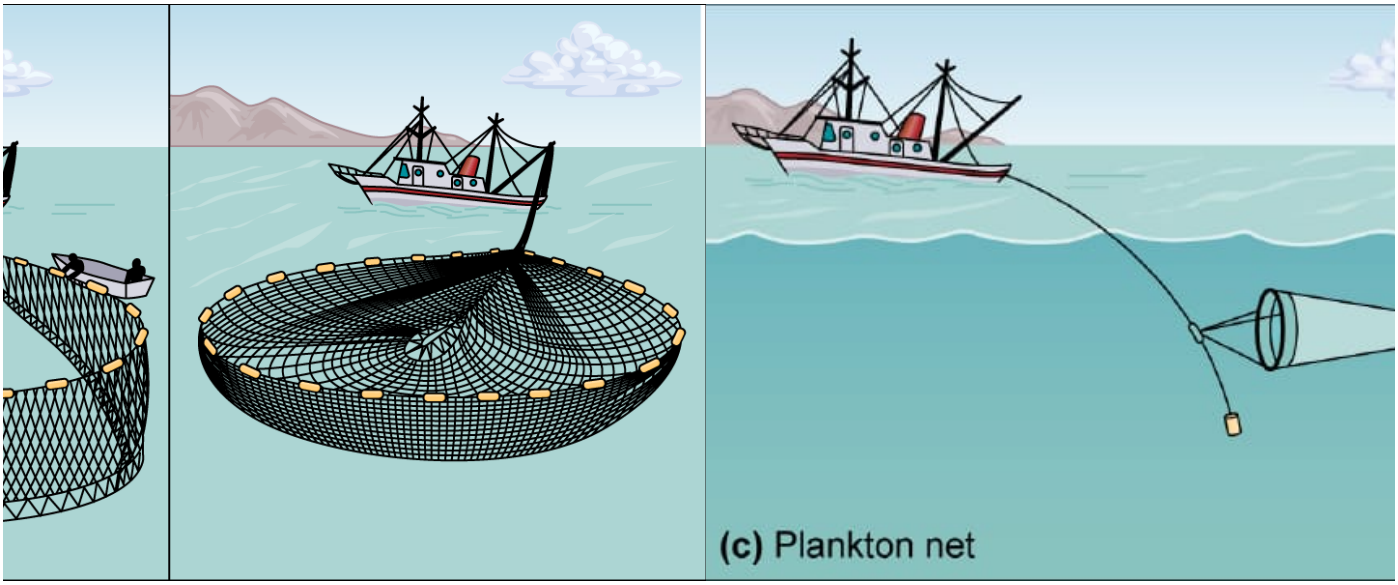


Figure 3-20. Many different types of nets are used to sample pelagic and benthic organisms. (a) The otter trawl uses two "otter boards" that "sail" in the current created by the vessel's movement, forcing the end of the net to stay open as it is towed. (b) The purse seine is towed quickly around a school of fish and then drawn up under the fish, preventing their escape. (c) Plankton nets are extremely fine mesh nets with a jar or bucket attached at the narrow end to collect plankton as the net is towed through the water. Some plankton nets can be opened and closed on command to sample only selected depths and locations, and some have current meters attached to measure the volume of water sampled as it flows through the net. (d) Drift nets may be deployed for many kilometers across the ocean, where they catch anything too large to swim through the net. Large animals, such as dolphins and sharks, may be inadvertently entangled and die.

Biologists need to know the net depth, how much water passed through the net, and the exact distance the open net is towed. Nets can be let out on a known length of cable over the stern of the ship, towed for a time, and then retrieved. However, without instrumentation, the biologist does not know the exact depth of the net tow, how much of the catch came from shallower depths as the net was let out or retrieved, or how much water passed through the net. Many sampling nets now have various types of gauges to measure the depth at which the net is towed. Biologists also need to know the amount of water passing through the net during a tow to calculate concentration or population density of the species in that area. To calculate the volume of water sampled by the net, the measured tow distance is multiplied by the cross-sectional area of the net opening. To provide an accurate length of the tow, many nets carry flow meters similar to impeller-driven current meters (**Fig. 3-18a**). Some modern nets can also be kept closed until reaching the desired sampling depth, and then opened, towed, and closed again before retrieval. The volume of water sampled by the net, the measured tow distance, is multiplied by the cross-sectional area of the net opening. Some nets are even equipped with video cameras to record species that evade the net.

The smallest pelagic organisms in the oceans (bacteria, archaea, viruses, and phytoplankton) can be sampled by collecting water in sampling bottles such as those described earlier. **Plankton** and microorganisms are often measured using surrogate parameters. For example, investigators often estimate the total **biomass** of **photosynthetic** plankton by measuring the total **chlorophyll** (**CC14**) concentration. Concentrations of some other groups of microorganisms are calculated by measuring the concentrations of other pigments similar to chlorophyll or by measuring the concentrations of specific types of genetic material such as **DNA**. Sharks, large fishes, and squid that avoid nets are sampled with fishing lines and lures. Small benthic organisms are collected by sediment collection devices such as grab samplers and box corers. Fishes and other large invertebrates that live on the seafloor are sampled with trawl nets towed across soft sediment bottoms (**Fig. 3-20a**) or by dredges similar to those used to collect rocks from the seafloor (**Fig. 3-10**).

Fishes and invertebrates that are particularly adept at eluding nets can be caught in a variety of traps. The animals are attracted to the traps with bait or lures. In the dark deeper waters or at night, they may also be attracted by light. Many types of traps have been used. Almost all of the traps allow the organism to enter the trap easily in pursuit of the bait or lure, but make it difficult or impossible for the organism to escape. **Figure 3-21** shows a commonly used lobster pot.

Figure 3-21. Lobster pots (or traps) are designed so the lobsters can crawl into the pot to eat the bait inside the pot, but then cannot escape.

Fragile Organisms

Fragile organisms that are easily damaged by nets are particularly difficult to collect and must be captured in closed bottles or jars. Such organisms are numerous in the oceans. Fragile organisms are especially common floating freely in near-surface open-ocean waters (**Chap. 14**), and in benthic **communities** at **hydrothermal** vents (**Chap. 15**).

Fragile organisms from the upper layers of the open ocean can be collected by divers that guide interesting organisms into an open jar that they seal immediately. Fragile or elusive small fishes and other organisms can also be sampled by a simple syringe-like device called a "slurp gun." The diver places the open mouth of the gun near the organism and simply sucks it into the slurp gun's body by withdrawing a plunger. Slurp guns are particularly useful for collecting small organisms from within cracks and holes in **coral reefs**, and they are the most environmentally sound means of collecting small tropical fishes for aquariums.

In waters deeper than **scuba** divers can reach, fragile organisms are collected from research submersibles, ROVs, or AUVs. The collection methods are the same as those used by divers, but the jars, slurp guns, or other collecting devices must be manipulated by remote control or complex autonomous systems.

Migrations and Behavior

Although the occurrence or concentration of a species within a given area is an important parameter, it tells us little about the organism's life cycle—including, for example, its migration, feeding, and reproduction. Observations by divers, submersibles, and remote cameras can provide some of this information, but only for species that remain relative local and stay in observable depths during the activity. However, many species, including whales and other **marine mammals**, turtles, and many fish species, swim large distances and dive to substantial depths to feed or reproduce, making them effectively impossible to directly observe. Some of these behaviors can be studied remotely. For example, **schools** of fish can sometimes be followed by sonar, and whales can be tracked by their songs, which can be heard at great distances. However, these are at best inadequate techniques for understanding their activities. Small sensing packages attached to an organism, for example, the skin of a whale, can gather data on time, depth, and temperature. The instrument package stays attached for some time, is released, floats back up to the surface, and is either recovered by researchers or transmits its data remotely. These instrument packages are becoming steadily more sophisticated and may include additional sensors or video cameras. A particularly interesting way to observe whales is a package that contains a microphone and motion recording device. Attached to whales, this device has been used to record the changes in diving behavior caused when a whale swims into the sound field generated by a surface sonar unit.

This page titled 3.6: Sampling Living Organisms in the Sea is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.7: Scuba, Manned, and Unmanned Submersibles

Direct observation is often the only or the best way to sample or measure many ocean processes. For example, a seafloor area may be characterized by jumbled rock formations and highly variable assemblages of organisms. In such areas, geological oceanographers can best identify and collect rocks and sediment of interest if they can visually inspect and select samples. Similarly, biologists can ensure that they are collecting the important species if they are able to see and select the organisms to be sampled. Direct observation in their natural **habitat** is also important for those species that do not survive transfer to an aquarium or laboratory, for biologists to study.

Scuba and Habitats

In shallow waters, sampling is often performed by scuba divers. However, because of the dangers of the bends and nitrogen narcosis, scuba divers cannot descend safely below about 90 m and they can remain underwater for only a short time.

Scuba equipment supplies a diver with air at the pressure that corresponds to the diver's depth. Thus, the pressure of the air breathed increases with depth. Divers are susceptible to two dangerous syndromes. The first of these, nitrogen narcosis, occurs when high-pressure air causes high concentrations of nitrogen to build up in the bloodstream. This causes symptoms very similar to those of alcohol intoxication—with possible consequences similar to drinking and driving. The second ailment, the bends, is a different life-threatening medical problem. It is caused by breathing high-pressure air at depth for too long and then returning to the surface too quickly. As the depth and immersion time of a dive increase, nitrogen continues to dissolve into the diver's bloodstream and then transfers into other body tissues. If the diver returns to the surface too rapidly, the excess nitrogen in the blood and tissues cannot escape quickly enough and forms damaging gas bubbles in the body.

One way of extending the length of time that scuba divers can remain at depth safely is to use an underwater habitat (**Fig. 3-22**). Scientists can live for days or weeks in such habitats, which are pressurized and anchored on the seafloor. The scientists can safely make multiple scuba excursions to research sites at approximately the same depth as the habitat, as long as they return to the habitat and not to the surface. Although underwater habitats have many valuable uses, particularly for behavioral and other biological studies, their utility is limited. They are expensive to maintain and operate, they cannot readily be moved to new research sites, and they cannot significantly increase the maximum depth at which scientists can work using scuba. In addition, to avoid the bends, scientists who live in habitats for a week or more must spend several days in a decompression chamber at the end of their stay, even if the habitat is only a few meters deep.

Figure 3-22. Underwater habitats, such as Aquarius, shown here in the Florida Keys, enable scientists to live and work up to several tens of meters underwater for several days without incurring decompression sickness. However, at the end of their stay they must undergo very long decompression periods in a decompression chamber.

Manned Submersibles

Most of the ocean floor and almost all of the ocean volume are too deep for scuba divers to reach. Marine scientists have used a variety of research submersibles to visit and work at greater depths. Manned oceanographic research submersibles are small submarines usually designed to carry no more than two or three scientists. They enable the scientists to observe and photograph organisms and the seafloor. Equipment is often attached to the outside of the submersible to collect organisms, rocks, sediment, and water samples that can be selected visually by the submersible's occupants.

The first recorded successful use of a submarine was in 1620. The vessel, built by a Dutchman, Cornelius Drebbel, had a waterproof outer skin of leather and was propelled by 12 oarsmen. It was reported to be able to stay as deep as 4 or 5 m underwater for several hours. Between 1620 and the 1930s, many different submarines were developed, but they were used primarily as warships. Today, the vast majority of submarines are still warships that are not suited to, or used for, most oceanographic research.

During their early development, research submersibles had very limited capabilities because they were built primarily to transport explorers who sought to dive to ever-greater depths. Such explorers usually performed scientific observations as only a secondary interest. Early submersibles, called "bathyscaphes" or "bathyspheres," provided little more than windows and lights for their occupants to view the oceans and seafloor, and most were not equipped to collect samples. In addition, most early submersibles could only sink to the seafloor and then drift with the currents or move short distances before being hauled back onto a ship. Therefore, positioning a submersible precisely at a previously selected site on the seafloor was not possible.

The exploration phase of research submersibles climaxed in 1960 when the bathyscaphe *Trieste* visited the deepest part of the ocean, the Mariana Trench, 10,850 m below the ocean surface. Since 1960, many new submersibles have been designed, but very few of these are capable of reaching the depths achieved by the *Trieste*. Most are designed for much shallower dives. Recent advancements in submersible design have centered on improving the submersible's ability to find precise locations on the seafloor, travel across the seafloor during a single dive, and collect samples at selected locations. Modern submersibles are strange-looking vessels with one or more protruding robot arms and a variety of baskets, other sample-collecting devices, video cameras, and measuring instruments that hang from the hull within reach of a robot arm (**Fig. 3-23**). They also carry powerful lights because sunlight does not penetrate more than a few hundred meters of seawater. Submersibles are now used for a variety of undersea observations, particularly observations of underwater volcanic features and of hydrothermal vents and their unique biological communities (**Chap. 15**).

Figure 3-23. A typical submersible, such as the Johnson Sea Link shown in this photograph, consists of a glass or plastic sphere inside a hull. The scientists descend within the sphere, which is provided with air and heat. Motors, articulated arms, and many different types of samplers and cameras are mounted outside the sphere. These are used to provide propulsion and collect samples.

Unfortunately, submersibles have disadvantages that limit their usefulness. Most submersibles must have a large surface vessel to carry or tow them to the research site, with a large crew to launch, retrieve, maintain, and repair them. Because submersibles are very expensive, the cost of each dive is extremely high in relation to the cost of other oceanographic research efforts.

Submersibles have very small interior crew spaces, so dives of several hours are an uncomfortable ordeal for the pilot and passengers. The discomfort, the need to carry air-recycling systems for the crew compartment, and the limited battery power available to operate the motors, life-support system, and scientific equipment are all factors that limit the maximum duration of each dive. Limited dive duration restricts the area that can be studied on each dive, particularly in deep waters where the submersible may spend several hours descending and ascending.

Remotely Operated Vehicles

Many of the limitations of submersibles can be overcome by use of a remotely operated vehicle (ROV). The basic ROV consists of a television camera mounted on a frame or sled. An electric motor, which drives a propeller, and electronically controllable steering devices are also mounted on the sled. The sled is attached to a surface ship by a cable through which power is supplied from the ship, video images are transmitted to the shipboard laboratory, and steering signals are sent to the sled. ROVs are much less expensive than submersibles and can spend many more hours underwater. They can be deployed from much smaller research vessels than the ones that carry submersibles; they do not require the difficult, expensive, and time-consuming procedures and equipment needed to protect the lives of submersible crews and passengers; and they can allow several scientists at the same time, rather than the one or two in a submersible, to view the seafloor and to direct sampling activities.

The sampling and measurement operations performed with submersibles involve remote manipulation of robot arms and other equipment located outside the submersible. Because nearly identical equipment can be mounted on ROV sleds, they can perform the same tasks as submersibles, but they are operated remotely by a pilot sitting comfortably in front of a video monitor in a ship's laboratory. In fact, if steerable video cameras are mounted on the ROV, the shipboard ROV operator has a better view of the environment than a submersible operator has, because the submersible operator's vision is often restricted to several small, fixed window ports. In addition, an ROV enables several scientists of different disciplines to take part in each "dive" and to participate, either from the research vessel or remotely by video link, in deciding when and where samples should be collected. Although they are less glamorous than research submersibles, ROVs may eventually replace almost all submersibles for ocean science.

Perhaps the most famous, although scientifically not very useful, achievement of modern research submersibles was the successful exploration of the wreck of the *RMS Titanic* and the recovery of some of its artifacts. What is not widely known is that the manned submersible dives to the *Titanic* were made possible by the work of an advanced ROV, the *Argo*. It performed the lengthy and difficult search for the wreck and the video exploration of the wreck and surrounding debris field. This preliminary work made it possible to send submersibles some months later to explore the wreck further and collect artifacts. Another ROV, *Jason Jr*., was used in the subsequent exploration of the *Titanic* by the submersible *Alvin*. *Jason Jr*. was attached to the *Alvin* rather than to a surface ship.

The *Jason II/Medea* system, developed since the exploration of the *Titanic*, is an example of sophisticated tandem ROV systems now used (**Fig. 3-24**). *Medea* is attached to a surface research vessel by a long cable. It carries lights, a video camera, and other instruments to survey the seafloor as it is positioned just above it. *Medea* can operate to depths of about 6500 m. *Jason II* is an ROV attached to *Medea* by a cable 35 m long. Once *Medea* has located an interesting area of seafloor, *Jason II* can be sent out to make a much more detailed survey of the area surrounding *Medea's* location. *Jason II* can be maneuvered very precisely because it is not connected directly to the cable attached to the ship above. Consequently, *Jason II* is isolated from the sometimes substantial ship motions that are partially transferred down the cable to *Medea*, which enables *Jason II* to more easily locate and collect samples on the seafloor. The *Jason II/Medea* system can remain at work continuously on the seafloor for a week or more. Such systems can locate or relocate themselves on the seafloor with a precision of a few meters by analysis of the travel times of sound pulses transmitted from the surface vessel, and from transponders mounted on the vehicles and anchored to the seafloor in a triangle surrounding the study area.

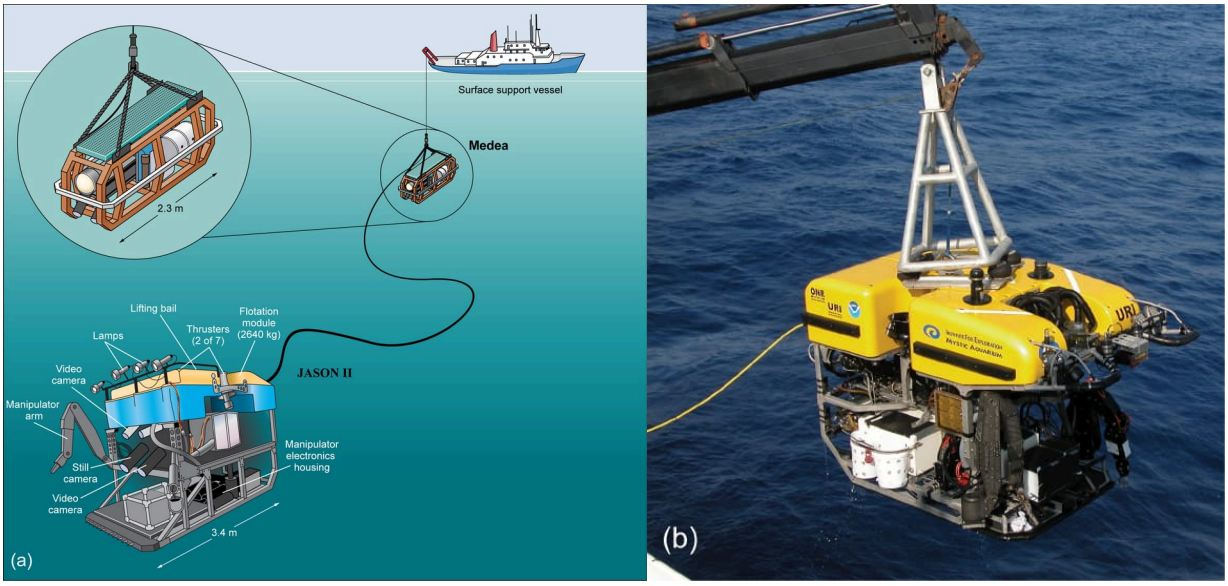


Figure 3-24. (a) The *Jason II/Medea* ROV system. The many motors, instruments, sampling arms, lights, and cameras that are mounted on *Jason II* and on *Medea* are operated remotely from a surface vessel. *Jason II* is connected to *Medea* by a 35-m-long neutrally buoyant cable and is thus isolated from any ship motions that are transmitted down the cable to *Medea*. (b) A typical ROV (*Hercules*) can do all the sampling tasks that can be performed by humans in a submersible but at a much lower cost and from smaller vessels.

Autonomous Floats and Gliders

Even ROVs may eventually become obsolete as the development of autonomous underwater vehicles (AUVs) is progressing rapidly. An AUV is a vehicle that moves through the ocean with no human occupants and with no direct connection to surface vessels or submersibles. It performs exploration and sampling tasks according to preprogrammed instructions or through limited communication of data and instructions to and from remote operators. The simplest and most widely used example of an AUV may be Argo (not the same as the ROV *Argo* mentioned earlier) (**Fig. 3-25**) and other similar floats. These are autonomous instrument packages programmed to alternately descend and ascend through the water column collecting data on the properties of **water masses**. They transmit their data to shore by radio each time they reach the surface. The floats are part of the Global Climate Observing System/Global Ocean Observing System (GCOS/GOOS), which now has about 4,000 of these floats deployed throughout the world's oceans. Some of these floats are now being equipped with additional sensors to measure parameters such as dissolved oxygen concentration, pH, nitrate concentration, chlorophyll-a concentrations, suspended particle concentration, and ambient light levels. These additional sensors are not yet proven to be as accurate and durable as the basic CDT but are expected to yield useful data and to be improved over time.

Figure 3-25. An Argo float system is deployed from a research vessel to begin its long unattended journey through the ocean depths.

One particularly interesting development of AUVs is the development of gliders. These look somewhat similar to the unmanned aerial drones now used by the military (**Fig. 3-26**). The most interesting feature of some of these gliders is that they can generate their own power by diving repeatedly between the warm surface water and colder water below and exploiting this temperature difference to generate electricity. This gives gliders the potential for deployments of virtually unlimited length of time and enables them to cover very large distances in the oceans in a single deployment. To become more useful to science, future AUVs, including gliders, will require the development of sophisticated new subsea robotics, artificial intelligence, and other technologies for underwater observation, sampling, and communications that can be incorporated in the AUV.

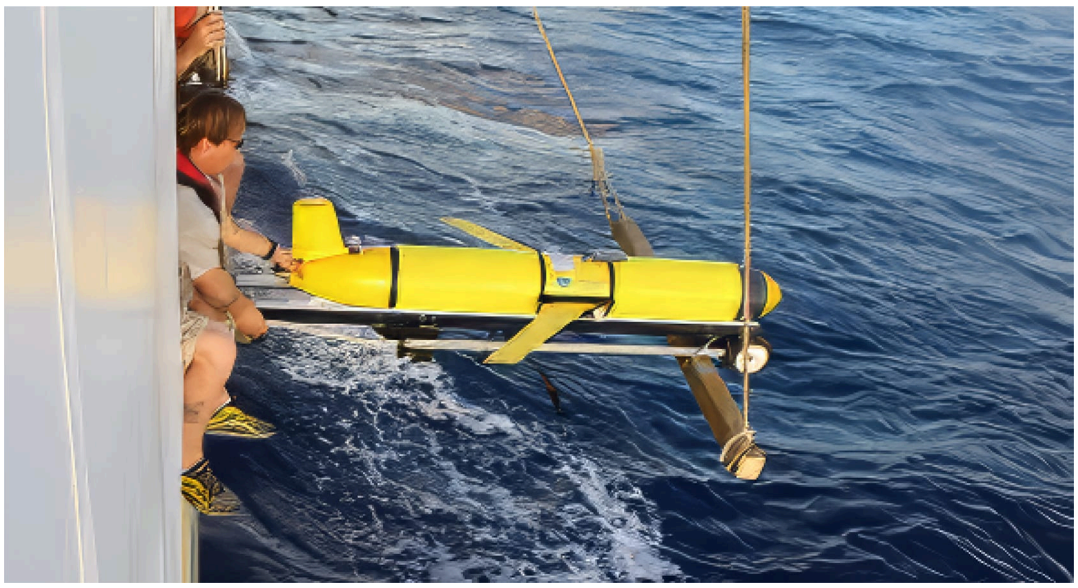


Figure 3-26. A Slocum Glider about to be deployed from a research vessel. Basic gliders can be small and easily deployed, while larger gliders can support longer missions and more complex instrumentation.

Accurate navigation and position fixing of data collected by AUV is a major problem. While there are a number of sophisticated technologies, such as inertial navigation that are adapted for AUV use, none of these systems currently provides the high degree of accuracy of positioning required for many current and potential uses of AUVs, such as returning to an exact location to resample and determining precise location for regional scale bathymetric maps. However, a relatively new and expensive technology based on a permanent or semi-permanent array of sonar beacons placed on the ocean floor (in an array similar to that in **Figure 3-18a**) can provide very accurate positioning. To use such systems, AUVs must be equipped to sense or interrogate such beacon arrays. This technology is, at present, used primarily for military AUVs used by a small number of nations that have built sonar beacon arrays in their coastal operating areas. However, this technology is expected to become less expensive and to

be used extensively for scientific study purposes in future.

This page titled 3.7: Scuba, Manned, and Unmanned Submersibles is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.8: Satellites

Satellite observations have truly revolutionized marine sciences. From a satellite, the entire surface of the world's oceans can be surveyed in just a few days (**Fig. 3-27**). In addition, large areas of the oceans can be surveyed comprehensively every few days or even hours. Such **synoptic** observations from satellites have enabled oceanographers to observe day-to-day, month-to-month, and year-to-year changes in ocean phenomena, including surface currents, surface water temperatures, waves, phytoplankton and suspended-sediment concentrations, and ice cover. Before satellites, such temporal changes could be observed only at fixed locations or within small regions because surface vessels cannot survey large regions quickly or often. Ships may take several weeks to complete a survey of several hundred square kilometers, but ocean conditions can change very rapidly, for example during storms. Hence, observations obtained at the beginning and end of a ship survey often describe very different states of the ocean. The degree to which satellite observations have provided a new view of the oceans is demonstrated by the number of figures in this text that present satellite data to describe important ocean processes.

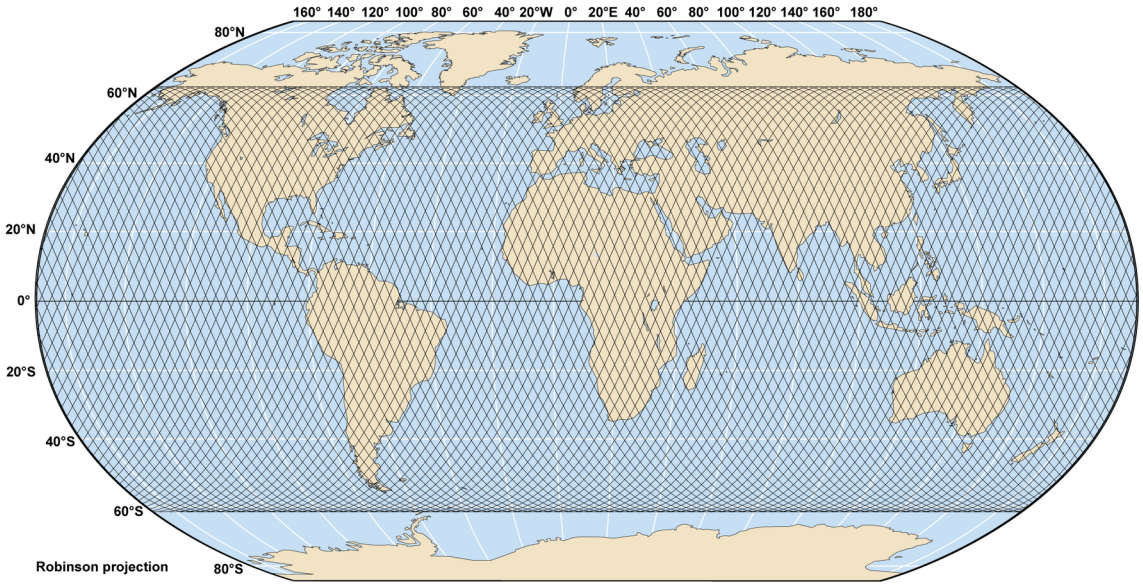


Figure 3-27. The *TOPEX/Poseidon* satellite travels in an orbit 1366 km high and inclined at 66° to the equator, which is typical of satellites used for ocean observations. This orbit allows coverage of 95% of the ice-free oceans every 10 days. The satellite's position is known to within 10 cm from the Earth's center. The network of curved lines on the map schematically illustrates the path that the satellite follows in each 10-day period.

Satellite observations of the oceans are made with a variety of devices, including radar, **lasers**, and color- and infrared-sensing scanners. Although satellites allow oceanographers to make rapid and comprehensive observations of the oceans, their value is limited because the sensors they carry rely primarily on **electromagnetic radiation**, which does not penetrate ocean water effectively. Therefore, most satellite observations record features and processes of only the upper few meters of the oceans and can provide only limited information about processes below the upper layer. However, satellites using radar can successfully monitor sea surface height with a precision that allows filtering their data on different frequencies to monitor ocean depth (**Chap. 4**), sea level rise, wave energy, tidal height variations, sea ice extent and thickness, and even internal waves and eddies well below the surface.

Satellites have made a special contribution to oceanography as aids to navigation. Until satellite navigation systems were developed, research ships far from land could not determine their locations with great accuracy. When clouds prevented star and sun sightings for several days, a vessel's position could be determined only with an uncertainty that often exceeded several kilometers. The most recent and accurate satellite navigation system is the Global Positioning System (GPS), which can determine a position anywhere on the Earth's surface to within a few meters. Precise navigation enables oceanographers to locate and map small features of the seafloor and to revisit precise locations on subsequent research cruises.

Satellites have also enabled high band width communications between land based research facilities and research vessels, platforms, and AUVs. This allows researchers to interactively participate in ocean sampling and other studies without ever leaving their home laboratory as they can monitor activities and communicate by video. This has reduced costs as fewer personnel are

needed on research vessels, but its more important impact may be that it makes it possible for a researcher who encounters an unknown species or an unusual occurrence to immediately bring in relevant experts, by video conference, wherever they may be in the world, so these experts can direct or suggest any additional study that might need to be done immediately.

Routine satellite observations of the oceans have many practical uses other than navigation and scientific research. For example, satellite measurements such as surface water temperature are used by fishers to locate the best fishing grounds and information about storms and currents are used by ships' captains to save fuel. They are also used to forecast weather, especially **hurricanes** and other storms, and to determine iceberg locations. The growing list of parameters that can be determined by satellites include currents and circulation patterns, internal waves and eddies, ocean surface temperature and salinity, surface wind patterns, **wave heights** and sea state, bathymetry, sea level rise, chlorophyll and plankton concentrations, distribution of river plumes, concentration of suspended sediment, beach **erosion**, and **shoreline** changes. Satellite observations of oceanic and atmospheric temperatures and other parameters will be particularly important in monitoring the enhanced **greenhouse effect** and its potential impact on world climate, sea level, and biological processes, and in improving our ability to understand and predict such phenomena.

This page titled 3.8: Satellites is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.9: Computers and Modeling

The development of computers revolutionized ocean sciences just as it has other aspects of human society. Ocean sciences are particularly dependent on computers for two important functions: analysis and display of the three-dimensional variations in important parameters, and data analysis using mathematical models.

Almost all studies conducted by ocean scientists involve parameters or processes that vary both geographically, with depth, over time. Data of importance to ocean processes must therefore be organized in a three-dimensional framework (four dimensions with time) and displayed visually for analysis and interpretation. Many of the illustrations in this text display data on maps or on vertical map sections. As we discussed earlier in this chapter, the relatively simple seafloor map that Heezen and Tharp generated by hand before the advent of computers was a massive undertaking. Computerized mapping now makes the development of such maps much simpler and quicker, even given the massive amounts of data that can be generated by satellites and automated sensors. Most computerized mapping is now done using Geographic Information Systems (GIS). GIS is designed to store data for any parameter that is collected at a specific location and time—for example, salinity, temperature, phytoplankton concentration, light intensity. The GIS system stores each data point referenced to its geographic coordinates (three-dimensional to include depth, where appropriate) and collection time. Simple-to-use software tools allow data for multiple parameters to be overlaid on each other (spatially and temporally) for comparison and analysis, and for these results to be displayed in a variety of ways.

Ocean sciences are heavily dependent on mathematical modeling, described in **CC10**, because the processes studied occur on very variable time and space scales, and many processes, especially biological parameters and interactions, are characterized by very limited data. Global climate models, used to investigate ocean-atmosphere interactions, greenhouse-induced global climate change, global weather patterns, and many other questions, rely particularly heavily on computers. As explained in **Chapter 8**, the scale of water motion in the ocean is one-tenth the scale of atmospheric motions that contribute to weather and climate. Furthermore, oceans are more complex vertically than the atmosphere. As a result, if the ocean part of the global model is to operate effectively, it must be split into smaller model cells and separated into more vertical layers than the atmosphere. Thus, the ocean component of these models can require as much as four orders of magnitude more computer time than the atmospheric component requires. For these models, this complexity is significant because it can take months to run the models through just a few years of simulated climate even on the fastest supercomputers.

This page titled 3.9: Computers and Modeling is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.10: CHAPTER SUMMARY

Difficulties of Studying the Ocean Environment.

Because water effectively absorbs all electromagnetic radiation, oceanographers can “see” through the ocean waters only by using sound waves, which travel through water with relatively little absorption.

The oceans average almost 4 km in depth. Observation and sampling instruments must be lowered through the water column—a time-consuming and expensive process. Until recently, precise navigation was extremely difficult away from sight of land, and resampling a specific location was nearly impossible.

Submersibles and instruments lowered through the ocean water column must be able to withstand very great pressure changes and the corrosiveness of seawater. They must also avoid fouling by organisms. Electrical equipment must be isolated from contact with seawater because seawater is an electrical conductor.

Oceanographers and their instruments must operate reliably on research ships that vibrate from engine noise, and that roll and pitch with wave action. Faulty equipment must be repaired at sea with available parts and personnel.

Mapping the Ocean Floor.

Until 1920, the only way to determine the depth of water under a ship was to lower a weight on the end of a line or wire until it touched bottom, and then measure the length of wire let out. This method was tedious and inaccurate, especially in deep waters. It was also prone to errors because it was difficult to determine when the weight reached the seafloor and because the path of the line or wire through the water column was distorted by currents.

In 1920, echo sounders were developed that measured the time taken for a sound ping to travel from ship to seafloor and back. Precision echo sounders were operated continuously on all research ships, and the data were used in the 1950s to produce the first detailed maps of seafloor topography.

Newer echo sounders send sound pulses spread over a wide area to either side of a research vessel, and the multiple echoes are processed by computer. The data are used to create detailed three-dimensional seafloor maps. Satellites can measure sea surface height so accurately that the data can be used to map the seafloor based on slight variations of sea surface height caused by gravity differences over the varying seafloor topography.

Seafloor Sediments.

Samples of seafloor sediment are obtained for several purposes by a variety of techniques. Samples of the upper few tens of centimeters of sediment are used to study recent sediment processes or the biology of the sediment. They are usually obtained with grab samplers or box corers.

Long vertical cores are obtained to study the history of sediment that is accumulated layer by layer. They are obtained by corers or by drilling. Drilling ships are very expensive to operate but can retrieve very long cores. Samples of rocks from the seafloor are obtained by dredges.

Indirect methods of studying sediments include gravity and magnetic-field strength measurements and seismic techniques in which sound pulses are echoed off layers of buried sediment.

Chemical and Physical Oceanography.

Water samples must be collected to determine most parameters of seawater chemistry and to study microscopic organisms. Samples are collected in bottles lowered to the desired depth, where they are closed and then retrieved. Sample contamination from the sampler’s materials and wire and from surface waters is often a problem because of the very low concentrations of some constituents. Some parameters, notably pressure (depth), conductivity (salinity), and temperature, can be determined in situ by

electronic sensors.

Currents are measured by tracking drifters, drogues, or floats as they move with the current; by the use of fixed current meters that measure the speed of rotation of a rotor or the tilt of the meter in the current; and by acoustic remote sensing of suspended sediment movement.

Living Organisms in the Sea.

Sampling marine organisms is difficult because pelagic species are widely dispersed, species vary greatly in size, some species avoid samplers, and some delicate species are damaged by samplers. Pelagic species are collected by nets, water-sampling devices, traps, and fishing lines. Benthic species are collected by grab samplers, dredges, and trawl nets. Some species can only be collected or observed using scuba or submersibles.

Scuba, Manned and Unmanned Submersibles.

Marine organisms in shallow waters, less than a few tens of meters deep, can be observed and sampled best by scuba divers. Scuba divers can spend only a few hours a day underwater unless they live in expensive underwater habitats. Marine organisms in deeper waters can be observed and sampled from manned submersibles, but these vessels are very expensive and can remain submerged for only a few hours. Unmanned remotely operated vehicles (ROVs) using cameras and robot arms are more economical and can remain submerged longer. Autonomous underwater vehicles (AUVs) are improving rapidly and are being used much more extensively.

Satellites.

Satellite-mounted sensors can be used to measure parameters and characteristics of near-surface waters, including temperature, concentrations of phytoplankton, concentrations of suspended sediment, and surface currents and waves. Satellite observations can cover large areas almost simultaneously and repeat these observations frequently, which is not possible with research ships or fixed monitoring systems. Seafloor topography can be inferred by very precise satellite radar measurements of sea surface elevation. Satellite navigation systems have dramatically improved oceanographers' abilities to map and return to specific features of the oceans.

Computers and Modeling.

Computers have contributed substantially to the development of ocean sciences, primarily by organizing and displaying geographic data and by performing mathematical modeling of complex systems.

3.10.1: Key Terms

3.10.2: Chapter Questions

3.10.3: References

This page titled 3.10: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.10.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- acoustic
- archaea
- bacteria
- barnacles
- bathymetry
- benthic
- biomass
- chlorophyll
- communities
- coral reefs
- contaminants
- continental drift
- CTD
- current
- density
- DNA
- electrical conductivity
- electromagnetic radiation
- environment
- erosion
- faults
- fracture zones
- frequency
- foul
- greenhouse effect
- habitat
- hurricanes
- hydrographic
- hydrophone
- hydrothermal vent
- invertebrates

- lava
- lasers
- manganese nodules
- marine mammals
- pelagic
- pH
- phosphorite nodules
- phytoplankton
- photosynthetic
- plankton
- reefs
- schools
- scruba
- seamounts
- seafloor spreading
- sediment
- shoreline
- sonar

- sounding
- submersible
- surface microlayer
- suspended sediment
- synoptic
- tomography
- topography
- turbidity
- viruses
- water masses
- wave heights
- weather
- zooplankton

This page titled 3.10.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.10.2: Chapter Questions

Study Questions

1. Why do we still know less, in some ways, about the floor of the oceans than we do about the surface of the moon?
2. What are the principal difficulties encountered in determining ocean depth and mapping the seafloor?
3. Why do oceanographers still use many different types of sampling equipment to obtain samples from the seafloor?
4. If you were designing a sampler to obtain samples of water from deep in the oceans, what factors would you need to consider?
5. Why do we believe there are many species of marine animals living in the deep sea that have never been seen or sampled?
6. Do you think manned submersibles are necessary for ocean study? Explain your answer. Describe the types of studies for which they are most useful.
7. Will satellites ever completely replace research vessels? Why or why not?
8. Why is accurate navigation important to ocean studies? Discuss the reasons.
9. What methods are used to obtain information about the sediments on and below the seafloor?
10. What methods are used to obtain information about the chemistry of and organisms living in the water column of the deep oceans?
11. Why is it easier to study ocean sediments than to study ocean water chemistry or organisms that live in the deep-ocean water column?

Critical Thinking Questions

1. The process of lowering and retrieving instruments to the ocean floor often takes much longer than is planned at a particular site. What might be some of the reasons for such delays?
2. You are planning the deployment, lowering, and retrieval of an instrument package to sample the seafloor in the deepest part of the Atlantic Ocean. How do you decide where to sample, and what do you need to do to ensure that the samples are taken at this precise location?
3. You are shown an instrument package that weighs several hundred pounds and is sitting on the deck of a research ship. You are told to deploy the package from the ship, lowering it to the seafloor, at a depth of 2 km. What do you need in order to do this, and how can you arrange to do this safely if the ship is in an area of strong wave action?
4. Before reading this chapter, did you know that the pressure changes associated with space travel are much less than those associated with travel into the depths of the oceans? Besides pressure changes, what other challenges can you think of that travel in space and in the deep oceans have in common?
5. Many oceanographers who first saw the Heezen and Tharp world ocean floor map in the 1970s and 1980s felt that they were being introduced to an entirely new planet. Why was this so? When did you first see a map of the ocean floor such as this? Why do you think many people are still completely unaware that there are long mountain chains on the ocean floor?
6. Suppose you are looking for the lost city of Atlantis, which, according to legend, sank beneath the waves. You have been told the approximate geographic location of Atlantis. What information do you need to select a sampling device that is capable of locating the city?
7. A research paper published in 1970 reported that the concentrations of several trace metals (zinc, copper, and iron) in North Atlantic Ocean waters appear to vary within a narrow range, but also appear to vary randomly with depth and location. Should you believe these data? What would you look for in the research paper to help you decide?
8. A grab sampler is lowered to the seafloor at a location that has never before been sampled, and it comes back closed but empty. What are the possible reasons for this result?
9. Undoubtedly, there are many species presently unknown to science that live in the deep oceans. It has been speculated that most of these are free-swimming in the water column and do not live sedentary lives in or on the sediment. Do you think this is a reasonable hypothesis? Why or why not?

This page titled 3.10.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

3.10.3: References

Figure 3-1a – Segar, D., and E. Stamman Segar, 1997: Diver wearing scuba rebreather system.

Figure 3-1b – Segar, D., and E. Stamman Segar, 1997: Backlit scuba diver.

Figure 3-1c – Segar, D., and E. Stamman Segar, 1997: Spectacular nudibranch.

Figure 3-4 – U.S. Naval Oceanographic Office/ National Imagery and Mapping Agency, U.S. Department of Defense; Digital Image by Dr. Peter W. Sloss, NOAA/NGDC.

Figure 3-5c – Lawrence, J. P., S. D. Ackerman, D. S. Foster, D. S. Blackwood, B. Butman, M.S. Moser, and H. F. Stewart: Sea-floor character and surface processes in the vicinity of Quicks Hole, Elizabeth Islands, Massachusetts. Woods Hole Science Center, U.S. Geological Survey Open-File Report 2006-1357, accessed 11 June 2025, Source.

Figure 3-6b,d,e – High resolution bathymetric imagery. Courtesy of Kraken Robotics Inc.

Figure 3-9d – Digital Elevation Model of a Jurassic reef from offshore Nova Scotia, Canada. Courtesy of Encana and M. Enachescu.

Figure 3-12 – General Oceanics, 1997: 1080 Go-Flo. Accessed 11 June 2025, Source.

Figure 3-13 – Adapted from Duxbury, A. C., and A. Duxbury, 1984: *An introduction to the world's oceans*. Addison-Wesley, 549 pp.

Figure 3-14 – Steve Hall, 1993: CTD deployment. Accessed 11 June 2025, Source, CC BY-SA 3.0.

Figure 3-21 – Man vyi, 2009: Cliavieaux Ste Cat'linne Jèrri. Accessed 11 June 2025, Source, Public Domain.

Figure 3-22 – Islands in the Sea 2002: 2002 Aquarius habitat. NOAA Ocean Explorer Gallery, Source.

Figure 3-23 – NOAA Ocean Explorer, 2002: The *Johnson-Sea-Link II* is launched off the stern of the *R/V Seward Johnson*. Islands in the Sea 2002, accessed 11 June 2025, Source.

Figure 3-4b – Brennanphillips, 2006: The science *ROV Hercules* (IFE/URI/NOAA) during a launch in 2005. Accessed 11 June 2025, Source.

Figure 3-25 – Polling, J., 2000: Float deployment from the *TS Golden Bear* with a rainbow in the background, NOAA PMEL Argo, accessed 11 June 2025, Source.

Figure 3-26 – NOAA Office of Ocean Exploration and Research, 2015:The Slocum Glider is deployed from NOAA Ship *Okeanos Explorer* in early March, accessed 11 June 2025, Source.

This page titled 3.10.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4: PLATE TECTONICS-EVOLUTION OF THE OCEAN FLOOR

Critical Concepts

- **CC1 Density and Layering in Fluids:** The Earth and all other planets are arranged in layers of different materials sorted by their density.
- **CC2 Isostasy, Eustasy, and Sea Level:** Earth's crust floats on the plastic asthenosphere. Sections of crust rise and fall isostatically as temperature changes alter their density and as their mass loading changes due to melting or to the formation of ice stemming from climate changes. This causes sea level to change on the coast of that particular section of crust. Sea level can also change eustatically when the volume of water in the oceans increases or decreases due to changes in water temperature or changes in the amount of water in glaciers and ice caps on the continents. Eustatic sea level change takes place synchronously worldwide and much more quickly than isostatic sea level changes.
- **CC3 Convection and Convection Cells:** Fluids that are heated from below, such as Earth's mantle, or ocean water, or the atmosphere, rise because their density is reduced. They continue to rise to higher levels until they are cooled sufficiently, at which time they become dense enough to sink back down. This convection process establishes convection cells in which the heated material rises in areas of upwelling, spreads out, cools, and then sinks at areas of downwelling.
- **CC7 Radioactivity and Age Dating:** Some elements have naturally occurring radioactive (parent) isotopes that decay at precisely known rates to become a different (daughter) isotope, which is often an isotope of another element. This decay process releases heat within the Earth's interior. Measurement of the concentration ratio of the parent and daughter isotopes in a rock or other material can be used to calculate its age, but only if none of the parent or daughter isotopes have been gained or lost from the sample over time.
- **CC11 Chaos:** The nonlinear nature of many environmental interactions makes complex environmental systems behave in sometimes unpredictable ways. It also makes it possible for these changes to occur in rapid and unpredictable jumps from one set of conditions to a completely different set of conditions.
- **CC14 Phototrophy, Light, and Nutrients:** Chemosynthesis and phototrophy (which includes photosynthesis) are the processes by which simple chemical compounds are made into the organic compounds of living organisms. The oxygen in Earth's atmosphere is present entirely as a result of photosynthesis.

Figure 4.1. Puʻuʻōʻō, the active vent of the Kilauea Volcano in Hawaii, is the result of a hot spot. This volcano often erupts relatively smoothly, but in 1997 it ejected cinders and ash into the atmosphere that "rained" on the airplane windshield like a hailstorm. The shiny surface below the vent mouth is a crust of solidified lava that conceals a rapidly flowing molten river of lava underneath. The river of lava extended several kilometers down the slopes of Kilauea, where it destroyed several homes shortly after this photograph was taken.

The ocean floor has irregular and complex **topographic** features—including mountain ranges, plains, depressions, and plateaus—that resemble topographic features on land (Fig. 3-4). During human history, those features have remained essentially unchanged aside from local modifications due to **erosion**, **sedimentation**, volcanic eruptions, and **coral reef** formation. However,

the human **species** has existed for only a very brief period of the planet's history (**Fig. 1-4**). In the billions of years before humans appeared, the face of the Earth was reshaped a number of times.

4.1: Structure of the Earth

4.2: Plate Tectonics

4.3: Plate Boundaries

4.4: Convergent Plate Boundaries

4.5: Divergent Plate Boundaries

4.6: Transform Faults and Fracture Zones

4.7: Hot Spots

4.8: Plate Interiors

4.9: Sea-Level Change and Climate

4.10: Present-Day Oceans

4.11: Chapter Summary

4.11.1: Key Terms

4.11.2: Chapter Questions

4.11.3: References

This page titled 4: Plate Tectonics-Evolution of the Ocean Floor is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.1: Structure of the Earth

The continents and ocean basins are not permanent. Instead, the location, shape, and size of these features are continuously changing, albeit imperceptibly slowly on the timescale of human experience. The global-scale processes that continuously reshape the face of the planet are, as yet, far from fully understood. However, we do know that these processes, called **plate tectonics**, originate deep within the Earth.

Layered Structure of the Earth

The Earth consists of a spherical central core surrounded by several concentric layers of different materials (**Fig. 4-2**). The layers are arranged by density, with the highest-density material at the Earth's center and the lowest-density material forming the outer layer, the Earth's crust. This arrangement came about early in the Earth's history, when the planet was much hotter than it is today and almost entirely fluid. The densest elements sank toward the Earth's center and lighter elements rose to the surface (**CC1**). The Earth's center is still much hotter than its surface. Heat is generated continuously within the Earth, primarily by the decay of radioisotopes (**CC7**).

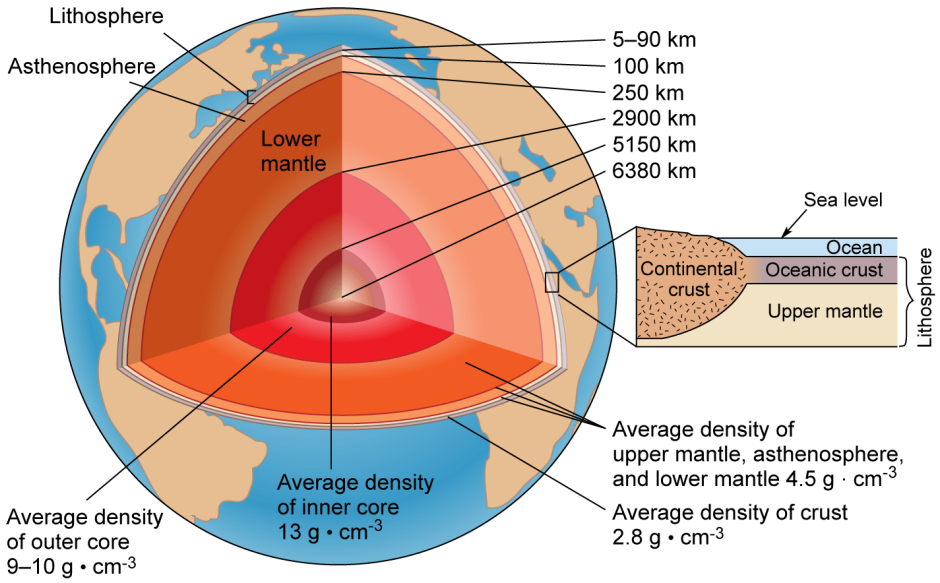


Figure 4-2. A cross section of the Earth showing its layers. Note that the thickness of the lithosphere has been greatly exaggerated in this diagram. If it were drawn to the correct scale, the lithosphere would appear as just a thin line at the Earth's surface.

The core, which is about 7000 km in diameter, is composed primarily of iron and nickel, and is very dense. It consists of a solid inner core and a liquid outer core. The **mantle**, which surrounds the core, is composed of material that is about half as dense as the core. Temperature and pressure both increase with depth below Earth's surface. As a result, the upper mantle, known as the **asthenosphere**, is thought to consist of material that is close to its melting point and is "plastic," and capable of flowing very slowly without fracturing. The best example of such materials in common experience may be **glacial** ice. Glaciers flow slowly in response to **gravity**. Within a glacier, pressure due to the weight of the overlying ice changes the properties of the ice deeper than about 50 m so that it too becomes plastic and can flow. Much of the deeper mantle is also thought to be capable of flowing very slowly. The ability to flow is a critically important factor in the tectonic processes that shape the Earth's surface.

The asthenosphere is surrounded by the **lithosphere**, the outermost layer of the Earth, which varies from just a few kilometers in thickness at the **oceanic ridges** to about 100 km in the older parts of the ocean basins and from about 130 to 190 km under the continents. The lithosphere consists of the mostly rigid outer shell of the mantle plus the solid crust that lies on the mantle. The lithosphere is less dense than the asthenosphere and essentially floats on top of the plastic asthenosphere. The oceans and atmosphere lie on top of the lithosphere. Pieces of lithosphere are rigid, but they move across the Earth's surface and in relation to each other as they float on the asthenosphere. They can be many thousands of kilometers wide, but they are generally less than 200 km thick. Because of these "platelike" qualities, they are called **lithospheric plates** (or "tectonic plates" or just "plates"). Processes that occur where the plates collide, move apart, or "slide" past each other are the principal processes that create the

mountains, trenches, and other surface features of the continents and the ocean floor.

Lithosphere, Hydrosphere, and Atmosphere

Relative to the Earth, the lithosphere is thin, rather like the skin on an apple (**Fig. 4-2**). At the top of this thin layer are the mountains, ocean basins, and other features of the Earth's surface. From the deepest point in the ocean to the top of the highest mountain is a vertical distance of approximately 20 km. This 20-km range is very small compared to the Earth's radius, which is more than 6000 km. Consequently, the planet is an almost smooth sphere (from space it looks smoother than the skin of an orange) on which the mountain ranges and ocean depths are barely perceptible.

There are two types of crust—oceanic and continental—both of which have a substantially lower density than the upper mantle material on which they lie. Oceanic crust has a higher density than continental crust (**Fig. 4-3**). According to the principle of **isostasy** (**CC2**), lithospheric plates float on the asthenosphere at levels determined by their density. Consequently, if the continental and oceanic crusts were of equal thickness, the ocean floor would be lower than the surface of the continents but not as deep as they actually are. However, oceanic crust is much thinner (typically 6–7 km thick) than continental crust (typically 35–40 km), causing an additional height difference between the surface of the continents and the ocean floor (**Fig. 4-4**).

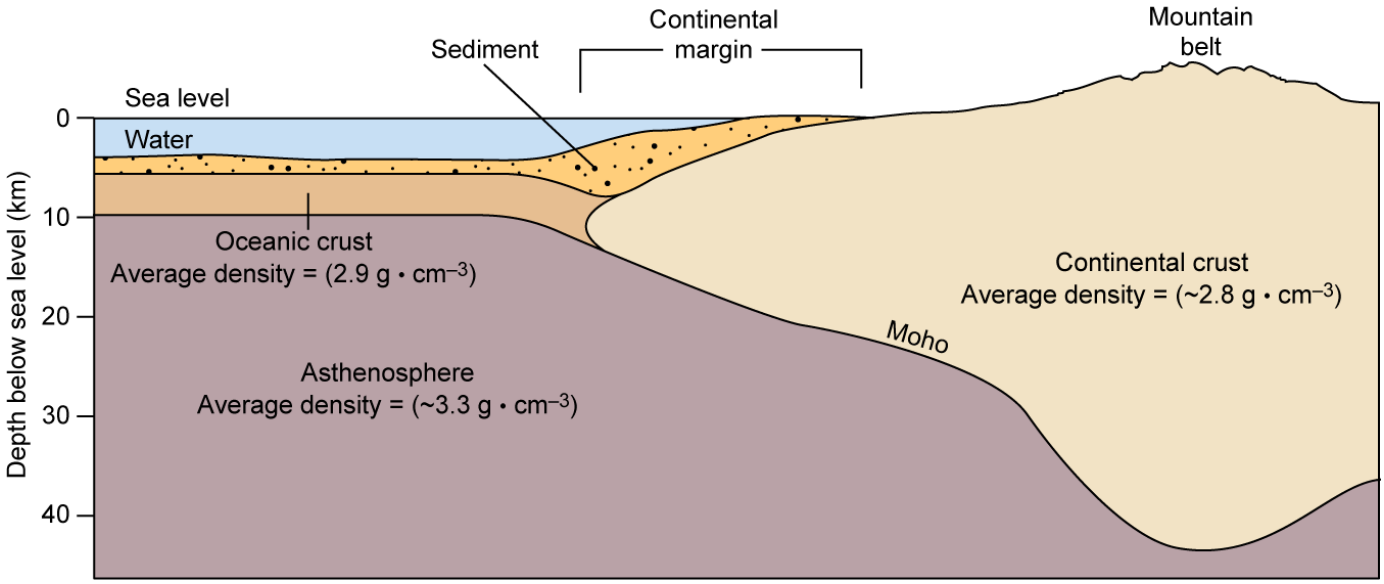


Figure 4-3. Structure of the lithosphere. Continental crust is typically 35 to 40 km thick, whereas oceanic crust is typically only 6 to 7 km thick. Continental crust has a lower density than oceanic crust, but both continental and oceanic crust have a considerably lower density than upper mantle material.

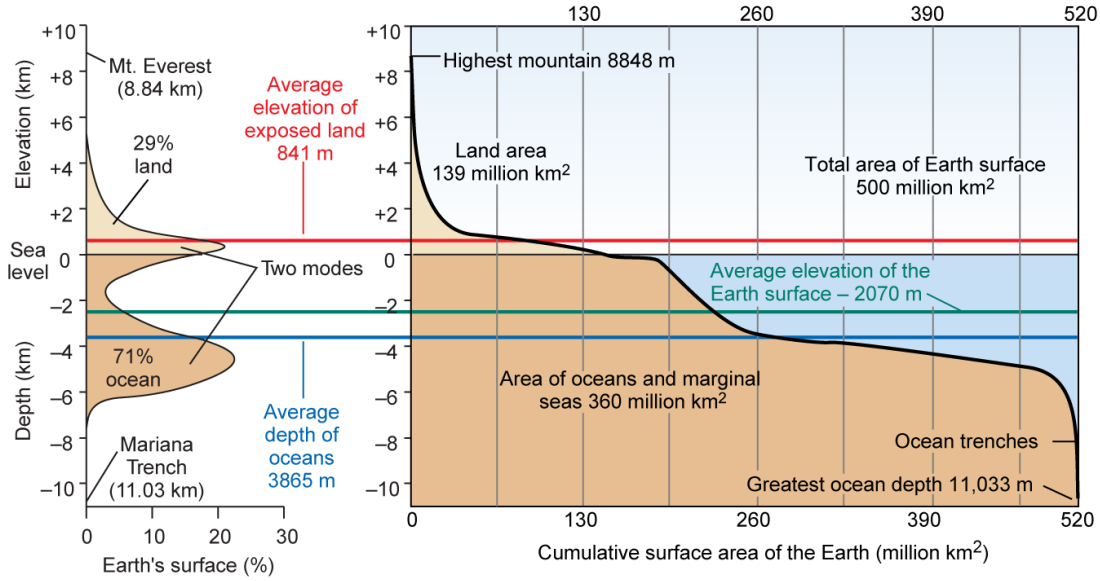


Figure 4-4. Relative heights of parts of the Earth's surface. The average elevation of the continents is 841 m above sea level, and the average depth of the oceans is 3865 m. Almost 30% of the Earth's surface is above sea level. Much of the surface of the continental crust is below sea level.

The density difference between continental and oceanic crust is due to differences in their composition. Both are composed primarily of rocks formed from cooling **magma** and consist mainly of silicon and aluminum oxides. However, continental crust is primarily granite, and oceanic crust is primarily gabbro or **basalt**, both of which have higher concentrations of heavier elements, such as iron, and thus a higher density than granite.

Surrounding each continent is a **continental shelf** covered by shallow ocean waters. The continental shelf is an extension of the continent itself, so this portion of the continental crust surface is submerged below sea level. Both the width and the depth of the shelf vary, but it is generally in waters less than 100 to 200 m deep and ranges in width from a few kilometers to several hundred kilometers. The continental shelf slopes gently offshore to the **shelf break**, where it joins the steeper **continental slope** (**Figs. 4-4, 4-5**). The continental slope generally extends to depths of 2 to 3 km. Seaward of the base of the slope, the ocean floor either descends sharply into a deep-ocean **trench** (**Fig. 4-5**) or slopes gently seaward on a continental rise that eventually joins the deep-ocean floor (**Figs. 4-4, 4-5**). Much of the deep-ocean floor is featureless flat **abyssal plain**, but other areas are characterized by low, rolling **abyssal hills**.

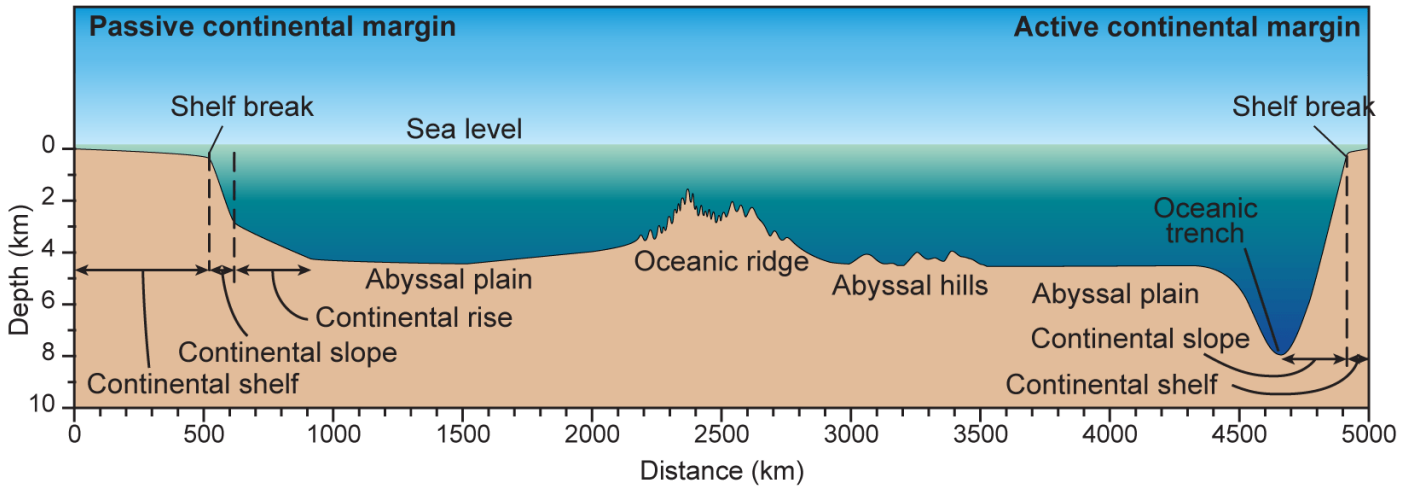


Figure 4-5. General features of the seafloor topography between the continents and the deep ocean floor.

A layer of **sediment** lies on top of the oceanic crustal rocks, constituting part of the crust. The thickness of the sediment is highly variable, for reasons discussed in this chapter and **Chapter 6.**

The **hydrosphere** consists of all water in the lithosphere that is not combined in rocks and minerals—primarily the oceans and

the much smaller volume of freshwater (**Table 5-1**). The oceans cover all the oceanic crust and large areas of continental crust around the edges of the continents; all of which total more than two-thirds of the Earth's surface area.

The atmosphere is the envelope of gases surrounding the lithosphere and hydrosphere and is composed primarily of nitrogen (78%) and oxygen (21%). Although these gases have much lower densities than liquids or solids, they are dense enough to be held by the Earth's gravity. Less dense gases, including hydrogen and helium, are so light that they tend to escape from the Earth's gravity into space. Although the light gases were present in large quantities when the Earth first formed, only trace concentrations are present in the atmosphere today. The atmosphere is discussed in **Chapter 7**.

Studying the Earth's Interior

The processes that occur beneath the lithospheric plates are very difficult to study because they occur deep within the Earth, beneath kilometers of crustal rocks and upper mantle. Studies of the processes below the crust rely primarily on indirect observations, such as examination of volcanic rocks, studies of earthquake-energy transmission through the Earth, and studies of meteorites. Meteorites are examined because they are believed to represent the types of material that make up the Earth's core and mantle. Until 2023, all samples of rocks thought to be composed of mantle material were likely chemically and mineralogically altered since no sample had ever been obtained from the mantle itself. However, in 2023, the *JOIDES Resolution* recovered a 1.268 m long drill core of nearly continuous mantle rock from a site on the Mid-Atlantic Ridge, where mantle rocks were exposed. Analysis of this core will continue for at least several years, but it has already revealed new information about mantle processes related to hydrothermal vent activity that occurs along the ridge.

A technique called "seismic **tomography**" enables scientists to use earthquake waves to study the Earth's internal structure in more detail than is possible by other means. It has yielded several intriguing findings about the Earth's interior. For example, features resembling mountains and valleys have been found on the core–mantle transition zone. The mountain-like features extend downward into the molten core and are as tall as those found on the Earth's surface. It has been hypothesized that some of these features may be sediment-like accumulations of impurities that float upward out of the liquid nickel/iron core.

This page titled 4.1: Structure of the Earth is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.2: Plate Tectonics

Over the geological timescales of the Earth's history, oceans and continents have formed and disappeared to be replaced by others through the movements of the lithospheric plates. We now have a basic understanding of how and why the lithospheric plates move and of how plate tectonics has continually shaped and changed the face of the planet.

Driving Forces of Plate Tectonics

The movements of the lithospheric plates are thought to be driven by heat energy transferred through the mantle by **convection** (**CC3**). The mantle has areas where its constituent material is **upwelled** and **downwelled**. As discussed later in this chapter, the convection processes in the mantle are complex, probably **chaotic** (**CC11**), and as yet not well understood. However, we do know that the upper mantle loses some of its heat by conduction outward through the Earth's cooler crust. As the upper material cools and contracts, its density increases. When its density exceeds the density of the mantle material below, the cooled material will tend to sink (downwell). This sinking mantle material is balanced by upwelling mantle material that has been heated by the Earth's core and **radioactive** decay within the mantle itself. As a result, this heated mantle material is slightly less dense than the material through which it rises. The precise nature and locations of the **convection cells** in the Earth are not yet known (**Fig. 4-6**). However, we believe that the convection process extends from the base of the lithosphere all the way to the core-mantle boundary, almost 3000 km below Earth's surface. We do know that flow of mantle material vertically does not take place at a uniform rate as there are discontinuities in the physical properties (that affect density) of mantle material at some depths, notably at about 650 km. Thus, sinking and rising mantle materials slows and speeds up along its path through the mantle and this causes a back up and lateral spreading of subducting slabs and ascending heated mantle material at various depths between the core-mantle boundary and the lithosphere (**Fig. 4-6**).

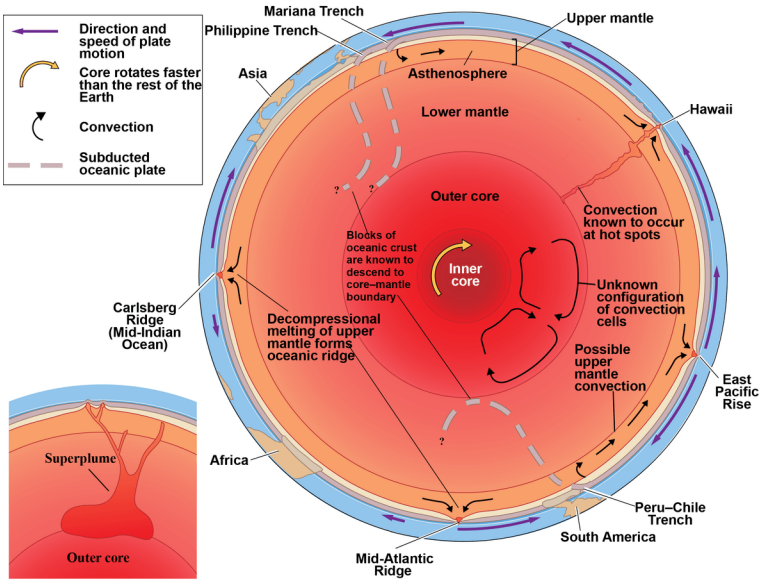


Figure 4-6. A generalized depiction of convection processes thought to occur in the Earth's mantle. The view here is from below the Southern Hemisphere. Not quite half of the Earth has been removed by slicing across the Earth at an angle to the equator. This slice removes much of the Southern Hemisphere, including Antarctica, Australia, much of the southern Atlantic and Indian Oceans, and most of the Pacific Ocean as far north as Hawaii and Indonesia. Look at a globe to see this better. It is known that the convection processes beneath some hot spots extend to the core–mantle boundary, and some subducted blocks of lithosphere may sink to this boundary, but the details of mantle convection circulation are not yet known. Convection cells must exist in the liquid outer core, but almost nothing is known about this circulation. The solid inner core rotates a little faster than the rest of the Earth, making an additional revolution about every 400 years. Scientists believe the faster rotation of the inner core may explain the periodic reversal of the Earth's magnetic field and plate tectonic processes that cause the continents to periodically gather together and then spread apart. However, much research is needed before these processes can be understood. Inset: We know little about superplumes but there are two superplumes at present; in the south Pacific Ocean south of Hawaii (in the part of Earth that is cut off in main figure); and under the African continent (further north than the tip of Africa visible in the main figure). These superplumes are thought to be a "blob" with plumes rising from it as depicted.

The existence of deep convection all the way down to the core-mantle boundary is known primarily from computer tomography.

For example, the plumes of higher-temperature mantle material beneath Hawaii and several similar **hot spots** (discussed later in this chapter; **Fig. 4-10**) have been traced through the entire mantle to the core-mantle transition. In addition, cool, rigid slabs of lithosphere, which apparently have been **subducted** at deep-sea trenches, have been detected deep in the lower mantle close to the core–mantle boundary. It is believed that the slabs of lithosphere sink into the lower mantle, where they are eventually heated and mixed with mantle material, and that some of this mixed material then rises back into the upper mantle at different locations. A single cycle of cooling, sinking, warming, and rising probably takes several hundred million years.

The areas where downwelling is thought to occur in the mantle are beneath the deep-ocean trenches, which surround most of the Pacific Ocean, and are found less extensively in other oceans (**Fig. 3-4**, **Fig. 4-10**). The areas under which mantle upwelling is thought to occur include large low shear velocity provinces (LLSVPs). In these areas, **superplumes** the size of continents "swell" more than 100 times Mount Everest's height upward from the core–mantle boundary into the mid-mantle. Two such areas are thought to lie under much of Africa and a large region of the southwestern and central Pacific Ocean, respectively (**Fig. 4-6**). There are also a number of volcanic hot spots in the crust, some of which are the locations of upwelling from deep within the mantle, and some of which may originate from superplumes. As we will see later in this chapter, there are also other hot spots and areas along the oceanic ridges where magma upwells through the lithosphere. However, this magma is thought to originate from shallow depths within the asthenosphere or upper mantle and be formed by melting of upper mantle material.

Oceanic ridges (also called "mid-ocean ridges") mark the boundaries between two tectonic plates that are moving apart. At these oceanic ridges, the lithosphere is pulled apart as the rigid tectonic plates are moved in response to subduction at other boundaries of each plate. As the plate edges separate, the weight of the lithosphere on the upper mantle at the ridge is reduced, which reduces the pressure on the upper mantle below. That upper mantle material is already hot and close to its melting point. Reduced pressure causes the melting point of the mantle material to become lower, and the upper mantle material melts and rises to fill the gap between the two plates. This results in volcanic eruptions along the ridge crest. The erupted material is cooled and solidifies as it erupts at the seafloor, and this solid material is added to the edges of both plates as new oceanic crust.

The forces that actually move the lithospheric plates across the Earth's surface act in response to the mantle convection, but they are far from fully understood. The plates are believed to be dragged across the mantle's surface as slabs of old, cold, dense lithosphere at the plate edges sink at the deep-sea trenches. The deep-sea trenches mark boundaries along which different plates meet and are called **subduction zones**. The mechanism that moves the plate is somewhat like that of an anchor (the old cold dense subducting edge of the plate) that, when dropped into the water, will drag down the line attached to it (the remainder of the rigid plate), even if, by itself, this line would float. The descending slabs of lithosphere create an additional effect similar to the vortex created as a ship sinks, which drags floating materials down with it. Finally, the plates may also move in response to gravity. Oceanic crust near the oceanic ridges, where new crust is formed, is less dense and floats higher than older oceanic crust. The oceanic crust cools, becomes more dense, and sinks as it moves away from the oceanic ridges toward the subduction zones. Thus, the oceanic crust on each plate lies on a slope, albeit a very shallow one, between the oceanic ridges and subduction zones. The plate thus has a tendency to "slide" downhill (in response to gravity), with newer, higher crust "pushing" older, lower crust down the slope.

Present-Day Lithospheric Plates

Seven major lithospheric plates are generally recognized—the Pacific, Eurasian, African, North American, South American, Indo-Australian, and Antarctic Plates—as well as several smaller plates (**Fig. 4-7**). However, some plate boundaries are not yet fully defined, and additional small plates that currently are considered to be parts of larger plates undoubtedly exist.

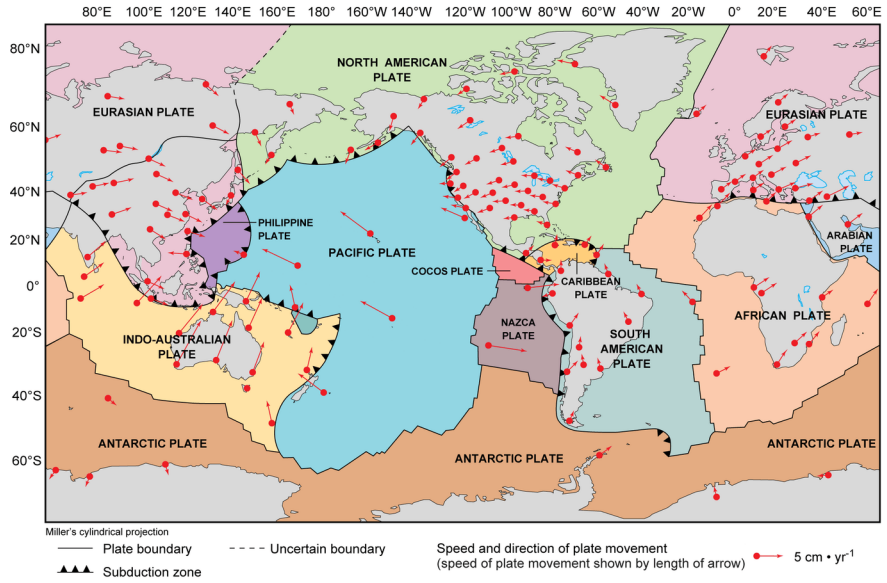


Figure 4-7. The major lithospheric plates and their current directions of motion relative to each other. Continental North America lies on the North American Plate, which extends to the center of the Atlantic Ocean. At the Pacific coast boundary of this plate, the North American and Pacific Plates are sliding past each other. The directions of motion show that the Atlantic Ocean is becoming wider, while the Pacific Ocean is becoming smaller.

All plates are in motion and, over geological time, both the direction and speed of movement can vary. **Figure 4-7** shows the directions in which the plates are moving relative to each other at present. The movement of the plates is extremely slow, just a few centimeters per year, or about the rate at which human fingernails grow. Despite their slowness, plate movements can completely alter the face of the planet in a few tens of millions of years. As they move, plates collide, separate, or slide past each other, and sometimes fracture to form smaller plates. At plate edges, the motion is not smooth and continuous; it occurs largely by a series of short, sharp movements that we feel as earthquakes. Most earthquakes occur at plate boundaries (**Fig. 4-8**).

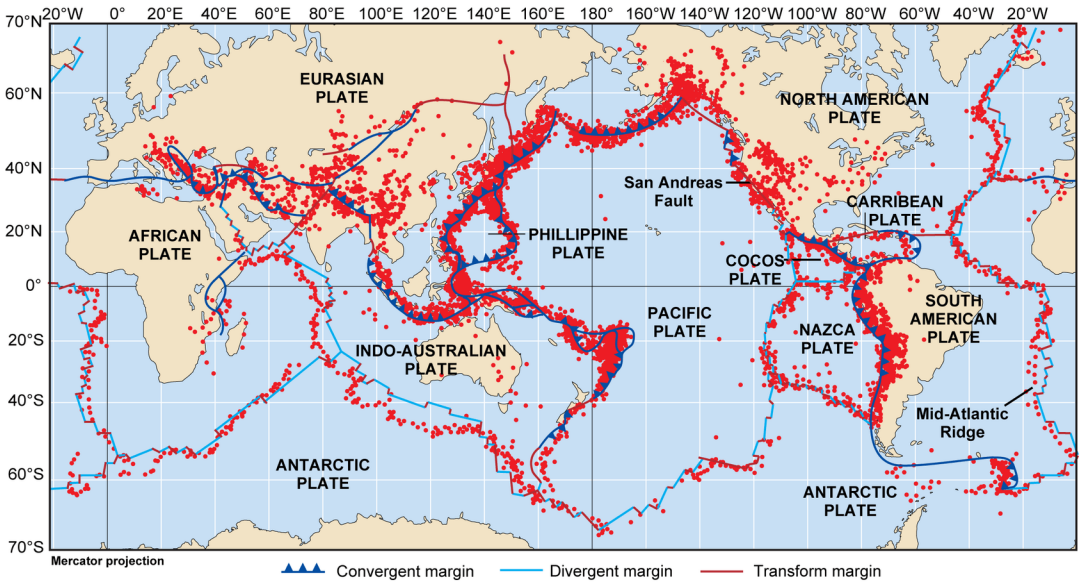


Figure 4-8. Locations of the world's earthquakes. Compare this figure with Figure 4-7. Earthquakes are concentrated along plate boundaries and appear to be more frequent at the plate boundaries where plates are converging.

Interactions between the moving plates are responsible for most of the world's topographic features. Most major mountain ranges, both on land and undersea, and all the deep-ocean trenches are aligned along the edges of plates. How plate movements create topographic features is examined later in the chapter.

Spreading Cycles

About 225 million years ago, most continental crust on the Earth was part of one supercontinent called “Pangaea.” Since then, Pangaea has broken up and the fragments have spread apart to their present locations. Continental crust breaks apart from a supercontinent, spreads out, and then rejoins to form a supercontinent in a process called a **spreading cycle**. The relative motions of the fragments of Pangaea during the past 225 million years, the current spreading cycle, have been investigated by studies of the magnetism and chemistry of rocks, **fossils**, and ocean sediments, which have revealed subtle clues as to when and where on the planet they were formed. From such studies we can determine from where and how fast the continents have drifted.

The history of the current spreading cycle is quite well known (**Fig. 4-9**). Initially the landmasses now known as Eurasia and North America broke away from Pangaea as a single block. North America later broke away from Eurasia, and South America and Africa separated from Antarctica, Australia, and India. Only much later were North and South America joined, as were Africa and Eurasia. Following the initial breakup of Pangaea, India and later Australia broke away from Antarctica. Since its break from Antarctica, India has moved northward to collide with Asia. This northward movement has been much faster than the rate of movement of other plates.

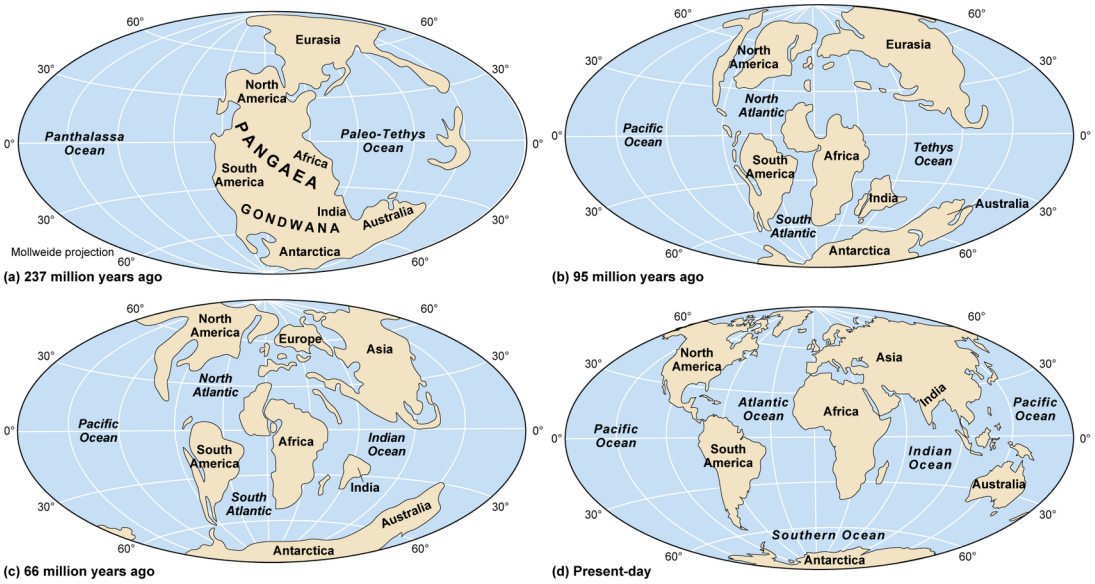


Figure 4-9. A reconstruction of the movements of the continents over the past 225 million years, which is the length of the present spreading cycle.

Although the history of the continents before the breakup of Pangaea is much less understood, Pangaea is known to have been assembled from a number of smaller continents that came together several hundred million years ago. The largest piece of Pangaea, called “Gondwanaland,” appears to have remained intact for more than a billion years. The other pieces of continental crust probably broke apart, spread out, and then re-joined to form a supercontinent several times before the formation of Pangaea. Therefore, we refer to the past 225 million years as the “current spreading cycle.”

As many as 10 spreading and assembling cycles may have occurred during the Earth’s history. Each continent is itself a geological jigsaw puzzle, consisting of pieces assembled and broken apart during previous spreading cycles. Today, for example, many areas in the interior of continents show geological evidence that they were subduction zones during earlier spreading cycles (**Fig. 4-10**). Subduction zones form at plate boundaries where oceanic crust is downwelled into the asthenosphere, so these ancient subduction zones could not have been formed where they now lie in the middle of continents. There are several types of plate boundaries, each with its own characteristic topographic features. The various types are identified in the next section, and each type is then further described in subsequent sections of this chapter.

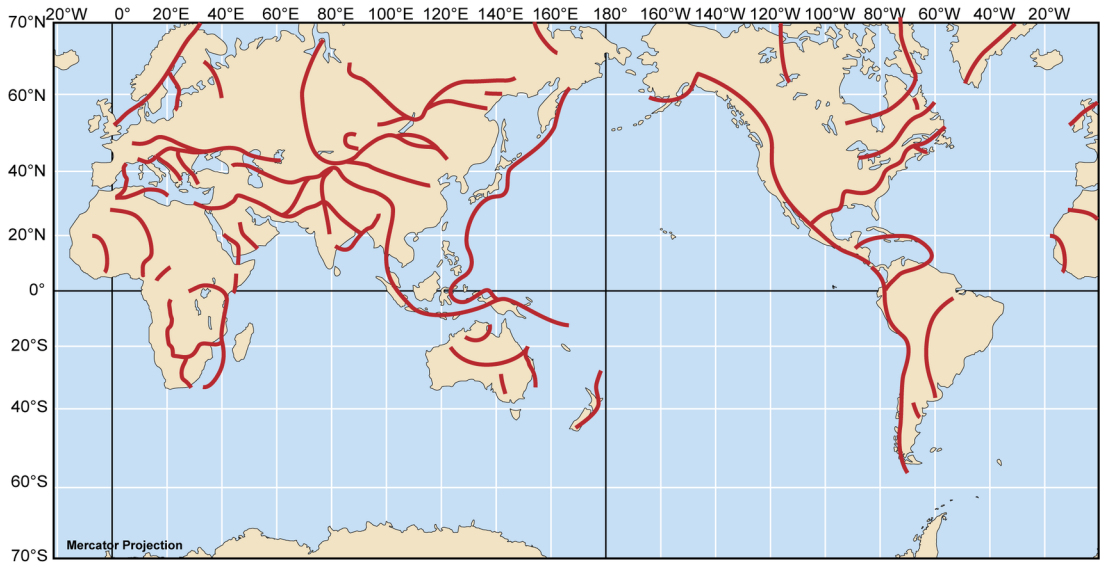


Figure 4-10. Sutures and active subduction zones. Sutures are the remnants of subduction zones where plates once converged. The oldest sutures that have been identified are in the interior of continents and are more than 570 million years old.

This page titled 4.2: Plate Tectonics is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.3: Plate Boundaries

One of three actions occurs at each plate boundary: two lithospheric plates collide with each other (**convergent plate boundary**), pull away from each other (**divergent plate boundary**), or slide past each other (**transform plate boundary**). Each action deforms the Earth's crust in a different way, creating characteristic topographic features. The behavior of the Earth's crust at a plate boundary depends on the type of crust at the edge of each of the adjacent plates and the directions of movement of the plates in relation to each other.

There are three types of convergent plate boundaries. In the first type, the edge of one **converging** plate is oceanic crust and the edge of the other is continental crust (e.g., the Pacific **coast** of South America); in the second type, both plate edges have oceanic crust (e.g., the Aleutian and Indonesian island arcs); and in the third type, both plate edges have continental crust (e.g., the Himalaya Mountains that divide India from the rest of Asia). At the first two types, where at least one of the plates has oceanic crust on its converging edge, oceanic crust is pushed under, or subducted, into the mantle (**CC3**, **Fig. 4-6**). These are called subduction zones.

Divergent plate boundaries are locations where lithospheric plates are moving apart. There are two types. The first type, called an **oceanic ridge**, is a plate boundary where both plates have oceanic crust at their edges (e.g., the Mid-Atlantic Ridge and the East Pacific Rise). The second type, called a **rift zone**, is a location where a continent is splitting apart (e.g., the East African Rift Zone). If the **divergence** at a rift zone continues long enough, a new ocean will form between the two now separated sections of the continent. The gap that would otherwise be created as two plates move apart at a divergent plate boundary is filled by magma upwelling from below.

Plate boundaries where plates slide past each other are called transform boundaries. There are few long sections of transform plate boundary on the present-day Earth. The best known stretch of transform boundary is on the Pacific coast of North America, primarily California, where the Pacific Plate slides past the North American Plate. Much shorter sections of transform plate boundary occur at intervals along all plate boundaries.

At certain locations along plate boundaries, three plates meet (**Fig. 4-7**) in areas called "triple junctions." There are two types of triple junction: stable and unstable. Stable triple junctions can persist for long periods of time, although their locations may migrate. Unstable triple junctions are those in which the relative motions of the plates cannot be sustained over time because of their geometry. The reason for this is a little complicated, but for a triple junction to be stable, the geometry must be such that each plate can continue to move at the same rate and in the same direction as a whole plate (the plate cannot "bend"). However, the rate at which the plate is subducted or added to by oceanic ridge spreading can be different along its boundaries with each of the other two plates it contacts. Triple junctions where three divergent plate boundaries intersect, as they do near Easter Island in the Pacific Ocean, are always stable. Most other triple junctions are unstable. Unstable triple junctions exhibit enhanced and complex tectonic activity, and their locations move as the interacting plates move relative to each other. An example of an unstable triple junction is the junction where three convergent plate boundaries (the Ryukyu, Japan, and Mariana trenches) intersect south of Japan (**Fig. 4-11**).

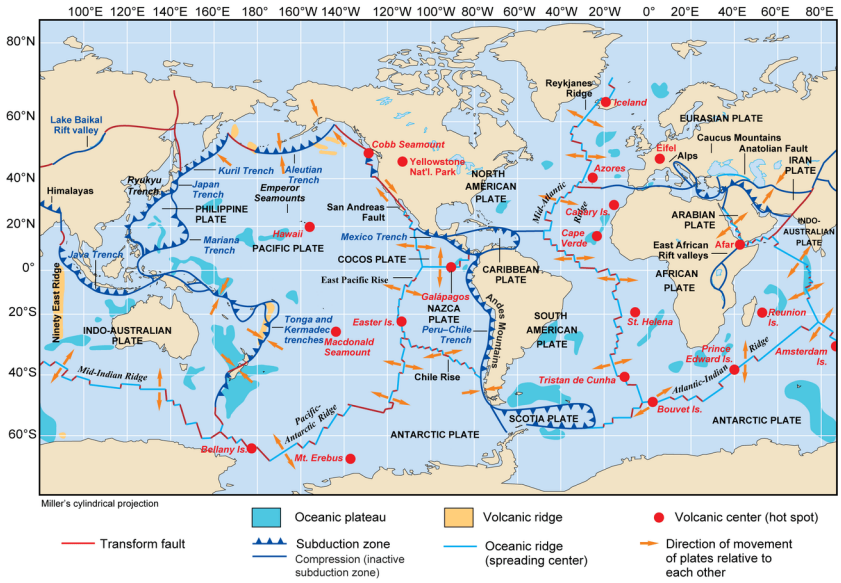


Figure 4-11. The locations of hot spots, subduction zones (trenches), and oceanic ridges. Hot spots occur both at plate boundaries and within plates. Subduction zones occur mostly in the Pacific Ocean, and oceanic ridges are interconnected throughout all of the world's oceans.

This page titled 4.3: Plate Boundaries is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.4: Convergent Plate Boundaries

At two of the three types of convergent plate boundaries, lithosphere is downwelled (subducted) at a subduction zone. Lithosphere (sediments, oceanic crust, and solid upper mantle layer) that enters a subduction zone was formed at an oceanic ridge, generally millions of years earlier (up to about 170 million years ago). Lithosphere slowly cools during the millions of years it travels horizontally at the top of the mantle away from where it was formed. As it cools, its density increases so that, by the time it enters a subduction zone, its density exceeds that of the mantle material beneath it and it can sink into (and through) the asthenosphere.

Why does the lithosphere sink through the asthenosphere only at subduction zones where two plates meet? It would seem that it should sink sooner, since, before it reaches the subduction zone, its density is already higher than that of the underlying asthenosphere. There are two reasons it does not sink sooner. First, the difference in density between lithosphere supporting old, cool oceanic crust and the warmer asthenosphere below is very small. Second, the resistance to flow is high because the lithosphere material is stretched out as a flat plate across the top of the asthenosphere (like a canoe paddle that meets strong water resistance when held flat for the thrust stroke but slices through the water easily when turned sideways). In a subduction zone, the edge of the subducting plate is bent downward (equivalent to turning the paddle blade sideways), thus allowing it to sink much more easily.

On today's Earth, most subduction zones surround the Pacific Ocean (**Fig. 4-11**). At each of these subduction zones, it is the Pacific Plate that is being subducted. The Pacific Ocean floor is being destroyed by subduction faster than new seafloor is created at its ridges, so the Pacific Ocean is becoming smaller. In contrast, the Atlantic and Indian Oceans are becoming larger. The oldest remaining seafloor sediment found in the Pacific Ocean (170 million years old) was retrieved from a hole drilled south of Japan by the International Ocean Drilling Program. All Pacific Ocean oceanic crust that existed prior to 170 million years ago, and the associated sediment, are believed to have entered subduction zones where they were either destroyed or added to the edges of continents. Most of the ocean floor is younger than the Pacific Ocean so almost no oceanic crust older than 170 million years old still exists on Earth because almost all older crust has entered subduction zones and been destroyed. The oldest oceanic crust so far discovered is in a small area called the Herodotus Basin between Cyprus, Crete and Egypt in the Mediterranean Sea dating to about 340 million years old. This area is thought to be a fragment of the Tethys Sea that once bordered the supercontinent Pangea (**Fig. 4-9a**).

Subduction Zones Next to Continents

The subduction zones that are located along the coastlines of continents mark convergent plate boundaries where a plate that has oceanic crust at its edge converges with a plate that has continental crust at its edge (**Fig. 4-12**). As the two plates move toward each other, the lithosphere of the plate supporting oceanic crust is thrust beneath the plate that supports the less dense continental crust. As the oceanic crust sinks beneath the continental crust, a subduction zone is formed with a characteristic deep trench parallel to the coast.

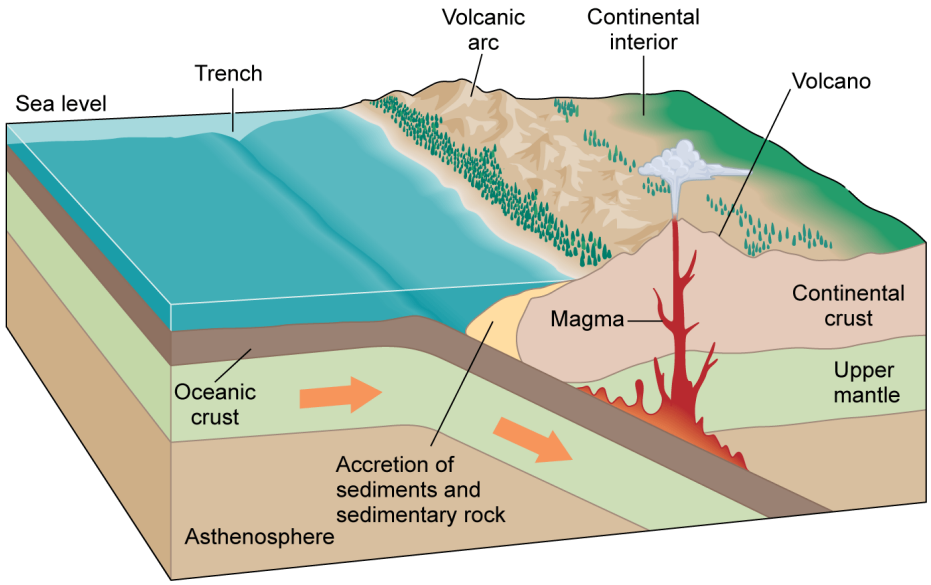


Figure 4-12. Subduction zone at a continental margin. A plate with oceanic crust is subducted beneath a plate with continental crust. This process forms a trench, a chain of volcanoes near the edge of the continent, and a chain of coastal hills or mountains, which are formed by compaction of the continental crust edge and accretion of compacted sediments and sedimentary rock from the subducting plate.

The edge of a continent that is carried into the plate boundary at a subduction zone is squeezed and thickened, forming a chain of coastal mountains along the edge of the continent. Because ocean sediments and sedimentary rocks have lower densities, like the continental rocks from which much of the sediment originated (**Chap. 6**), they are relatively **buoyant**. As a result, these materials tend to be "scraped off" the oceanic plate as they are dragged downward into the subduction zone rather than subducted with it. The sedimentary materials are further compressed, folded, and lifted as more such material accumulates from continuing subduction of additional oceanic crust. Thus, some of the marine sediments from the subducting oceanic crust are collected in the subduction zone and contribute to the formation of the coastal mountains. This process explains why Darwin found marine fossils high in the Andes Mountains (**Chap. 2**).

Volcanoes

Another process that occurs at subduction zones is the formation of a line of volcanoes on the continental crust located a few tens or hundreds of kilometers inland from the plate boundary. The volcanoes are formed when oceanic crust and some associated sedimentary material are subducted beneath the edge of the continental plate. As this oceanic crust and sediment are subducted deeper into the Earth, they are heated by the **friction** of their movement, by the hotter mantle material below, and by increasing pressure. This causes water and other volatile constituents to be released by the subducting plate and its associated sedimentary material as fluids, which migrate upwards and combine with the mantle material above the plate. This causes physical and chemical changes in the mantle material, lowering its melting point, and it melts to form magma with a high concentration of water and other volatile constituents. The resulting magma rises through the overlying continental crust and toward Earth's surface, where it erupts to form volcanoes near the edge of the continental plate (**Fig. 4-13a,c,d**). The erupting magma is rich in silica, which makes it more viscous (resistant to flow). As the magma erupts, pressure decreases, and water and other volatile constituents become gaseous and expand rapidly. The result is that these eruptions are often explosive, ejecting large amounts of ash and other volcanic material.

Figure 4-13. (a) Mt. St. Helens, in Washington State, is a subduction zone volcano that erupted violently in 1980. The top of the mountain was blown away during the eruption, leaving behind a large crater. This explosive eruption released about 24 megatons of thermal energy and ejected about 0.1 km^3 of ash. (b) The eruption of Mt. Pinatubo in the Philippines in 1991 was the largest volcanic eruption worldwide between 1912 and 2022 when the underwater Hunga Tonga eruption, shown here, was of similar magnitude. The Mt. Toba eruption 73,000 years ago was very much larger: Mt. Pinatubo erupted about 4 km^3 of ash, Toba produced at least 800 km^3. (c) Mt. St. Helens, May 1980. (d) Mt. Saint Helens September 10th, 1980.

Most active subduction zones in which the two plates have continental and oceanic crust at their respective edges are in the Pacific Ocean, and therefore, they are often called “Pacific-type margins.” Such margins include the west coast of Central and South America, the west coast of North America from northern California to southern Canada, and the coast of South-central Alaska. These areas have a well-developed ocean trench, coastal mountain range, and inland chain of volcanoes that characterize Pacific-type plate margins (**Fig. 4-14**).

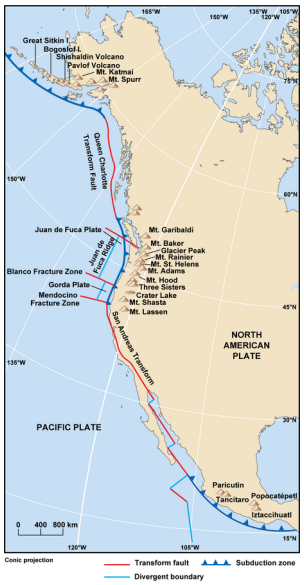


Figure 4-14. Stretches of the west coasts of Mexico, California, Oregon, Washington, Canada, and Alaska are characterized by trenches and continental subduction zone volcanoes. The active volcanoes associated with the subduction of the Juan de Fuca and Gorda Plates under the North American Plate do not extend farther south than Mount Lassen in northern California. To the south, the San Andreas Fault, which can be traced from Cape Mendocino through southern California and into northern Mexico, is a transform fault and not a subduction zone. Continuing southward, active volcanoes are again found in central Mexico, where the transform faults of California end and the subduction zone between the Cocos Plate and the North American Plate begins.

Exotic Terranes

Throughout the oceans are small areas where the seafloor is raised a kilometer or more above the surrounding oceanic crust. Called **oceanic plateaus**, these areas constitute about 3% of the ocean floor (**Fig. 4-11**). Some oceanic plateaus are extinct seafloor volcanoes, others are old volcanic ridges, and still others are fragments of continental crust called "microcontinents." Small extinct undersea volcanoes can be broken up and subducted with oceanic crust. However, the larger oceanic plateaus are too thick to be subducted. Therefore, when sections of oceanic plateau enter a subduction zone, their rocks become welded onto the edge of the continent to form **exotic terranes** (**Fig. 4-15**). Exotic terranes may also be formed when islands, including those originally formed at **magmatic arc** subduction zones (see the next section), are carried into a subduction zone where they impact the edge of a continent. Much of the Pacific coast of North America consists of exotic terranes.

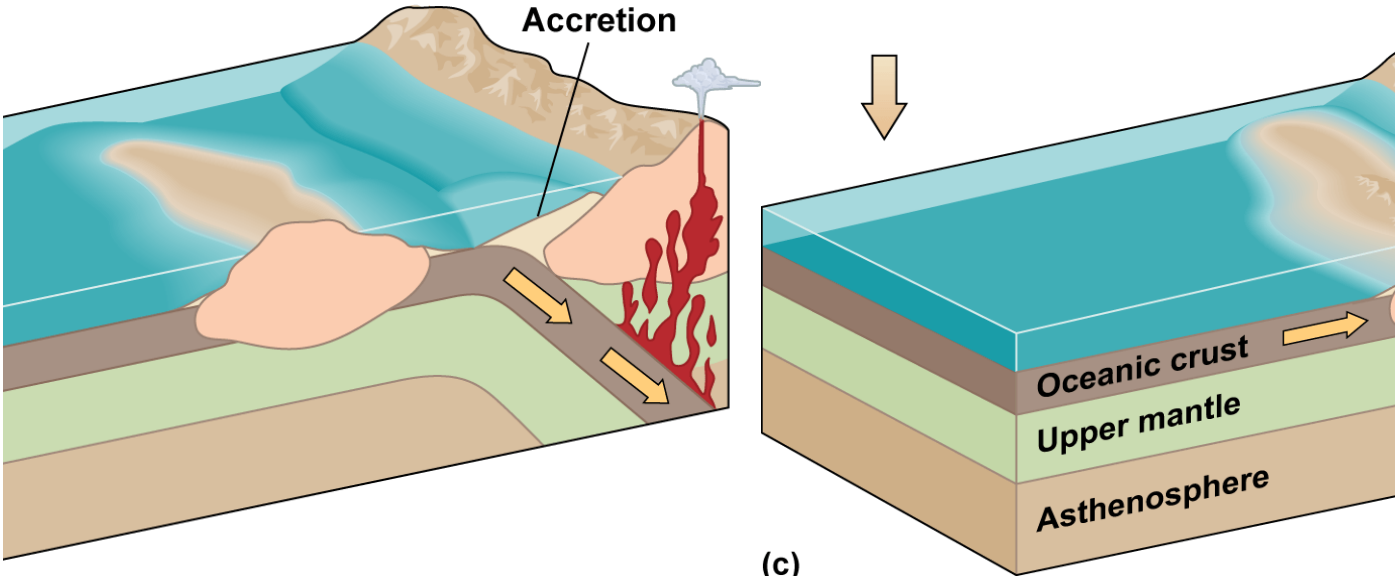


Figure 4-15. Formation of exotic terranes. Oceanic plateaus, inactive oceanic ridges, and volcanoes are scraped off the oceanic crust as it enters a subduction zone. The scraped-off material forms new continental crust that is welded onto the edge of the continent and is called an exotic terrain.

Subduction Zones at Magmatic Arcs

At **oceanic convergent plate boundaries** where both plates have oceanic crust at their edges, a chain of volcanoes erupts magma to form an island chain parallel to the subduction zone. Because the most prominent examples of this type of plate boundary on the present-day Earth are curved (arced), these boundaries are called magmatic arcs (or "island arcs"). Processes that occur at such plate boundaries are similar to processes that occur at subduction zones at a continent's edge. However, because no continental crust is present, the oceanic crust of one plate (which has higher density) is subducted beneath the oceanic crust of the other plate (which has lower density). Because oceanic crust cools with age, and density increases with decreasing temperatures, older oceanic crust has a higher density and is subducted beneath younger crust. For example, at the Aleutian Islands plate boundary, the older Pacific Plate is being subducted below the younger lithosphere of the Bering Sea portion of the North American Plate.

The subduction of oceanic crust at magmatic arcs forms a trench system that parallels the plate boundary (**Fig. 4-16**). Sedimentary materials from the subducting plate accumulate on the edge of the nonsubducting plate, just as they do at subduction zones at a continent's edge. They may form a chain of low **sedimentary arc** islands joined by an underwater ridge called an "outer arc ridge." Behind the ridge is an outer arc basin, an area of the non-subducting plate where the crust is affected little by the subduction processes.

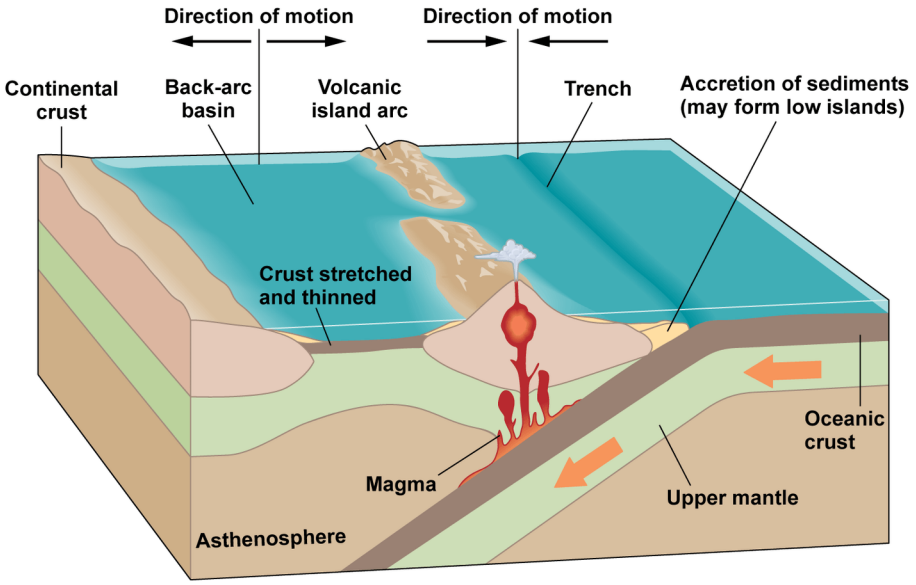


Figure 4-16. At magmatic arc subduction zones, as the plate with the higher-density oceanic crust is subducted, a chain of volcanic islands (magmatic or island arc) is formed along the edge of the nonsubducting plate. Sometimes low sedimentary islands also form between the island arc and the trench. At some magmatic arcs, such as shown here, the oceanic plate subducts fast enough that the adjacent plate (shown with a continental crust edge in this example) does not move toward the plate boundary as fast as the subduction occurs. In this situation, a trench and island arc form outside a back-arc basin, which is created by stretching and thinning of the non-subducting plate edge (the continent edge in this example). The directions of plate motion shown are relative to each other. In some cases, the trench may migrate seaward, while the non-subducting plate migrates more slowly in the same direction so that a back-arc basin is still formed.

Most present-day magmatic arc subduction zones are around the Pacific Ocean. They include Indonesia (Fig. 4-17), the Mariana Islands, the Aleutian Islands, and Japan (Fig. 4-6). Each of these areas has the characteristic trench and magmatic arc, some have the low outer arc islands, and some have shallow, sediment-filled **back-island basins**, as described in the next section.

Figure 4-17. The outer arc ridge, volcanoes, and back-arc basins of Indonesia and the Philippines. The numerous active volcanoes on the major islands of Indonesia, including Sumatra and Java, are evidence of the very active nature of this oceanic convergence. The eruptions of Krakatau in 1883 and Tambora in 1815 were two of the largest eruptions of the past several centuries.

Magmatic Arc Volcanoes

On the nonsubducting plate of a magmatic arc subduction zone, a line of volcanic islands (the magmatic arc) forms parallel to the plate boundary (Fig. 4-16). The islands are constructed by rising magma, which is produced by the sinking, heating, and subsequent melting of subducted oceanic crust and mantle material. These volcanic islands are equivalent to the chain of

volcanoes formed on the continental crust of subduction zones at the edges of continents. Like their continental counterparts, magmatic arc volcanoes often erupt explosively because subducted sediments and their associated water are also heated and erupted with the magma. One of the best known of these explosive eruptions is the 1883 eruption of Krakatau (Krakatoa) in Indonesia (**Fig. 4-17**). This eruption altered the Earth's **climate** for several years afterward, and the eruption and resulting **tsunami** killed an estimated 36,000 people. More recently, in 2022, Hunga Tonga, a large volcano on the Tonga subduction zone whose peak was about 150 m underwater at the time, erupted explosively. The force of the eruption was hundreds of times more powerful than the atomic bomb dropped on Hiroshima in World War 2. The volcano ejected an estimated 10 km3 of rock ash and sediment and created a massive tsunami that caused damage thousands of kilometers away. Hunga Tonga also caused the largest explosive shock wave that has ever been recorded. This shock wave travelled around the world several times. The eruption also ejected a huge amount of water vapor to the atmosphere, equivalent to about 10% of all the water vapor in Earth's atmosphere at any one time, or enough to fill 58,000 swimming pools. Fortunately, the explosion occurred in a sparsely populated area of the Pacific and caused only 6 known deaths.

At all convergent plate boundaries, the distance between the plate boundary and the line of volcanoes is shorter where the subducting plate's angle of descent is greater. Because the lithosphere cools and its density increases as it ages, older lithosphere tends to sink more steeply (faster) into the asthenosphere than younger lithosphere. Thus, the distance between the trench and the volcano chain is less where the subducting lithosphere is older.

Back-Arc Basins

Sometimes a subducted plate's rate of destruction can exceed the rate at which the two plates are moving toward each other, particularly where old lithosphere is sinking steeply into the asthenosphere. Under these circumstances, the edge of the nonsubducted plate is stretched, which causes a thinning of the lithosphere at its edge. The thinning may create a **back-arc basin** (sometimes called a "back-island basin") behind a magmatic arc (**Fig. 4-16**). In extreme cases, the lithosphere is stretched and thinned so much that magma rises from below to create new oceanic crust in the basin.

Back-arc basins are generally shallow seas with a floor consisting of large quantities of sediment eroded by wind and water from the newly formed mountains of the magmatic arc and from nearby continents. The Mariana Trench subduction system, where the Pacific Plate is being subducted beneath the Philippine Plate, provides an example of a back-arc basin. The back-arc basin lies between the Mariana Trench–Island subduction zone and the Philippine Trench. About 200 km to the west of the Mariana subduction zone is a now-inactive oceanic ridge that is the former location of back-arc spreading (**Fig. 3-4**).

Collisions of Continents

Continental collision plate boundaries form at the convergence of two plates that both have continental crust at their edges. When the lithosphere of one plate meets the edge of the other plate, neither is sufficiently dense to be dragged into the asthenosphere and subducted. Therefore, the continental crusts of the two plates are thrust up against each other. As the collision continues, more continental crust is thrust toward the plate boundary, and the two continents become compressed. One continent is generally thrust beneath the other, lifting it up. The forces created and the energy released by such a collision are truly immense, raising a high mountain chain along the plate boundary. The effect of such collisions is not unlike the effect of a head on collision of two cars, in which each car's hood is compressed and crumpled upward, and one car may ride partially under the other.

Collision plate boundaries are relatively rare on the Earth today, but the geological record indicates that they were more common in the past, such as when Pangaea was formed from preexisting continents. The most prominent continental collision plate boundaries today are the ones between India and Asia and between Africa and Eurasia (**Fig. 4-11**). Older examples, formed before the present spreading cycle, include the Ural Mountains of Russia and the Appalachian Mountains of North America.

The ongoing India–Asia collision is a particularly vigorous collision. It began about 40 million years ago when the edges of the continental shelves of the two continents first collided (**Fig. 4-18**). The movement of India toward Asia before the collision was very fast in relation to the speed at which other plates are known to move. The high speed explains the extreme violence of the India–Asia collision, during which India has been thrust under the Asian continent, creating the Himalaya Mountains and compressing and lifting the vast high steppes of Asia. The Himalaya Mountains continue to be uplifted, and the many powerful earthquakes felt throughout the interior of China and Afghanistan are caused by earth movements that release the compressional stress continuously built up by the India–Asia collision. This collision may also be compressing and fracturing the Indo-Australian Plate across its middle, as its northern section is slowed by the collision and its southern section continues northward.

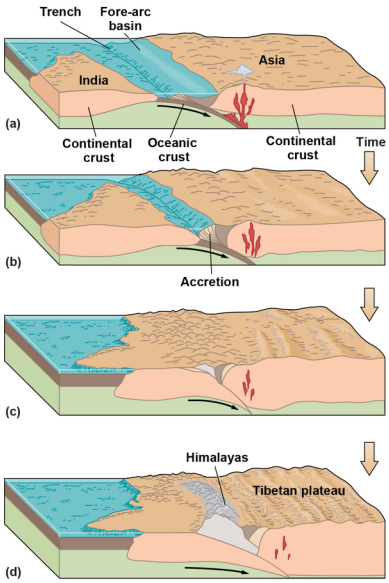


Figure 4-18. The continental collision boundary between India and Asia. About 50 million years ago, the northern margin of the Indian continental crust began a collision with the southern margin of the Asian continental crust. India is being thrust under the Asian continent as the collision continues. The Himalayas and the high Tibetan Plateau were both created by the collision. The tops of the Himalaya Mountains are formed from sedimentary rocks scraped off the oceanic crust of the ocean floor destroyed as the continents first came together.

A similar continental collision began about 200 million years ago between Africa and Eurasia. The convergence is slower than the India–Asia convergence, so the collision is less violent. Nevertheless, the Africa–Eurasia collision is responsible for building the Alps and for the numerous earthquakes that occur on the Balkan Peninsula (parts of Turkey, Greece, Albania, Bosnia and Herzegovina, Bulgaria, Croatia, Macedonia, Romania, Slovenia, Serbia, and Montenegro) and the adjacent region of Asia (parts of Turkey, Armenia, Georgia, Azerbaijan, and Iran).

This page titled 4.4: Convergent Plate Boundaries is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.5: Divergent Plate Boundaries

The processes by which divergent plate boundaries are first formed are not fully understood. It is believed that rift zones in the continents may be at least partially caused by upwelling of magma accumulated under the continents, and that this magma may accumulate because the continental crust acts as an insulator and causes heat to build up in the mantle below it. However, the most prominent rift zone on the Earth today, the East African Rift Zone, is thought to have been caused by upwelling associated with the superplume that is believed to exist in the lower mantle beneath the African continent.

At oceanic ridges, new oceanic crust is continuously formed by upwelling of magma from the asthenosphere as the plates move apart, but this upwelling is not the cause of the plate divergence. Instead, the magma upwelling occurs in order to fill the gap created when sinking (downwelling) of lithosphere at a subduction zone elsewhere on one (or both) of the plates moves the entire plate toward this subduction zone, "pulling" the plate away from the oceanic ridge and creating a divergent boundary.

Oceanic Ridges

When two plates are pulled apart (diverge), the gap between them is filled by upwelled magma. The reduction of weight on the upper mantle that occurs as the plate edges are drawn apart causes solid mantle material to melt as pressure is reduced. This hot, low-density magma rises through the splitting crust and spreads on the seafloor from a series of volcanoes aligned along the gap. This is the primary process responsible for building the connected undersea mountain chains of the oceans (**Fig. 3-4**) and is often called **seafloor spreading**. The mountain chains are known as oceanic ridges and oceanic rises. Most are in the approximate center of their ocean basins and follow the general shape of the coastlines on either side of the ocean (**Fig. 4-11**). An example is the Mid-Atlantic Ridge, which bisects the Atlantic Ocean from near the North Pole to the southern tip of South America (**Fig. 4-19**, **Fig. 3-4**).

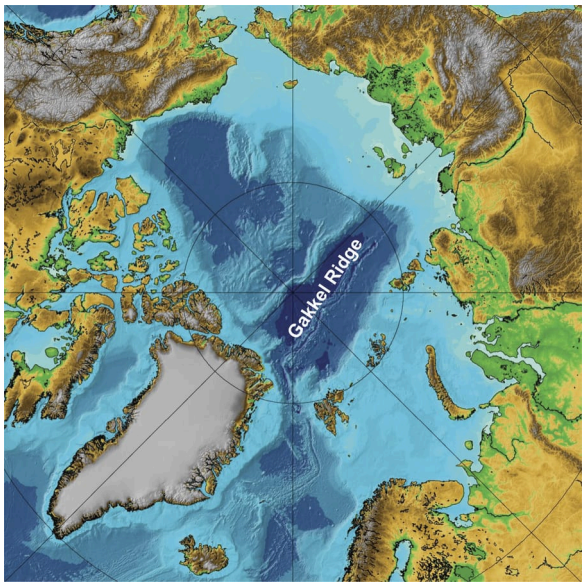


Figure 4-19. Topography of the Arctic Ocean floor. The Mid-Atlantic Ridge extends into the Arctic Ocean. This extension is named the Gakkel Ridge. Shallow sills restrict water circulation between the Arctic Ocean and other oceans. The sill between the Arctic Ocean and the Bering Sea is especially shallow, and the circulation between the Arctic and Pacific Oceans is more restricted than the circulation between the Arctic and Atlantic Oceans.

As the plate edges separate at an oceanic ridge, new oceanic crust is formed by a sequence of processes that cause oceanic crust to be layered. Molten magma upwells from below and forms a chamber of molten magma below the seafloor. Some magma cools and solidifies on the sides of this chamber to form a rock called "gabbro." This becomes the lowest of four layers that are found consistently in almost all oceanic crust. The next layer up is formed by magma that rises through vertical cracks and solidifies in wall-like sheets called "dikes." The third, and initially upper, layer is composed of pillow basalt, which is formed when erupting **lava** is cooled rapidly by seawater. The final layer of oceanic crust is composed of sediments deposited from the ocean water column.

New crust formed at an oceanic ridge or rise is hot and continues to be heated from below as long as it remains above the divergence zone. Because it is hot, it is less dense than older, cooler crust. Beneath the new crust, the solid mantle portion of the lithosphere is also thin. Therefore, the new crust floats with its base higher on the asthenosphere (**CC2**), and consequently, the seafloor is shallower. As the new crust is moved progressively farther from the divergent plate boundary, it cools by conduction of heat to the overlying water and becomes denser. In addition, the lithosphere slowly thickens as mantle material in the asthenosphere just below the base of the lithosphere cools, solidifies, and is added to the lithospheric plate. Therefore, the aging crust on the plate sinks steadily to float lower on the asthenosphere. Thus, oceanic ridge mountains slowly move away from the divergence zone and sink, while new mountains form at the divergence zone to take their place.

Oceanic Ridge Types

There are three basic types of oceanic ridges distinguished by spreading rate, the rate at which the two plates are being pulled apart. Two of these types constitute a majority of the length of the world's oceanic ridge system. The third, where the spreading rate is very slow, is less well studied.

The two faster-spreading types of oceanic ridges have certain common characteristics. Each has a string of volcanoes aligned along the axis of the ridge. The volcanoes create a continuously spreading layer of new oceanic crust on the ocean floor. The most common type of oceanic ridge, exemplified by most of the Mid-Atlantic Ridge, has a well-defined, steep-sided **central rift valley** extending down its center (**Fig. 4-20**, **Fig. 3-4**). The valley is the gap formed by the two plates pulling apart. Numerous earthquakes occur beneath the valley, and volcanic vents erupt magma to create the new oceanic crust. Central rift valleys are generally 25 to 50 km wide and 1 to 2 km lower than the surrounding peaks. The oceanic ridge mountains stand 1 to 3 km above the surrounding ocean floor and are extremely rugged.

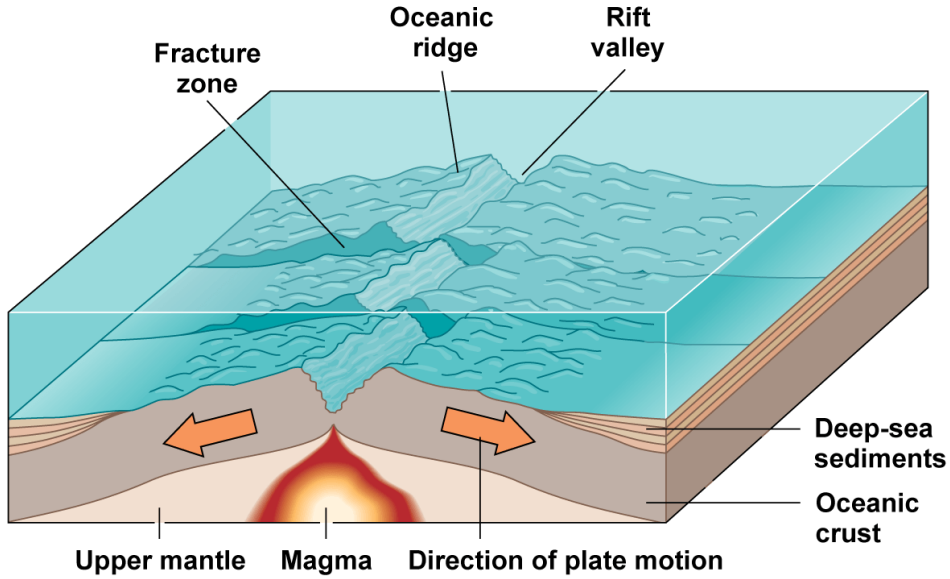


Figure 4-20. Oceanic ridge divergent plate boundaries are characterized by a ridge with a central rift valley and by fracture zones. On slowly spreading ridges, such as the Mid-Atlantic Ridge, the rift valley may be 30 km wide and up to 3 km deep, and the ridge is rugged with steep slopes. On rapidly spreading ridges, such as the East Pacific Rise, the rift valley is typically 2 to 10 km wide and only 100 km or so deep, and the ridge is broad with low-angle slopes and relatively smooth topography.

The second type of oceanic ridge, exemplified by the East Pacific Rise, is much broader and less rugged than the Mid-Atlantic type, and the central rift valley is absent or poorly defined (**Fig. 3-4**). The lack of a central rift valley is thought to be due to a faster spreading rate. The less rugged topography of this type of ridge is reflected in the seafloor of the surrounding plate, which is flatter than the seafloor surrounding more slowly spreading ridges.

The third type of oceanic ridge occurs where the spreading rate is extremely slow. This type of oceanic ridge has very few and widely spaced volcanoes. In addition, much of the new seafloor generated at these ridges is missing the characteristic basalt layer found elsewhere. This layer is thought to be missing because magma at these slowly spreading ridges has time to cool and solidify before it rises to the seafloor. The entire process appears to be different at these very slowly spreading ridges than at other oceanic ridges. At very slowly spreading ridges, the seafloor is apparently just cracked apart between volcanoes, and warm but solid rock then rises to become new seafloor. The Gakkel Ridge beneath the Arctic Ice Cap, northeast of Greenland, and the

Atlantic-Indian Ridge as it snakes around the southern tip of Africa are examples. By some estimates, up to 40% of the oceanic ridge system worldwide may be of this extremely slowly spreading type.

Oceanic Ridge Volcanoes

Where the peaks of oceanic ridge volcanoes approach the ocean surface, as they do near Iceland, eruptions can be violent. In such cases, numerous small steam explosions are created at the volcano vent when hot erupting magma and seawater come in contact (you may have seen video footage of such eruptions). However, most eruptions at oceanic ridge volcanoes are quiet and smooth-flowing, despite the magma–seawater contact. There are several reasons for the lack of explosiveness. First, unlike the magma of subduction zone volcanoes, the magma rising into oceanic ridge volcanoes does not include sediments and water. Second, water movements carry heat away from the erupting magma, minimizing or preventing the production of steam. Third, the boiling point of water is several hundred degrees Celsius at the high pressures present at the depths of most oceanic ridge volcanoes.

Because they are not explosive, and they occur deep within the ocean, most oceanic ridge volcanic eruptions occur without being noticed. Nonetheless, they can be at least as large and as frequent as eruptions on land. For example, a sidescan **sonar** survey of part of the East Pacific Rise in the early 1990s revealed a huge lava field formed by an eruption believed to have occurred within the previous 20 years. The lava field covers an area of 220 km^2 and has an estimated average thickness of 70 m. The volume of lava is estimated to be about 15 km^3— enough to cover 1000 km of four-lane highway to a depth of 600 m, or to repave the entire U.S. interstate highway system 10 times. The lava field is the largest flow known to be generated during human history (much larger lava fields from prehistoric eruptions are known). Only a small fraction of the 60,000 km of oceanic ridge has been studied by sidescan sonar surveys or by **submersibles**. Hence, the huge young lava field on the East Pacific Rise may be dwarfed by undiscovered flows from other undersea eruptions.

In 2009, the ROV *Jason II* captured video of an eruption of the West Mata volcano, located 1100 m deep on the Tonga Ridge. Several more underwater volcano eruptions have now been captured on video and studied extensively by scientists (watch an eruption here). It appears these underwater volcanic eruptions are far more frequent and numerous than volcanic eruptions on land.

Rift Zones

Rift zones occur where a continent is being pulled or is breaking apart. If the rifting continues long enough, an ocean basin is created between two separate, smaller continents. A rift zone may be created when a continental crustal block remains in one place for a prolonged period. In such instances, heat beneath the continent is partially trapped by the insulating effect of the continental rocks. The mantle temperature may increase until the rock at the base of the lithosphere melts, at which time hot magma could rise and split the continental block. Rift zones may also be formed by variations in the convection processes operating within the mantle that cause the mantle upwelling to become located under the continent (**CC3**). As yet, we have no detailed understanding of the relative roles of mantle convection and the insulating effect of continental crust, or of how these processes interact to split continents and create new oceans. In contrast, the processes that occur once a rift zone has been initiated are much better understood.

New Oceans

A new ocean may form in a location where the temperature of the mantle below the continent becomes elevated (**Fig. 4-21**). The process may occur as follows: As the mantle temperature increases, its density decreases and it rises isostatically (**CC2**), causing the crust to be thrust upward to form a dome. The upward thrust stretches the crust, causing fractures that extend outward from the center of the dome. As such a fracture widens, the lithosphere beneath the bulge begins to melt. Eventually, blocks of crust break off and slip down into the rift valley formed by the fractures. This occurs unevenly along the length of the rift, with some sections of the rift valley being deeper than others.

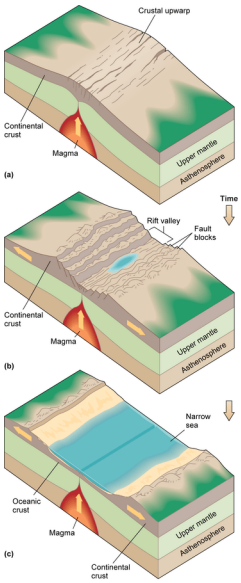


Figure 4-21. The history of new ocean formation at a rift zone. (a) Rising magma beneath a continent pushes the crust upward and creates cracks and fractures. (b) The crust is stretched and thinned. Collapses occur along a series of fault blocks, leading to the formation of a rift valley. Lakes may be formed in the rift valley. (c) The rift widens, allowing ocean water to enter and creating a narrow sea, such as the Red Sea. The rift continues to widen, and an oceanic ridge is formed down its center, eventually creating a new ocean.

Initially, the rift valley floor may be well above sea level. In wet climates, the deeper areas of the valley may fill with rainwater, forming long, narrow lakes. Volcanoes form in the rift valley and on its sides as magma upwells through cracks left by the fracturing and slipping blocks of crust. The sides of the rift move steadily apart, and magmatic rocks accumulate at the bottom of the valley. Because magmatic rocks, when they cool, are denser than the continental rocks being displaced to the sides of the rift, the rift valley floor sinks until it is eventually below sea level. At this point, the rift may fill with seawater and become an arm of the ocean. However, new volcanoes and landslide debris falling from the rift valley flanks may fill and temporarily increase the elevation of the rift valley floor. Thus, the connection with the ocean may be made and broken numerous times before the rift valley becomes a “permanent” **marginal sea**.

As the rift continues to widen, the original continent splits into two separate continents, separated by a widening ocean with an oceanic ridge at its center. The rift valley provides a route through which the excess heat built up below can be released. If the quantity of heat is limited, the rifting process may stop before an ocean is formed. However, if excess heat is supplied at a high rate, the spreading may continue, and a new ocean may be formed.

Of the several rift zones on the Earth today, the East African Rift is among the newest (**Fig. 3-4**). The African continent appears to have remained at its present location for more than 100 million years. It is elevated several hundred meters higher than most of the other continents and is believed to be directly above a mantle superplume. The East African Rift displays a rift valley, rift valley lakes, and rift volcanoes typical of newly formed continental divergent plate boundaries. The Red Sea is probably a more advanced rift zone, where a connection with the ocean, perhaps a permanent one, has already been made.

This page titled 4.5: Divergent Plate Boundaries is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.6: Transform Faults and Fracture Zones

Transform faults are formed where two plates slide past each other. Numerous earthquakes occur along such faults as the edges of the two plates periodically lock and then break loose and slide past each other. The best-known transform fault is the San Andreas Fault in California, where a long section of the Pacific Plate is sliding northward past the North American Plate (**Fig. 4-14**). Most transform faults are much shorter than the San Andreas Fault.

Motions of plates at transform faults do not create the dramatic trenches, volcanoes, and mountain ranges of other types of plate boundaries. However, the movement of the two adjacent plates creates complicated stresses where the two plates meet, produces faults along which earthquakes occur, and forms low hills or mountains. Mountain building may be enhanced at bends on the plate boundary. One such bend on the San Andreas Fault is at the Santa Cruz Mountains south of San Francisco and was the site of the October 1989 Loma Prieta earthquake.

Numerous short transform faults are formed at divergent and convergent plate boundaries. As two plates move apart or together, they also rotate in relation to each other because the plates are moving on a spherical Earth (**Fig. 4-22**). The rate of spreading and creation of new oceanic crust will vary along the plate boundary when such rotation occurs. To accommodate the varying rates of spreading or subduction between two rigid plates, new plate material must be added or destroyed at different rates in adjacent segments of the plate boundary. This is accomplished by the creation of transform faults between short sections of the plate boundary (**Fig. 4-23**). In the transform fault region, the edges of the two plates slide past each other, one side moving slightly faster in one direction than the other, which is moving in the opposite direction.

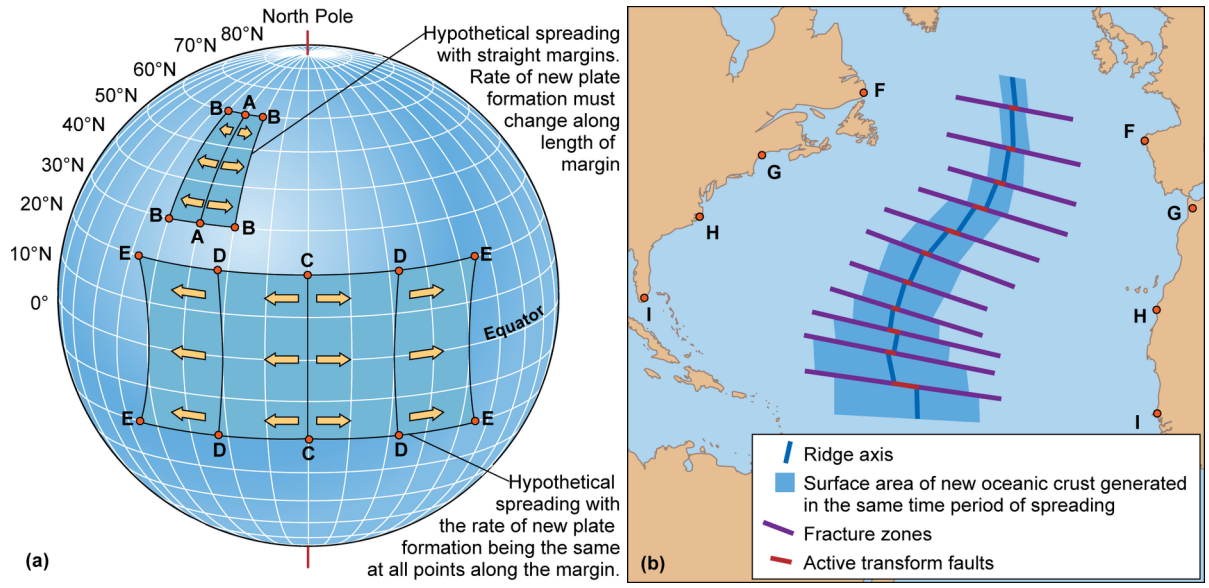


Figure 4-22. (a) Because lithospheric plates have rigid shapes and move on a spherical surface, they must rotate with respect to each other, and the rate of formation or destruction of crust at any plate boundary must vary along the plate boundary's length as shown by the spreading from A to B in this figure. The principle is similar to the movement of your eyelids as you open and close your eyes. The eye is exposed or covered more rapidly at the center than at the two corners. The rate of spreading cannot be equal along the plate boundary because the plates would need to bend and deform. This is shown by the vertical lines through the sequence C–D–E , which represent where the plate divergent boundary would be if spreading were uniform along the plate boundary. (b) Each of the points labeled with the same letter on either side of the Atlantic Ocean were joined at one time. Since the pairs are now separated by different distances, crust must have been formed at different rates along the oceanic ridge. Because the lithosphere is rigid and cannot stretch along its edge, variation in the rate of formation or destruction of crust is accommodated by the formation of transform faults between segments of the plate boundary.

You can demonstrate why this is necessary with a simple test using a book. With the book open on a flat surface, place one hand on the lower right corner of the right-hand page and press down so that it cannot move. With your other hand, lift all the pages (without the hard cover) at the top right corner slightly and grasp them between your finger and thumb. Hold these pages tightly together so that they cannot slide across each other, and try to lift and rotate these pages toward the lower left corner. Now loosen your grip on the top corner of the pages and try the same movement, allowing the pages to slide across each other. The resistance you felt and the distortion of the page edges is no longer there. Although the motions in this test are different from

the motions of lithospheric plates on a sphere, the principle is the same. The book pages, when allowed to, slide across each other—the equivalent of transform faults between the pages.

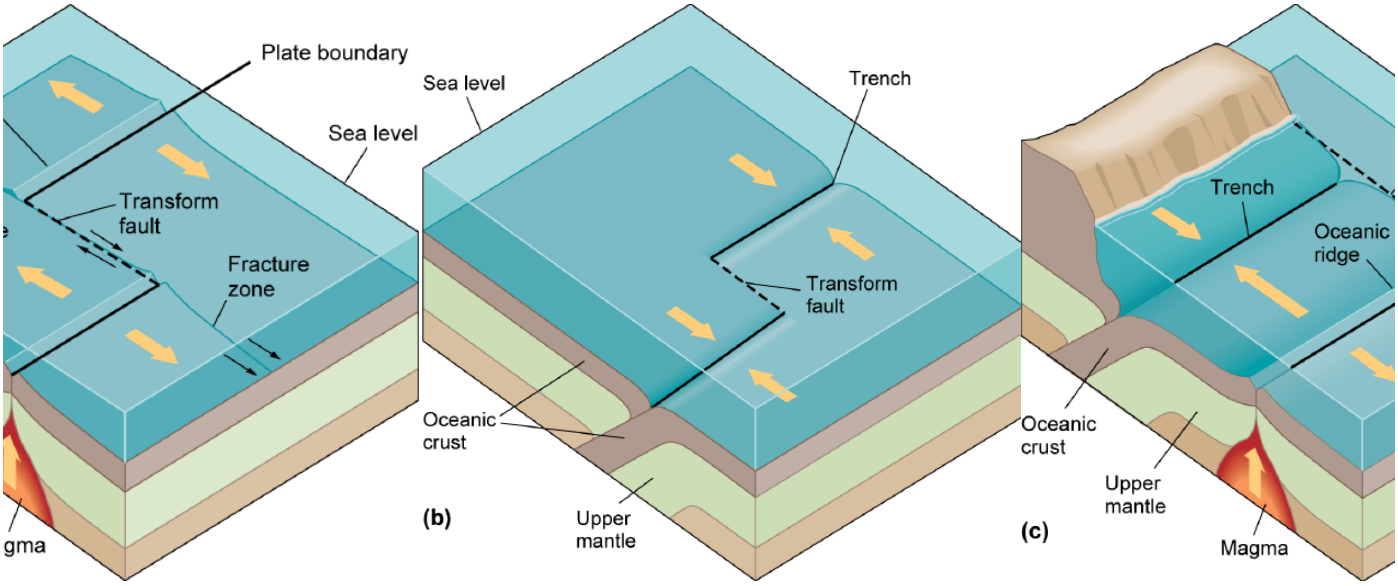


Figure 4-23. Adjacent plates move past each other at a transform fault. (a) Most transform faults connect two segments of an oceanic ridge. The plates slide past each other in the region between the two ridge crests. This is the transform fault. Beyond this region, the two ridge segments are locked together and form a fracture zone. A transform fault can also connect (b) two adjacent segments of a trench or two trenches, or (c) an oceanic ridge and a trench.

Transform faults not only accommodate the changing rates of plate creation or destruction but also affect seafloor topography. To examine how transform faults work and how they influence topography consider the oceanic ridge transform fault depicted in **Figure 4-23a**. The transform fault is the segment of the plate boundary where the line of the oceanic ridge is offset. Along this segment, the sections of crust newly formed on each of the two plates slide past each other. However, once the spreading motion has transported the new oceanic crust past the central rift valley of the adjacent segment, it is no longer adjacent to the other plate. Instead, it is adjacent to another piece of its own plate. Because the two sides of this joint are now parts of the same plate and moving in the same direction, the two sides do not slide past each other. These areas are called **fracture zones**. The topographic roughness formed at the transform fault remains after the fracture zone has moved away from the plate boundary. **Figure 3-4** shows numerous transform faults and associated fracture zones that cut perpendicularly through the oceanic ridges. At each transform fault, the line of the oceanic ridge crest is offset.

Some transform faults connect divergent and convergent plate boundaries, and others occur in subduction zones. The seafloor at subduction zones is covered by deep sediments that flow into any depressions and blanket the topography. Therefore, transform faults in subduction zones are less well defined by topography than their oceanic ridge counterparts are.

This page titled 4.6: Transform Faults and Fracture Zones is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.7: Hot Spots

Scattered throughout the world are locations where heat from the mantle flows outward through the crust at a much higher rate than in the surrounding crust. Some of these hot areas are within plates, and some are at plate boundaries. Most of them are beneath oceanic crust, but others are under continental crust (**Fig. 4-11**). Most hot locations occupy only a very limited area at the Earth's surface and are, therefore, called hot spots.

Many, but not all, of Earth's hot spots are known to be located where convection plumes upwell from the deep mantle (**Fig. 4-6, CC3**). Some of these may be plumes that extend all the way to the boundary between the core and mantle. Others may originate from a superplume in the lower mantle. Yet other hot spots, such as that at Yellowstone Park, seem to be linked to shallow convection processes in the upper mantle only. Finally, some locations where the hot area under the lithosphere is more extensive may not themselves be hot spots (although hot spots may occur within such regions). Instead, they may be locations where continents have remained for prolonged periods of time. At such locations, the continents may act as a blanket to trap the heat flowing upward through the upper mantle, and the heat may melt and thin the crust. The East African Rift may be an example. However, a superplume is believed to be located beneath the region in which the East African Rift occurs, and this rift may be the result of several interacting processes.

Most oceanic hot spots are characterized by active volcanoes that rise through the ocean floor to form islands. Examples of islands located at hot spots are Iceland and the island of Hawaii. Both have active volcanoes that generally erupt relatively quietly, without explosions, and bring copious amounts of magma to the surface (**Fig. 4-24**). The magma solidifies and steadily builds the island.

Figure 4-24. The 1983 eruption of Pu'u'ō'ō, the active vent of Kilauea Volcano in Hawaii. Kilauea is a hot-spot volcano that ejects small amounts of ash and large amounts of lava in a relatively smooth, non-explosive manner. This eruption of Kilauea continued for 35 years, ending in 2018.

The quantity of lava produced by hot-spot volcanoes is so large that such volcanoes are the tallest topographic features on Earth. For example, the island of Hawaii rises about 5500 m from the seafloor to the ocean surface and then rises another 4205 m above the ocean surface to its highest mountain peak. Its total elevation, about 9700 m, is gained over a horizontal distance of less than 200 km and far exceeds the 8848-m elevation of Mount Everest. The enormous weight of Hawaii and the adjacent islands has depressed the Pacific Plate through **isostatic leveling** (**CC2**). As a result, the seafloor around the islands has a broad moat-like depression some 500 m deeper than the surrounding seafloor. The downward deformation of the Pacific Plate extends about 300 km from Hawaii. Surrounding the "moat" region is another broad area where the seafloor is deformed upward as the crust is compressed outward by the island rising through its center.

Until recently, it was thought that hot spots remain fixed in place with respect to the Earth's rotational axis for tens or hundreds of millions of years as the lithospheric plates move over them. However, it is now known that at least some hot spots do migrate

independently within the mantle, so two independent processes change the location of individual hot spots, movement of the hot spot plumes and movement of the lithospheric plates.

As a lithospheric plate moves over (or in relation to) an oceanic hot spot, the most recently formed volcanic island moves away from the hot spot. With the migration of the island from the hot spot, another segment of plate is brought over the hot spot, new volcanic vents open in it, and eventually another island may be formed. Continued plate movement takes this island away from the hot spot, and yet another island is formed, and so on. The results of that process can be seen in the seafloor topography as a trail of islands and **seamounts** (undersea cone-shaped mountains). In the Pacific Ocean, the chains of islands and seamounts that align northwest of Macdonald Seamount, Easter Island, and Hawaii are all remnants of their respective hot spots (**Fig. 4-11**).

The island and seamount trails provide a history of the rate and direction of movement of the plate relative to the hot spot. Of the Hawaiian Islands, the island of Hawaii is the youngest and has several active volcanoes, including Kilauea (**Fig. 4-24**) and Mauna Loa. About 20 km southeast of Hawaii, a 3.5 km tall seamount called Loihi has been built on the sea floor and continues to grow by volcanic action (**Fig. 4-25**). Its current peak is about 970 m below sea level, but if volcanic activity continues at its present rate, Loihi may become the next Hawaiian island 10,000 to 40,000 years from now.

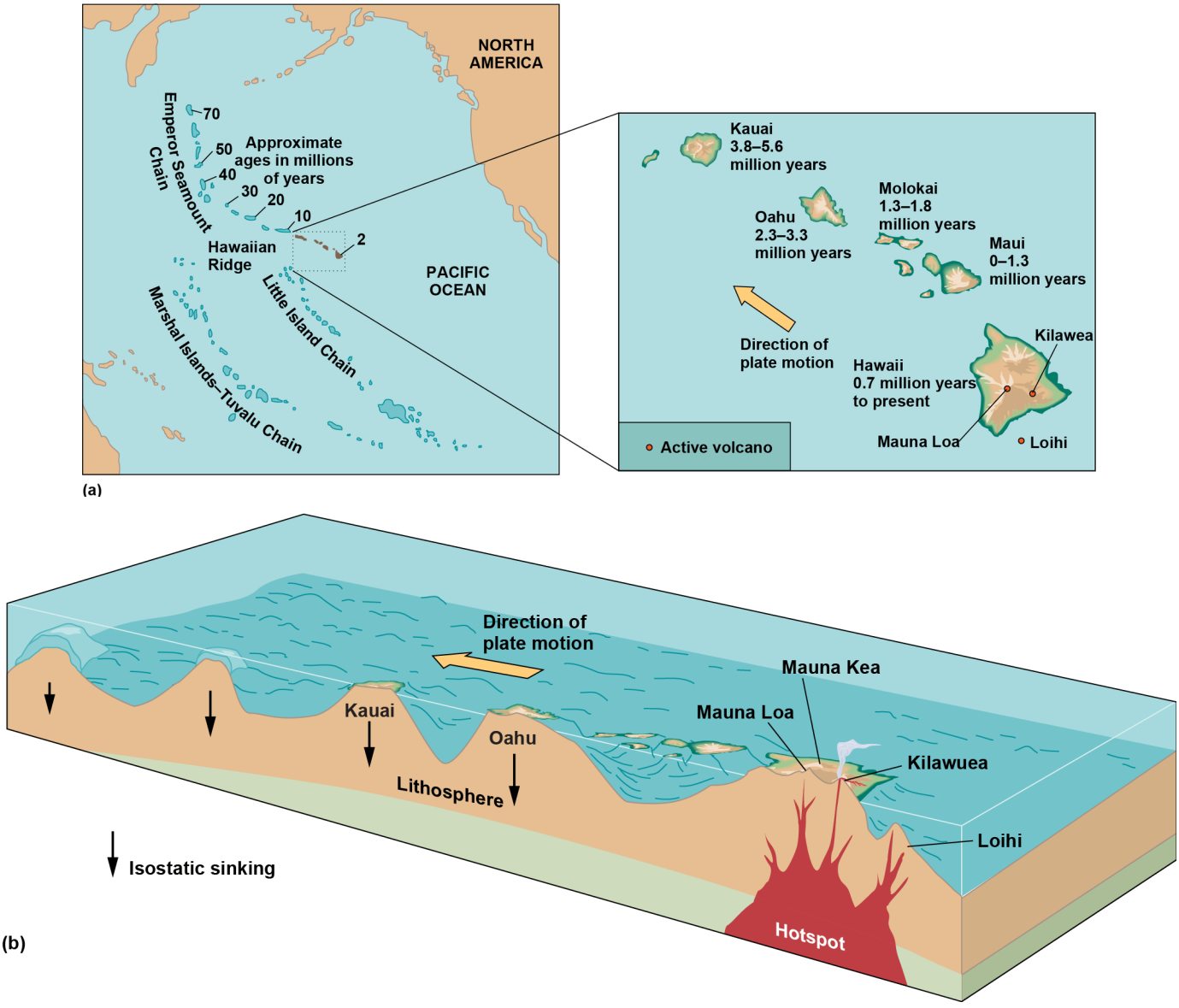


Figure 4-25. (a) The Hawaiian Island–Emperor Seamount chain. (b) The islands and seamounts are formed at the hot spot. As the plate moves relative to the hot spot, new islands are formed, and the older islands move off the hot spot, cool, sink isostatically, and are eroded.

Radioisotope dating (**CC7**) of the volcanic rocks in the Hawaiian Island–Emperor Seamount chain shows each island or seamount

to be progressively older with distance northwest from Hawaii (**Fig. 4-25a**). For example, Oahu was formed about 2 to 3 million years ago and only has inactive, although not yet necessarily extinct, volcanoes. As hot-spot volcanic islands migrate away from the hot spot, their volcanoes become inactive, the islands cool and sink isostatically, and they are subjected to erosion (**Fig. 4-25b**).

Some hot spots lie on divergent plate boundaries where two plates are moving apart. A volcanic island formed at such a location is steadily broken apart as the plates diverge. Iceland is a good example. Each side of the island, with its cooling volcanoes, migrates away from the hot spot on its respective plate. Evidence of this process can also be seen in the ocean floor topography. For example, the Icelandic Ridge stretching between Greenland and Europe consists of sediment-buried remnants of volcanoes that occupied the Icelandic hot spot when the Atlantic Ocean was narrower. Similarly, the Walvis Ridge and Rio Grande Rise in the South Atlantic Ocean are remnants of the Tristan da Cunha hot-spot volcanoes (**Fig. 4-11**).

A Lesson about Science

The rate of Pacific Plate movement appears to have varied relatively little during the past 50 million years. However, a distinct change in the direction of the Hawaiian Island–Emperor Seamount chain occurs at islands formed about 50 million years ago (**Fig. 4-25**). Seemingly identical changes of direction are also seen in the Easter Island and Macdonald Seamount chains.

Studies of the change in direction of the Hawaiian Island–Emperor Seamount chain provide a lesson for this author and the ocean and geological science community. This story provides you, the readers of this text, with a perfect example of why critical thinking skills are so important. It also illustrates that, while the material in this book represents the most recent scientific consensus regarding what is known about our ocean world, that consensus is always evolving.

As recently as 1998 (when the first edition of this text appeared), the consensus view of the scientific community was that hot spots remained fixed in place for tens or hundreds of millions of years as the lithospheric plates moved over them. At that time, the scientific consensus was also that the distinct change in the direction of the Hawaiian Island–Emperor Seamount chain indicated an abrupt (in geological time) change in the Pacific Plate's direction of motion. Experts in the field agreed that the available data fit this conclusion. For example, the ages of the islands in the chain had been measured and did increase with distance from the hot spot. Other hot-spot trails were also found on the Pacific Plate, and they mirrored both the direction and the change in direction of the Hawaiian Island chain.

Fortunately, some scientists continued to think critically, even when the available data seemed to fit these explanations very well. These scientists were uncomfortable that nobody could explain how or why the direction of the Pacific Plate movement changed so abruptly. Thus, when the opportunity presented itself, they examined cores drilled into three seamounts of the Emperor chain that were formed before the "change in direction of motion." They analyzed these cores for many different parameters, but the key data were obtained by examination of the **paleomagnetic** signatures of the magmatic rocks that form these seamounts. These data allowed the researchers to determine the rocks' paleolatitudes (the **latitudes** at which these rocks had formed by cooling and solidification of liquid magma).

To explain how paleolatitude was measured, we have to look carefully at Earth's magnetic field characteristics. We normally hold a compass horizontally so that its needle can rotate to align north and south. However, if we turn the compass on its side and orient it toward the pole, we find that the Earth's magnetic field is not parallel to the ground (except at the magnetic equator). In the Northern Hemisphere, the compass needle points at an angle toward the ground (the **dip angle**) in the direction of the north magnetic pole (**Fig. 4-26**). The dip angle increases as we move toward the magnetic pole. At the magnetic pole, the needle points at a 90° angle directly into the ground. At the magnetic equator, the dip angle is zero and the needle is parallel to the ground. Because the magnetic and geographic poles are always reasonably close together, the dip angle of the compass needle is a measure of latitude. Larger angles correspond to higher latitudes.

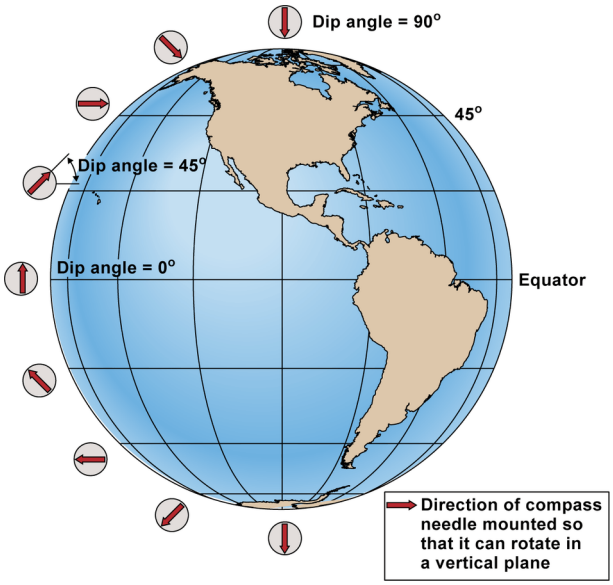


Figure 4-26. The principles of paleomagnetic dip angle. The dip angle is the angle between the rock's direction of magnetization and the Earth's surface (if the rock has not been tilted since being formed). Because the dip angle changes with latitude, the original dip angle of a rock's magnetization reveals the latitude at which it was formed.

Magnetic materials in magma orient both horizontally and vertically to align with the Earth's magnetic field when the magma solidifies. Therefore, rocks carry paleomagnetic information about both their horizontal orientation with respect to the pole at the time they were formed, and the latitude at which they were formed, the paleolatitude.

What the measured paleolatitudes of the three Emperor chain seamounts revealed was that these islands had not been formed at the same latitude that Hawaii now occupies. Instead, each older seamount had been formed progressively farther north. How could this be if hot spots are fixed in place? Of course, it cannot. Most scientists had accepted the hypothesis that hot spots remain fixed in place relative to the Earth's axis of rotation. That hypothesis now appears to be wrong. This hot spot must have moved south during a period about 50 to 80 million years ago, when the Emperor seamounts were formed. The lesson here is that, when a new hypothesis is proposed, there are almost always a few pieces of data missing, or data that do not exactly fit the "facts" of the hypothesis. These minor inconsistencies are the clues to how a hypothesis might be wrong.

Where does this new evidence leave our understanding of hot spots and how they fit into the jigsaw puzzle of plate tectonics? First, it raises many more questions. For example, why are there almost identical changes in direction in other Pacific hot-spot trails? Did all the Pacific hot spots move in the same direction at the same time? Scientists must answer these questions, and many more, if we are to better understand plate tectonic processes.

This page titled 4.7: Hot Spots is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.8: Plate Interiors

Oceanic crust is formed and destroyed by the tectonic processes that occur at divergent and convergent plate boundaries and hot spots. Its topography is modified by various processes over the tens or hundreds of millions of years between its creation and destruction. The edges of the continents, which are formed at divergent plate boundaries, are modified by a variety of processes between the time they are formed and the time they enter convergent plate boundaries.

Oceanic Crust

As it moves away from an oceanic ridge, the lithosphere cools, becomes denser, and sinks steadily deeper into the asthenosphere. Therefore, the seafloor becomes lower with increasing distance from the oceanic ridge. As the cooling crust sinks, it is progressively buried by a continuous slow "rain" of solid particles through the water column that accumulate as sediments on the seafloor (**Figs. 4-20, 4-27**). Because sediments tend to accumulate faster in topographic lows, the original topography of the rugged oceanic crust that is formed at oceanic ridges is progressively buried and smoothed as the crust moves away from the oceanic ridge. The effect is very similar to that of a snowfall. If left undisturbed, a few centimeters of snow obscures features, such as street curbs and potholes, and softens larger features by mounding around them. As more snow accumulates, larger features are buried, and even cars in a parking lot may be difficult to find. Similarly, the lower topography of the oceanic crust is completely obscured after it has traveled a few hundred kilometers away from the ridge, though the oceanic crust takes millions of years to move such distances. The higher topography of the oceanic crust survives the sedimentation as rounded hills or mountains rising above the surrounding flatter areas.

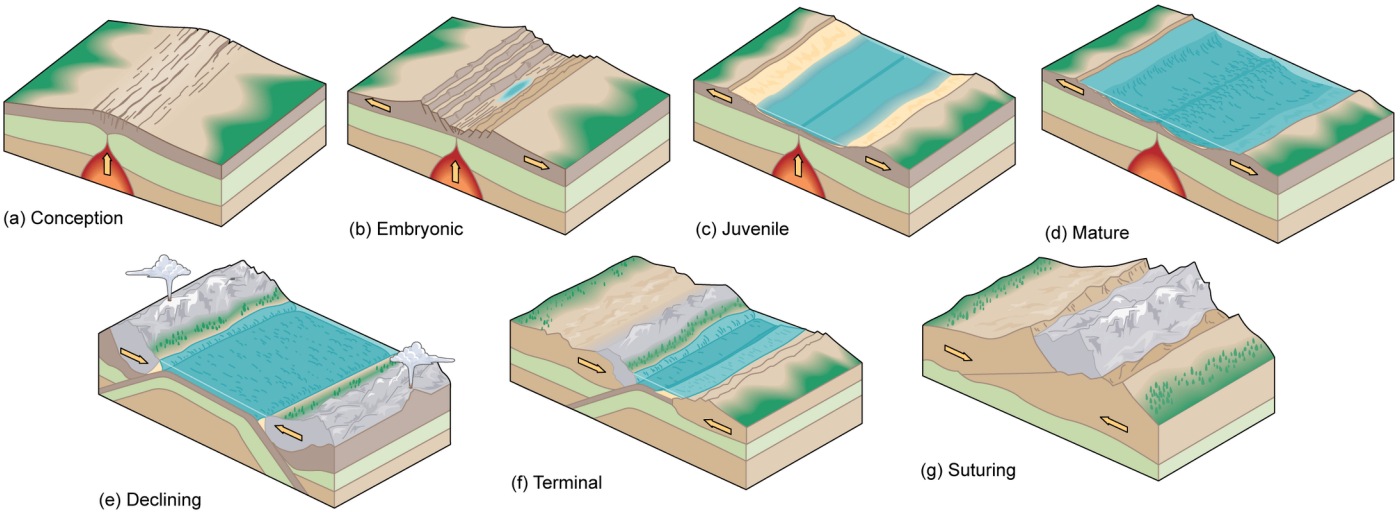


Figure 4-27. The history of an ocean from its creation at an oceanic ridge to its disappearance at a continental collision.

The largest topographic features of oceanic hot spots and convergent boundaries where both plates have oceanic crust (e.g. magmatic arcs) are commonly cone-shaped volcanoes. Their conical form is preserved and even enhanced by sediment accumulation. Therefore, much of the deep-sea floor is characterized by generally cone-shaped abyssal hills and mountains (called seamounts), even in regions far from the oceanic ridges or hot spots (**Fig. 3-4**).

Some seamounts have flat tops and are called "tablemounts" or **guyots**. Volcanic mountain cones that have sufficient elevation when first formed at the oceanic ridge or at hot spots emerge above sea level as islands. The island tops are eroded by wind, water, and waves much faster than the volcano can cool and sink isostatically. Thus, before such volcanoes sink isostatically below the surface, their tops are totally eroded away. The eroded flat tops are preserved once the volcanoes are completely submerged because erosion is extremely slow under the ocean surface, far away from winds and surface waves. The Hawaiian Island–Emperor Seamount chain exemplifies the various stages of this process (**Fig. 4-25**).

Some seamounts form the submerged base of nearly circular coral reefs called atolls (**Fig. 4-28**). Reef-building corals grow only in shallow water (less than several hundred meters), and most species inhabit the warm tropical oceans (**Chap. 15**). In such regions, a coral reef may become established around islands that are formed when the tops of oceanic ridge or hot-spot volcanoes extend

to or above sea level (Fig. 4-29). The coral reef continues to build upward from the flanks of the volcano as the volcano sinks isostatically. If the upward growth rate of the reef is fast enough to match the sinking rate of the volcano, the top of the live coral remains in sufficiently shallow water to continue growing.

Figure 4-28. Satellite image of a coral atoll with its shallow interior lagoon (Hull Island, Kiribati).

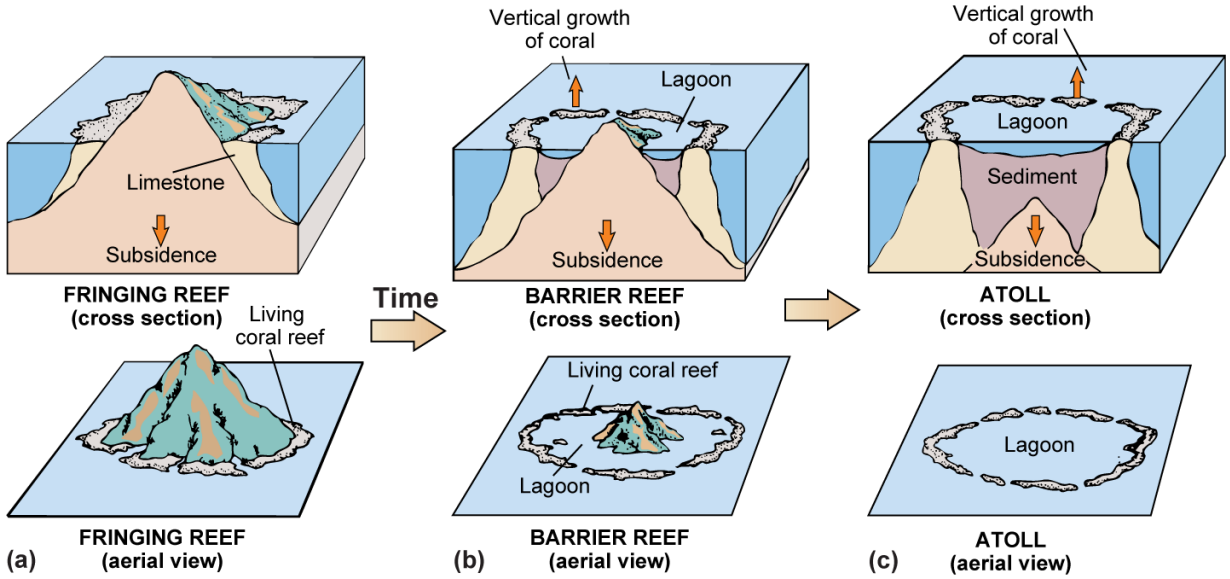


Figure 4-29. Coral reefs form in shallow water in the tropics and subtropics. Reefs are well developed around volcanic islands created at hot spots because there is little runoff of freshwater and sediment, both of which inhibit healthy growth of coral reefs. Coral reefs evolve as volcanic islands are formed and then sink isostatically as they move off the hot spot and cool. (a) Fringing reefs are formed around the perimeter of rising or static volcanic islands. (b) When the island sinks, a barrier reef is formed as the fringing reef grows upward. (c) Eventually, the island sinks completely beneath the surface, leaving an atoll where the barrier reef continues to grow upward.

The coral first forms a **fringing reef** adjacent to the island (Fig. 4-29a). As the volcano continues to sink, this reef continues to grow upward and becomes a barrier reef separated from the island by a lagoon (Figs. 4-29b, 4-30). Eventually, the volcano completely sinks and leaves only an atoll (Figs. 4-28, 4-29c). On the historic Beagle voyage of 1831 to 1836, Charles Darwin visited Keeling Atoll and several reef-fringed islands of the South Pacific. Even though Darwin had no knowledge of plate tectonics, he proposed an explanation for the formation of atolls that is very similar to our understanding of that process today.

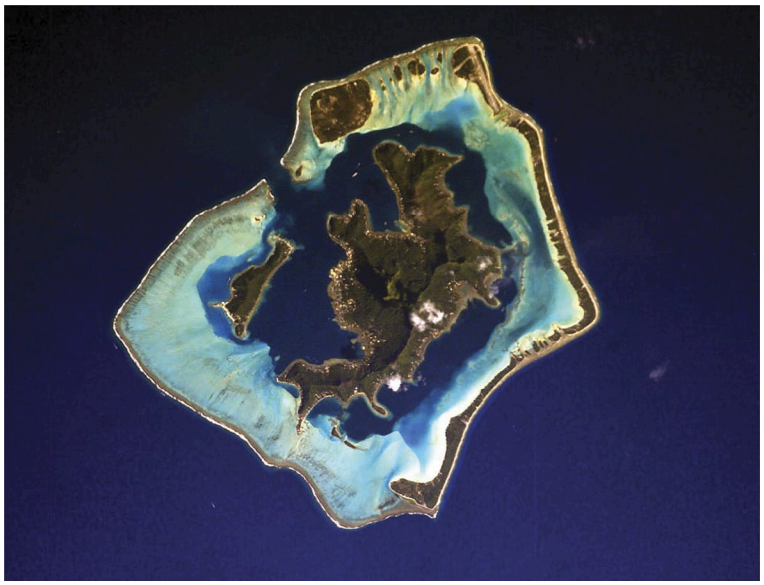


Figure 4-30. This volcanic island, Bora Bora in the Leeward group of the Society Islands of French Polynesia, is surrounded by a lagoon and a well-developed barrier reef.

Continental Edges

We have seen how the processes at convergent, divergent, and transform fault plate boundaries form and shape the edges of continents. However, fewer than half of the coasts (or margins) of today's continents lie at plate edges. Most are located in the middle of lithospheric plates. Most of these coastlines were formed initially at the rift zones created as Pangaea broke apart. Because there are few earthquakes or volcanoes at such **continental margins**, they are called **passive margins**. The Atlantic coasts of North America and western Europe are examples.

Passive Margins

Development of a passive margin begins as the edge of a new continent is formed at a mature continental rift valley (**Fig. 4-31**). The new edge is isostatically elevated because of heating in the rift zone. Consequently, rivers drain away from the edge toward the interior of the continent. Shallow seas form in the rift as blocks of the continental crust edge slide down into the rift zone. Because few rivers carry sediment into the seas, the **turbidity** of the water in these seas is low. Consequently, the light needed for **photosynthesis** (**CC14**) penetrates deep into the clear waters, and **primary productivity** (growth of marine organisms that create organic matter from carbon dioxide and an energy source; **Chap. 12**) is high. High productivity leads to large quantities of organic matter that may accumulate in the sediment of the new marginal seas. Because the seas are shallow and the rift valley is narrow, they may be periodically isolated from exchange with the large ocean basins. Under these conditions, thick salt deposits may be formed as seawater evaporates and its dissolved salts precipitate (**Chap. 6**).

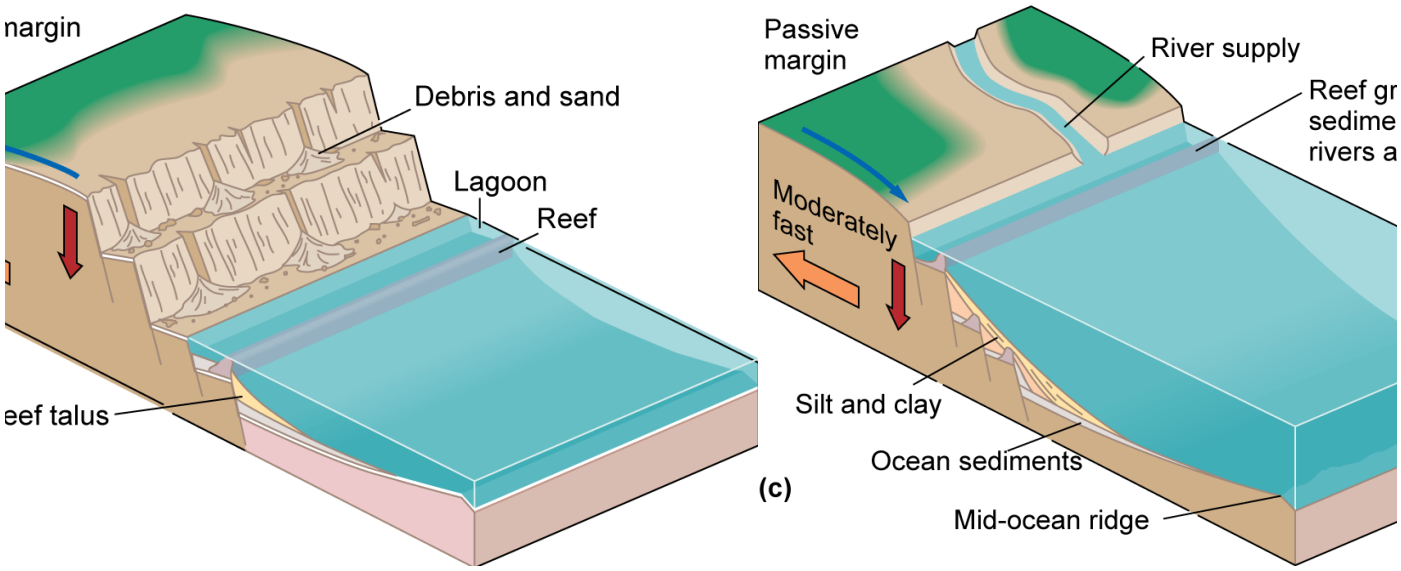


Figure 4-31. A passive margin develops after the initial stages of the creation of a new ocean. (a) A rift forms at a continental margin. Layers of volcanic ash, desert sands, and evaporites form as lakes develop and dry up in the rift valley. (b) As the new ocean forms, reefs and high-productivity lagoons form along the ocean edge because rivers flow inland and turbidity is low in such young oceans. (c) As the margin moves away from the new oceanic ridge, it cools and subsides. Rivers start to flow toward the new ocean, depositing lithogenous sediments in the previously productive lagoons and reefs. Productivity in the coastal waters, especially coral reef growth, is suppressed by high turbidity caused by the river inputs of suspended sediment. (d) Sediment layers build up on the passive margin to form a broad sediment-covered continental shelf. The buried lagoon and reef sediments often form oil reservoirs.

As the passive margin moves away from the divergent plate boundary, which by that time has developed an oceanic ridge, both continental crust and oceanic crust cool and subside isostatically. Eventually, the edge of the continent subsides sufficiently that the direction of the slope of the landmass is reversed. Then the rivers that flowed away from the margin during its early history instead flow into the ocean at the margin. This **runoff** brings large quantities of sediment from the land, which increases turbidity and siltation in the coastal ocean, thus reducing light penetration and primary productivity. Both the marginal seas and their deposits of organically rich sediment eventually are buried deeply as they subside farther below sea level. When organic sediments are sufficiently deep to be heated and subjected to high pressure, the organic matter may be converted into **hydrocarbon** compounds, such as those found in oil and gas. If the overlying rocks are permeable, the oil and gas migrate toward the surface. Where the overlying rocks are not permeable, they trap the oil and gas, forming reservoirs. Reservoirs formed at passive margins provide most of the world's oil and gas.

The characteristic feature of a passive margin is a **coastal plain**, which may have ancient, highly eroded hills or mountain chains inland from it. In addition, the coastal region is generally characterized by **salt marshes** and many shallow estuaries. Offshore, the continental shelf is wide and covered by thick layers of sediment. These features have been modified in many areas by changes both in sea level and in the distribution of glaciers that, in turn, were caused by climate changes.

The Fate of Passive Margins

The depth and width of the continental shelf at passive margins can be influenced by isostatic changes caused by the heating of relatively immobile continental blocks (**CC2**). For example, the continental shelf of much of the west (Atlantic) coast of Africa is narrower than those of many other passive margins. The reason is that the entire continent has been lifted by isostatic leveling in response to the heat accumulation that is causing the East African Rift to form. Africa's Atlantic coast may eventually become a new subduction zone if the Atlantic Ocean crust near Africa cools sufficiently and if the East African Rift continues to expand. Another possibility is that the rifting in East Africa will simply stop. Many factors will influence the outcome, including how much heat is available in the mantle to drive the East African rifting process and to continue spreading at the Mid-Atlantic Ridge, how cold and dense the Atlantic Ocean crust is near West Africa, and the driving forces on other plates.

Whatever happens in Africa, the probable ultimate fate of some passive margins is that they will become subduction zones. This will be the outcome if the oceanic crust at the passive margin cools sufficiently to subduct into the mantle, causing the plate movements to change and the ocean to begin to close. Eventually, the ocean could be destroyed as two continents meet at a collision margin.

This page titled 4.8: Plate Interiors is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.9: Sea-Level Change and Climate

Although tectonic processes and sedimentation are the principal agents that shape seafloor topography, changes in climate also affect certain features, particularly the continental shelves and coasts. Substantial climate changes have been observed during recorded human history. For example, much of North Africa and the Middle East, which is now a region of desert and near-desert, was a fertile region with ample rainfall during Greek and Roman times. However, such short-term changes are small in comparison to changes that occur during the hundreds of millions of years of a tectonic spreading cycle.

Climate Cycles

Over the past 1000 years, the Earth's global average surface air temperature has varied by about 1.5°C. During the past 10,000 years, it has varied within a range of about 2°C to 3°C (Fig. 4-32). However, in the immediately preceding 2 to 3 million-year period, the Earth's temperatures were as much as 10°C below what they are today. These colder periods are called **ice ages** or glacial periods. During the most recent ice age, the Earth's climate alternated between periods of glacial maxima (when temperatures were about 10°C below those of today) and interglacial periods (when temperatures were close to those of today). During glacial maxima, the polar ice sheets extended to much lower latitudes than they do now. Several earlier major ice ages each lasted about 50 million years. Longer periods with warmer climates have occurred between ice ages. The most recent ice age may not yet be over, and the relatively warm climate of the past 10,000 years may simply represent a geologically short interglacial period.

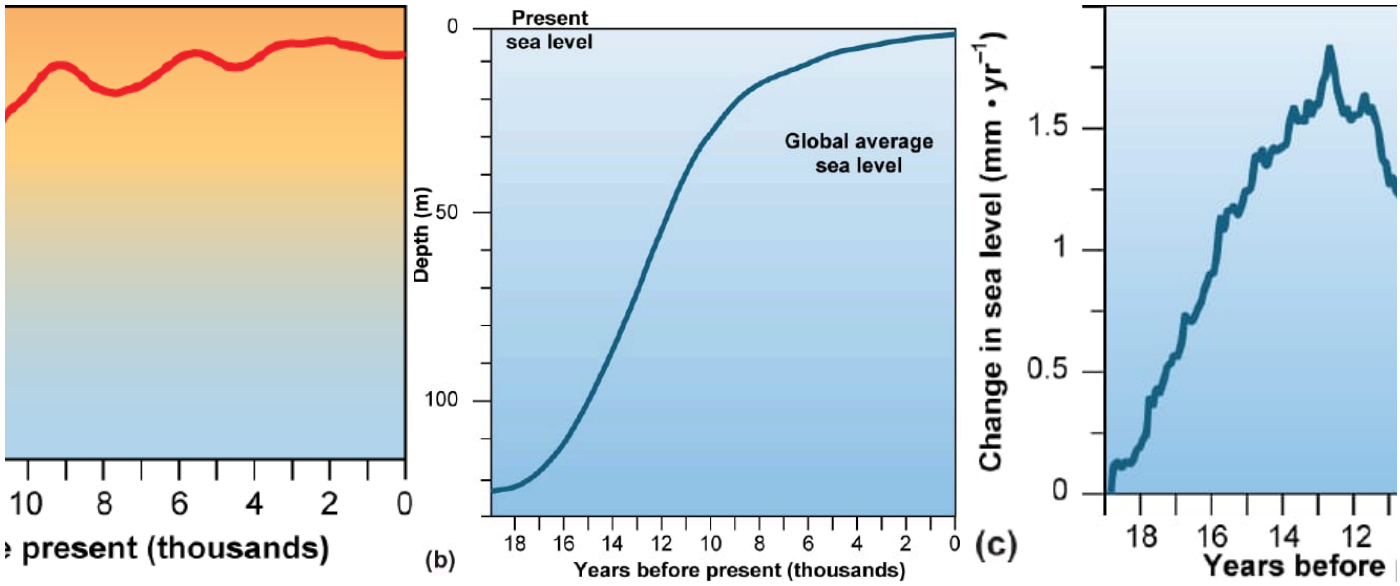


Figure 4-32. The history of climate and sea-level changes during the past 18,000 years. (a) The global average surface air temperature was about 7–8°C lower than today until about 18,000 years ago and rose close to its present value by about 8000 years ago. (b) In response to the global temperature increase, sea level rose rapidly (about 120 m in total) as global climate warmed from about 18,000 to about 8000 years ago. More recently, sea level continued to rise slowly as global temperatures remained relatively stable. (c) The rate of sea level change increased to nearly 2 mm•yr^{-1} about 12,000 years ago and has decreased since then.

Climate varies on timescales ranging from a few years to tens or hundreds of millions of years. Variations over the past 18,000 years are shown in Figure 4-32. These shorter timescale variations are superimposed on longer-term variations. Understanding the causes and consequences of the historical variations is important. Such knowledge may provide the key to understanding the effects and consequences of the extremely rapid (on geological timescales) global warming that has been predicted to occur and that appears to already have begun as a result of human enhancement of the atmospheric **greenhouse effect**.

Eustatic Sea-Level Change

When the Earth's climate cools for a long time, ocean water cools and contracts and polar ice sheets expand as water is

transferred from oceans to land in the form of glaciers and snow. Both of these processes cause the sea level to drop. Conversely, warm climate periods tend to heat and expand ocean water, and melting continental ice returns water to the oceans, raising the sea level. Such changes of sea level, called eustatic changes, occur synchronously throughout the world, although not all locations experience exactly the same amount of rise or fall. In contrast, sea-level changes caused by isostatic movements of an individual continent affect only that continent (**CC2**).

During the initial breakup of Pangaea, the climate was relatively warm, and it remained so until about 10 to 15 million years ago. During the warmest part of this period, about 75 million years ago, sea level was considerably higher than it is today, and as much as 40% of the Earth's present land area was below sea level. For example, a shallow sea extended from the Gulf of Mexico far north into Canada and covered what is now the land between the Rocky Mountains and the Appalachian Mountains. In contrast, at the peak of the most recent glacial period, about 20,000 years ago, sea level was at least 100 m below its present level, exposing much more land area. Sea level has risen and fallen by various amounts many times as Pangaea has broken apart in the present spreading cycle, and coastlines have migrated back and forth accordingly.

Sea-Level Change and Continental Margin Topography

Erosion by rivers carves out valleys, and the rivers transport the eroded sediment downstream, where it is deposited in lower-lying areas or in the coastal oceans. Erosion by waves and winds at coastlines (**Chap. 11**) also tends to reduce topography and deposit the eroded sediment in the shallow waters of the continental shelves. In contrast, at water depths of more than a few meters, erosional forces in the oceans are generally reduced and sediment accumulation is dramatically reduced, except in proximity to rivers that transport massive sediment loads to the ocean (**Chap. 6**).

As a result of sea-level oscillations, the area between the edge of the continental shelf and an elevation several tens of meters above the present sea level has been subjected to alternating cycles of wind, river, and wave erosion at some times and sediment deposition at others. These processes have substantially modified the topography along the coasts. The effect is most apparent at passive margins. On most such margins, there is evidence of sea incursion and erosion throughout the area between the continental shelf edge and areas far inland from the current coastline. This evidence includes buried deposits that contain freshwater and shallow-water marine organisms and flat or low-relief topography.

The continental shelf, which has been progressively covered by the rising sea during the past 19,000 years, is cut across by numerous **shelf valleys**. Most of the valleys were carved out by rivers during the last ice age, when sea level was lower. Many **submarine canyons** are extensions of shelf valleys (**Fig. 4-33**).

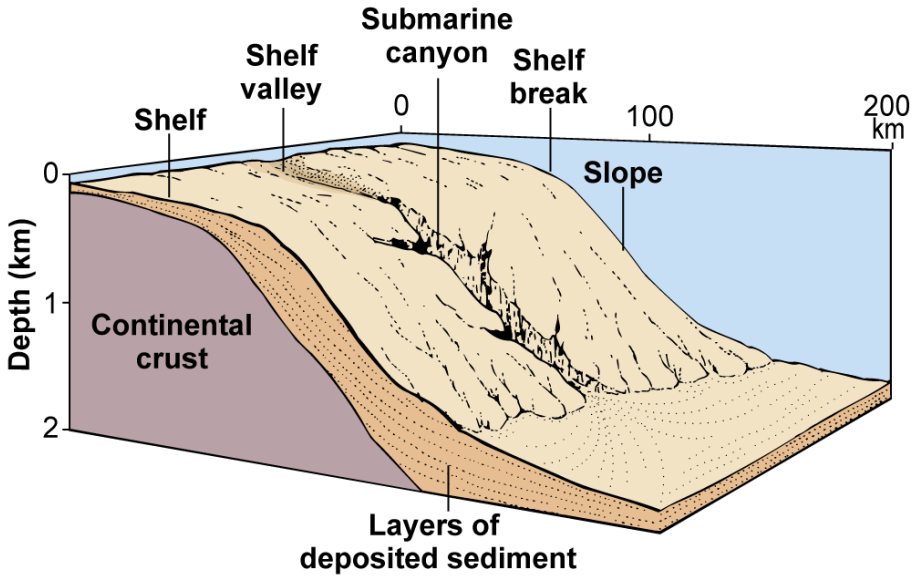


Figure 4-33. The continental shelf is often cut across by shelf valleys, some of which connect to steep-sided submarine canyons that cut across the continental slope.

Isostatic Sea-Level Change

In addition to eustatic processes, isostatic processes induced by changing climate can affect coastal topography. During an ice age, massive ice sheets accumulate over many parts of the continental crust. The weight of the ice forces the continental crust to sink lower into the asthenosphere (**CC2**). This process also depresses (lowers the level of) the continental shelves. For example, Antarctica's continental shelf is currently depressed about 400 m lower than most other shelves by the weight of its ice sheets. At the end of an ice age, when the ice melts, a depressed section of a plate slowly rises until it again reaches isostatic equilibrium. However, isostatic leveling is much slower than eustatic sea-level change. Many areas of continental crust, including the coasts of Scandinavia and the northeastern U.S. and Canada, are still rising in response to the melting of glaciers that occurred several thousand years ago.

Glaciers shape topography in other ways as well. They often cut narrow, steep-sided valleys. Many valleys left after glaciers melted have been submerged by rising sea level to become deep, narrow arms of the sea known as **fjords** (Chap. 13).

Sea Level and the Greenhouse Effect

Although oscillations of sea level are normal occurrences in geological time, human civilization has emerged during a long period of relatively stable sea level (Fig. 4-34).

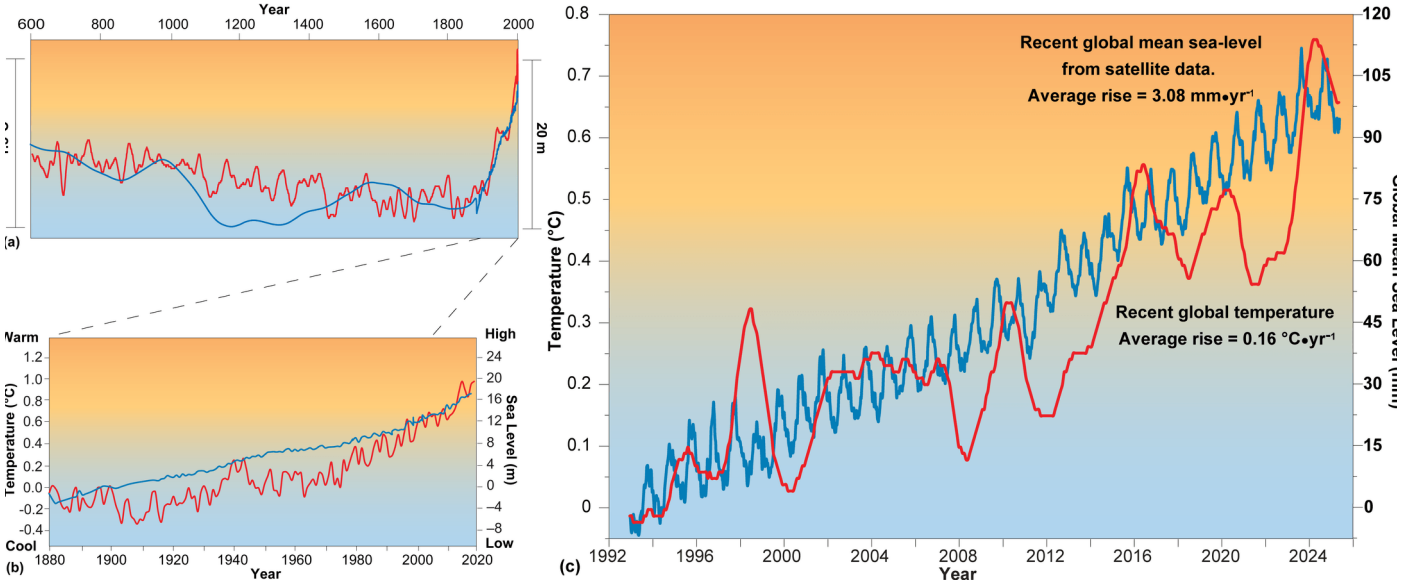


Figure 4-34. Global average surface air temperature and sea level variations within the past 1400 years. (a) Since 600 CE, global average surface air temperature has varied by about 1.5°C, but is currently rising at about 1°C per century. Sea level has varied by about 20 m and appears to follow the trend of temperature changes. (b) Since about 1920, air temperature has started to rise more quickly, as has sea level. However, there is much short-term variability, and the apparent upward trend may not be sustained. (c) Since 1993, instrumental records of temperature show large variations in temperature, but a general increasing trend of 0.16°C•yr^{-1}. During the same period, highly-accurate satellite measurements show that sea level rise may be accelerating and is now estimated to be about 3 mm•yr^{-1}. It is not known how far temperature and sea level will rise in the future. It is known that the temperature increase and sea level rise are due to a combination of natural change and the enhanced greenhouse effect but it is not clear what the relative contributions of these two factors has been. This raises the question as to whether the rate of sea level rise could accelerate substantially if the current rising temperature trend is sustained.

If the predicted greenhouse warming of the planet by as much as 2°C during the first half of this century does indeed occur, the higher average global temperature will cause the ocean water to continue to warm and expand, and the current polar ice sheets to continue to melt. As a result, sea level could rise by several meters or more. Warming and melting of ice sheets are slow processes that will continue possibly for centuries even if we were to completely stop adding to the concentration of carbon dioxide in the atmosphere. However, the most recent predictions based on the best available scientific information are that sea level will rise by about 1 m, and likely more, by the end of this century. A sea level rise of only a few tens of centimeters would inundate vast areas, including major coastal cities. Therefore, the history of coastal modification during periods of sea-level rise is of more than academic interest.

This page titled 4.9: Sea-Level Change and Climate is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.10: Present-Day Oceans

Having learned about processes that create, shape, and destroy ocean floor topography, we can look with greater understanding at the present-day oceans, which are all connected. The Atlantic Ocean connects with both the Indian and Pacific Oceans near Antarctica. The Arctic Ocean connects with the Pacific Ocean only where shallow water covers the continental shelf in the Bering Strait between Alaska and Siberia. In contrast, the Arctic Ocean is connected with the Atlantic Ocean by the deeper and much wider passage through the Nansen Fracture Zone between Greenland and Spitsbergen, Norway (**Fig. 4-18**). The somewhat deeper connection between the Arctic and Atlantic Oceans allows somewhat restricted water transfer between the two basins (**Chap. 8**).

Pacific Ocean

The Pacific Ocean is the world's largest and, on average, deepest ocean. It is almost completely surrounded by narrow continental shelves and deep trenches of subduction zones. It has many islands, including volcanic islands and atolls formed at hot spots, islands in magmatic and sedimentary arcs, and islands that appear to be small pieces of previous continents.

Aside from hot spots and limited areas of new spreading, notably the East Pacific Rise, the Pacific Ocean crust is generally old, cold, and dense. Hence, it floats low in the asthenosphere. However, the oldest oceanic crust found in the Pacific Ocean is not substantially older than the oldest oceanic crust in other oceans. This is because all of the Pacific Ocean's older crust has been subducted. Passive margins of continents that once surrounded the Pacific developed into subduction zones long ago. Oceanic crust that once was present near the passive margins also has been subducted. The sediment that once covered the subducted crust was either subducted or compacted into rock and added to island arcs or continents as exotic terrane.

When the Pacific Ocean margins changed from passive margins to subduction zones as the ocean stopped expanding hundreds of millions of years ago, a ring of coastal mountains formed on the surrounding continents. The mountains prevented most rivers on these continents from draining into the Pacific. In addition, trenches in which river-borne sediments accumulated were formed around much of the Pacific. In other areas, magmatic arcs formed, creating marginal seas where sediments discharged by rivers were trapped. All oceanic crust now found in the Pacific Ocean, other than crust formed in marginal basins that are not part of the Pacific Plate, was created after these mountain chains, trenches, and magmatic arcs were formed. Hence, since its formation, the oceanic crust of the Pacific Plate has remained remote from any major rivers and the large amounts of sediment that they can contribute (**Chap. 6**), and the **sedimentation rate** on this crust has been slow for hundreds of millions of years. Because it is not significantly older than the crust in other oceans and because the sedimentation rate is low, the Pacific Ocean floor has a relatively thin cover of sediment. As a result, the abyssal plains are not as flat as parts of the Atlantic Ocean's abyssal plains.

Atlantic Ocean

The Atlantic is a long, narrow ocean that is often considered to include the Arctic Ocean because the Mid-Atlantic Ridge stretches essentially from the North Pole to Antarctica. The Atlantic Ocean is expanding in an east–west direction. An oceanic ridge, the Mid-Atlantic Ridge, runs down its center. Most of the Atlantic coasts of Europe, Africa, and North and South America are passive margins. The Atlantic's oldest oceanic crust is at two subduction zones: the Puerto Rico–Cayman Trench at the entrance to the Caribbean Sea, and the South Sandwich Trench located northeast of the Antarctic Peninsula.

The Atlantic Ocean has fewer island arcs and hot spots than the Pacific has. Greenland, the world's largest island, is part of the North American continent and is joined to Canada by the submerged continental shelf of Baffin Bay. Continental shelves are broader in the Atlantic Ocean than in the Pacific Ocean, and they constitute a larger percentage of the total area of the Atlantic than of the Pacific. The Atlantic also has several shallow marginal seas, including the Baltic, Mediterranean, and Caribbean Seas and the Gulf of Mexico. Thick salt deposits and oil and gas reservoirs typical of developing passive margins are present in many locations around the edges of the Atlantic Ocean. Such deposits formed early in the ocean's history, during periods when the new ocean and its marginal seas were isolated from the rest of the world ocean.

The Atlantic Ocean floor has fewer seamounts and a smoother abyssal plain than the Pacific Ocean floor, in part because of the relative dearth of hot-spot volcanoes in the Atlantic. In addition, topography is buried by very large quantities of sediment deposits that are found on the Atlantic Ocean crust, except on the newly formed oceanic ridge mountains. The large quantities of sediments are derived from the great quantities of sediment that rivers have transported into the Atlantic, as the passive margins on both sides of the ocean have eroded. The Atlantic continues to receive freshwater runoff and some sediment from

vast drainage areas of Europe, Africa, and the Americas, particularly in equatorial regions. Two large rivers, the Amazon and the Congo, empty into the equatorial Atlantic. They contribute about one-fourth of the total worldwide river flow to the oceans.

Indian Ocean

The Indian Ocean is the youngest of the three major ocean basins; it was formed only during the past 125 million years by the breakup of Gondwanaland. The plate tectonic features and history of the Indian Ocean are complex. For example, it is not yet known how and why the prominent Ninety East Ridge (**Figs. 4-11, 3-4**) that divides the Indo-Australian Plate was formed.

The northern part of the Indian Ocean is dominated by the collision plate boundary between India and Eurasia. The newly formed Himalaya Mountains are readily eroded, and large quantities of sediment are transported into the Arabian Sea and the Bay of Bengal by many rivers. Those rivers include three that are among the world's largest: the Indus, which empties into the Arabian Sea; and the Ganges and Brahmaputra, both of which empty into the Bay of Bengal. The enormous quantities of sediment flowing from India and the Himalayas have accumulated to form massive **abyssal fans** and extensive, relatively shallow abyssal plains in large areas of the northern Indian Ocean.

To the east of the India–Eurasia collision, along the northeastern edge of the Indian Ocean, is the very active Indonesian subduction zone. Its trenches and island arcs extend between mainland Asia and Australia. To the west of the India–Eurasia collision, the northern Red Sea is a rift zone that becomes an oceanic ridge system at its southern end. This oceanic ridge extends southward into the central Indian Ocean, where it divides. One of the ridges that originates at that point extends southwest around Africa, and the other extends southeast around Australia. The Indian Ocean is opening in a complex manner as the African, Antarctic, and Indo-Australian plates move apart.

Passive margins are present along most of West Africa, most of Australia, and the coast of India. There are few islands in the Indian Ocean other than the many islands that form the Indonesian arc, where subduction is occurring along an oceanic convergent plate boundary. Madagascar, which is now part of Africa, appears to be a fragment of Pangaea that broke away from India when India began to move rapidly northward toward Asia, long after the initial breakup of Pangaea itself.

Marginal Seas

Several arms of the major oceans are partially isolated from the major ocean basins by surrounding landmasses. Such marginal seas are of four types:

One type consists of shallow submerged areas of continental crust. Examples include the Baltic Sea, the North Sea, Baffin Bay, and Hudson Bay.

A second type is formed in back-arc basins behind subduction zones, which often contain deep areas where oceanic crust is present. The marginal sea is separated from the open oceans by islands of the magmatic arc and a submerged ridge that connects the islands. Many such seas contain thick sediment deposits derived primarily from erosion of the newly formed islands of the arc. Examples include the Java Sea behind the Indonesian Arc, the South China Sea behind the Philippines, and the Aleutian Basin in the Bering Sea behind the Aleutian Island chain.

The third type of marginal sea is a narrow remnant of an old closing ocean. The Mediterranean Sea, which lies between the converging African and Eurasian Plates, is a good example. Such seas may have thick sediments.

Finally, some marginal seas are the long and narrow arms formed as a continental divergent plate boundary develops into a new ocean. The primary example is the Red Sea.

This page titled 4.10: Present-Day Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.11: CHAPTER SUMMARY

The Earth and Plate Tectonics.

The Earth consists of a solid inner core, liquid outer core, plastic mantle, and solid overlying lithosphere. The mantle, especially the upper mantle or asthenosphere, is close to its melting point and can flow like a fluid, but very slowly. The thin lithosphere consists of continental or oceanic crust overlying a layer of solidified mantle material and is separated into plates that float on the asthenosphere. About 225 million years ago, all the continents were joined. Since then they have been separated by plate tectonic movements.

Plate Boundary Processes.

Lithospheric plates may pull away from (diverge), collide with (converge), or slide past each other. Oceanic crust is created at divergent plate boundaries and destroyed at convergent plate boundaries. Divergent plate boundaries are the oceanic ridges and areas where continents are being pulled apart. Oceanic ridges are undersea mountain chains with many active volcanoes, and they are offset at transform faults.

Convergent plate boundaries are downwelling zones where old oceanic crust is subducted. Subduction zones at the edges of continents are characterized by an offshore trench and coastal mountains formed by compression of the continental crust plate and accumulation of sediment scraped off the subducting oceanic crust plate. Subducted and heated crust melts and magma rises to form volcanoes on the continental crust plate. Subduction zones at which oceanic crust is at the edge of both plates are characterized by a trench and a magmatic arc (and sometimes a separate sedimentary arc) of islands on the nonsubducting plate. A high subduction rate may stretch the nonsubducting plate forming a back-arc basin. A collision where two continents meet at a convergent plate boundary is characterized by mountain chains created by compression of the continental crusts of the two colliding plates.

Hot Spots.

Hot spots cause persistent volcanic activity. Some are situated over zones where upwelling convection extends throughout the mantle. Lithospheric plates move independently of any movements of most hot spots. As the lithospheric plate, hot spot, or both move with respect to each other, hot-spot trails of islands and seamounts are formed.

Plate Interiors.

As new oceanic crust moves away from a divergent plate boundary, it cools, sinks isostatically, and is buried by sediment. Edges of continents that are not at plate boundaries are known as passive margins and are characterized by a flat coastal plain, shallow estuaries and swamps, and a wide, heavily sediment-covered continental shelf.

Sea-Level Change and Climate.

The Earth's climate is naturally variable. When the average surface temperature changes, eustatic changes of sea level occur globally. When the Earth warms, sea level rises as ocean water expands thermally, and vice versa. At the Earth's warmest temperatures, the oceans covered as much as 40% of the present land surface area. At its lowest temperatures, sea level was at least 100 m lower than it is today, and most of the continental shelves were exposed.

Isostatic leveling causes sea level to change in relation to the local coast. If continental crust is weighted by ice during a glacial period, or if its temperature falls (density increases), it sinks. If crust loses weight (as it does when ice melts during warm periods) or warms, it rises. However, isostatic leveling is very slow.

Present-Day Oceans.

The Pacific is the largest and oldest ocean. It is ringed by subduction zones and has many volcanic islands and atolls formed at hot spots and magmatic arcs. Because few rivers drain directly into it and sediments are trapped in subduction zones and marginal seas, its seafloor has a relatively thin sediment cover.

The Atlantic Ocean is widening as lithospheric plates move apart at the Mid-Atlantic Ridge. It has few islands and broad continental shelves. Compared with the Pacific Ocean, it has more rivers and thicker average sediment cover.

The Indian Ocean is the youngest ocean. It has a complex oceanic ridge system, few islands, and thick sediment cover, especially in the north, where major rivers empty from the new, easily erodible Himalaya Mountains created at the India–Eurasia continental collision.

There are four types of marginal seas: shallow seas where continental crust is submerged, long narrow seas where continents are breaking apart, seas between continents that are moving toward a future collision, and back-arc basins behind subduction zones.

4.11.1: Key Terms

4.11.2: Chapter Questions

4.11.3: References

This page titled 4.11: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.11.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- abyssal fans
- abyssal hill
- abyssal plain
- asthenosphere
- atoll
- back-arc basin
- barrier reef
- basalt
- buoyant
- central rift valley
- chaotic
- coastal plain
- continental collision
- plate boundary
- continental divergent
- plate boundary
- climate
- coast
- continental drift
- continental margins
- continental shelf
- continental slope
- convection
- convection cell
- convergence
- convergent plate boundary
- crest
- crust
- deposition
- dip angle
- divergence
- divergent plate boundary
- downwell
- erosion
- eustasy
- exotic terrane
- fjords
- fossils
- fracture zone
- fringing reef
- friction
- glacial
- gravity
- greenhouse effect
- guyot
- hot spot
- hydrocarbon

- hydrosphere
- ice age

- isostasy
- isostatic leveling
- latitudes
- lithosphere
- lithospheric plate
- magma
- magmatic arc
- mantle
- marginal sea
- oceanic convergent plate boundary
- oceanic plateau
- oceanic ridge
- paleomagnetic
- passive margin
- photosynthesis
- plate tectonics
- primary productivity
- radioactive
- rift zone
- runoff
- salt marshes
- seafloor spreading
- seamount
- sediment
- sedimentary arc
- sedimentation
- sedimentation rate
- shelf break
- shelf valleys
- sonar
- species
- spreading cycle
- subduct
- subduction zone
- submarine canyons
- superplume
- tomography
- topographic
- transform fault
- transform plate boundary
- trench
- tsunami
- turbidity
- upwelled

This page titled 4.11.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.11.2: Chapter Questions

Study Questions

1. Describe the Earth's mantle. How do we know what the mantle is made of and how it behaves?
2. What are the differences between continental crust and oceanic crust, and why are these differences important?
3. What is a hot-spot trail? How do hot-spot trails show that lithospheric plates move across the Earth's surface?
4. What three types of motion occur at plate boundaries?
5. List the types of convergent plate boundaries. Describe the characteristics and locations of volcanoes associated with convergent plate boundaries. Why are there few or no volcanoes at convergent plate boundaries where two continents collide?
6. What processes occur at oceanic ridges to form their mountainous topography and the fracture zones that cut across them?
7. Describe the changes in seafloor depth and sediment cover with increasing distance from an oceanic ridge. What causes these changes?
8. Why are passive margins described as "passive"? What are their characteristics?
9. Distinguish between isostatic and eustatic processes that cause sea level to change. How do these processes complicate efforts to measure changes in global sea level by measuring sea-level heights at various coastlines?
10. If there were no ice cap on the Antarctic continent, which coast of the United States—the California coast, the Pacific Northwest coast, or the Mid-Atlantic coast—would the Antarctic coast and continental shelf resemble? Why?
11. Describe and explain the principal differences in geography and seafloor topography between the Atlantic Ocean and the Pacific Ocean.

Critical Thinking Questions

1. The structure of the Earth and the processes of plate tectonics depend on the way that fluids and solids of different density interact. List examples of everyday situations in which the difference in density between substances or within a substance determine their behavior. One such example is oil floating on top of water.
2. If continental crust and oceanic crust were able to flow like the material in the asthenosphere, how might the Earth be different? How would the processes of plate tectonics be different?
3. Most ice ages in the Earth's history have lasted about 50 million years. The most recent ice age began 2 to 3 million years ago. About 10,000 years ago, the glaciers that extended into much lower latitudes than they do today began to melt and the Earth's climate warmed. Does this evidence indicate that the most recent ice age is over? Explain the reasons for your answer.
4. Is it possible that plate tectonic motions occur at present on any planet or moon in our solar system other than the Earth? Why or why not?
5. Are there planets or moons on which there were probably active plate tectonic motions in the past but not at present? If so, which planets or moons are the most likely to have had such motions? Why?
6. Using the surface topography maps of the planets and moons that are now available, how would you investigate whether, in fact, there had been plate tectonic activity on these planets?
7. Volcanoes occur at intervals along the length of divergent plate boundaries, but not along the length of transform faults such as the San Andreas Fault. Why?
8. Evidence exists that there are extinct volcanoes in California arranged roughly parallel to the San Andreas Fault. What are the possible explanations for their origin? How would you determine which of the possible explanations was correct?
9. If the Atlantic Ocean stopped expanding and started to contract, what would happen to New York City? Where would it be located a few million years after this change of direction occurred? Would it still be on land? Where would the nearest mountains be? Where would the coastline be?
10. There have apparently been several spreading cycles in the Earth's history. What might cause the continents to repeatedly collect together, then break apart, only to collect together again?
11. One of the effects of the enhanced greenhouse effect is that sea level is rising and will continue to rise. What are the physical processes that will cause sea-level change as climate changes? How many of these causal factors can you list? Which of these factors would be the most important? Would any of them tend to lower sea level? What effects would a sea-level rise of 1 m or 10 m have on humans?
12. In the 1995 movie Waterworld, melting of all the ice in glaciers and the polar ice caps caused the oceans to expand and cover all the land surface on the Earth except for a few small islands. Is this possible? Why or why not?

This page titled 4.11.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

4.11.3: References

Figure 4-1 – Segar, D., and E. Stamman Segar, 1997: An eruption of Pu‘u‘ō‘ō.

Figure 4-2 – Adapted from Robertson, E. C., 2011: The Interior of the Earth. USGS, accessed on 18 June 2025, Source.

Figure 4-3 – Adapted from E. J. Tarbuck and F. K. Lugens, 1991, *Earth Science, 6th Edition*, Macmillan Publishing Company, 528 pp.

Figure 4-7 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, Source.

Figure 4-8 – Adapted from Isacks, B., J. Oliver and L. R. Sykes, 1968: Seismology and the new global tectonics. *J. Geophys. Res.*, **73**, 5855–5899, Source.

– Adapted from Barazangi, M., and J. Dorman, 1969: World seismicity maps compiled from ESSA, Coast and Geodetic Survey, epicenter data, 1961-1967. *Bull. Seismol. Soc. Am.*, **59**, 369–380, Source.

Figure 4-9 – Adapted from Pipkin, B. W., 1994: *Geology and the Environment*. West Publishing Company, 466 pp.

Figure 4-9 – Adapted from Scotese, C. R. and N. M. Wright, 2018: PALEOMAP Paleodigital Elevation Models (PaleoDEMS) for the Phanerozoic. EarthByte, Source.

Figure 4-10 – T. H. Van Andel, 1985: *New Views of an Old Planet: Continental Drift and the History of the Earth 1 Ed*. Cambridge University Press, 324 pp.

Figure 4-11 – T. H. Van Andel, 1985: *New Views of an Old Planet: Continental Drift and the History of the Earth 1 Ed*. Cambridge University Press, 324 pp.

Figure 4-12 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, Source.

Figure 4-13a – Doukas, M., 1980: July 22, 1980 eruption of Mount St. Helens sent pumice and ash 6 to 11 mi (10-18 km) into the air, and was visible in Seattle, Washington (100 mi/160 km north). USGS Volcano Hazards Program, accessed 18 June 2025, Source.

Figure 4-13b – Harlow, D., 1991: The June 12, 1991 eruption column from Mount Pinatubo taken from the east side of Clark Air Base. USGS Volcano Hazards Program, accessed 18 June 2025, Source.

Figure 4-13c – Glicken, H., 1980: Mount St. Helens on May 17, 1980, one day before the devastating eruption. The view is from Johnston's Ridge, six miles (10 kilometers) northwest of the volcano. USGS Volcano Hazards Program, accessed 18 June 2025, Source.

Figure 4-13d – Glicken, H., 1980: MSH80 st helens from johnston ridge 09-10-80. USGS Cascades Volcano Observatory, Wikipedia, Source, Public Domain.

Figure 4-16 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, Source.

Figure 4-19 – Arctic Exploration 2002, 2002: International bathymetric chart of the Arctic Ocean, with locations of proposed expeditions of the International Survey of Arctic Ridges and Basins. NOAA/OER, accessed 24 June 2025, Source.

Figure 4-20 – Adapted from Kious, W. J., and R. I. Tilling, R.I., 1996: This dynamic earth—The story of plate tectonics (online ed., ver. 1.20, 2015). U.S. Geological Survey General Interest Publication, Source.

Figure 4-24 – Segar, D., and E. Stamman Segar, 1997: The 1983 eruption of Pu‘u‘ō‘ō.

Figure 4-25a – Clague, D. A., and G. B. Dalrymple, 1987: The Hawaiian-Emperor volcanic chain. Volcanism in Hawaii, Part 1, R. W.

Decker, T. L. Wright, and P. H. Stauffer, Eds. 5–100, Source.

Figure 4-27 – Adapted from Dewey, J. F. and J. M. Bird, 1970: Plate tectonics and geosynclines. *Tectonophysics*, **10**, 625–638, Source.

Figure 4-28 – NASA World Wind, 2006: Orona Atoll, Source.

Figure 4-29 – Adapted from P.R. Pinet, 1992: *Oceanography, an introduction to the planet Oceanus*. West Publishing Company, 350 pp.

Figure 4-30 – NASA, Johnson Space Center, 2003: Bora Bora ISS006, Source.

Figure 4-31 – Adapted from Van Andel, T. H., 1985: *New views on an old planet, continental drift and the history of earth*. Cambridge University, 324 pp.

Figure 4-32a – Data from Cross-Chapter Box 2.1, Figure 1 in IPCC, 2021: Chapter 2. In: Climate Change 2021: The Physical Science Basis. V. Masson-Delmotte, and Coauthors, Eds., Cambridge University Press, 287–422, Source.

Figure 4-32b – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, Source.

Figure 4-32c – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, Source.

Figure 4-34a – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, Source.

Figure 4-34b – Data from Bintanja, R., R.S.W. van de Wal, and J. Oerlemans. 2005. Modelled atmospheric temperatures and global sea levels over the past million years. *Nature*, **437**, 125–128, Source.

Figure 4-34c – Data from GISTEMP Team, 2025: GISS Surface Temperature Analysis (GISTEMP), version 4. NASA Goddard Institute for Space Studies. Dataset accessed 24 June 2025, Source.

Figure 4-34c – Data from Willis, J. K., and Coauthors, 2025: NASA-SSH Global Mean Sea Level from Simple Gridded Sea Surface Height. Ver. 1. Dataset accessed 24 June 2025, Source.

This page titled 4.11.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5: WATER AND SEAWATER

Critical Concepts

- **CC1 Density and Layering in Fluids**: Water in the oceans is arranged in layers according to the water density. Many movements of water masses in the oceans are driven by differences in water density.
- **CC3 Convection and Convection Cells**: Fluids, including ocean water, that are cooled from above or heated from below, sink or rise because their density is increased or reduced respectively. This establishes convection processes that are a primary cause of vertical movements and the mixing of ocean waters.
- **CC5 Transfer and Storage of Heat by Water**: Water's high heat capacity allows large amounts of heat to be stored in the oceans and released to the atmosphere without much change of ocean water temperature. Water's high latent heat of vaporization allows large amounts of heat to be transferred to the atmosphere in water vapor and then transported elsewhere. Water's high latent heat of fusion allows ice to act as a heat buffer reducing climate extremes in high latitude regions.
- **CC6 Salinity, Temperature, Pressure, and Water Density**: Seawater density is controlled by temperature, salinity, and to a lesser extent pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Movements of water below the ocean surface layer are driven primarily by density differences.
- **CC7 Radioactivity and Age Dating**: Seawater contains naturally occurring and man-made radioisotopes of many elements at very low concentrations. These are useful for tracing movements of the elements through biogeochemical cycles.
- **CC8 Residence Time**: The residence time of an element in seawater is the average length of time atoms of the element spend in the oceans and it depends, to a large extent, on the rate of processes that remove them from solution. Residence time is a major factor in determining the concentration of elements in seawater.
- **CC9 The Global Greenhouse Effect**: The oceans play a major part in studies of the greenhouse effect because the oceans store large amounts of carbon dioxide and are a source of other greenhouse gases including methane.
- **CC14 Phototrophy, Light, and Nutrients**: Phototrophy, that includes photosynthesis, the major processes in the production of living matter, depend on the availability of light. Thus, the transmission of light through ocean water limits the depth to which phototrophy may occur.
- **CC18 Toxicity**: Many dissolved constituents of seawater become toxic to marine life at concentrations above their natural levels in seawater. A number of these constituents are nutrients that can limit primary production if their concentrations are too low. Thus, life depends on the relatively invariable concentrations of dissolved chemicals in the oceans.

Figure 5-1. All known life forms on Earth depend on water and its unique properties. Although it is well known that plants and animals must take up or ingest water to survive, water is also important in many other ways. For example, the seawater in this photograph at Deacon's Reef, Papua New Guinea, is transparent to visible light. As a result, light energy can penetrate into the ocean, where it is used for photosynthesis. Many of the chemical elements and compounds invisibly dissolved in the seawater are essential to life. Water is important in many other subtle ways as well. For example, the organisms living in this reef ecosystem, and the terrestrial plants and animals that live on nearby land, thrive in a climate in which the temperature changes little between day and night. If water did not have a high heat capacity, these organisms would experience daily temperature variations and extremes comparable to those that occur in the middle of the largest deserts.

Although most of us are aware that water is essential to life as we know it, the range of unique properties of water that make life possible is less well known. For example, the properties of water:

- Help to create and control climate and weather
- Influence the formation and modification of the land and seafloor topography
- Enable essential chemicals to be transported to and within living organisms
- Control many features of our physical environment, such as rain, snow, and the waves on oceans and lakes
- Underlie the functioning of many aspects of modern society, ranging from cooling systems for automobile engines and power plants, to ice cubes that keep drinks cold in summer

The dissolving power of water, the composition of seawater, and the processes that add or remove dissolved substances in ocean water are important to the studies of **sediments** (**Chap. 6**), life in the oceans (**Chaps. 12, 14, 15**), and **pollution** (**Chap. 16**).

Water's unique physical properties are critically important to processes discussed in many chapters in this text, especially ocean–atmosphere interactions (**Chap. 7**) and the circulation of water in the oceans (**Chap. 8**). The behavior of sound and of **electromagnetic radiation**, including visible light, as they pass through seawater is important to ocean life (**Chap. 12**) and contributes to the difficulties faced by oceanographers in studying the oceans (**Chaps. 2, 3**). Light and sound transmission also influence the survival and reproduction strategies of marine **species** (**Chap. 14**).

In this chapter, we will investigate the extraordinary properties of water that permit it to support life on our planet, we will discuss how water's dissolving ability affects the chemical composition of ocean water, and we will examine the most important physical properties of water and some of the ways that they influence the Earth and our **environment**.

5.1: Origins and Distribution of the Earth's Water

5.2: The Water Molecule

5.3: The Dissolving Power of Water

5.4: Sources and Sinks of Chemicals Dissolved in Seawater

5.5: Salinity

5.6: Dissolved Chemicals in Seawater
5.7: Heat Properties of Water
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
5.9: Ice Formation
5.10: Surface Tension and Viscosity
5.11: Transmission of Light and Other Electromagnetic Radiation
5.12: Transmission of Sound
5.13: Chapter Summary
5.13.1: Key Terms
5.13.2: Chapter Questions
5.13.3: References

This page titled 5: Water and Seawater is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.1: Origins and Distribution of the Earth's Water

Early in its history, the Earth was hot and mostly molten. Heavy elements such as iron and nickel migrated toward the Earth's center, while lighter elements, such as silicon, aluminum, and oxygen, moved upward toward the surface. The lightest elements, which included hydrogen and oxygen, and some compounds of light elements, including carbon dioxide, methane, and water vapor, migrated upward to form an atmosphere. The lightest gaseous elements, hydrogen and helium, were largely lost to space. After the Earth cooled, the **crust** solidified and water vapor in the atmosphere condensed into liquid water.

In the billions of years since the Earth's water first condensed, the temperature of the atmosphere apparently has changed relatively little. Hence, in addition to water vapor in the atmosphere, liquid and solid water were probably present in oceans and ice sheets throughout those billions of years. Volcanic activity released water that was trapped deep within the Earth's interior as the Earth solidified and continues to do so. However, the rate at which water has been released by volcanic activity since the Earth solidified is believed to be small in comparison with the volume of the oceans and ice sheets. Therefore, the amount of water at the Earth's surface probably has remained relatively unchanged for many millions of years.

Uniqueness of the Earth

The Earth is the only planet in the solar system with liquid water on its surface. The outer planets—Saturn, Jupiter, Uranus, and Neptune—are cold, and any water present is vapor or ice. The surface of Mars is currently too cold to have liquid water, although it may have been warmer in the past. Certain features of the Martian surface suggest that liquid water was present at one time in its history, and we have direct evidence from instruments on robotic vehicles on the Martian surface that this was almost certainly the case. The surfaces of Mercury and Venus are hot and could have water only as a vapor. In addition, the mass and therefore the **gravity** of Mercury, Venus, and Mars are sufficiently small that much of the water formed during their early history probably escaped into space.

There is evidence that liquid water does exist on other bodies in our solar system but not on their surface. Evidence suggests that there is liquid water under the frozen surfaces of three of Jupiter's four larger satellites: Callisto, Europa, and Ganymede. In the case of Ganymede, the water appears to be between 145 and 193 km below the surface. Water ice has also been detected on all of the major satellites of Saturn and Uranus.

Water is the only known substance that is present in all three physical forms—liquid, solid (ice), and gas (atmospheric water vapor)—within the range of temperatures and pressures found on the Earth's surface and in its atmosphere. The presence of water in all three forms and the conversion of water from one form to another are important to the maintenance of climatic conditions within the range we consider acceptable for human life and for the stability of the Earth's **ecosystems**.

Distribution of the Earth's Water

More than 96% of the world's water is in the oceans (**Table 5-1**). However, the tiny fractions of water in the atmosphere and in freshwater systems (lakes, rivers, streams, and **groundwater**) are disproportionately important to humanity. Freshwater may be the most precious and scarce natural resource supporting human civilization. Transport of water vapor from the oceans through the atmosphere to land as rain and snow determines the availability of the freshwater resource. The transfer of water between oceans and the atmosphere determines critical aspects of our **climate**. In **Chapter 7** we learn more about the importance of the hydrologic cycle whereby water is exchanged among the oceans, atmosphere, freshwater, groundwater, **glaciers** and polar ice sheets.

Table 5-1a. *The World's Water*

Location	Percentage of Total $2.0 \cdot 10^9$ km^3
Oceans	96.5

Location	Percentage of Total $2.0 \cdot 10^9$ km^3
Freshwater	3.5

Table 5-1b. *The World's Freshwater*

Freshwater	Percentage of Total $2.0 \cdot 10^9$ km^3
Ice caps and glaciers	1.78
Surface water (lakes, rivers, streams)	0.013
Air and soil moisture	0.002
Groundwater	0.69

This page titled 5.1: Origins and Distribution of the Earth's Water is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.2: The Water Molecule

To understand the unusual properties of water, we need to understand the structure of atoms and molecules, particularly the water molecule.

Atoms and Electrons

Atoms consist of negatively charged electrons orbiting in shells that surround a nucleus. The nucleus contains neutrons that have no charge and protons that are positively charged. The electrical charge of a proton is equal and opposite to that of an electron, and the number of protons in each atom is equal to the number of electrons. Hence, atoms are electrically neutral. The actual charge on every electron is $-1.602 \bullet 10^{-19}$ coulombs and on every proton $+1.602 \bullet 10^{-19}$ coulombs. For convenience, when describing ions (atoms that have lost or gained an electron), their values are usually designated as $^{-1}$ or just $^{-}$, and $^{+1}$ or just $^{+}$ (or $^{-2}$, $^{+2}$, etc. for more than one electron or proton charge).

Each electron shell is capable of holding only a certain number of electrons. For example, the innermost shell holds a maximum of two electrons, and the second shell holds a maximum of eight. Elements that have their outermost shell filled are noble gases (helium, neon, argon, krypton, xenon), which are inert (that is, they do not react chemically with most atoms).

The atoms of most elements have an outermost electron shell that is not completely filled with the maximum number of electrons. However, it is energetically favorable for each shell to be completely filled. Therefore, atoms of some elements, such as chlorine or oxygen, that have a nearly full outer electron shell, have a tendency to gain electrons in order to fill the shell. Atoms of other elements, such as sodium, that have outer electron shells less than half full, have a tendency to lose one or more electrons, producing an empty outer shell.

Chemical Bonds between Atoms

To fill their outer electron shells, two or more atoms of different elements can combine to create a molecule of a chemical compound. This can happen in two ways. First, one atom can donate one or more electrons to an atom of another element that has an incomplete outer shell. For example, a sodium atom can lose an electron, and a chlorine atom can gain this electron. Both now have completed outer electron shells. Sodium having lost an electron becomes a positively charged sodium **ion**, and chlorine becomes a negatively charged chloride ion. Because the two ions have opposite electrical charges, they are **electrostatically** attracted to each other and are bonded together by this attraction (Fig. 5-2). This type of bond is called an **ionic bond**.

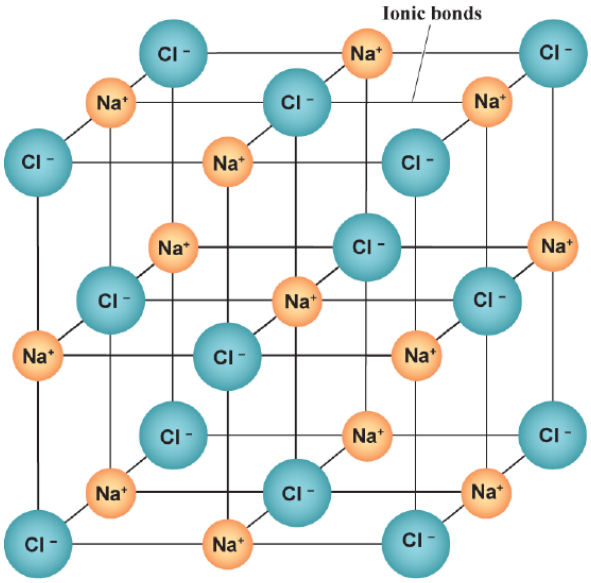


Figure 5-2. The sodium and chlorine atoms of common salt are held together in a salt crystal by ionic bonds. The ionic bond is an electrostatic attraction between the positively charged sodium ion (Na^+) and the negatively charged chlorine ion (Cl^-).

The second way in which atoms of different elements can share electrons is called a **covalent bond**. One or more outer-shell electrons spend part of their time in the outer shell of each of the two atoms that are bonded. For example, each of two hydrogen atoms in the water molecule can share its single electron with an oxygen atom. To describe such electron sharing simply, each hydrogen electron orbits both its own hydrogen atom nucleus and within the outer shell of the oxygen atom electron cloud. At the same time, two electrons from the outer shell of the oxygen atom orbit within their own shell and around the hydrogen atoms (**Fig. 5-3a**). In this way, the oxygen has the extra electrons it needs to fill its outer electron shell for part of the time, and each of the two hydrogen atoms has two electrons to fill its outer electron shell for part of the time. Because the outer shell of each atom in the molecule is full at least part of the time, the bonded atoms form a stable molecule. Most covalent bonds are stronger than most ionic bonds. Consequently, atoms in a covalently bonded molecule usually are more difficult to break apart than those that are bonded by ionic bonds.

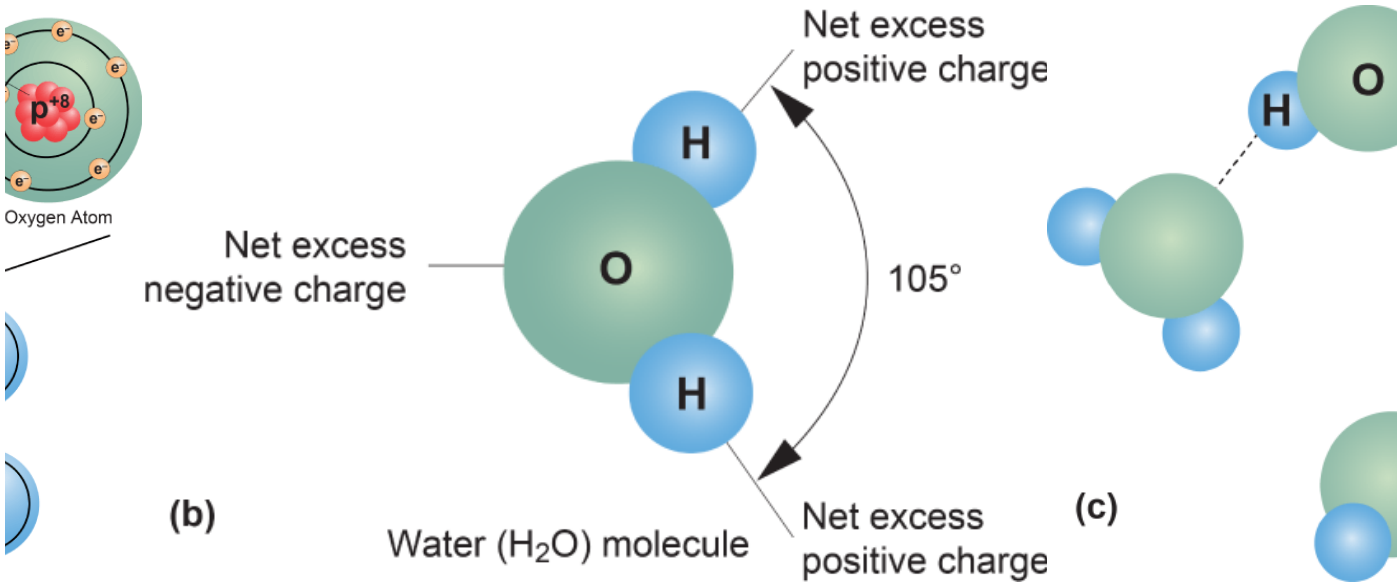


Figure 5-3. The water molecule consists of two hydrogen atoms and one oxygen atom. (a) The oxygen atom has six electrons in its outer shell. It lacks two electrons that would be needed to complete this shell. Each hydrogen atom has one electron and needs one more to complete its single shell. In a water molecule, the electrons from the hydrogen atoms and the six electrons from the outer shell of the oxygen atom are shared and orbit around both the oxygen and the hydrogen atoms. This is a covalent bond in which electron clouds from the bonded atoms partially overlap. (b) Both hydrogen atoms in a water molecule are on one side of the oxygen atom and are separated by an angle of 105°. The hydrogen electrons orbit partially around the oxygen atom, leaving the positive nucleus of each hydrogen atom partially "exposed" and giving this side of the water molecule a slight net positive charge. Similarly, the electrons orbiting the opposite side of the oxygen atom give that side of the molecule a slight net negative charge. Hence, water is a polar molecule. (c) Hydrogen bonds are formed between water molecules when the positive side of one molecule is attracted to the negative side of another.

Van der Waals Force and the Hydrogen Bond

Molecules are electrically neutral. However, there is a weak attractive force between all molecules called the **van der Waals force**. It is caused by the attraction of the protons of one atom to the electrons of another. An additional attractive force or bond, called the **hydrogen bond**, is present between the molecules of water. In other chemical compounds, the hydrogen bond is either absent or weaker than it is in water. The strength of the hydrogen bond is what gives water most of its anomalous properties (**Table 5-2**).

Table 5-2. *Anomalous Properties of Water and Their Importance*

Property	Special Characteristic of Water	Importance
Heat capacity	Higher than that of any solid or liquid other than ammonia.	Water moderates climate in coastal regions. Curre large amounts of heat. Ocean water temperatures invariable in comparison with terrestrial temperatu

Property	Special Characteristic of Water	Importance
Latent heat of fusion	Higher than that of any substance other than ammonia.	When ice forms, most of the energy lost is released atmosphere, and ice absorbs large amounts of hea therefore acts as a thermostat to keep high-latitud atmosphere near the freezing point all year.
Latent heat of vaporization	Higher than that of any other substance.	Heat is transported from the low-latitude ocean by atmospheric circulation and released to the atmos precipitation at higher latitudes.
Thermal expansion	Pure water has a density maximum at 4°C. The temperature at which the maximum occurs decreases as salinity increases. There is no maximum at salinity >24.7.	Freshwater and low-salinity seawater stay unfrozen winter in lakes and estuaries.
Solvent property	Can dissolve more substances than any other liquid.	Water dissolves minerals from rocks and transport oceans. Water is the medium in which the chemica support life occur.
Surface tension	Higher than that of any other liquid.	Surface tension controls the formation of droplets atmosphere and bubbles in the water. Some organ surface tension to anchor themselves to or walk or
Physical states	The only substance present as a gas, liquid, and solid within the temperature range at the Earth's surface.	Water vapor evaporated from the ocean helps trans warm low latitudes to cold high latitudes. Liquid wa contributes to that transport through ocean curren regions, the presence of both ice and water moder

Like other molecules, the water molecule is electrically neutral. However, the arrangement of atoms and electrons in the water molecule is such that the side of the molecule away from the hydrogen atoms has a small net negative charge, whereas the two areas where the hydrogen atoms are located have a small net positive charge (**Fig. 5-3b**). Molecules that behave as though they have a positive and a negative side are called "polar molecules." The reasons for the polarity are complicated, but they can be understood through a simplified model (**Fig. 5-3a**). The six electrons in the outer shell of the oxygen atom are arranged in pairs. The two hydrogen atoms are covalently bonded to the electrons in one of these pairs. This leaves two pairs of unshared orbiting electrons on the side of the oxygen atom opposite the hydrogen atoms, causing this side of the oxygen to have a negative charge bias. On the hydrogen side of the molecule, hydrogen electrons are shared with, and actually spend more time on, the oxygen atom, leaving the positively charged nucleus of each hydrogen atom "exposed." Hence, this side of the molecule has a positive charge.

Because the water molecule is polar, the negatively charged side of one molecule is attracted to the positively charged side of an adjacent molecule. This attractive force is the hydrogen bond (**Fig. 5-3c**). It is relatively strong, but not as strong as ionic or covalent bonds. The relative strengths of bonds between atoms and molecules are listed in **Table 5-3**. The relative strength of a bond is an indication of how much energy is needed to break that bond. Substantially more energy is needed to break hydrogen bonds than is needed to counter van der Waals forces between molecules. The relatively high strength of the hydrogen bond is responsible for the anomalous properties of water.

Table 5-3. *Relative Strengths of Bonds between Atoms and Molecules*

Type of Bond	Approximate Relative Strength
Van der Waals force (between molecules)	1
Hydrogen bond (between molecules)	10
Ionic bond (between atoms)	100
Covalent bond (between atoms)	Usually >1000

This page titled 5.2: The Water Molecule is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.3: The Dissolving Power of Water

Water can dissolve more substances and greater quantities of these substances than any other liquid. Its unique dissolving power is related to the polar nature of the water molecule. Many inorganic chemical compounds—for example, sodium chloride (NaCl)—have ionic bonds. In such compounds, the two ions (Na^+ and Cl^- in the case of NaCl) are held together by the attraction of the two opposite electrical charges. If such a compound is placed in water, the electrostatic attraction is greatly reduced and the ionic bonds are broken by a process called **hydration**. Hydration occurs when a positive **ion** (e.g., Na^+) is surrounded by water molecules oriented with their negative sides toward the positive ion (**Fig. 5-4**). Conversely, a negative ion (e.g., Cl^-) is surrounded by water molecules oriented with their positive sides toward the negative ion. Thus, each ion becomes free to move independently of the other ion, and the compound dissolves. A variety of ionically bonded inorganic compounds, or "salts," are dissolved in seawater. Many organic compounds also can be ionized and dissolved. Covalent compounds are generally less soluble in water than ionic compounds, although many covalent compounds, such as silica (SiO_2, quartz, sand), dissolve in small quantities.

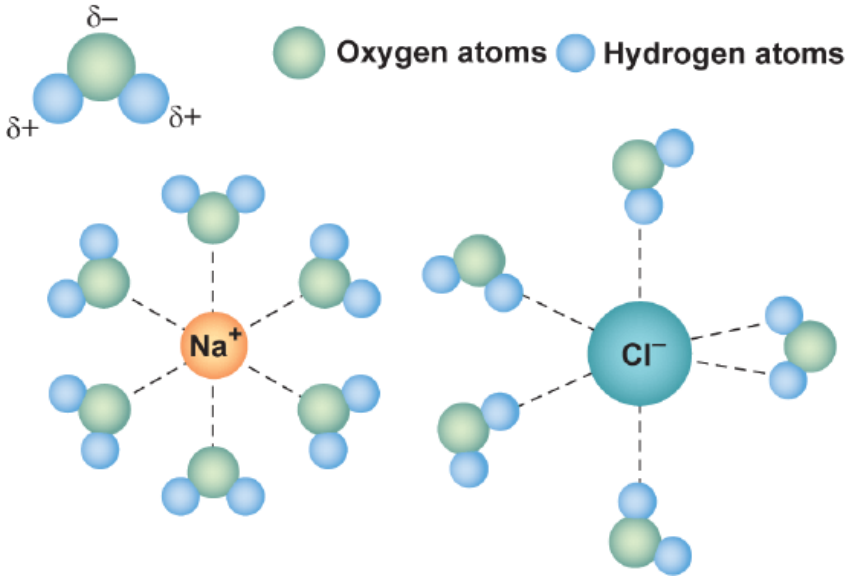


Figure 5-4. Compounds that are ionically bonded, such as common salt, are readily dissolved in water because both the positively charged ion (the sodium cation, Na^+) and the negatively charged ion (the chloride anion, Cl^-) are hydrated. Each ion is surrounded by polar water molecules oriented so that their positive sides (hydrogen atoms) face the anion and their negative sides (oxygen atom opposite the hydrogen atoms) face the cation. Hydration reduces the attraction between the cation and the anion and promotes their dispersal (solution) in the water.

Water's exceptional dissolving power is the reason why most elements are present in seawater, even though many are in very low concentrations. Most of the dissolved substances occur as either cations (positively charged ions) or anions (negatively charged ions), each of which has a surrounding sphere of properly oriented water molecules. Ions dissolved in seawater are the source of elements needed for the growth of phototrophic marine bacteria and algae, on which most ocean life depends (**Chap. 12**). Note that **phototrophy** in the oceans is dominated by microscopic bacteria, archaea, and **algae**. Plants exist only in some shallow ocean areas and account for a very tiny fraction of ocean life.

The dissolving power of water is also important to terrestrial life. Land plants obtain many of their needed elements from solution in water through their root systems. Animals, including humans, obtain many of the elements and other chemicals needed for their **biochemical** systems by dissolving these substances from their food through the digestion process. In fact, almost all processes that support the growth and function of living organisms on our planet depend on the dissolving power of water. Life as we know it would not exist if water did not have the ability to dissolve and separate the ions, enabling them to be moved in solution into and within living tissue.

This page titled 5.3: The Dissolving Power of Water is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.4: Sources and Sinks of Chemicals Dissolved in Seawater

Seawater is a solution of many different chemical compounds: cations, anions, and both organic and inorganic compounds that are not ionized. Concentrations of the compounds in ocean water are determined by their behavior in global **biogeochemical cycles** (Fig. 5-5) and by their abundance in the materials of the Earth's crust. Compounds are both added to ocean water from sources such as river **runoff** and removed from seawater to sinks such as the seafloor sediments.

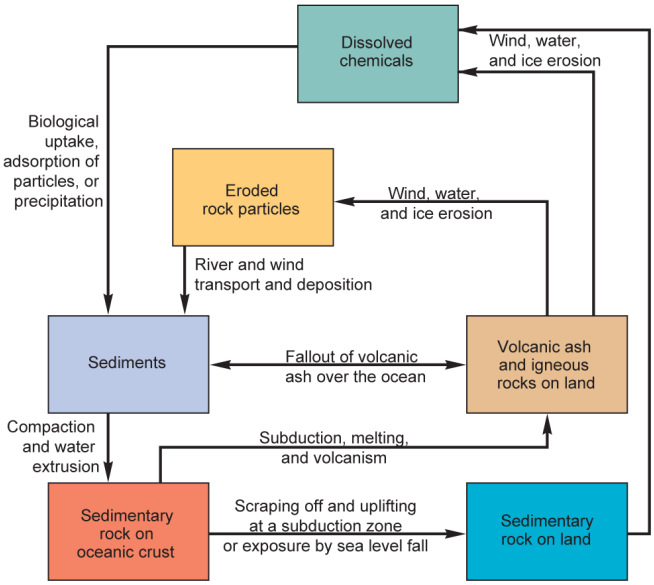


Figure 5-5. A simplified biogeochemical cycle.

Biogeochemical Cycles

Continental rocks are **weathered** and transported to the oceans both as particles and as dissolved ions. The particles are deposited as ocean sediment. Many dissolved elements are used in biological processes and subsequently are incorporated in seafloor sediment in the form of the **hard parts** and **detritus** of marine organisms. Elements dissolved in seawater also can be precipitated directly or on particulate matter or they may be **adsorbed** by (adhere to) mineral grains and detritus and thus be removed to the sediment.

Over millions of years, ocean sediments are compacted by the weight of overlying sediments. Water is squeezed out and minerals precipitate between the grains, cementing the grains together to form sedimentary rock. Vast volumes of sedimentary rock and sediment eventually enter a **subduction zone**. Some of this material is scraped off the descending **lithospheric plate** edge and raised to form **sedimentary arc** islands, **continental margin** rocks, and **exotic terranes** (Chap. 4), where it is again weathered and eroded, thus continuing the biogeochemical cycle. Much of the sedimentary rock and sediment is **subducted** into the **mantle**, where some of it is melted and ejected through volcanoes as ash or **lava**. This material either reenters the oceans and returns to the ocean sediment directly, or collects on the land, where it is weathered and eroded to start a new cycle (Fig. 5-5).

Biogeochemical cycles are actually much more complicated than Figure 5-5 indicates. For example, many elements used by marine organisms are rapidly recycled to seawater solution by the decomposition of detritus. Similarly, some elements from oceanic crustal rocks enter seawater in fluids discharged by **hydrothermal vents** at **oceanic ridges** (Chap. 15).

Steady-State Concentrations

The concentration of each element in ocean water is determined by the rate at which it enters the ocean water and the rate at which it is removed. If the rate of input exceeds the rate of removal, the concentration will rise, and vice versa. Although global biogeochemical cycles include processes that take hundreds of millions of years, these cycles have been ongoing for billions of years and are thought to be at an approximate **steady state**. Steady state is achieved when the total quantity of an element

in each of the compartments (square boxes) in **Figure 5-5** remains approximately the same over time (its input rate equals its removal rate).

To see how a steady state is achieved, consider what would happen if an element's rate of input to the oceans were increased. The total quantity of an element within the global biogeochemical cycle does not change. Therefore, as the input to the oceans increases and the total quantity of the element in the oceans increases, the total quantity of the element in the other components of the cycle (sediments, rocks, etc.) must decrease (**Fig. 5-5**). Consequently, less of the element is in the components of the cycle that provide inputs to the oceans, and their inputs to the oceans must decrease. If an element's input to the oceans increases, so do the rates of the various removal processes in the biogeochemical cycle.

Here's a simple analogy: If we pour orange juice (the input) into a glass full of water and stir continuously, the glass will overflow. At first, the overflowing liquid (the output) is almost pure water, with only traces of orange juice mixed in it. The input of orange juice has increased, but neither the concentration in the glass nor that in the output reaches a new steady level instantly. As we continue to pour orange juice into the glass, the concentration of juice in the glass increases progressively to full concentration. The concentration of juice in the output also increases progressively in response to the change in concentration in the glass. When the concentrations of orange juice in the glass and in the output match the concentration of orange juice in the input, the system has reached a new steady state. Subsequently, unless we change the concentration of orange juice in the input, or alter the output in some way (such as by allowing the orange juice pulp to settle and accumulate at the bottom of the glass), the input equals the output and the concentration of orange juice within the glass does not vary. Biogeochemical cycles are more complicated than our simple analogy because they have many inputs and outputs, and multiple "glasses" that empty into each other. However, they reach steady state in a similar way, and this steady state can be disturbed only by changes in one or more inputs or outputs.

When the global biogeochemical cycles are approximately at steady state, the rate of input of most elements to the oceans is approximately equal to the rate of removal. However, during Earth's long history, changes such as **plate tectonic** movements and climate changes have altered the rates of input or output of the various elements to the oceans so the concentration of elements in the oceans has varied over long timescales.

Residence Time

The concentrations of elements in seawater are determined largely by the effectiveness of the processes that remove them from solution. Concentrations are high if the element is not removed rapidly or effectively from solution in ocean waters and if it is abundant in the Earth's crust. The effectiveness of removal is expressed by the **residence time** (**CC8**), which is a measure of the mean length of time an atom of the element spends in the oceans before being removed to the sediment (or atmosphere).

The relationships of crustal abundance, residence time, and concentration in seawater for several elements are shown in **Table 5-4**. Note that the effectiveness of the removal processes, as expressed by residence time, is the principal determinant of concentration. Concentrations in this table are expressed both as mg•kg–1 the traditional units, that express weight of element per kg of seawater and, **molality** units that reflect the relative number of atoms of each element in seawater rather than the total weight of the elements.

Table 5-4. *Crustal Abundance, Oceanic Residence Time, and Seawater Concentration of Several Elements*

Element	Crustal Abundance (%)	Ocean Residence Time (years)	Concentration ($mg \cdot kg^{-1}$)	Concentration ($pmol \cdot kg^{-1}$)
NA (sodium)	2.4	55,000,000	10,780	$4.69 \cdot 10^{11}$
Cl (chlorine as Cl^{-1})	0.013	87,000,000	19,360	$5.46 \cdot 10^{11}$
Mg (magnesium)	2.3	13,000,000	1280	$5.27 \cdot 10^{10}$

Element	Crustal Abundance (%)	Ocean Residence Time (years)	Concentration ($mg \cdot kg^{-1}$)	Concentration ($pmol \cdot kg^{-1}$)
K (potassium)	2.1	12,000,000	399	$1.02 \cdot 10^{10}$
S (sulfur as sulfate, SO_4^{2-})	0.026	8,700,000	898	$2.80 \cdot 10^{10}$
Ca (calcium)	4.1	1,100,000	412	$1.02 \cdot 10^{10}$
Fe (iron)	2.4	200 to 500	0.00003	540
Al (aluminium)	6.0	200	0.00003	1110
Mn (manganese)	0.5	60	0.00002	360
Pb (lead)	0.001	80	0.0000027	13

This page titled 5.4: Sources and Sinks of Chemicals Dissolved in Seawater is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.5: Salinity

Many of the properties of water are modified by the presence of dissolved salts. The total quantity of dissolved salts in seawater is expressed as **salinity**. Until the early 1980s, salinity was expressed in grams of dissolved salts per kilogram of water or in parts per thousand, for which the symbol is ‰ (note that this is different from the percent symbol, %). The symbol ‰ is read as “per mil.” Open-ocean seawater contains about 35 g of dissolved salts per kilogram of seawater and thus has a salinity of 35‰.

The original approach to measuring salinity was to evaporate water and weigh the salt residue. This tedious procedure was inaccurate because some dissolved ions, such as bromide and iodide, decomposed in the process and the elements were lost as gases. Various other methods have been used to determine salinity, including measuring the chloride concentration (which is closely related to total dissolved solids because seawater follows the principle of constant proportions, as discussed in the next section).

The most precise and widely used method of salinity determination is the measurement of **electrical conductivity**. Salinity is measured by comparison of the conductivity of two solutions, one of which has a precisely known salinity. Until the early 1980s, the comparison was made with a standard seawater whose salinity was determined by a reference laboratory in Copenhagen, Denmark, and later in England. This method worked well but became very difficult as oceanography grew and the reference laboratory had to supply standard water samples to hundreds of laboratories worldwide. For this reason, and to improve the precision of salinity measurements, salinity has been redefined as a ratio of the electrical conductivity of the seawater to the electrical conductivity of a standard concentration of potassium chloride solution.

Because salinity is now defined as a ratio of electrical conductivities, it is no longer measured in parts per thousand but is expressed in “practical salinity units” (PSU). The average seawater salinity is now expressed as 35 without the ‰ symbol. However, seawater with a salinity of 35 PSU does have a concentration of almost exactly 35 g of dissolved salts per kilogram. All salinity values in this text are PSU values and are stated without the ‰ symbol, as is consistent with currently accepted practice. However, the symbol is still used in some publications.

Very small salinity changes can significantly alter seawater **density**. Consequently, salinity often must be measured to ±0.001, and to achieve that precision, conductivity must be measured to ±1 part in 40,000. Such precision is readily achievable in the laboratory, and compact rapid reading conductivity sensors achieve nearly that precision when mounted in the **CTD** probes described in **Chapter 3**.

This page titled 5.5: Salinity is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.6: Dissolved Chemicals in Seawater

The chemicals dissolved in seawater include most elements, a variety of naturally occurring and human-made **radionuclides** (**radioisotopes**), and numerous organic compounds. Elements in solution are generally ionized. Many of these ions are compound ions, such as nitrate (NO^{3-} and phosphate (PO_4^{3-}), in which atoms of one element are combined with atoms of other elements. Hence, although we talk about dissolved nitrogen or phosphorus, we generally use the term “dissolved constituents,” not “dissolved elements.” Dissolved constituents are designated major, minor, or trace according to their concentration. Trace constituents include organic compounds, radionuclides, and trace elements. The term “trace elements” is widely used, although it is not accurate, because dissolved trace elements are present in seawater as simple or compound ions. Dissolved gases are generally considered separately.

Concentrations of dissolved constituents of seawater are usually measured in parts per million ($mg \bullet kg^{-1}$) or parts per billion ($\mu g \bullet kg^{-1}$). One part per million is roughly equivalent to one teaspoonful mixed into 5000 liters of water, or enough water to fill more than 14,000 cans of soda. One part per billion is roughly equivalent to mixing one teaspoonful into 5,000,000 liters of water, enough water to fill about five Olympic-size swimming pools. Some dissolved trace metal and organic constituents of seawater occur at concentrations in the parts per trillion range. One part per trillion is equivalent to one cent in 10 million dollars or one second in 31,700 years. Oceanographers now routinely express concentrations in molality units, discussed in **Chapter 1**.

Major Constituents

The major dissolved constituents have concentrations greater than 100 parts per million by weight. The six major constituents (**Table 5-5**) are chlorine, sodium, magnesium, sulfur (as sulfate), calcium, and potassium. They occur as the ions identified in **Table 5-5**. Together, these six ions constitute 99.28% of all dissolved salts in the oceans (**Fig. 5-6**), and sodium and chloride (the constituents of table salt) alone constitute more than 85%.

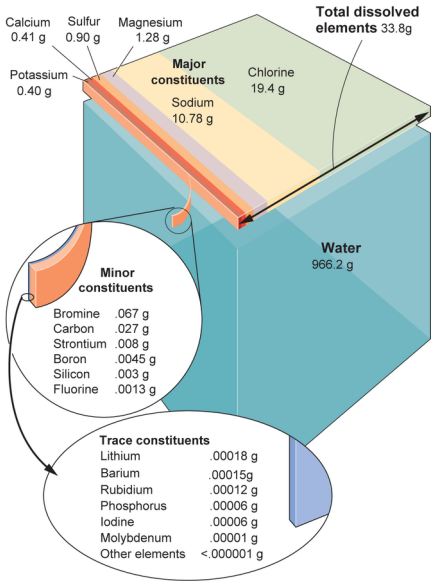


Figure 5-6. The dissolved elements in 1 kg of seawater. The total mass of dissolved elements is less than the total mass of dissolved salts (salinity) because the mass of dissolved salts includes the H^+ and OH^- ions associated with the dissolved species of some elements in seawater (see **Table 5-5**).

Table 5-5. *Concentrations and Speciation of the Elements in Seawater with a Salinity of 35*

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Chlorine	Cl	$1.936 \cdot 10^4$	Cl^-
Sodium	Na	$1.078 \cdot 10^4$	Na^+
Magnesium	Mg	$1.28 \cdot 10^3$	Mg^{2+}
Sulfur	S	$8.98 \cdot 10^2$	SO_4^{2+}, $NaSO_4^-$
Calcium	Ca	$4.12 \cdot 10^2$	Ca^{2+}
Potassium	K	$3.99 \cdot 10^2$	K^+
Bromine	Br	67	Br^-
Carbon	C	27	HCO^{3-}, CO_3^{2-}, CO_2
Nitrogen	N	8.3	N_2 gas, NO^{3-} , NH_4
Strontium	Sr	7.8	Sr^{2+}
Boron	B	4.5	$B(OH)_3$, $B(OH)_4^-$, H_2BO_3
Oxygen	O	2.8	O_2 gas
Silicon	Si	2.8	$Si(OH)_4$
Fluorine	F	1.3	F^-, MgF^+

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Argon	Ar	0.62	Ar gas
Lithium	Li	0.18	Li^{+}
Rubidium	Rb	0.12	Rb^{+}
Phosphorus	P	$6.2 \cdot 10^{-2}$	HPO_4^{2-}, PO_4^{3-}, H_2PO_4
Iodine	I	$5.8 \cdot 10^{-2}$	IO^{3-}, I_2
Barium	Ba	$1.5 \cdot 10^{-2}$	Ba^{2+}
Molybdenum	Mo	$1 \cdot 10^{-2}$	MoO_4^{2-}
Uranium	U	3.2•10–3	$UO_2(CO_3)_2^{4-}$
Vanadium	V	$2.0 \cdot 10^{-3}$	$H_2VO_4^{-}$, HVO_4^{2-}
Arsenic	As	$1.2 \cdot 10^{-3}$	$HAsO_4^{2-}$, $H_2AsO_4^{-}$
Nickel	Ni	$4.8 \cdot 10^{-4}$	Ni^{2+}
Zinc	Zn	$3.5 \cdot 10^{-4}$	$Zn(OH)_2$, Zn^{2+}, $ZnCO_3$
Krypton	Kr	$3.1 \cdot 10^{-4}$	Kr gas
Cesium	Cs	$3.1 \cdot 10^{-4}$	Cs^{+}

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Chromium	Cr	$2.1 \cdot 10^{-4}$	$Cr(OH)_3$, CrO_4^{2-}
Antimony	Sb	$2.0 \cdot 10^{-4}$	$Sb(OH)^{6-}$
Neon	Ne	$1.6 \cdot 10^{-4}$	Ne gas
Copper	Cu	$1.5 \cdot 10^{-4}$	$CuCO_3$, $CuOH^+$
Selenium	Se	$1.55 \cdot 10^{-4}$	SeO_3^{2-}
Cadmium	Cd	$7 \cdot 10^{-5}$	$CdCl_2$
Xenon	Xe	$6.6 \cdot 10^{-5}$	Xe gas
Aluminium	Al	$3 \cdot 10^{-5}$	$Al(OH)_4^-$
Iron	Fe	$3 \cdot 10^{-5}$	$Fe(OH)_2^+$, $Fe(OH)_4$
Manganese	Mn	$2 \cdot 10^{-5}$	Mn^{2+}, $MnCl^+$
Yttrium	Y	$1.7 \cdot 10^{-5}$	$Y(OH)_3$
Zirconium	Zr	$1.5 \cdot 10^{-5}$	$Zr(OH)_4$
Thallium	Tl	$1.3 \cdot 10^{-5}$	Tl^+
Tungsten	W	$1 \cdot 10^{-5}$	WO_4^{2-}

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Niobium	Nb	$5 \cdot 10^{-6}$	Not known
Rhenium	Re	$7.8 \cdot 10^{-6}$	ReO_4^-
Helium	He	$7.6 \cdot 10^{-6}$	He gas
Titanium	Ti	$6.5 \cdot 10^{-6}$	$Ti(OH)_4$
Lanthanum	La	$5.6 \cdot 10^{-6}$	$La(OH)_3$
Germanium	Ge	$5.5 \cdot 10^{-6}$	$Ge(OH)_4$
Hafnium	Hf	$3.4 \cdot 10^{-6}$	Not known
Neodymium	Nd	$3.3 \cdot 10^{-6}$	$Nd(OH)_3$
Lead	Pb	$2.7 \cdot 10^{-6}$	$PbCO_3$, $Pb(CO_3)_2^{2-}$
Tantalum	Ta	$<2.5 \cdot 10^{-6}$	Not known
Silver	Ag	$2 \cdot 10^{-6}$	$AgCl_2^-$
Cobalt	Co	$1.2 \cdot 10^{-6}$	Co^{2+}
Gallium	Ga	$1.2 \cdot 10^{-6}$	$Ga(OH)_4$
Erbium	Er	$1.2 \cdot 10^{-6}$	$Er(OH)_3$

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Ytterbium	Yb	$1.2 \cdot 10^{-6}$	$Yb(OH)_3$
Dysprosium	Dy	$1.1 \cdot 10^{-6}$	$Dy(OH)_3$
Gadolinium	Gd	$9 \cdot 10^{-7}$	$Gd(OH)_3$
Praseodymium	Pr	$7 \cdot 10^{-7}$	$Pr(OH)_3$
Scandium	Sc	$7 \cdot 10^{-7}$	$Sc(OH)_3$
Cerium	Ce	$7 \cdot 10^{-7}$	$Ce(OH)_3$
Samarium	Sm	$5.7 \cdot 10^{-7}$	$Sm(OH)_3$
Tin	Sn	$5 \cdot 10^{-7}$	$SnO(OH)_3^-$
Holmium	Ho	$3.6 \cdot 10^{-7}$	$Ho(OH)_3$
Lutetium	Lu	$2.3 \cdot 10^{-7}$	$Lu(OH)_3$
Beryllium	Be	$2.1 \cdot 10^{-7}$	$BeOH^+$
Thulium	Tm	$2 \cdot 10^{-7}$	$Tm(OH)_3$
Europium	Eu	$1.7 \cdot 10^{-7}$	$Eu(OH)_3$
Terbium	Tb	$1.7 \cdot 10^{-7}$	$Tb(OH)_3$

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Mercury	Hg	$1.4 \cdot 10^{-7}$	$HgCl_4^{2-}$, $HgCl_2$
Indium	In	$1 \cdot 10^{-7}$	$In(OH)^{2+}$
Rhodium	Rh	$8 \cdot 10^{-8}$	Not known
Palladium	Pd	$6 \cdot 10^{-8}$	Not known
Platinum	Pt	$5 \cdot 10^{-8}$	Not known
Tellurium	Te	$7 \cdot 10^{-8}$	$Te(OH)_3$
Bismuth	Bi	$3 \cdot 10^{-8}$	BiO^+, $Bi(OH)^{2+}$
Thorium	Th	$2 \cdot 10^{-8}$	$Th(OH)_4$
Gold	Au	$9.9 \cdot 10^{-9}$	$AuCl_2^-$
Ruthenium	Ru	$5 \cdot 10^{-9}$	Not known
Osmium	Os	$2 \cdot 10^{-9}$	Not known
Radium	Ra	$1.3 \cdot 10^{-10}$	Ra^{2+}
Iridium	Ir	$1.3 \cdot 10^{-10}$	Not known
Protactinium	Pa	$5 \cdot 10^{-11}$	Not known

Constituent Elements	Chemical Symbol	Concentration ($mg \cdot kg^{-1}$)	Some Probable Dissolved Species
Radon	Rn	$6 \cdot 10^{-16}$	Rn gas

**Note: Even for the more abundant constituents, concentration may vary slightly. For the rarer elements, the listed concentrations are uncertain and may be revised as analytical methods improve.*

With the exception of calcium, which is used by many marine organisms to build calcium carbonate hard parts of their bodies (**Chap. 6**), the major constituents are not utilized significantly in biological processes and do not interact readily with inorganic particles. Therefore, the primary process by which they are removed from the oceans is by the precipitation of salt deposits. This process occurs in shallow, partially enclosed **marginal seas** or embayments where the rate of removal of water by evaporation far exceeds its replacement by precipitation, river flow, and mixing with the open ocean (**Chap. 6**).

Inputs of major constituents from rivers and other sources are very small in comparison with the quantities in the oceans. Because concentrations of these constituents are not affected significantly by inputs or removal processes, their residence time is very long and the relative concentrations of these major constituents do not vary significantly. This principle of constant proportions is a cornerstone of chemical oceanography. The ratios of major constituent concentrations in seawater only vary significantly in enclosed seas, where evaporation leads to salt precipitation, and in **estuaries**, where the ocean water is mixed with substantial quantities of river water. River water has a much more variable composition of the constituents.

Minor Constituents

The minor constituents of seawater include bromine, carbon, strontium, boron, silicon, and fluorine. They have concentrations between 1 part per million and 100 parts per million (**Table 5-5**, **Fig. 5-6**). Nitrogen and oxygen are not considered to be among the minor constituents, because they are present in seawater primarily as dissolved molecular oxygen (O_2) and nitrogen (N_2) gases. Dissolved gases are considered separately later in the chapter. Together, the six major and six minor constituents constitute more than 99.6% of all the dissolved solids. Several minor constituents, notably carbon and silicon, are utilized extensively in biological processes. Therefore, the concentrations of these constituents in ocean waters are variable geographically and with time, changing in response to uptake by marine organisms and release during decay of organic matter. Such variations are discussed in **Chapter 12**.

Trace Elements

Other than the 12 major and minor constituents and the dissolved gases, all the elements listed in **Table 5-5** have concentrations in seawater of less than 1 part per million. Together, these trace elements constitute only about 0.4% of the total dissolved solids in seawater. Most trace elements are used extensively in biological processes or are absorbed by or adsorb easily to particles that remove them from seawater. Many are introduced in significant quantities by hydrothermal vents, undersea volcanoes, decomposition of organic matter, atmospheric sources such as volcanic gases, river outflows, and release from seafloor sediment when it is disturbed. Concentrations of the various trace constituents vary substantially in different parts of the ocean in response to variations in the input and removal processes.

Many trace elements are essential minerals for marine life, others are toxic, and many are essential for marine life at low concentrations but toxic at higher concentrations (**CC18**). Measurements of the concentrations of trace metals and observations of their **spatial** and **temporal** variations in the oceans are important to marine biological and pollution studies.

Many trace elements that are essential (e.g., iron, zinc, and copper) or toxic (e.g., lead and mercury) have seawater concentrations of about one part per billion (10^9) down to less than one part in 10^{13} (**Table 5-5**). Marine chemists have great difficulty measuring such extremely small concentrations. For many elements, the amount that dissolves into the water sample from surfaces of samplers and sample bottles or is contributed by dust in laboratory air can be greater than the quantity in the original water sample itself. This necessitates water sampling bottles designed to minimize or eliminate such contamination for the specific element being analyzed and clean laboratories to avoid contamination during sample analysis.

Radionuclides

A variety of naturally occurring and human-made radionuclides are present in seawater at extremely low concentrations (**CC7**). **Radioactivity** emanating from individual **isotopes** can be measured at exceedingly low levels, and sample contamination is generally a less critical problem than it is for trace metals. Therefore, determining even extremely small concentrations of radionuclides in seawater is relatively easy. In addition, many radionuclides are introduced to the oceans at known locations, and thus radionuclides can be used to trace the movements of ocean water (**Chap. 8**).

Radionuclides behave in biogeochemical cycles in ways that are virtually identical to those of the stable (nonradioactive) isotopes of their elements. Therefore, radionuclide distributions in seawater, sediments, and marine organisms, and movements of radionuclides among them, are used to infer movements of stable isotopes through the marine biogeosphere.

Organic Compounds

Thousands of different dissolved organic compounds are present in seawater. They include naturally occurring compounds such as proteins, carbohydrates, **lipids, amino acids**, vitamins, and petroleum hydrocarbons, as well as synthetic contaminants, including **DDT** and **PCBs** (polychlorinated biphenyls). The number of organic compounds is so large that probably less than 1% of them have been identified. Most organic compounds are very difficult to study because they are present in the parts per trillion concentration range (one teaspoonful in 5000 Olympic-size swimming pools). However, we believe that marine algae and perhaps some animals cannot grow successfully without dissolved compounds, such as vitamins (**Chaps. 12, 13**). We also know that certain organic compounds are highly toxic, even at concentrations in the parts per trillion range.

Most organic compounds in seawater are naturally occurring compounds. Many are produced by marine organisms and released to seawater, either through **excretions** (similar to urine) or by being dissolved after the death and decomposition of the organism. Organic compounds are also transported from land to the ocean in rivers and through the atmosphere. In some areas, especially where runoff from **mangrove** swamps or **salt marshes** carries large quantities of organic matter, the concentrations of colored dissolved organic compounds (sometimes called “gelbstoff”) are high enough to make the water appear brownish yellow.

Some dissolved organic compounds are removed from solution by attachment to particles that sink to the seafloor. However, most such compounds are taken up by marine organisms or decomposed to their inorganic constituents in the water column, primarily by **bacteria** and **archaea** (**Chap. 12**).

Dissolved Gases

Gases are free to move between the atmosphere and the oceans at the ocean surface. The net direction of the exchange is determined by the **saturation solubility** and concentration of the gas in seawater. The saturation solubility is the maximum amount of the gas that can be dissolved in water at a specific temperature, salinity, and pressure. If seawater is undersaturated, a net transfer of gas molecules into the water occurs. If seawater is oversaturated, the net transfer is from the water into the atmosphere.

The atmosphere is composed primarily of nitrogen (78%) and oxygen (21%) and contains several other minor gases. Carbon dioxide constitutes about 0.037% of all atmospheric gases. The distribution of gases dissolved in ocean waters is very different (**Table 5-6**). The oceans have proportionally more oxygen and less nitrogen than the atmosphere because of differences in the saturation solubility of these gases. In addition, the ratios of carbon dioxide concentration to oxygen and nitrogen concentrations are much higher in seawater than in the atmosphere because carbon dioxide reacts with water in a complicated way to produce highly soluble carbonate (CO_3^{2-}) and bicarbonate (HCO^{3-}) anions.

Table 5-6. *Distribution of Gases by Volume in the Atmosphere and Dissolved in Seawater*

Gas	Atmosphere (%)	Surface Ocean (%)	Total Ocean (%)
Nitrogen (N_2)	78	48	11

Gas	Atmosphere (%)	Surface Ocean (%)	Total Ocean (%)
Oxygen (O_2)	21	36	6
Carbon Dioxide (CO_2)*	0.037	15	83

*CO_2 in the atmosphere, CO_2 plus HCO^{3-} plus CO_3^{2-} in the oceans.

Oxygen and Carbon Dioxide

Gases can be exchanged between the ocean and the atmosphere only at the ocean surface. In the water column beneath the ocean surface, the proportions of dissolved gases are changed primarily by biochemical processes. In the shallow **photic zone** where light penetrates, carbon dioxide is consumed, and oxygen released during **photosynthesis**. However, in the much larger volume of deep-ocean waters where no photosynthesis occurs, the dominant process affecting dissolved gas concentrations is the consumption of oxygen through **respiration** and decomposition. The excess carbon dioxide produced by respiration and decomposition can escape to the atmosphere only at the ocean surface, and there is no mechanism for resupplying dissolved oxygen to the water below the photic zone. Therefore, deep-ocean water is depleted of oxygen and stores large quantities of carbon dioxide (Table 5-6).

The total quantity of carbon dioxide dissolved in the oceans is about 70 times as large as the total in the atmosphere. The processes that control oxygen and carbon dioxide concentrations at different depths and locations within the oceans are discussed in more detail in Chapter 12. The role of the oceans in absorbing and storing carbon dioxide is critical to the fate of carbon dioxide that has been, or will be, released by industrialized civilization, and to the severity of global climate change due to the enhanced **greenhouse effect** (**CC9**).

The saturation solubility of gases varies with pressure, temperature, and salinity. It increases with increasing pressure and decreasing temperature. Therefore, seawater at depth in the oceans can dissolve much higher concentrations of gases than surface seawater can. There is continuous movement of water from ocean surface layers to the deep layers and eventually back to the surface (Chap. 8). As water moves through the ocean depths, concentrations of dissolved gases are changed by biochemical processes and by mixing with other water masses that have different concentrations of the gases. Geological processes, including the release of gases from undersea volcanoes and from decaying organic matter in sediments, can also change the concentrations of some gases, but those processes are generally of minor significance.

When water that is saturated with carbon dioxide sinks below the surface layer, gases can no longer be exchanged with the atmosphere. However, carbon dioxide is released into deep-ocean waters by respiration and by decay of organic matter. The added gas remains dissolved because the saturation solubility is increased by the higher pressures.

The increase in solubility caused by increased pressure is what keeps carbon dioxide dissolved in carbonated sodas. The carbon dioxide is dissolved in the soda under increased pressure at the bottling plant and then sealed in its container at the higher pressure. When the container is opened, the internal pressure is released, and the carbon dioxide bubbles out because the concentration exceeds the saturation solubility at the lower pressure.

The saturation solubility of gases generally increases as temperature decreases and is generally lower at ocean water salinities than in pure water (Fig. 5-7). Therefore, oxygen concentrations are higher in cold surface waters near the polar regions than in tropical waters. In tropical waters, the oxygen concentration is low enough that, under certain circumstances, it can be inadequate for the respiration needs of some marine species. For the same reason, tropical waters are more vulnerable to marine pollution by oxygen-consuming waste materials, such as sewage (Chap. 16).

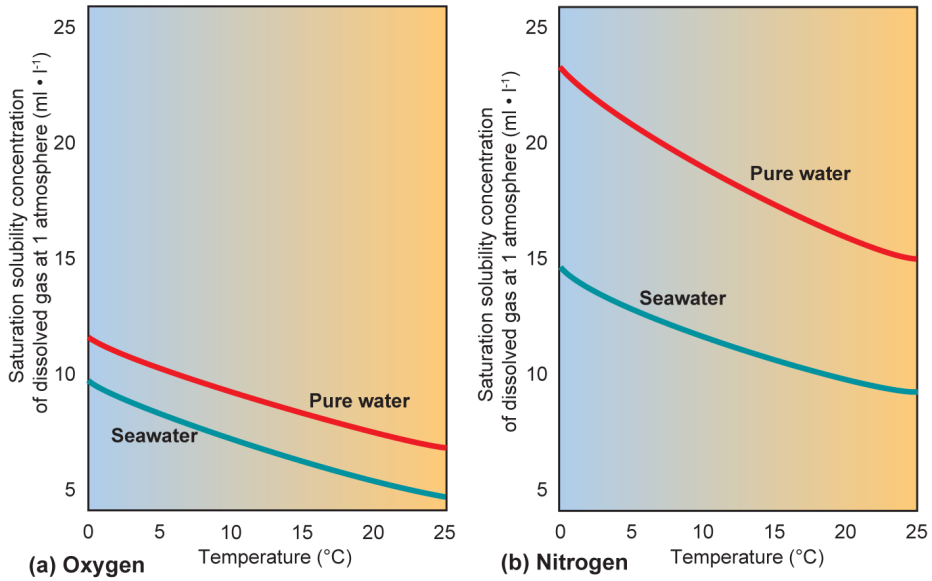


Figure 5-7. Solubility of (a) oxygen and (b) nitrogen in pure water and seawater. Note that the solubility of each gas is reduced as temperature increases and is lower at seawater salinity than in freshwater.

Other Gases

In addition to the major atmospheric gases, several other gases are present in seawater (**Table 5-7**). With the exception of sulfur dioxide, these gases are produced primarily by marine organisms, so surface waters are oversaturated and net movement of the gases is into the atmosphere. The quantities of such gases supplied to the atmosphere by the oceans are relatively small in comparison with those from other sources. However, the ocean concentrations of methane, for example, must be taken into account in global climate change studies because atmospheric methane contributes significantly to the greenhouse effect (**CC9**). Atmospheric sulfur dioxide comes primarily from **fossil fuel** burning, industrial processes, and volcanoes. Sulfur dioxide can be converted to sulfuric acid, the principal component of **acid rain**. The oceans act as a sink for this air contaminant by absorbing sulfur dioxide and the sulfate ions present in the runoff from acid rain.

Table 5-7. *Estimated Flux of Gases between Oceans and Atmosphere*

Gas	Total Transfer ($g \cdot yr^{-1}$)	Direction of Net Transfer
Sulfur dioxide (SO_2)	$1.5 \cdot 10^{14}$	Atmosphere to ocean
Nitrous oxide (N_2O)	$1.2 \cdot 10^{14}$	Ocean to atmosphere
Carbon monoxide (CO)	$4.3 \cdot 10^{13}$	Ocean to atmosphere
Methane (CH_4)	$3.2 \cdot 10^{12}$	Ocean to atmosphere
Methyl iodide (CH_3I)	$2.7 \cdot 10^{11}$	Ocean to atmosphere

Gas	Total Transfer ($g \cdot yr^{-1}$)	Direction of Net Transfer
Dimethyl sulfide $(CH_3)_2S$	$4.0 \cdot 10^{13}$	Ocean to atmosphere

pH and Buffering

The **acidic** or **alkaline** property of water is expressed as **pH**, which is a measure of the concentration of hydrogen ions (H^+). The pH increases as the hydrogen ion concentration decreases, and it is measured on a logarithmic scale of 0 to 14: pH 0 is the most acidic, pH 14 is the most alkaline, and pH 7 is neutral. Pure water (free of dissolved carbon dioxide) is neutral, pH 7. Seawater normally is about pH 8, mildly alkaline. The near-neutral pH of natural waters is very important to aquatic biology. For example, persistent acid rain, which can be about pH 5, can severely damage or destroy aquatic life in lakes by reducing the natural pH of the lake water. In addition, the numerous marine species with calcium carbonate hard body parts cannot construct these parts if the water is even mildly acidic.

Seawater pH is buffered (maintained within a narrow range of pH 7.5 to 8.1) through the reactions of dissolved carbon dioxide. Dissolved carbon dioxide combines with water to form carbonic acid (H_2CO_3). The carbonic acid partially dissociates (separates into ions) to form a hydrogen ion and a bicarbonate ion (HCO_3^-), or two hydrogen ions and a carbonate ion (CO_3^{2-}). Carbon dioxide, carbonic acid, bicarbonate, carbonate, and hydrogen ions coexist in equilibrium in seawater:

$$CO_2 \text{ (gas)} + H_2O \diamondsuit H_2CO_3 \diamondsuit H+ + HCO_3^- + 2H^+ + CO_3^{2-}$$

The addition of acid to seawater increases the number of hydrogen ions, which reduces the pH of the seawater and forces the equilibrium to shift so that less carbonate and bicarbonate are present. This shift reduces the number of hydrogen ions present, which offsets the reduction of pH. The opposite shift occurs if alkali is added to seawater. Such buffering capacity partially protects the ocean waters from pH changes that otherwise might result from acid rain or from acidic or alkaline industrial **effluents**. However, the amount of carbon dioxide released by humans is so large that ocean pH is indeed changing slowly. Some of the possible consequences of this change were discussed in **Chapter 1**.

This page titled 5.6: Dissolved Chemicals in Seawater is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.7: Heat Properties of Water

Of the properties of water, perhaps none are more important than the unique responses of water, ice, and water vapor to the application and removal of heat.

Heat Energy and Phase Changes

All substances can exist in three different phases: solid, liquid, or gas. The 3 phases of water and their properties are shown in **Figure 5-8**. In a solid, attractive forces between molecules (van der Waals forces, and hydrogen bonds if present) are strong enough to ensure that the molecules stay firmly fixed in place relative to each other, even though the molecules (or atoms in a pure element) vibrate. Heating the solid increases the strength of these vibrations. When the heat energy of each molecule is sufficient to overcome most of the attractive forces, the solid melts and becomes liquid. In a liquid, the molecules have enough energy to vibrate, rotate, and translate (temporarily move about in relation to each other). However, they do not have enough energy to escape completely from the attractive forces of their neighbors. If more heat is added to a liquid, each molecule eventually has enough energy to break free of the attractive forces. The compound then becomes a gas, in which each molecule is free to move about by itself. If heat is removed, the process is reversed; gas becomes liquid, and then solid, as heat is progressively lost.

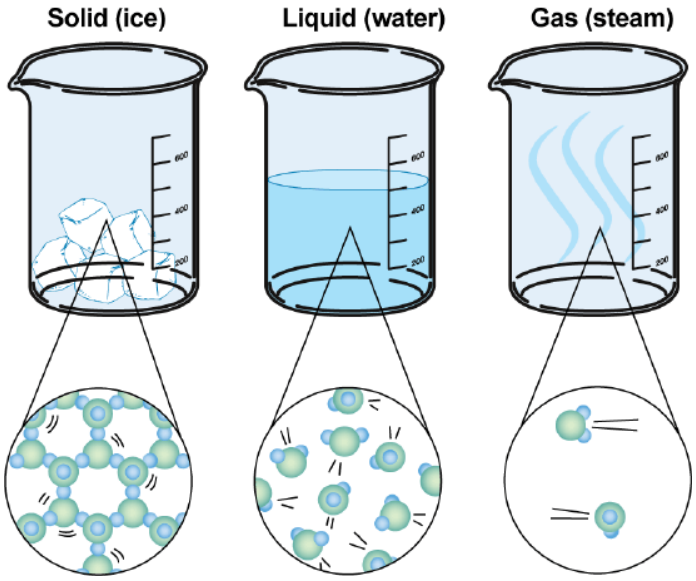


Figure 5-8. Characteristics of solid, liquid, and gaseous water.

Table 5-8. *Characteristics of Water*

Property	Solid (ice)	Liquid (water)	Gas (steam)
Motion	Limited to vibrations in almost fixed positions	Molecules moving around, can flow	Complete freedom of motion, can flow
Shape	Fixed, definite, rigid	Same as that of the lower part of the container, variable	Same as that of the closed container, highly variable

Property	Solid (ice)	Liquid (water)	Gas (steam)
Volume	Almost constant, variest only slightly with temperature	Almost constant, varies slightly with temperature	Same as that of the closed container, variable

Molecules of different chemical compounds have van der Waals forces of different strengths. As a result, each compound requires a certain characteristic quantity of heat energy to convert from solid to liquid or from liquid to gas. The stronger the attractive force between the molecules, the more heat energy each molecule must have to break free of this force. Therefore, the temperature at which a solid melts (the melting, or freezing, point) and the temperature at which a liquid vaporizes (the boiling, or condensation, point) increase as the van der Waals attractive force increases.

Freezing and Boiling Points

To convert solid water (ice) to liquid water and liquid water to gaseous water (water vapor), the heat energy supplied must overcome both the van der Waals attractive force and the much stronger attractive force of the hydrogen bond. Therefore, the amount of heat energy that each molecule of water must have to become free to rotate and move in relation to adjacent molecules (that is, to change from ice to water) is much greater than it would be if the hydrogen bonds were not present. Similarly, the amount of heat energy that each water molecule must have to free itself completely and enter the gaseous state is much greater than it would be without the hydrogen bond.

The extra energy needed to break the hydrogen bond causes water's freezing point and boiling point temperatures to be anomalously high. If there were no hydrogen bond, water would freeze at about –90°C and boil at about –70°C, and all water on the Earth would be gaseous. There would be no oceans or life as we know it. Comparison of the boiling and freezing points of water with those of other hydrogen compounds that are formed with elements that, like oxygen, have two electrons short of a full outer electron shell illustrates the anomalous nature of water (**Fig. 5-9**). Molecules in which hydrogen is combined with elements whose atoms are similar in size to oxygen but have a different number of outer-shell electrons (hydrogen fluoride, HF; ammonia, NH_3) are also polar. They, too, have hydrogen bonds and anomalously high melting and boiling points. As discussed later in this chapter, adding salt to water raises the boiling point and lowers the freezing point.

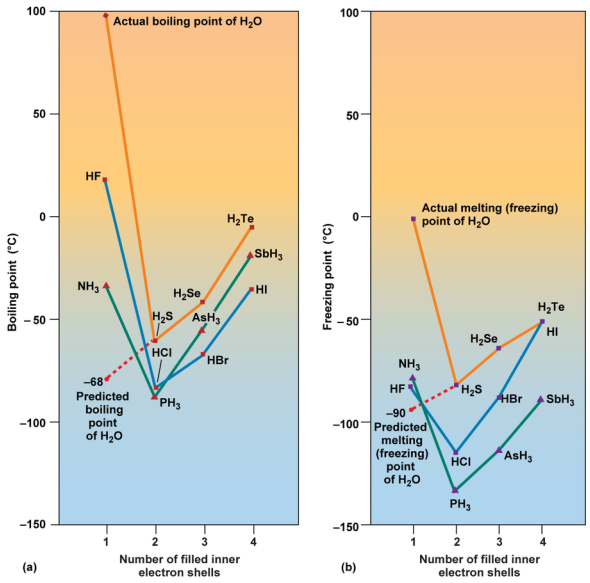


Figure 5-9. Freezing and boiling points of hydrogen compounds. The lines on the chart connect compounds of hydrogen with elements that have the same number of electrons in their outer shell but a different number of filled inner shells. If there were no hydrogen bonds, both the boiling point and the melting point of those compounds would be expected to increase progressively along these lines as the number of filled shells increased. However, water (H_2O), hydrogen fluoride (HF), and ammonia (NH_3) all have strong hydrogen bonds between their polar molecules, whereas there is no appreciable hydrogen bond in the compounds of hydrogen that are formed with sulfur (S), selenium (Se), tellurium (Te), chlorine (Cl), bromine (Br), iodine (I), phosphorus (P), arsenic (As), and antimony (Sb). In this latter group of compounds, the hydrogen bond is virtually absent because the atom combined with the hydrogen is much larger, and, therefore, the polarity of the molecules is much less. Since the hydrogen bond must be broken to convert solid to liquid or liquid to gas, the boiling points (a) and melting (freezing) points (b) of water, hydrogen fluoride, and ammonia are much higher than would be expected if they were not held by hydrogen bonds.

Heat Capacity and Latent Heat

In addition to the freezing and boiling points, the numerical values of three other related heat properties of water are anomalously high because of the hydrogen bond: the **heat capacity,** the **latent heat of fusion**, and the **latent heat of vaporization**. These properties express the quantity of heat per specified quantity of a substance needed to raise the temperature, to convert solid to liquid, and to convert liquid to gas, respectively. Heat is a form of energy, and as such, it is quantified by a unit of energy. The SI unit of energy is the **joule** (abbreviated J). This SI unit is not yet in common usage in the U.S., although it is widely used elsewhere. Many students will be more familiar with the **calorie** (abbreviated cal) widely used as the unit of energy in the U.S. In this text, we will use the calorie, and we will also identify the equivalent value in joules.

To understand the concepts of heat capacity and latent heat, we can consider the sequence of events that occurs when we add heat to ice (**Fig. 5-10**). Within the solid ice, the individual molecules vibrate, but the vibrations are suppressed by the hydrogen bonds. Adding heat energy to the ice increases the intensity of vibration of each water molecule and increases the temperature of the ice. The temperature of 1 g of ice is increased by 1°C when approximately 0.5 cal (2.1 J) is added. As heat continues to be added, the temperature continues to rise until the ice reaches its melting point. Heat that raises (or lowers) the temperature of a substance to which it is added (or removed) is called **sensible heat**. The quantity of heat needed to increase the temperature of a specified quantity of a substance is called the "heat capacity." For ice, the heat capacity is approximately 0.5 cal•g^{-1}•$°C^{-1}$ (2.1 J•g^{-1}•$°C^{-1}$). That is, one-half of a calorie (2.1 J) is needed to raise the temperature of 1 g of ice by 1°C.

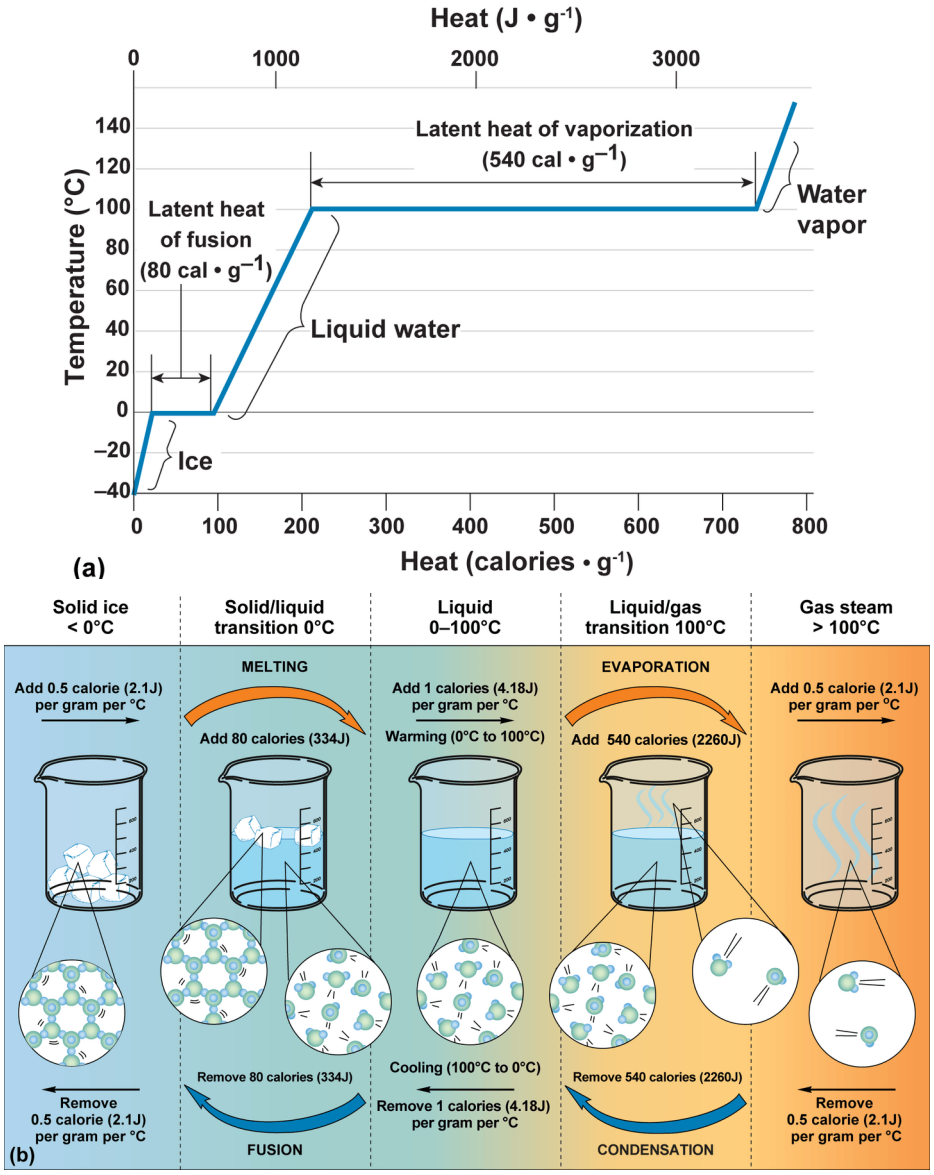


Figure 5-10. Water's latent heat of fusion (melting) is the amount of heat that must be added to convert ice to water at the freezing (melting) point temperature. Water's latent heat of vaporization is the amount of heat that must be added to water to convert water to water vapor at boiling point temperature. Both of these are high because of the need to provide energy to overcome the hydrogen bond attraction. As latent heat is added to change the phase, there is no change in temperature. In the reverse processes, the same amounts of latent heat that were added are released when water vapor condenses and when water freezes. The heat capacity of water is the amount of heat required to raise the temperature of 1 g of liquid water by 1°C and is measured in cal•g^{-1}•°C^{-1} (J•g^{-1}•°C^{-1}). The same amount of heat that was absorbed during warming is released as water cools.

After the ice has reached the melting point temperature, the temperature of the solid-liquid (ice-water) mixture does not increase again until an additional 80 calories (334 J) of heat energy has been added for each gram of ice (**Fig. 5-10**). That is, as heat is added, the ice is progressively converted to water. Heat added to a substance that does not raise its temperature but instead changes its state (solid to liquid or liquid to gas) is called "latent" heat. The heat added to melt a specified quantity of ice is the latent heat of fusion (melting). The term latent is used because the heat is stored in the molecules of the liquid water and is released when the water is refrozen (fused). Hence, the latent heat of fusion of ice is 80 cal•g^{-1} (334 J•g^{-1}).

If we continue to add heat after all the ice is converted to water, the heat is again taken up as sensible heat and the temperature of the water rises. The heat capacity of water is about twice that of ice, or 1 cal•g^{-1}•°C^{-1}. That is, 1 cal (4.2 J) of heat energy must be added to raise the temperature of 1 g of water by 1°C.

After 100 cal•g–1 (418 J•g–1) of heat energy has been added to the water (starting as melted ice water at 0°C), the water

temperature reaches the boiling point. With the continued addition of heat, the temperature does not rise again until an additional 540 cal•g^{-1} (2260 J•g^{-1}) of heat energy has been added and all the water has been converted to a gas. Hence, the latent heat of vaporization of water is 540 cal•g^{-1} (2260 J•g^{-1}). Finally, if we add more heat to the gaseous water (water vapor), its temperature rises by about 1°C for each additional 0.5 cal•g^{-1} (2.1 J•g^{-1}), so the heat capacity of water vapor is about 0.5 cal•g^{-1}•°C^{-1} (2.1 J•g^{-1}•°C^{-1}).

The latent heats and heat capacity of water are very high primarily because hydrogen bonds are stronger than van der Waals forces. More heat energy is needed to overcome the attraction between molecules in solid, liquid, and gaseous forms of water than in substances whose molecules are bound only by van der Waals forces. Consequently, the heat capacity of liquid water is the highest of all liquids other than liquid ammonia and is higher than that of all solids. The latent heat of fusion of water is the highest of all substances other than ammonia; and the latent heat of vaporization of water is the highest of all known substances.

Implications of the High Heat Capacity and Latent Heats of Water

The anomalously high heat capacity and latent heats of water have many implications. In our everyday experience, we rely on the high latent heat of fusion to keep our iced drinks cold. When we add ice cubes to a cold drink, the drink remains cold for a long time, until all the ice has melted. The drink, a mixture of ice and water, remains cold as it gains heat from its surroundings, until it has absorbed 80 calories (334 J) for each gram of ice, converting the ice to water. Subsequently, when all the ice has melted, the drink warms much faster because each calorie (each 4.2 J) gained can raise the temperature of 1 g of water by 1°C.

The high heat capacity of water is illustrated by the behavior of hot drinks. A hot drink cools much more slowly than, for example, an empty cup or glass taken out of a hot dishwasher. The reason is that more heat must be lost per unit weight of water than of any other substance (other than liquid ammonia) to lower its temperature by a given number of degrees.

The high heat capacity of water allows large amounts of heat energy from the sun to be stored in ocean waters without causing much of a temperature change. The heat is released to the atmosphere when atmospheric temperatures fall. Because water has this heat-buffering capability, coastal locations have milder climates than inland locations (**Chap. 7**). The heat-buffering capacity of the oceans and the transfer of latent heat between ice, oceans, and atmosphere are important to global climatic control and to studies of the effects of releases of gases that contribute to the **greenhouse effect** (**Chap. 7**).

In the polar ocean and coastal regions, water's high latent heat of fusion acts to control the air and water temperatures in much the same way that ice cubes cool an iced drink. During winter, latent heat is released to the atmosphere as water freezes to form more sea ice. Conversely, in summer, heat added to the polar regions is used to melt the ice. Because the heat lost or gained is latent heat, not sensible heat, the ocean surface water and sea ice remain at or close to the freezing point throughout the year. Because air temperatures are partially controlled by ocean temperatures, the annual climatic temperature ranges in polar ocean regions are small (**Chap. 7, CC5**).

Evaporation

Liquids, such as water, can be converted to a gas at temperatures below their boiling point by a process known as "evaporation." For example, puddles of cold water can evaporate from the ground after a rainstorm. Although the average energy level of water molecules remains the same unless there is a change of temperature, the energy level of individual molecules varies around this average. Some molecules that temporarily possess higher amounts of energy can overcome their hydrogen bonds and escape completely from the liquid phase (evaporate).

Water evaporating from the oceans carries its latent heat of vaporization into the atmosphere and releases it when the water subsequently condenses as rain. Because the latent heat of vaporization is high, water can transport large quantities of heat energy from ocean to atmosphere, where that energy can be redistributed geographically. For example, areas of the Earth where evaporation is slow and precipitation is high tend to be warmed, while areas where evaporation is fast tend to be cooled (**Chap. 7, CC5**).

Water's high latent heat of vaporization facilitates cooking and can be observed in a kitchen. If we heat a pot of water on a stove, the water can be brought to a boil fairly quickly, but then, even if we do not change the heat setting, it takes much longer for the pot to boil dry. It takes only 100 cal•g^{-1} (418 J•g^{-1}) to heat water from its freezing point to its boiling point, but more than five times as much, 540 cal•g^{-1} (2260 J•g^{-1}), to vaporize or convert the water to steam. Water's high latent heat of vaporization is also apparent when we bake food for substantial lengths of time at temperatures well above 100°C and find that the food is still moist.

Because evaporating molecules gain the energy to escape by colliding with molecules that remain behind in the liquid, the average energy level (temperature) of the water molecules that remain behind is reduced. This is why we feel colder when we first climb out of a swimming pool or shower. While we are wet, water molecules evaporate using heat energy gained from our skin, and the air and leaving behind colder skin. After toweling off, we have less water on our skin and the amount of evaporation is reduced. Another example of this phenomenon is the wind chill factor used in **weather** forecasts. Increasing the speed of wind blowing across a wet or damp surface, such as the skin, increases the evaporation rate and thus increases the rate of surface cooling.

At temperatures below the boiling point, the average water molecule has less energy than it does at the boiling point. Therefore, more heat must be supplied to evaporate each molecule of water than is needed to vaporize it at the boiling point, and the latent heat of vaporization increases with decreasing temperature. For example, the latent heat of vaporization at 20°C is 585 cal•g^{-1} (2449 J•g^{-1}) which is 45 cal•g^{-1} (189 J•g^{-1}) higher than at 100°C.

This page titled 5.7: Heat Properties of Water is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater

Many of the movements of water masses in the oceans are driven by differences in density. Solid objects that have higher density than water sink, and those that have lower density rise and float. Liquid water can also rise or sink if its density is different from that of the surrounding water (**CC1, CC3**). Water density is controlled by changes in pressure, temperature (**CC6**), and concentration of dissolved constituents (salinity).

Pressure

Because water molecules in a liquid form can be forced together only slightly as pressure increases, seawater is virtually incompressible, and its volume decreases and density increases only very slightly with pressure. Therefore, pressure changes are not as important to controlling water density in the oceans as changes in temperature and salinity are. However, the pressure at the greatest depths in the oceans is more than 1000 times as great as atmospheric pressure. This pressure change causes seawater density to be approximately 2% higher than it is in shallow water at the same temperature and salinity. Some research studies must take this small difference into account. Like all other gases, water vapor is compressible, and its density varies substantially with pressure.

Temperature

Increasing temperature adds energy that enables the molecules of solids and liquids to vibrate, rotate, and/or translate more vigorously. Thus, the average distance between the molecules generally increases. As the same number of molecules occupies a larger volume (the material expands), increasing the temperature causes the density to decrease. Ice is no exception to this rule. However, pure water (but not seawater, as discussed later) behaves anomalously, because liquid water has a density maximum at 4°C (rounded from precisely 3.98°C). Between 0°C and 4°C, water density actually increases with increasing temperature (**Fig. 5-11**). In the rest of liquid water's temperature range (4°C to 100°C), pure water behaves normally and density decreases with increasing temperature.

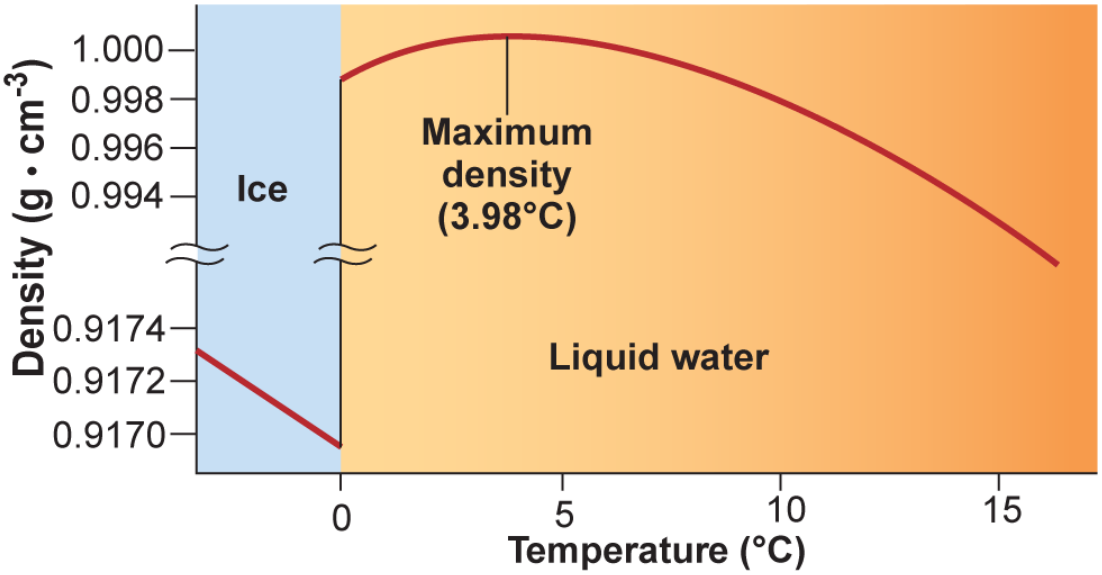


Figure 5-11. Density of pure water plotted against temperature. Water with no dissolved salts has a maximum density at 3.98°C. At lower temperatures, water density decreases until the freezing point is reached. At the freezing point, there is a discontinuity in the density scale. The density of ice is much less than that of liquid water because of the open structure of the ice crystal lattice. When solid ice and liquid water occur together at 0°C, the ice floats on the water because its density is less than that of the water.

The reason for the anomalous effect of temperature on water density is the hydrogen bond. Water molecules form clusters in

which the molecules are arranged in a lattice-like structure. The atoms in the cluster are held in place by hydrogen bonds (**Fig. 5-12**). The structure is similar to that of ice, and the molecules of water in a cluster occupy a larger volume than molecules that are not clustered. The ordered clusters remain together for only a few ten-millionths of a second, but they are continuously forming, breaking, and re-forming. Both the number of clusters present at any time and the number of molecules in each cluster increase as the temperature decreases (more unbroken hydrogen bonds are present). Because clustered molecules occupy a greater volume than unclustered molecules, an increase in the number of clusters and in the number of molecules per cluster decreases the density of the water. Above 4°C, there are too few clusters to counteract the normal temperature effect on density. Below 4°C (actually 3.98°C), however, clustering increases fast enough that the decrease in density caused by the clustering is faster than the increase caused by the normal temperature effect.

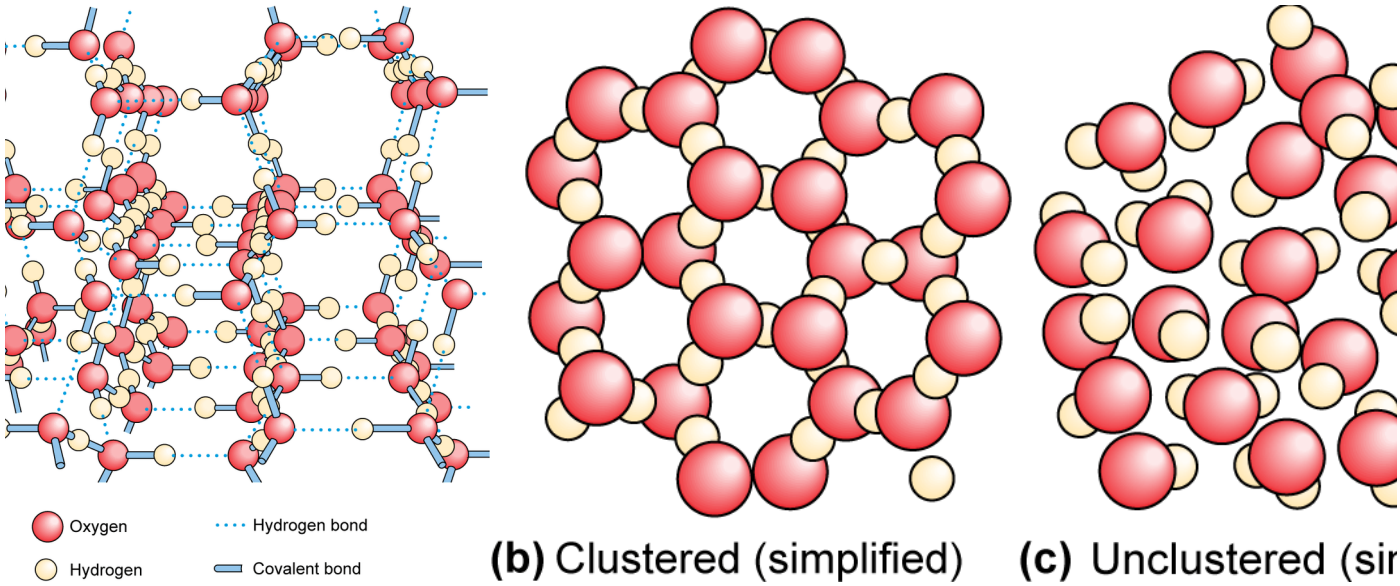


Figure 5-12. The hydrogen bond plays a major role in the properties of ice and water. (a) Molecules in the ice crystal lattice are bonded to each other by hydrogen bonds and arranged in a hexagonal lattice. The resulting structure is very open, which explains the low density of ice. (b) Water molecules constantly form clusters that are temporarily held together by hydrogen bonds. The clusters consist of chains or networks of water molecules arranged in hexagons. The representation here shows this hexagonal form but, for simplicity, does not show the hydrogen bonds. (c) When water molecules are unclustered, the molecules are closer to each other than they are in clusters. Clusters are more likely to form and may persist longer at low temperatures, which is why water's density decreases below 3.98°C.

Dissolved Salts and Density

Salts dissolved in water increase water density for several reasons. First, the ions or molecules of most substances dissolved in seawater have a higher density than water molecules. Dissolved substances also reduce the clustering of the water molecules, further increasing the density, particularly at temperatures near the freezing point.

Combined Effects of Salinity and Temperature

The density of ocean waters is determined primarily by salinity and temperature. **Figure 5-13a** shows the relationships among the salinity, temperature, and density of seawater. Raising the temperature of freshwater from 4°C to 30°C (the range between the temperature of maximum density and the highest temperature generally found in ocean surface waters) decreases its density by about 0.0043 (from 1.0000 to 0.9957), or about 0.4%. At a constant temperature, changing the salinity from 0 to 40 (approximately the range of salinity in surface waters) changes the density by about 0.035, or about 3.5%. These observations suggest that salinity is more important than temperature as a determinant of density. This is often true in rivers and estuaries where the water has a wide range of salinity, but the range of salinity in open-ocean waters is much smaller. In fact, 99% of all ocean water has salinity between 33 and 37, and 75% has salinity between 34 and 35 (**Fig. 5-14**). Similarly, 75% of ocean water has a temperature between 0°C and 5°C, and the rest has a much wider temperature range, between about –3°C and 30°C.

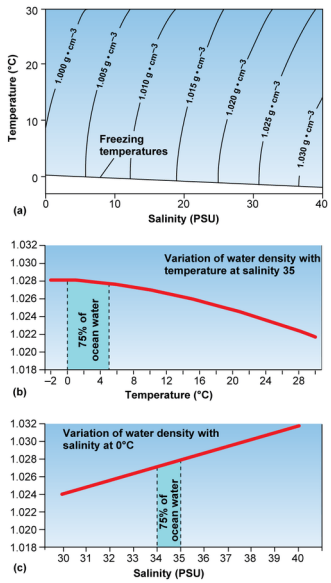


Figure 5-13. The effects of temperature and salinity on the density of water. (a) The relationships among salinity, temperature, and dens complex. The effects of temperature and salinity individually are more obvious if we examine (b) the relationship of temperature (–3°C to to density at a fixed salinity, and (c) the relationship of salinity (30 to 40 PSU) to density at fixed temperature. The shaded sections o plots in (b) and (c) represent the narrow ranges of salinity and temperature present in most ocean water.

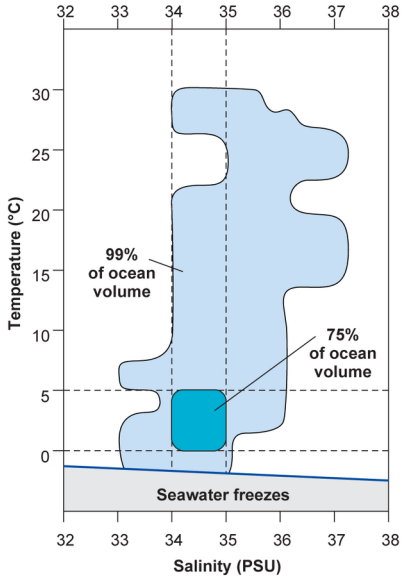


Figure 5-14. Temperature and salinity characteristics of ocean waters. Ninety-nine percent of all ocean water has salinity and temperature within the range shown by the larger shaded area. Seventy-five percent of all ocean water has the much narrower range of salinity and temperature within the inner darker-shaded area.

With the exception of water discharged by **hydrothermal vents**, the highest temperatures in the oceans are in surface waters in tropical regions. **Figure 5-13b,c** relates salinity, temperature, and seawater density, and shows that, in most of the ocean, temperature and salinity are of approximately equal importance in determining ocean water density. However, their relative importance varies with location and depth. For example, temperature changes are more important to density variations in the tropical water column, where salinity variation is relatively small but temperature variation with depth is relatively large. In contrast, salinity is more important in some high-**latitude** regions, where salinity variations are relatively large as a result of high volumes of freshwater **runoff** and the formation and melting of ice, but temperatures are generally uniform and near the freezing point.

This page titled 5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.9: Ice Formation

When surface water is cooled to the freezing point, the behavior of ice and water in lakes and oceans is determined by two factors: the relative density of ice and water and the variations of water density with salinity at temperatures near the freezing point.

Dissolved Salts and Freezing

Increasing the salinity lowers the freezing point of water (**Fig. 5-15**). This is why salt is used to de-ice roads. Increasing the concentration of dissolved salts also inhibits the clustering of water molecules and therefore lowers the temperature at which the density maximum occurs (**Fig. 5-15**). As salinity increases, the temperature of the water density maximum decreases more rapidly than the freezing point temperature. Therefore, as salinity increases, the difference between the temperature of maximum density (4°C in pure water) and the freezing point are the same temperature (–1.36°C). At any salinity above 24.7, the freezing point is reached before a density maximum occurs. Since open-ocean water generally has a salinity above 24.7, seawater does not have an anomalous density maximum.

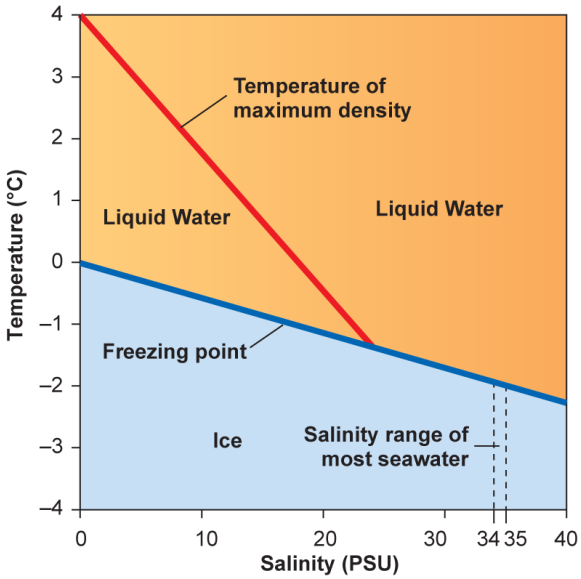


Figure 5-15. Increasing the concentration of dissolved salts in water reduces both the freezing point and the temperature at which the anomalous density maximum occurs. However, the temperature of maximum density is reduced more rapidly than the freezing point as salt concentration increases. Consequently, seawater (99% of which has salinity greater than 33) has no anomalous density maximum.

Because freshwater has a maximum density at 4°C and seawater has no maximum density, lakes and rivers behave differently from the oceans when freezing. As winter begins, the surface water of a freshwater lake cools and its density increases. This cold, dense surface water sinks and displaces bottom water upward to be cooled in its turn. This **convection** process is called "overturning" (**CC3**). Convection continues until all the water in the lake is 4°C, the maximum density. As the surface layer of water cools below 4°C, its density no longer increases. Instead, it actually decreases. Therefore, this water does not sink. The water in this surface layer continues to cool until it reaches 0°C and then the water freezes. The resulting ice has a lower density than water and floats. With further cooling, the lake water continues to lose heat, but because of the density maximum there is no convection, so the heat is lost by **conduction** through the overlying ice layer. As heat continues to be lost, the water immediately under the ice cools to 0°C and freezes. Therefore, the ice layer grows progressively thicker from underneath. Ice is not a very good conductor of heat and a large amount of heat must be lost to convert water to ice because of the high latent heat of fusion. Consequently, the ice layer on a lake surface will grow to a thickness of at most only a meter or two, depending on the length and severity of the winter. If the lake is deep enough, a pool of liquid water at 4°C will remain below the ice of the lake throughout the winter. Many freshwater organisms that cannot tolerate being frozen survive the winter in such pools of water.

In contrast to freshwater, as seawater (salinity greater than 24.7) cools, its density increases continuously until it reaches its freezing point of about –2°C. Cooling water sinks away from the surface until the whole upper layer approaches freezing and ice forms. This is why most of the deep water in the oceans is cold, generally below 4°C (**Chap. 8**). As seawater freezes, most of

the dissolved salts are excluded from the ice and remain in the unfrozen water. Therefore, salinity increases in the remaining water, and as a result, the freezing point decreases (**Fig. 5-15**). If freezing occurs quickly, or if winds break up the surface ice and splash water onto the ice surface, isolated pockets or pools of liquid water can be present within the ice. The water in the pockets has high salinity, and its temperature may be several degrees below the freezing point of seawater with normal salinity. Some microscopic algae, bacteria, and animals survive the Arctic and Antarctic winters in such isolated pockets of liquid inside the sea ice.

Density of Ice

Most substances have higher density in the solid phase than in the liquid phase. Water is an exception to this rule. Ice at 0°C has a density of 0.917 g•cm^{-3}, about 8% less than the density of water of the same temperature. Hence, water expands when it freezes. The reason is that the ice crystal has an open lattice structure, and fewer water molecules are packed into the volume that they would occupy as water (**Fig. 5-12**). In the winter, expansion of water as it freezes bursts water pipes, breaks up rocks, and causes potholes in street surfaces when small cracks are filled with water and then freeze.

Ice floats because it is less dense than water. If ice did not float, it would sink to the bottoms of ponds and lakes as they froze, and surface water would continue to freeze and sink until no liquid water was left. Hence, aquatic organisms would have no liquid water refuge below the ice in which to survive the winter. In addition, sea ice formed in the polar regions would sink to the floor of the oceans and accumulate. Because there would be no way to bring the ice to the surface to remelt it, the oceans would always have a warm-water surface layer over a solid ice layer extending to the seafloor.

This page titled 5.9: Ice Formation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.10: Surface Tension and Viscosity

As we have seen, hydrogen bonding plays an important role in the thermal properties of water. Hydrogen bonding also contributes to the high **surface tension** of water and its relatively high **viscosity**. Surface tension and viscosity are major factors in determining the behavior of waves and controlling the processes by which water molecules are transported across biological membranes.

Surface Tension

Molecules within a liquid are subject to attractive forces that pull them toward all adjacent molecules. Moreover, molecules at the surface of a liquid are attracted much more strongly to their neighboring liquid molecules than to the more remote gas molecules above the surface (**Fig. 5-16**). Therefore, all molecules on the surface are pulled toward the interior of the liquid. The attractive force tends to minimize the number of molecules at the surface and creates a surface tension that pulls the liquid surface into a configuration with the minimum possible surface area. Because of the strong hydrogen bond attraction between water molecules, the surface tension of water is higher than that of any known liquid other than mercury.

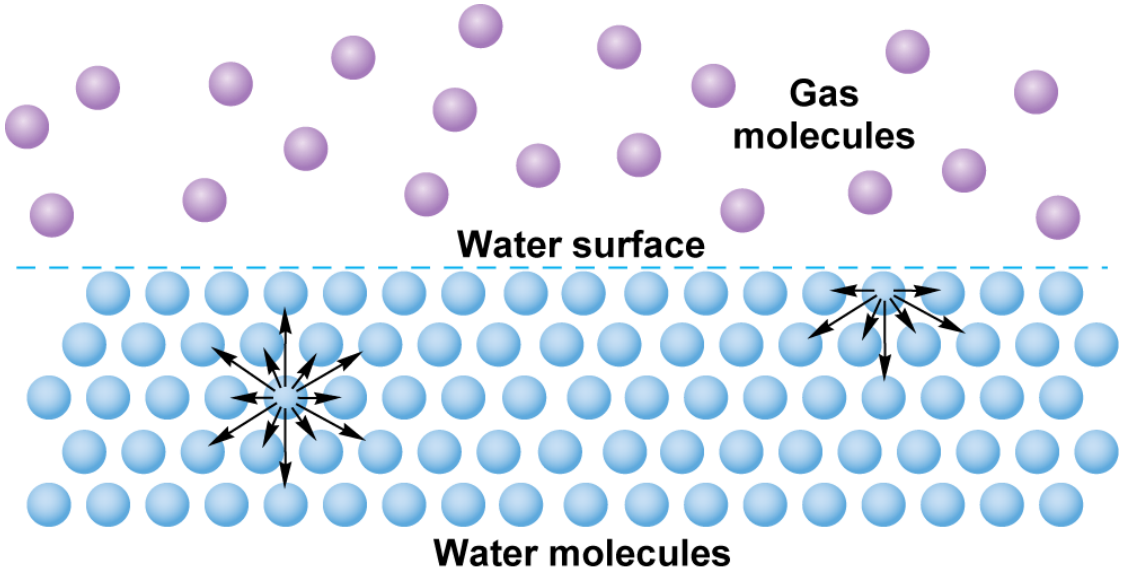


Figure 5-16. All molecules in a liquid are attracted to each other by van der Waals forces. Liquids have surface tension because the molecules at the surface are attracted more strongly to other molecules of the liquid than to the gas molecules above, which are more dispersed and thus, on average, farther away. Molecules at the surface have a net inward attraction that makes the surface behave somewhat like the rubber of an inflated balloon. The surface tension of water is particularly high because the water molecules are attracted to each other by both van der Waals forces and the much stronger hydrogen bond.

Surface tension tends to pull water droplets into a spherical shape because a sphere has the lowest possible ratio of surface area to volume. We can see the approximately spherical shape in beads of water on a newly waxed automobile, and in drips from a leaky faucet.

If we pour water into one side of a saucepan (or ocean), gravity distributes the water so that all parts of the water surface are at the same level. Similarly, if we disturb a surface—say, by sloshing the saucepan—the water will quickly return to its original state, in which its surface is flat and horizontal. Surface tension "pulls" the surface into the minimum area possible, which in the saucepan is a flat horizontal plane. For this reason, the high surface tension of water critically affects the generation and dissipation of small waves on the ocean surface (**Chap. 8**).

If an object is to break through a water surface, it must be heavy enough to push aside the water molecules and break the surface tension. Any object with a density greater than that of water can be made to float on the water surface if its weight per unit of water surface area is too low to break the surface tension. For example, a steel sewing needle can be positioned to float on the water surface. Similarly, insects such as the water strider can walk on water because of its high surface tension.

The high surface tension of water has significant effects on the formation and behavior of gas bubbles and spray, which are created in the oceans primarily by breaking waves. Once they have formed in water, small gas bubbles cannot easily break through the ocean surface to escape to the atmosphere. When bubbles do break through the surface, a large number of small water droplets are ejected into the air. Because of their high surface tension, the droplets do not readily combine into larger drops that would fall back to the water surface faster. The high surface tension of water helps to retain gas bubbles in the water and water droplets in the air, enhancing the efficiency of gas transfer between oceans and atmosphere. Because small droplets have large surface areas in relation to their volume, evaporation from the oceans is also enhanced.

Because of the high surface tension of water, a very thin surface water layer called the **surface microlayer**, just a few molecules thick, is always present on oceans and lakes. Many small aquatic organisms, such as mosquito **larvae**, can anchor themselves from below on this layer.

Viscosity

Viscosity is a measure of a liquid's internal resistance to flow or the resistance of the liquid to the movement of an object through it. A low-viscosity liquid flows easily, whereas a high-viscosity, or "viscous," liquid flows slowly. Water has a lower viscosity than many liquids, such as honey or motor oil. However, the viscosity of water is high enough to provide substantial resistance to minute organisms that swim or sink in water. To these tiny organisms, moving through water is equivalent to our swimming through ketchup. In fact, many microscopic organisms rely on the viscosity of water to prevent them from sinking out of the upper layers of ocean water where they live.

Water viscosity varies with salinity and temperature. Viscosity increases as temperature decreases (slightly less than 1% per 1°C within the normal range of ocean water temperatures) because more structured water molecule clusters are present at lower temperatures. The clusters do not move out of the way as easily as individual molecules do. Hence, microscopic organisms in warm tropical waters have more difficulty in avoiding sinking than organisms in cold polar waters do. Many tropical microorganisms have compensated by evolving elongated spines or frilled appendages that increase "drag" as they move through the water. Related polar species generally do not have these ornate features.

Viscosity also increases as salinity increases, because the component ions of the dissolved salts are very effective in surrounding themselves with water molecules to form clusters that resemble water molecule clusters (**Fig. 5-4**). However, the change in viscosity caused by dissolved salts is small within the range of salinity in freshwater and the oceans. Viscosity increases by less than 1% from freshwater (0 psu) to 35 psu seawater at the same temperature.

This page titled 5.10: Surface Tension and Viscosity is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.11: Transmission of Light and Other Electromagnetic Radiation

Light is a form of electromagnetic radiation emitted by the sun and stars. However, visible light occupies only a small segment of the very wide electromagnetic radiation spectrum (Fig. 5-17). Ultraviolet light, X-rays, and gamma rays have progressively shorter **wavelengths** than visible light, whereas infrared light, microwaves, and radio waves have progressively longer wavelengths.

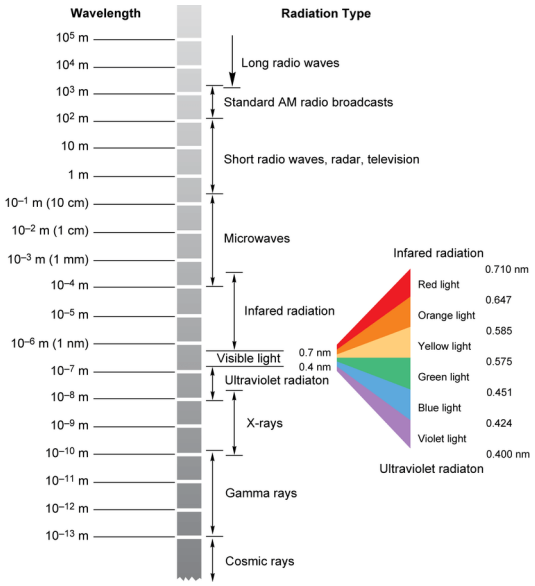


Figure 5-17. The electromagnetic radiation spectrum.

Absorption

Water absorbs electromagnetic radiation, but the depth of water penetrated varies with the intensity of the radiation and with wavelength. At most wavelengths, absorption is so effective that, even at very high intensity, the radiation can penetrate only a few centimeters or meters of water. Absorption is less effective within the narrow range of wavelengths of visible light and for very long-wavelength radio waves. However, even at the wavelengths of lowest absorption, the most intense radiation (e.g., lasers) cannot penetrate more than a few tens or hundreds of meters of water. For this reason, oceanographers cannot use electromagnetic radiation, such as radar, radio waves, or laser light, to “see” through the oceans in the same way that radar can see through clouds to the Earth’s surface or even the surface of Venus.

Because radio waves and other forms of electromagnetic radiation do not penetrate far through water, they also cannot be used for underwater communication. For example, long-range communication with submarines is a considerable problem. Very low-**frequency** radio waves can penetrate a few tens of meters into the ocean and are used to communicate with submarines near the surface. Because intense visible light can penetrate up to several hundred meters below the ocean surface, satellite-mounted lasers also are used to communicate with submarines at shallow depths. The inability of electromagnetic radiation, such as radar, to penetrate ocean water is the principal reason why submarines are among the best places to hide strategic missiles. Detection of submarines relies primarily on sound waves, which are transmitted easily through ocean water.

The greater transmissibility of visible light through water (compared to other electromagnetic wavelengths) is critically important to life in the oceans (**Chaps. 12, 13**). Like life on the land, most non-**microbia**l life in the oceans depends on algae and bacteria that need light to grow by **phototrophy** (**CC14**). Since light energy decreases rapidly with depth in the oceans, phototrophic activity also decreases with depth until the light energy is insufficient to sustain growth and reproduction. To define this depth, ocean scientists use the depth at which the light intensity is reduced to 1% of that at the ocean surface. Water above this level is referred to as the **photic zone** and water below as the **aphotic zone**.

Some wavelengths of visible light are better absorbed by water than others (Fig. 5-18a). Wavelengths at the red end of the

visible spectrum are the most rapidly absorbed, and violet wavelengths also are absorbed relatively quickly; blue and green light penetrate farthest. Outside the visible light spectrum, infrared radiation (wavelengths longer than red light) and ultraviolet radiation (wavelengths shorter than violet light) are strongly absorbed by water.

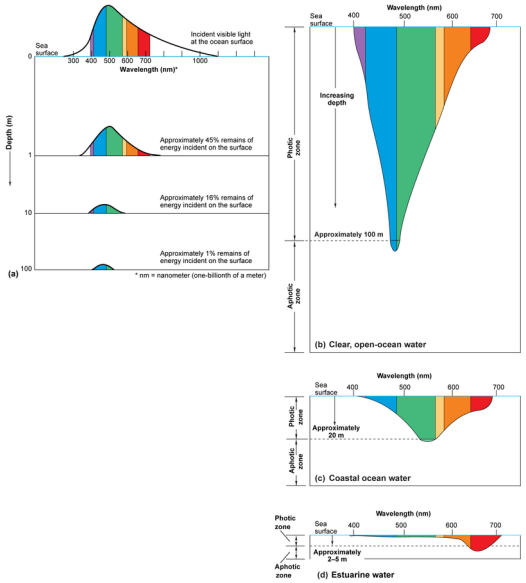


Figure 5-18. Light penetration and absorption in ocean waters. (a) Absorption of electromagnetic energy by ocean water. Note that energy in the infrared (about 700–1000 nm) and ultraviolet (about 300–400 nm) wavelengths is absorbed very rapidly compared to energy in the visible light wavelengths (400–680 nm). Also, energy at longer wavelengths of visible light (red/orange) is absorbed more quickly than energy at shorter (blue/green) wavelengths. The remaining parts of the figure show the depth at which light of visible wavelengths is reduced to less than 1% of the intensity at the surface, the demarcation between the photic and aphotic zones in (b) clear open-ocean water, (c) coastal-ocean water, and (d) estuarine water. Note that the photic zone depth is much reduced in coastal and estuarine waters. Below the photic zone, light is insufficient for phytoplankton to produce enough organic matter by photosynthesis to survive. Note also that the wavelength range of light reaching the lower part of the photic zone moves from the blue/green part of the spectrum toward the red end of the spectrum in coastal and estuarine waters. This is the reason why open-ocean waters usually appear blue while coastal and estuarine waters do not.

As light penetrates the ocean, it is absorbed not only by water molecules, but also by suspended particles. Accordingly, light penetrates deeper in clear ocean waters than in high-turbidity coastal waters with high **suspended sediment** loads or high plankton concentrations (**Fig. 5-18b-d**).

Scattering and Reflection

We are all familiar with the idea that light is reflected by smooth “shiny” surfaces such as the silvered surface of a mirror. Light is also reflected when it encounters a rough surface. However, individual light rays meet different parts of the rough surface at different angles, and as a result, they are reflected in different directions. When this occurs, the light is said to be **scattered**.

As light penetrates ocean water, scattering occurs when some of the photons of light bounce off water molecules, molecules of dissolved substances, or suspended particles. The light scatters in all directions, but some is directed upward and is said to be **backscattered**. In very clear ocean waters with few suspended particles or plankton, most of the red and yellow light is absorbed rather than scattered (**Fig. 5-18b-d**). Light of blue wavelengths has a greater probability of being backscattered because blue light travels through a greater length of the water column. Therefore, most of the backscattered light in clear ocean waters is of blue wavelengths, and it is this backscattered light that gives the ocean its deep blue color.

In waters with more suspended particles or plankton, there is a higher probability of backscattering of all visible wavelengths. Hence, in more **turbid** waters, the backscattered light comprises a much wider range of wavelengths, and the water appears not blue, but green or brownish. The specific color is determined by the nature of the suspended particles. All particles absorb light more effectively at some wavelengths, and their color is determined by which wavelengths of the light they do not absorb. Light of these nonabsorbed wavelengths is scattered and, therefore, constitutes the wavelengths we see when looking at the particles. Most particles in the oceans are either green, such as many **phytoplankton** cells, or brown, such as sand and other mineral particles. Ocean waters with large phytoplankton populations (**Chap. 12**) tend to appear greenish, whereas coastal waters with large loads of suspended sediment (mineral grains) appear muddy brown. Therefore, color variations of the ocean surface indicate the quantity of particulate material and phytoplankton in the near-surface water. Ocean surface color measured by sophisticated

sensors aboard satellites (**Fig. 5-19**) is used to investigate distributions of the living and nonliving particles in the oceans (**Chaps. 6, 12**).

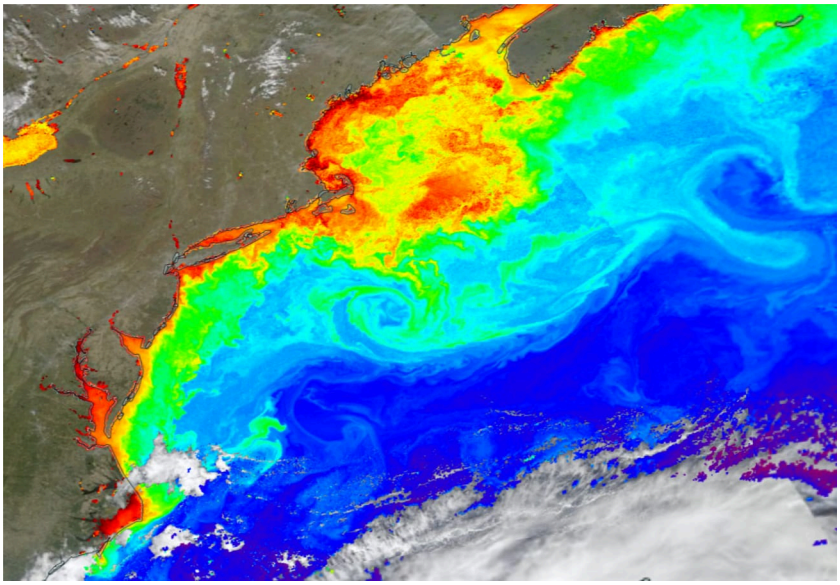


Figure 5-19. This image of the Northwest Atlantic Ocean shows data obtained from a satellite-mounted instrument called the Ocean Color Instrument, which measures the intensity of light reflected or backscattered by the Earth's surface. The data are from a narrow wavelength band, in which the intensity of the light received by the scanner from the ocean surface increases with increasing concentrations of chlorophyll-containing phytoplankton in the photic zone. Intensities of light in this wavelength band are depicted by the false colors in this image. Red colors show the areas of highest chlorophyll and phytoplankton concentration; and progressively lower concentrations are depicted by the orange, yellow, green, and blue shades. Note that the highest concentrations of chlorophyll are found in the coastal region and that the image reveals the complex patterns associated with different water masses.

The underwater world seen in movies or video is a blaze of colored creatures. However, **scuba** divers see such colors only in very shallow water. Just a few meters below the surface, the red and yellow wavelengths of natural sunlight are absorbed, so bright red fishes appear black and most objects appear blue or bluish green. The true colors are revealed only when lights with a full spectrum of visible wavelengths are shone on the marine life. For this reason, underwater photographers or videographers must carry powerful lights.

By noting changes in the visibility and lighting of the water, scuba divers can observe scattering and absorption of light by particles. In turbid water with more particles, daylight does not penetrate very far. In addition, at shallow levels to which some natural light penetrates, horizontal visibility is reduced. Just as daylight is blocked from penetrating vertically into the water by scattering and absorption, light traveling from an object horizontal to the diver is absorbed and scattered. If the absorption and scattering are substantial, the light from distant objects does not reach the diver and the objects cannot be seen. When light is scattered equally in all directions, a scuba diver may not be able to visually distinguish between up and down.

Refraction

The speed of light in seawater is less than that in air. As light waves pass from air into water or from water into air, they are **refracted** (their direction is changed because of the change in speed) (**Fig. 5-20a**). An easy way to see refraction is to sight down the length of a ruler and dip the end of the ruler into a tumbler of water (**Fig. 5-20b**). The tip of the ruler appears to be bent upward as it passes into the water. Refraction causes fishes in an aquarium to appear bigger and closer than they really are. Similarly, fishes and other objects seen through a scuba mask appear larger and closer than they are.

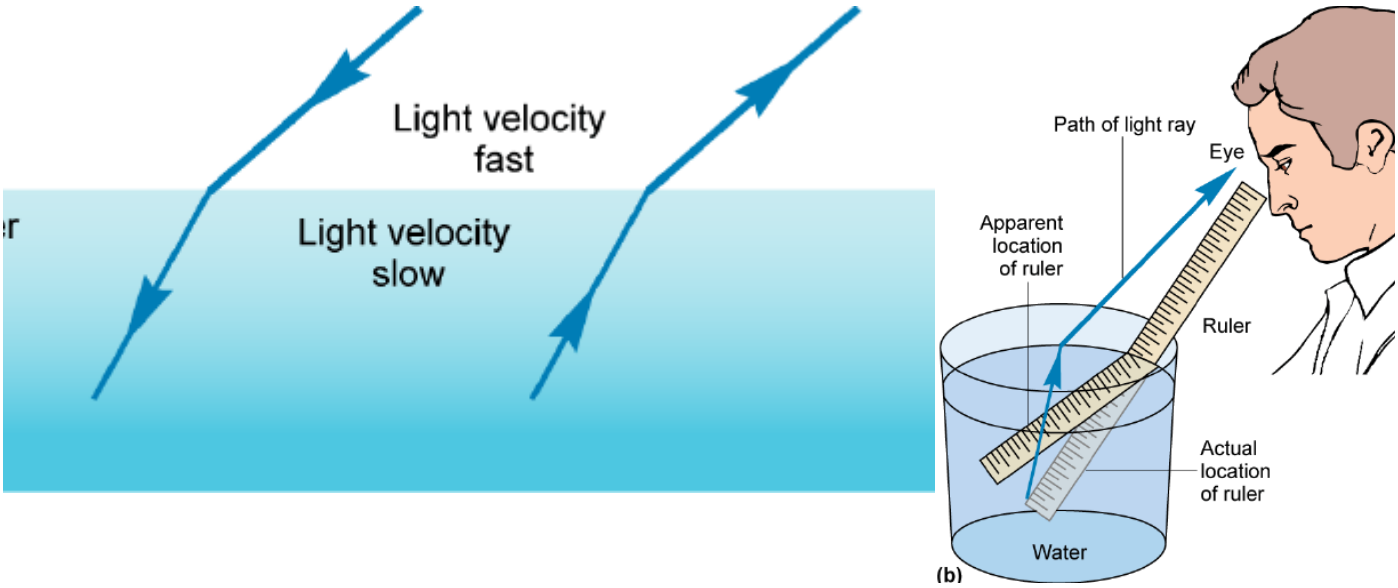


Figure 5-20. (a) Light rays are bent, or refracted, when they pass between air and water. (b) Inserting a ruler or stick through a water surface at an angle illustrates the distortion in the way refraction makes objects appear. The ruler looks as if it is bent at the point it enters the water.

Refraction between air and water increases as salinity increases. Therefore, measurements of refraction are often used to determine approximate salinity.

This page titled 5.11: Transmission of Light and Other Electromagnetic Radiation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.12: Transmission of Sound

Electromagnetic radiation travels both through a vacuum and through substances such as air and water. Sound is fundamentally different because it cannot be transmitted through a vacuum. The sound effects in space movies are pure fantasy, because sound cannot travel across the vacuum of space. Sound is transmitted as a vibration in which adjacent molecules are compressed in sequence. One molecule is pushed into the next, increasing the "pressure" between the two molecules. The second molecule pushes on a third, increasing the pressure between the second and third molecules and relieving the pressure between the first two molecules, and so on. Hence, sound waves are pressure waves transmitted through gases, liquids, and solids. The pressure waves can range from very high frequencies, which the human ear cannot hear (ultrasound), to low frequencies below the audible range that we sometimes feel as vibrations from our surroundings.

Sound waves are absorbed, reflected, and scattered as they pass through water much less than electromagnetic radiation is. Therefore, sound travels much greater distances in water, and sound waves are the principal tools of communication and remote sensing for both oceanographers and marine animals.

Sound Velocity

The speed of sound is about four times greater in water than in air. In seawater, the speed of sound increases with increases in salinity, pressure, and temperature (Fig. 5-21a,b). However, the changes due to salinity variations are relatively small within the range of salinities in ocean waters. Thus, sound velocity generally decreases with depth in the upper layers of the oceans, where the temperature change is large, but then increases again with depth in the deep layers, where temperature variations are small and pressure changes are more important (Fig. 5-21c). Sound velocity in the oceans can be determined if the salinity, temperature, and depth (pressure) are known.

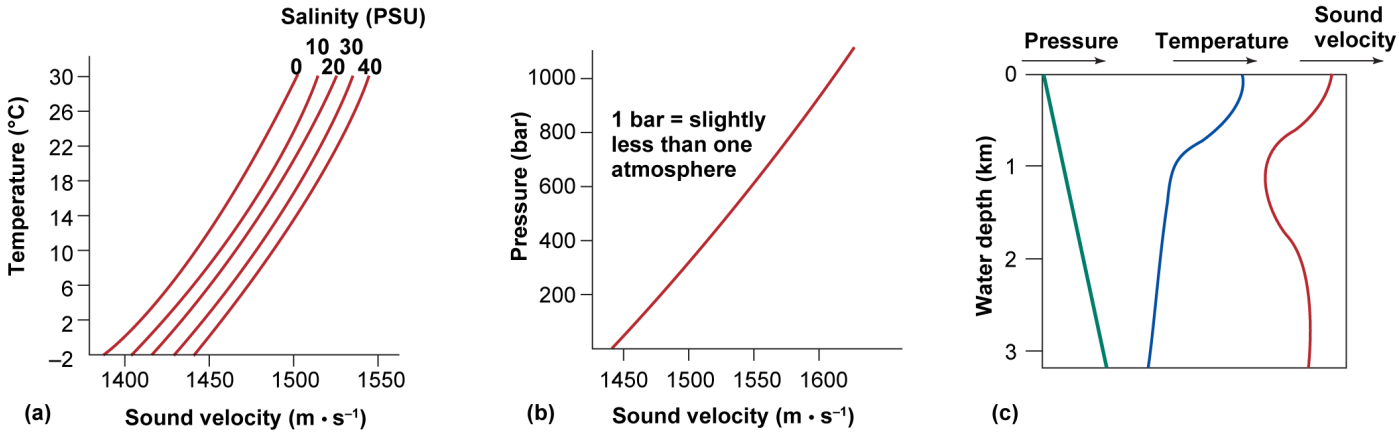


Figure 5-21. The velocity of sound in ocean water varies with temperature, pressure, and salinity. Sound velocity increases with (a) incre temperature or salinity and (b) increasing pressure. Salinity is comparatively less important in determining sound velocity than tempera pressure. (c) Variations of pressure and temperature with depth in the oceans produce changes in the sound velocity that are someti complex. Generally, sound velocity decreases with depth until it reaches a minimum at about 1000 m (1 km) and then increases as de increases.

Sonar

Sound velocity is important because sound waves are used to measure distances in water. Sound pulses sent through water bounce off objects and the seafloor. If the sound velocity at all points within the sound path is known and the time it takes for the sound to travel to and from an object is measured, the distance to the object (most often the seafloor) can be calculated. This is the principle of **sonar**.

Sonar systems send and receive sound pulses to measure ocean floor depth or to measure the distance and direction of objects that reflect sound, such as submarines. **Schools** of fish, concentrations of tiny animals called **zooplankton** on which fishes feed, and concentrations of suspended sediment can also be detected with sonar.

Different frequencies are employed for different applications. For example, low frequencies are used for deep-ocean **sounding**. Higher frequencies allow smaller particles (or animals) to be detected, but higher-frequency sound is more effectively absorbed by water. Therefore, higher frequencies are used in applications involving small objects in shallow water, such as finding fishes and tracing plumes of particles from river discharges or ocean dumping. Dolphins use their natural sonar to locate objects such as fishes, and they can tune the sound frequencies they use: low frequencies for long distance, and higher frequencies for more detailed echo "vision" at shorter distances.

Sound Refraction and the Sound Channel

Just as light is refracted as it passes between air and water, sound waves are refracted as they pass between air and water (**Fig. 5-22a**), or through water in which the sound velocity varies. At any specific ocean depth, the parameters that determine sound velocity are fairly constant over great distances. Hence, sound waves that travel horizontally are little affected by refraction. Sound waves that travel vertically are also not affected significantly by refraction, because they pass perpendicularly to the horizontal layers of varying sound velocity. In contrast, sound waves that pass through the ocean at any angle other than horizontal or vertical are refracted in complicated curving paths. The paths are determined by the initial direction of the sound waves and their varying velocity in the waters through which they pass (**Fig. 5-22b**).

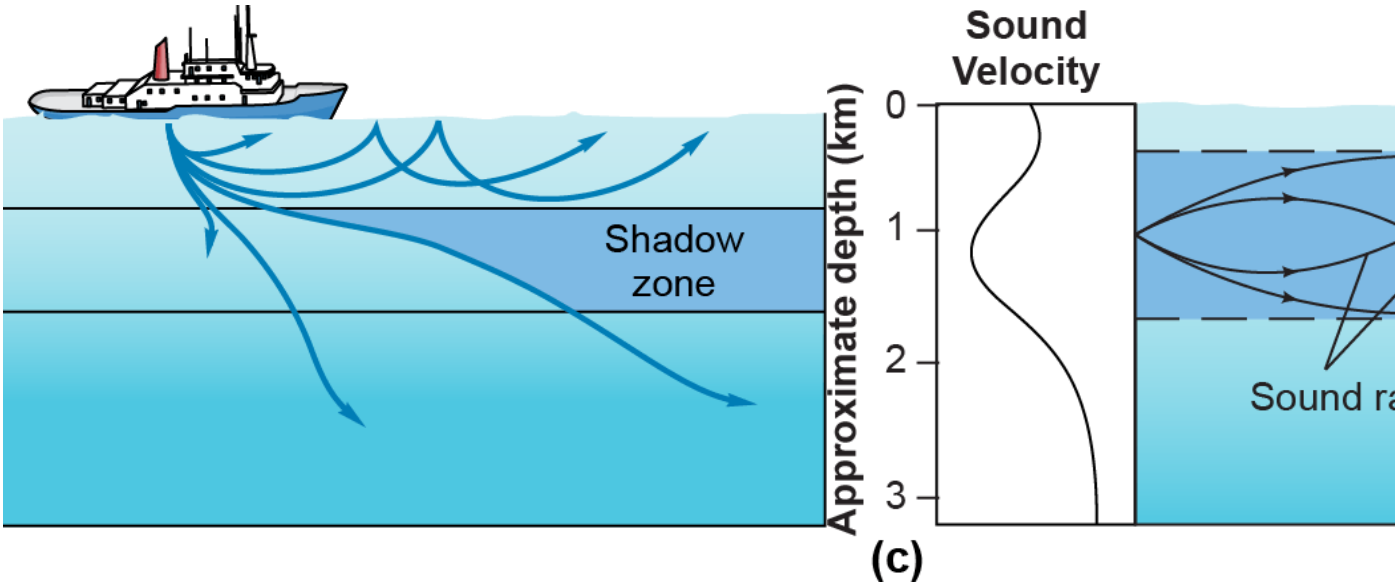


Figure 5-22. (a) Sound of certain wavelengths may be reflected when it encounters an interface between two water layers in which the sound velocity is different. Sound of other wavelengths can pass through such an interface and be refracted. (b) Sound can be reflected and refracted in complex ways. A sonar pulse sent from the surface can be refracted so that there is a shadow zone within the depth range around the sound velocity minimum. Because sound emitted by a surface vessel cannot enter the zone, submarines can hide in this shadow zone. (c) A sound generated at the depth of minimum sound velocity can be refracted back and forth within the low-velocity sound layer as it travels out from the source. This layer is called the "sound channel." Sound can be transmitted very long distances horizontally within the sound channel with little loss of intensity. (d) Outside the sound channel, sound spreads spherically, and its intensity is reduced in proportion to the square of the distance from the source. Within the sound channel, sound spreads cylindrically and loses intensity only in proportion to the distance from the source.

Two important consequences of sound refraction in the oceans are of interest. First, sonar sound pulses sent out by a vessel at any downward angle not close to the vertical are refracted away from a zone called the "shadow zone" (**Fig. 5-22b**). The zone starts at a depth where sound velocity is at a maximum, being slower at both shallower and deeper levels. Such a maximum is present within submarine operating depths in many parts of the oceans. Hence, submarines can hide from sonar detection if they can remain in the shadow zone. This is one reason why the navies of the world continuously monitor changes in the temperature and salinity of the upper few hundred meters of the ocean throughout their operating areas. Small changes in these characteristics can change the location and effectiveness of the shadow zone. Because sound travels in complicated curved paths between the sources and echoing objects, knowledge of salinity and temperature distributions is essential in determining the sound velocity distribution and estimating the location and depth of an object detected by sonar.

The second important consequence of sound refraction in the oceans is that sound emitted at or close to the depth of a sound velocity minimum is focused in what is called the "sound channel" (**Fig. 5-22c**). With the exception of sound waves traveling out

vertically (or nearly so) from such a source, all sound waves are refracted back and forth as they pass alternately up and down across the depth of the velocity minimum. A sound velocity minimum, and therefore a sound channel, is present at a depth of several hundred meters throughout most of the world's oceans. Sound normally spreads out spherically (in all directions), and the loss in intensity (attenuation) due to the spreading is proportional to the square of the distance from the source (**Fig. 5-22d**). In contrast, sound is focused in the sound channel and spreads cylindrically, so that the attenuation is proportional only to the distance from the source. Hence, sound can travel very large distances within the sound channel with relatively little attenuation. For example, sounds of small underwater explosions in the sound channel near Australia have been detected as far away as Bermuda in the North Atlantic Ocean.

Acoustic Thermometry

The vast distance over which sound can travel in the sound channel is the basis of an experiment, begun in 1991, that provides a sensitive means to determine average ocean temperature.

Determining whether the Earth's average surface and/or atmospheric temperature has recently changed or is now changing as a result of greenhouse effect enhancement is very difficult. Changes of only one-tenth of a degree per year in the global average temperature could drastically change the Earth's climate in just a few decades. However, because of the high temporal and **spatial** variability of temperatures in the ocean, land surface, and atmosphere, the best measurements currently possible from conventional thermometers or satellites are not accurate enough to identify changes in the global average temperature of less than about a tenth of a degree. In contrast, changes of only a few thousandths of a degree per year in the average temperature of ocean water can be identified by measurement of the travel times of sounds over distances of thousands of kilometers in the sound channel. This method of monitoring is known as "**acoustic** thermometry."

The principle of acoustic thermometry is simple and was proven with a preliminary experiment, the Acoustic Thermometry of Ocean Climate (ATOC) program, in 1991. A large underwater transducer (a sound producer like a loudspeaker) is lowered into the sound channel. In the preliminary experiment, the transducer was placed in the ocean near Heard Island in the southern Indian Ocean (**Fig. 5-23**). The transducer transmits low-frequency sound pulses into the sound channel. At several locations tens of thousands of kilometers away, sensitive **hydrophones** detect the incoming sound and precisely record its arrival time. The travel time measured to within a fraction of a second is used to compute the average speed of sound between transducer and receiver.

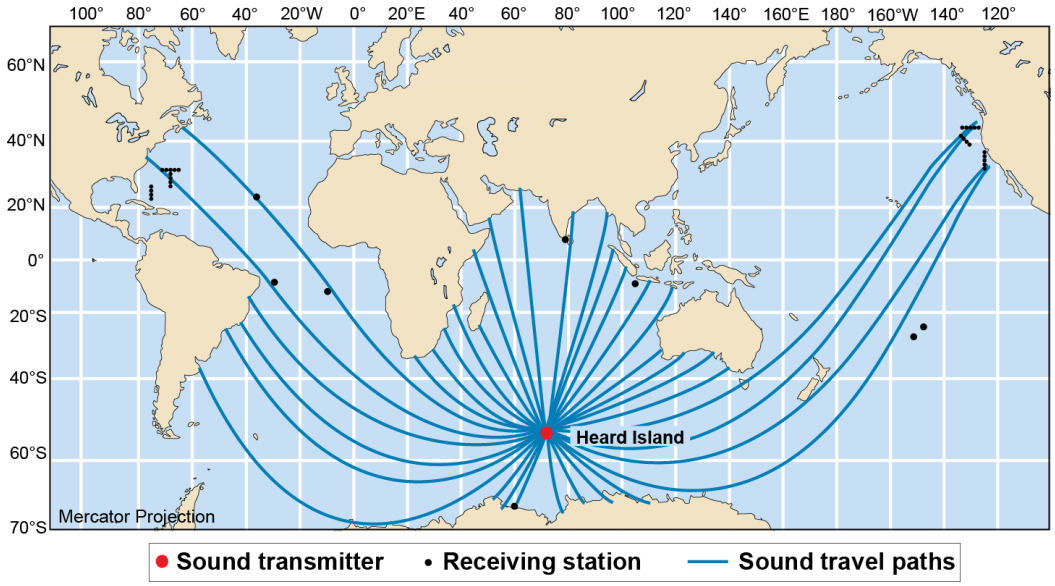


Figure 5-23. The Heard Island experiment. Sound pulses were emitted from a source in the sound channel at Heard Island, and the time of travel to the various receiving sites (shown by the black dots) was measured very precisely. Because the travel time of sound would change if the average temperature of the ocean water in the sound channel changed, arrays like this could be used to monitor changes in the average temperature of ocean waters at the depth of the sound channel.

The speed of sound in seawater increases with increasing temperature. Therefore, if the travel time between a transducer and a listening point in the system decreases from year to year, it indicates that the average temperature within the sound channel between those two points has increased (unless the average salinity or average depth of the sound channel has changed). An experimental acoustic array with the sound source located just north of the Hawaiian Islands and receiving stations surrounding

the Pacific Ocean basin at distances of 3000 to 5000 km from this source has demonstrated that average ocean temperature measurements can be made with a precision of about 0.006°C. Results from this program have revealed surprisingly large seasonal and other short-term variations of average temperatures. The program operated between 1996 and 2006 but did not continue long enough to reveal any long-term trends. The program was abandoned, at least in part, because of controversies concerning the possible effects of the sounds produced on marine species, especially whales, dolphins, and other **marine mammals**.

Recently the idea of monitoring ocean temperatures by acoustic thermometry has been revived by studies of acoustic waves that are created by earthquakes and that travel in the sound channel. Results obtained using this technique have identified temperature fluctuations in the East Indian Ocean on timescales of 12 months, 6 months, and approximately 10 days, and have enabled an estimate of the decadal warming trend that is consistent with but significantly higher than the trend estimated from Argo float data. As with all environmental measurements, the existence of two independent techniques for measuring trends and assessing temporal variability promises to substantially improve the accuracy and detail of ocean water temperature change monitoring studies.

Ocean Noise

Scuba divers know that the undersea world is not silent. Besides the sounds of breathing regulators and boat propellers, an observant scuba diver hears a variety of other noises. They often sound like a well-known snap-crackle-pop cereal. The natural sources of sound in the oceans are many and varied, but they generally fit into three categories: noises from breaking waves and bursting air bubbles, noises from vessels and other human mechanical equipment, and biological noises. Each of the sources produces sounds across a wide range of frequencies.

Biological sounds are produced by many species and through various means. For example, whales and dolphins use sound to locate objects, such as prey, and to communicate with each other. Dolphins may also use intense bursts of sound to confuse or stun their prey. Certain **crustaceans** (shrimp, crabs, and lobsters) make clicking noises as they close their claws, and certain fishes make sounds by inflating and deflating the **swim bladder** (a gas sac used to control **buoyancy**).

Natural sounds and the sounds of human activities, both of which are always present in the ocean, constitute a background noise above which the sonar sound probes used by oceanographers or navies must be strong enough to be heard. Submarines can be detected easily if they send out sonar signals to locate vessels that are searching for them, so instead they use sensitive directional microphones to detect the noise of ships' propellers.

Because sound can travel very long distances in water, sensitive microphones can pick up sounds made hundreds or even thousands of kilometers away. Using sensitive microphone arrays, navies obtain data that are analyzed by supercomputers to separate the sounds of submarine or ship engines and other ship noises from background noise. This technology has become highly sophisticated. In some cases, the listening arrays can identify a specific submarine or surface vessel by its own specific sound "fingerprint," determine the vessel's location precisely, and track its movements across oceans. This capability has been extended so that the system can now identify and track some species in the oceans including whales. Individual whales can even be identified and tracked as each whale makes its own unique sounds.

This page titled 5.12: Transmission of Sound is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.13: CHAPTER SUMMARY

Origins and Distribution of the Earth's Water.

The Earth is the only planet known to have water in all three states: solid (ice), liquid, and gas (atmospheric vapor). The oceans contain 97% of the Earth's water.

The Water Molecule.

Water has many unusual properties related to the polar nature of its molecule and the resulting hydrogen bonds between water molecules.

The Dissolving Power of Water.

Water can dissolve more substances than any other known liquid. Water molecules can surround and hydrate negatively and positively charged ions, allowing the oppositely charged ions of a compound to move independently within the surrounding water molecules.

Salinity.

Various properties of water depend on the concentration of dissolved salts, expressed as salinity. Salinity is usually determined by the measurement of electrical conductivity.

Composition of Seawater.

Seawater is a complex solution containing almost all naturally occurring elements, many radionuclides, numerous organic compounds, and dissolved gases. Elements are present primarily as ions and are classified as major, minor, or trace constituents according to concentration. Dissolved gases, including oxygen, nitrogen, and carbon dioxide, can move freely between surface seawater and the atmosphere. Concentrations of elements in seawater are in an approximate steady state. These concentrations are determined by the crustal abundance of the elements and their behavior in biogeochemical cycles. Ratios of major ions are generally almost constant, although the total dissolved constituent concentration (salinity) varies.

Dissolved carbon dioxide is present as carbonate, bicarbonate, and molecular carbon dioxide in a complex equilibrium that allows more carbon dioxide to be dissolved than would otherwise be the case. This is one reason why the oceans contain about 70 times as much carbon dioxide as the atmosphere. Carbon dioxide is produced by respiration and decomposition of organic matter throughout the ocean depths. Most organic matter in the oceans is produced through photosynthesis, which uses carbon dioxide and releases oxygen. Photosynthesis requires light energy and occurs only in upper layers of ocean waters.

Heat Properties of Water.

Water has anomalously high freezing and boiling points. Hence, water can be present in solid, liquid, and vapor phases in the range of temperatures found at the Earth's surface. Water also has anomalously high heat capacity, latent heat of fusion, and latent heat of vaporization. These anomalous properties enable water to store and transport large amounts of heat through the oceans and atmosphere. Without these properties, the Earth's climate would be much more extreme.

Effects of Pressure, Temperature, and Dissolved Salts on Seawater Density.

Pure water has an anomalous density maximum at 4°C, but the maximum disappears with increasing salt concentration and is

not present in seawater of normal salinity. Seawater density increases with increasing salinity and with decreasing temperature. Water expands as it freezes, so ice is less dense than water and floats. As seawater freezes, the salts are excluded and left in solution.

Surface Tension and Viscosity.

Water has an anomalously high surface tension. This property affects the behavior of the sea surface, including the creation and dissipation of small waves and the formation and behavior of gas bubbles and spray droplets in breaking waves. Water has a high viscosity relative to many other liquids. This high viscosity allows microscopic organisms that are more dense than water to remain in the surface layers of the oceans, because water provides a relatively high resistance to their motion including sinking.

Transmission of Light and Other Electromagnetic Radiation.

Water effectively absorbs electromagnetic radiation. Absorption is less effective at the wavelengths of visible light, which can penetrate up to several hundred meters into clear ocean waters. Blue and green wavelengths of light are absorbed less than red, yellow, and violet wavelengths. This light is the energy source for phototrophy, which includes photosynthesis.

Suspended particles absorb, scatter, and reflect light. The depth to which light penetrates to support phototrophy is reduced as the concentration of suspended particles increases.

Transmission of Sound.

Sound is absorbed much less effectively by water than electromagnetic radiation is. Sound travels long distances through water with little loss of intensity, and it is used by oceanographers and marine animals for communication and remote sensing in the oceans. The principal use of sound by oceanographers is to measure ocean depth and seafloor topography with sonar.

Sound velocity varies slightly with salinity, temperature, and depth (pressure). Very accurate measurements of ocean depth by sonar require detailed knowledge of salinity and temperature variations in the water column. Sound pulses can be used to make very precise measurements of year-to-year changes in average ocean water temperature based on the time sound takes to travel from a source to receiving microphones thousands of kilometers away.

5.13.1: Key Terms

5.13.2: Chapter Questions

5.13.3: References

This page titled 5.13: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.13.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- acid rain
- acoustic
- adsorbed
- algae
- alkaline
- amino
- amino acids
- anion
- archaea
- backscattered
- bacteria
- biogeochemical
- biogeochemical cycle
- buoyancy
- calorie
- cation
- conduction
- contaminant
- continental margin
- convection
- covalent bond
- crust
- crustaceans
- CTD
- DDT
- density
- detritus
- dissolved
- effluents
- electrical conductivity
- electromagnetic radiation
- electrostatically
- environment
- eroded
- estuaries
- estuary
- excretion
- exotic terranes
- fossil fuel
- frequency
- gravity
- greenhouse effect
- groundwater
- hard parts
- heat capacity
- hydration
- hydrogen bond
- hydrophone
- hydrothermal vent
- ion

- ionic bond
- isotope
- joule
- latent heat of fusion
- latent heat of vaporization
- latitude
- lipids
- lithospheric plate
- mangrove
- mantle
- marginal seas
- marine mammals
- microbial
- molality
- nutrient
- oceanic ridges
- PCBs
- pH
- photic zone
- photosynthesis
- phototroph
- phototrophy
- phytoplankton
- plankton
- radioactivity
- radionuclide (radioisotope)
- refracted
- respiration
- residence time
- runoff
- salinity
- salt marsh
- saturation solubility
- scattered
- schools
- scuba
- sediment
- sedimentary arc
- sensible heat
- solubility
- sonar
- sounding
- spatial
- species
- steady state
- subducted
- subduction zone
- surface tension
- surface microlayer
- suspended sediment
- swim bladder
- turbidity
- van der Waals force
- viscosity
- water mass
- wavelengths
- weathered

- zooplankton

This page titled 5.13.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.13.2: Chapter Questions

Study Questions

1. What are the principal sources and sinks of elements dissolved in seawater? Explain why the totals of these sources and sinks must be approximately balanced.
2. Why are some elements present in high concentrations in seawater, whereas others are present in much lower concentrations?
3. Why is the solubility of carbon dioxide in seawater so high?
4. The water molecule is polar. Why does this property give water a high latent heat of vaporization?
5. In Florida citrus groves, growers often try to encourage ice crystals to grow on the trees and fruit during unusually cold nights when the air temperature drops below freezing. Explain how the physical properties of water and ice make this an effective method for protecting the citrus fruit from being frozen and damaged.
6. Why is it easier for a person to float in the ocean than in freshwater?
7. If ice were dense enough that it always sank, how would lakes behave in winter? How might the oceans be different?
8. What is surface tension, and why is this property of water important in the ocean? Why is the high surface tension of water important in our everyday life? The surface tension of water is reduced by the addition of detergents. How and why is that effect useful to us?
9. Why is it dark deep in the oceans? Why can we not easily see the bottoms of streams and rivers?
10. What do we need to know about the properties of the ocean water beneath a ship if we want to use sonar to measure the depth accurately? Why does sound travel long distances in ocean water?

Critical Thinking Questions

1. Describe all the ways that you can think of in which the Earth would be different if the water molecule were not polar.
2. Continental rocks and soils are constantly leached as freshwater dissolves ions and gases in rocks and transports them to the oceans. However, it is thought that the concentration and composition of dissolved salts in the oceans has remained approximately the same for hundreds of millions of years. Explain how this stability is possible and explain why the hundreds of millions of years of leaching have not depleted the concentrations of such soluble elements as sodium and potassium in terrestrial rocks.
3. In this chapter the concept of steady-state processes is illustrated by an analogy of pouring orange juice into a glass filled with water. (a) List some processes that occur in our homes, industry, or economy that are steady-state processes in the sense that the outputs change in response to changes in the inputs (or vice versa). (b) How quickly does this response lead to a new steady state in the examples you have given, and how would changes in inputs or outputs in your example produce temporal changes equivalent to the changes in “concentration” of the “substance” in the “box” discussed in **CC8**?
4. If water’s heat capacity and latent heat of vaporization were much lower, would we still have rain, and would puddles of rainwater behave differently? Explain your answer. Describe some of the things that would be different in your everyday life.
5. If we look through the water of a swimming pool, we can see the bottom. If the water surface is calm, the tiles and lane markings appear clear and sharp. However, if the surface is disturbed, even if only by tiny ripples, these markings appear indistinct and seem to shimmer. Why?

This page titled 5.13.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

5.13.3: References

Figure 5-1 – Segar, D., and E. Stamman Segar, 1997: Deacon's Reef, Papua New Guinea.

Figure 5-14 – Adapted from Montgomery, R. B., 1958: Water characteristics of Atlantic Ocean and of world ocean. *Deep Sea Res.*, **5**, 134–148, Source.

Figure 5-19 – NASA Worldview, 2025: Chlorophyll a (L2) 22 October 2024. Terra/MODIS, NASA Earth Science Data and Information System (ESDIS), accessed 17 October 2025, Source.

> We acknowledge the use of imagery from the NASA Worldview application, part of the NASA Earth Science Data and Information System (ESDIS).

Figure 5-21 – Adapted from Kinsler, L. E., and A. R. Frey, 1962: *Fundamentals of Acoustics*, John Wiley & Sons, 524 pp.

Figure 5-23 – National Research Council. 1991. Heard Island Experiment. The National Academies Press. Accessed 24 June 2025, Source.

David R. Lide, Ed., 2005: *CRC Handbook of Chemistry and Physics, 86th Edition*. CRC Press, 2544 pp.

This page titled 5.13.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6: OCEAN SEDIMENTS

Critical Concepts

- **CC1 Density and Layering in Fluids**: Fluids, including the oceans, are arranged in layers sorted by their density. Heat sources under the seafloor can heat and reduce the density of any water present, and this heated water rises through the seafloor and up into the water column.
- **CC2 Isostasy, Eustasy, and Sea Level**: Earth's crust floats on the plastic asthenosphere. Sections of crust rise and fall isostatically as temperature changes alter their density. Oceanic crust cools progressively after it is formed and sinks because its density rises. Thus, the seafloor becomes deeper with increasing distance from an oceanic ridge.
- **CC3 Convection and Convection Cells**: Fluids that are heated from below, including water within sediments and cracks in the seafloor, rise because their density is decreased. This establishes convection processes. Heated water rising through the seafloor is replaced by colder water that seeps or percolates downward into the seafloor to replace it.
- **CC4 Particle Size, Sinking, Deposition, and Resuspension**: Suspended particles (either in ocean water or in the atmosphere) sink at rates primarily determined by particle size: large particles sink faster than small particles. Once deposited, particles can be resuspended if current (or wind) speeds are high enough. Generally large particles are more difficult to resuspend, although very fine particles may be cohesive and also difficult to resuspend. Sinking and resuspension rates are primary factors in determining the grain size characteristics of beach sands and sediments at any given location.
- **CC6 Salinity, Temperature, Pressure, and Water Density**: Sea water density is controlled by temperature, salinity, and to a lesser extent pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Heated water discharged by hydrothermal vents has a high enough salinity in some areas of very limited mixing that it is more dense than the overlying seawater and collects in a layer next to the seafloor.
- **CC7 Radioactivity and Age Dating**: Some elements have naturally occurring radioactive (parent) isotopes that decay at precisely known rates to become a different (daughter) isotope. Measuring the concentration ratio of the parent and daughter isotope can be used to calculate the age of the various materials since they were first formed, but only if none of the parent or daughter isotope are gained or lost from the sample during this time period. This condition is usually not met in sediments and radioisotope age dating is difficult so other dating methods including variations in fossil assemblages, and magnetic field properties are used extensively.
- **CC8 Residence Time**: The residence time of seawater in a given segment of the oceans is the average length of time the water spends in the segment. In restricted arms of the sea or where residence time is long, very fine grained particles from glaciers and rivers can become concentrated in the water. Residence time is also a major factor in determining the change in concentration of an element in seawater if its inputs to the oceans change.
- **CC9 The Global Greenhouse Effect**: The oceans play a major part in studies of the greenhouse effect as the oceans store large amounts of carbon dioxide both in solution and as carbonates in sediments, formed at shallow enough depths that the carbonates are not dissolved.
- **CC14 Phototrophy, Light, and Nutrients**: Chemosynthesis and phototrophy, which includes photosynthesis) are the two processes by which simple chemical compounds are made into the organic compounds of living organisms. Many sediments have high concentrations of particles that originate from photosynthetic organisms or species that consume these organisms.

Figure 6-1. This beautiful cove and beach on the south shore of Hawaii is known locally as the "Green Sand Beach." The shoreline consists of volcanic ash that is easily eroded. The eroded cliffs are the source of the sand on this beach and those nearby, as well as on the seafloor on the narrow continental shelf. The minerals that are not weathered and have a distinctly greenish hue that, unfortunately, does not show well in photographs but that can be seen more easily in the inset close-up image of the sand grains.

Why should we be interested in marine **sediments**? Approximately 70% of the Earth's surface is covered by the oceans, and almost all of the seafloor beneath these oceans is covered by sediments. These sediments provide a home for many living species and contain a wealth of information about the history of the Earth's climate, oceans, and continents. Sediments accumulate layer by layer and include the remains of dead organisms. By studying changes in the type and chemical composition of the sediment and organisms in each layer beneath the sediment surface, we can learn a great deal about the Earth's history and the evolution of life.

This chapter reviews the origin and types of particles present in ocean sediments and the mechanisms by which they enter the oceans and are transported to the seafloor. It also describes how the type of sediment that accumulates at any location can change with changes in climate and plate tectonics and examines how the sediments can be used to investigate the Earth's history.

6.1: Sediments and Biogeochemical Cycles
6.2: Classification of Sediments
6.3: Lithogenous Sediments
6.4: Biogenous Sediments
6.5: Hydrogenous Sediments
6.6: Cosmogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments
6.9: Distribution of Surface Sediments
6.10: The Sediment Historical Record
6.11: Chapter Summary
6.11.1: Key Terms
6.11.2: Chapter Questions
6.11.3: References

This page titled 6: Ocean Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.1: Sediments and Biogeochemical Cycles

Sediments are important components of **biogeochemical** cycles (**Chap. 5**), and they act as both sinks and sources of elements dissolved in seawater. Human civilization has now altered the **steady state** (**Chap. 5**) of the biogeochemical cycles of many elements by mining rocks and burning **fossil fuels**; by deforestation, which increases erosion; and by releasing **contaminants** that cause **acid rain** and promote **leaching**. Many human activities have increased the rate at which some elements enter the ocean and atmosphere, which will probably cause concentrations of these elements in seawater to rise even if the rate of removal to the sediment increases to match the increased input rate (**CC8**).

For many elements, the rate of input to the oceans has varied over millions of years. Periods of intense volcanic activity or changes in land erosion rates related to sea-level and **glaciation** changes, for example, affect the rate of input to the oceans. Without a full understanding of these processes, we cannot understand or predict how releases of elements due to human activities will affect the ocean environment. Biogeochemical cycles are difficult to study because of their many complex interactions and because some of the key processes occur on timescales of tens of millions of years. Fortunately, seafloor sediments preserve a history of the biogeochemical cycles and their variations over the past 170 million years (the age range of almost all surviving seafloor; **Chap. 4**) and perhaps as far back as 340 million years ago. Sedimentary rocks on land preserve even older history. Studying the jigsaw puzzle of information contained in ocean sediment is essential to understanding the consequences of human activities, including pollution (**Chap. 16**).

This page titled 6.1: Sediments and Biogeochemical Cycles is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.2: Classification of Sediments

Sediments are separated into categories to facilitate comparisons with each other and communication among scientists. There are a number of classification schemes. The most widely used schemes are based either on the range of sizes of the particles (grains) that make up the sediment or on the origin of the predominant particles in the sediment. Each type of scheme has its advantages, as we shall see in this chapter.

Classification by Grain Size

The size of the grains in marine sediments varies from boulders tens of centimeters or more in diameter to grains that are so small (less than 0.001 mm in diameter) that they cannot be distinguished except under the most powerful electron microscopes. Various grain size classification schemes have been used, but generally sediments are described as clay, silt, sand, or gravel (**Fig. 6-2**). Sediments with the finest grain size ranges—silt and clay—are often classified together as mud.

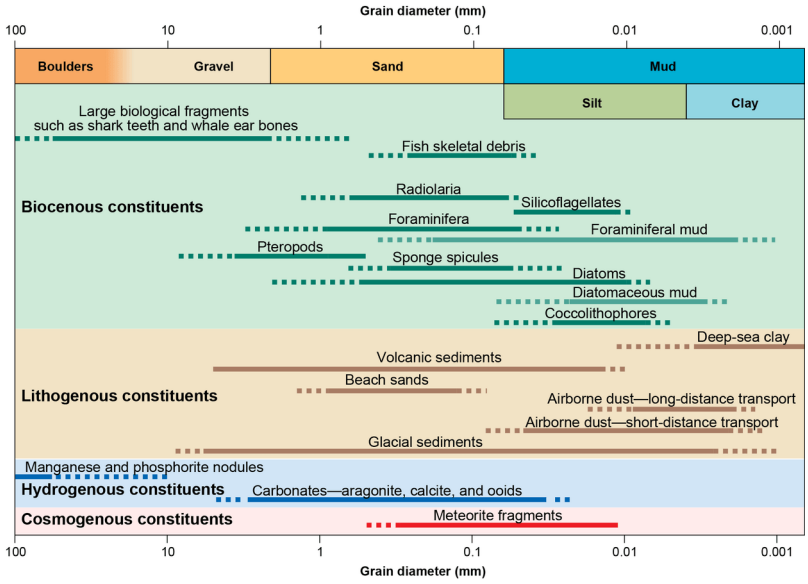


Figure 6-2. Various types of particles, each of which has its own characteristic size range, are commonly found in marine sediments. The particles in foraminiferal and diatomaceous muds are, on average, smaller than the sizes of the foraminifera and diatoms from which they are derived because the hard parts of these organisms are partially dissolved and broken mechanically during transport to, and burial in, the sediments.

Sediments with a wide range of grain sizes can be found in different areas of the oceans or at different depths within the layers of sediment at any given location. However, most marine sediments that have accumulated at the same time and in the same area of the oceans consist of only grains that are within a narrow range of grain sizes. Sediments that consist mostly of grains that are within a narrow range of sizes are said to be well sorted. For example, most sediment on the deep-ocean floor consists of well-sorted mud, and many areas near beaches consist of well-sorted sand. Sediments that consist of a wide range of grain sizes are said to be poorly sorted. For example, some undersea landslide deposits are poorly sorted, consisting of a mixture of mud, sand, and gravel. Poorly-sorted sediments are found in only limited areas in the oceans. Transport mechanisms that cause sediment sorting are discussed in **CC4**, and sorting of beach sand is discussed in **Chapter 11**.

Classification by Origin

Sediments can be classified as **lithogenous, biogenous, hydrogenous**, or **cosmogenous** to indicate the origin of the dominant type of particle. Lithogenous particles consist of fragments of rocks and minerals **weathered** and **eroded** from rocks on land, and fragments of ash and rock from terrestrial and undersea volcanic eruptions. Biogenous particles are the remains of organisms. Because most organic matter is decomposed rapidly, biogenous particles are predominantly the mineralized **hard parts** (shells or skeletal material) of marine organisms. They include, for example, the hard parts of certain **phytoplankton** and **zooplankton**

species, **mollusk** shells, fish bones, and whale and shark teeth. Hydrogenous particles are formed by precipitation of dissolved inorganic chemicals from seawater to form solids. Cosmogenous particles are meteorite fragments that have passed through the atmosphere.

Each category of particles can be subdivided into several different types. For example, biogenous particles can be classified according to the types of organisms from which they are derived (**Fig. 6-2**).

The distribution of sediment particles in the oceans is determined by the size range and origin of each type of particle, the susceptibility of each type of particle to decomposition or dissolution in seawater, and the mechanisms transporting the particles. The distribution of sediments is discussed later in this chapter. However, to understand the distribution of sediments, we need first to understand the origins and sources of the particles that make up sediments and the processes that transport and transform the particles in the ocean.

This page titled 6.2: Classification of Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.3: Lithogenous Sediments

Most lithogenous sediment particles are the products of weathering and erosion of terrestrial rocks by water and wind. For this reason, sediments with a high proportion of lithogenous particles are often called **terrigenous** sediments. Rock fragments are continuously broken off of solid rock by the action of running water or waves, as a result of freezing and thawing of ice, or by the actions of plant roots or animals. Once they are broken into fragments, rock particles are further weathered to smaller particles and transported by rivers and streams, **glaciers**, waves, and winds.

Rocks of the landmasses are composed of a large variety of minerals. The minerals are slowly altered and partially dissolved by reactions with oxygen, carbon dioxide, and water. During such chemical weathering, many minerals partly or wholly dissolve, leaving behind resistant minerals including quartz, feldspars, and clay minerals, which are all siliceous minerals. Clay minerals, which are less resistant to weathering, are layered structures of silicon, aluminum, and oxygen atoms. Some clay minerals also contain iron or other elements.

Five natural transport mechanisms bring lithogenous sediment to the oceans: freshwater runoff, glaciers, waves, winds, and landslides. Humans have added another mechanism: ships. Until the mid to late 20th century, all vessels simply threw trash and garbage overboard. Cans, bottles, plastics, and clinker (cinders from coal-burning ships) can now be found in sediments throughout the oceans, particularly under major shipping lanes (**Chap. 16**). The various transport mechanisms introduce different size ranges of sediment particles to the oceans, and each mechanism has a different distribution of input locations.

Transport by Rivers

Rivers vary substantially in terms of their volume of flow, **current** speed, and **turbulence**. In the fast-flowing upper parts of rivers where they cut through steeply sloped mountain valleys, strong turbulence can be created, and as a result, substantial quantities of rock can be eroded. Such turbulent rivers carry a wide size range of particles. As rivers reach flatter land near the coast, turbulence and current speed decreases, and the largest particles are deposited in this region (**CC4**). Consequently, under normal flow conditions, rivers transport mostly fine-grained particles to the oceans. In contrast, when heavy rains flood rivers, both the volume of the river discharge and the river speed increase dramatically, and great quantities of deposited sediment, including larger particles, are resuspended from the riverbed. River inputs are continuous, but the rate of input, particularly of the larger particles, peaks during major flood events. The peaks can be dramatic. Rivers may transport more sediment to the ocean in a few days after an unusually massive storm, such as a hurricane, than they do during several years or longer of normal conditions.

Not all rivers discharge significant amounts of sediment to the oceans. Flat coastal plains often form when valleys, carved by rivers or glaciers when sea level was lower, are later flooded as sea level rises. Rivers that flow in these recently exposed valleys deposit sediment that gradually fills them and a flat plain is formed. Even after the plain forms, rivers often continue to follow the paths of the original valleys. Most particles transported by rivers collect in **estuaries** on coastal plains and are not transported to the ocean. For example, many of the rivers of the Atlantic coast of North America pass through such estuaries and do not transport much sediment to the ocean. Unless sea level changes, the estuaries will eventually fill with sediment, and **deltas** may be formed by accumulation of sediment discharged at the river mouths.

Rivers that discharge onto active continental margins at **subduction zones** are generally short and drain limited land areas between the **coastline** and the ridges of the coastal mountain ranges created by the subduction process. The Pacific coasts of North and South America are examples (**Fig. 6-3**). Because they drain limited land areas, such rivers generally carry only a small load of eroded rock particles. The load may be greater where there are gaps in the coastal mountains, but even so, in some river systems, only small quantities of suspended sediment are transported into the oceans. For example, the considerable suspended sediment load from the rivers draining the Sierra Nevada in California is deposited in the northern part of San Francisco Bay and does not reach the Pacific Ocean.

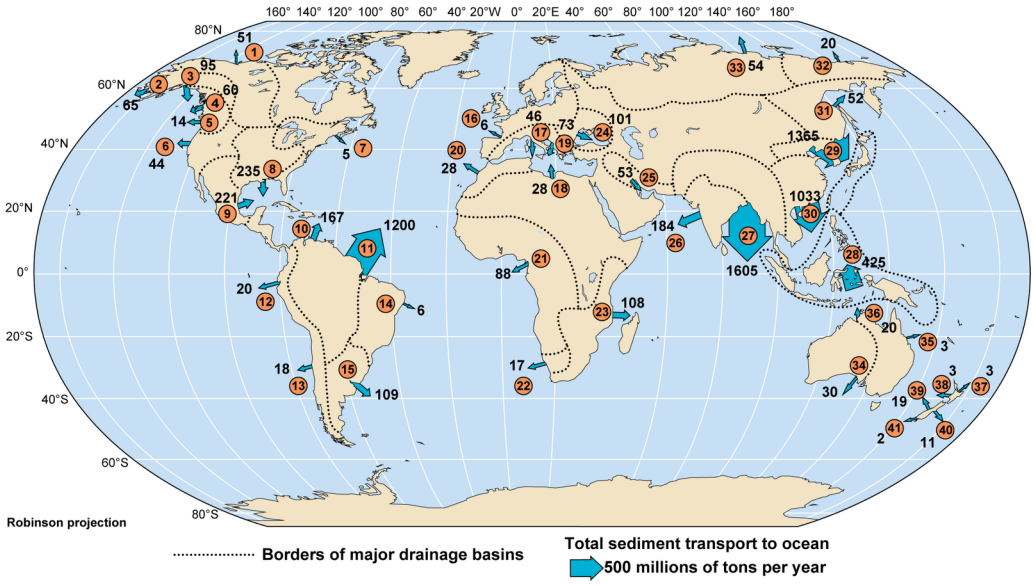


Figure 6-3. Estimated inputs of lithogenous sediments to the oceans by major rivers. These inputs are mostly to marginal seas and are dominated by runoff from the Asian continent to the marginal seas of Southeast Asia. The Amazon is the only very large source of input to the Atlantic Ocean. Inputs to the Pacific Ocean are mostly trapped in the subduction zone trenches that surround it.

*List of rivers from **Figure 6-3***

Number	River
1	Mackenzie, Babbage, Colville
2	Yukon, Kuskokwim
3	Susitna, Copper
4	Fraser, Stekie, Skeena
5	Columbia
6	Russian, Eel, Santa Clara
7	Hudson, Saint Lawrence
8	Mississippi, Brazos
9	Rio Grande, Magdalena
10	Orinoco
11	Amazon
12	Chira, Numerous small
13	Numerous small
14	Sao Francisco
15	Plata, Negro, Colorado, Parana, Uruguay
16	Rhine, Seine, Loire, Elbe, Garonne
17	Ebro, Tiber, Rhone, Po, Lamone
18	Nile, Medjerda
19	Semani, Drini
20	Several small
21	Niger, Zaire

Number	River
22	Orange
23	Zambezi, Limpopo, Rufiji, Tana
24	Danube, Dnieper, Don, Coruh, Sakarya
25	Tigris, Euphrates
26	Indus, Namanda
27	Ganges, Brahmaputra, Irrawaddy
28	Sepik, Fly, Purani, Numerous small
29	Huang He (Yellow), Choshui, Kaoping
30	Mekong, Chang (Yangtze)
31	Amur
32	Kolyma, Indigirka
33	Lena, Ob, Yenisey
34	Murray, Darling
35	Burdekin
36	Ord
37	Motu
38	Wanganui
39	Haast, Hokitika
40	Waimkariri
41	Waiau

Approximately 90% of all lithogenous sediments reach the ocean through rivers, and 80% of this input is derived from Asia. The largest amounts are from four rivers that flow into the Indian Ocean (**Fig. 6-3**). The Ganges, Brahmaputra, and Irrawaddy discharge into the Bay of Bengal, and the Indus discharges into the Arabian Sea. Most of the other rivers that transport large amounts of suspended sediment into the ocean empty into **marginal seas**. Examples are the Chang (Yangtze) River (to the East China Sea), the Huang (Yellow) River (to the Yellow Sea), and Mekong (to the South China Sea)—all in Asia—and the Mississippi (to the Gulf of Mexico) in the U.S. With the exception of the Amazon's input to the central Atlantic, river discharges of lithogenous sediment to the Pacific and Atlantic Oceans are very limited.

Erosion by Glaciers

Ice is squeezed into cracks in the rocks over which glaciers flow, causing pieces of rock to break off of the floor and sides of the glacial valley. This process of erosion is extremely effective, as can be seen from the deep fjords that the abundant glaciers cut during the last ice age (**Fig. 11-6**). The rock eroded by a glacier is bulldozed, dragged, and carried down the glacial valley with the glacier, then deposited at its lower end. At present, glaciers reach the sea only at certain places in high latitudes, including Antarctica, Alaska, Greenland, and Patagonia. These glaciers release most of the eroded rock into the water close to where they enter the sea. Some icebergs that break off the ends of glaciers may be transported by ocean currents hundreds of kilometers or more before they melt. Rock particles are transported with the ice and can be released far from the glacier, but the amount of sediment transported in this way is small.

Glaciers can transport rock fragments that range from the finest grains to house-sized boulders. The larger boulders, pebbles, and sand grains are almost all deposited in ocean sediments close to the end of the glacier, but finer-grained material can be transported by currents to more remote deposition sites.

Some glaciers release large amounts of "glacial flour," particles so finely ground that they remain suspended for weeks or months,

even in lakes or slowly flowing streams where turbulence is very low. Where glaciers discharge into lakes or fjords with long **residence times** (**CC8**), glacial flour can become so concentrated that the water is distinctly milky (**Fig. 6-4**).

Figure 6-4. The Cavell Glacier flows down Mt. Edith Cavell in Jasper National Park, Alberta, Canada and enters a lake. Note the milkiness of the water due to the large quantities of fine particles (glacial flour) released from the melting glacial ice. Unlike in this lake, when glaciers enter the oceans, much of the fine-grained material coagulates because of the high salinity and is deposited quickly. Note also the rock debris on the surface of the glacier itself. This and the dark colored inclusions you can see in the glacial ice consists of particles of many different sizes. The larger particles will be deposited very close to where the glacier ends, while the smaller particles may be transported long distances.

Erosion by Waves

Waves continuously erode many coastlines (**Chap. 11**). The eroded rock particles are similar to those introduced to the oceans by rivers. However, particles eroded by waves generally have a larger proportion of unweathered mineral grains, unless the eroding coastline is composed of sedimentary rock (**Chap. 11**).

Wave erosion of coastlines creates particles of all sizes, from large boulders that fall from undercut cliff faces to the finest clay particles. Waves sort the particles, transporting small ones offshore while leaving larger ones on or close to the shore.

Transport by Winds

Dust particles can be transported very long distances through the atmosphere by winds before they are deposited on the ground or ocean surface. Dust particles in the atmosphere fall to the ground and are resuspended by air currents in much the same way that particles are deposited and resuspended by ocean currents (**CC4**). Normal winds can transport only very fine particles, but, as we can observe on a windy day at a dry sand beach, high winds can also transport larger grains.

When dust particles fall on the ocean surface, they become suspended sediment in the water. Consequently, there is a continuous flux of dust particles from the land to the oceans. The flux of airborne dust to the oceans is particularly strong in some locations, such as the northern subtropical Atlantic Ocean, where the prevailing winds blow from the Sahara Desert out across the ocean. Dust clouds that stretch thousands of kilometers over the ocean can be seen in this area in satellite images (**Fig. 6-5**). Airborne particles in Florida and the Bahamas are often rust-red because they consist mainly of large quantities of the red, clay-sized particles from the Sahara Desert that are transported across the Atlantic Ocean by winds. Windblown dust also is transported from the Gobi Desert in Asia over the North Pacific Ocean. These lithogenous clay-sized particles form a significant fraction of the deep-ocean sediment between 20°N and 30°N in the Pacific Ocean, where the accumulation rate of other particles is very low.

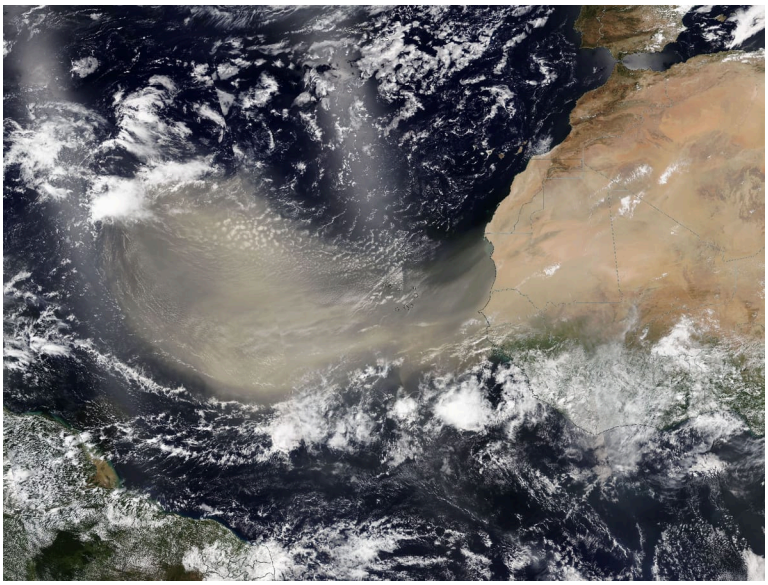


Figure 6-5. This June 2020 satellite image shows a huge yellowish-brown, windblown dust plume from the Sahara Desert. The plume extends westward from the African continent. After traveling approximately 1500 km over the Atlantic Ocean, the plume turns northwestward toward Nova Scotia, Canada.

Although the amount of dust in the atmosphere is usually small, the continuous flux of small quantities through the atmosphere adds up to a substantial input to the oceans over geologic time. Consider how much dust collects on a piece of furniture in only a few days. Multiply that amount by billions to see how much dust can reach the ocean surface over a period of millions or tens of millions of years.

Fine particles are deposited on all parts of the ocean surface at a relatively uniform rate, although rates of deposition are greater in areas downwind from deserts. High storm winds carry larger particles, up to fine sand size, over the oceans. Such particles are deposited relatively close to their coast of origin because most storm winds do not persist for long periods or blow over great distances.

Although winds generally transport only smaller particles, larger lithogenous particles do enter the atmosphere as a result of volcanic eruptions, especially the explosive eruptions of **convergent plate boundary** volcanoes (**Chap. 4**). Explosive eruptions can instantaneously fragment and blast upward large quantities of rock, which are then carried by winds as ash and cinders so that they rain out over a very large area. The principal fallout occurs immediately around and downwind of the volcano, where almost all of the larger particles fall. Eruptions produce particles of all sizes, including tiny particles that can be carried thousands or tens of thousands of kilometers in the atmosphere before being deposited. The most violent eruptions throw large quantities of ash into the upper atmosphere, where swift winds, such as the **jet stream**, can transport them one or more times around the planet before they finally fall. Volcanic ash from the largest eruptions can remain in the upper atmosphere for years and affect the Earth's climate by reducing the amount of solar energy that passes through the atmosphere to the Earth's surface (**CC9**). Such ash can also reduce the ozone concentration in the Earth's ozone layer (**Chap. 7**).

Geologically recent explosive eruptions that have ejected large quantities of ash into the atmosphere include the 1980 Mount St. Helens eruption in Washington, 1991 Mount Pinatubo eruption in the Philippines, and the 2022 Hunga Tonga eruption in the South Pacific Ocean. The Pinatubo eruption completely buried the then U.S. Clark Air Base with ash, even though the base was tens of kilometers from the volcano. The 1980 eruption of Mount St. Helens is estimated to have ejected about 1 cubic kilometer (km^3) of ash into the atmosphere, and Mount Pinatubo and Hunga Tonga about four or five times that amount. These eruptions were dwarfed by two Indonesian eruptions: Krakatau in 1883 and Tambora in 1815, which ejected an estimated 16 and 80 km^3 of ash, respectively. The eruption that created the Long Valley caldera, one of the Earth's largest calderas, in California 700,000 years ago may have ejected 500 km^3 of ash. Much of the ash ejected by these eruptions would have fallen on the ocean's surface or been washed into the ocean in runoff and then transported to the sediments. Distinct ash layers corresponding to these and other eruptions are found in some ocean sediments.

Transport by Landslides

Landslides occur when loose soil or rock moves down a slope under the force of gravity in a process known as “mass wasting.” Water, ice, and winds may be involved in loosening the soil or rock, but the rock or soil is primarily transported by gravity, not by the water, ice, or wind. Some landslides occur on the slopes of a shoreline and may carry rock and soil particles of a wide range of sizes directly into the ocean. Most such material is initially deposited, but it then becomes subject to erosion and transport by waves. Mass-wasting processes also occur on steep slopes on the ocean floor. Slumps or **turbidity currents** (discussed in more detail later in this chapter) can occur as a result of accumulation of sediment on a continental shelf or at the head of a **submarine canyon**, or accumulation of volcanic rock on the side of a submerged or partly submerged volcano. These events can transport substantial amounts of lithogenous sediments into the deep-ocean basins.

This page titled 6.3: Lithogenous Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.4: Biogenous Sediments

Most non-microbial life in the oceans is sustained by the conversion of dissolved carbon dioxide to living organic matter by photosynthetic organisms. **Phototrophy** takes place only in the upper part of the water column (at most a few hundred meters deep) where sufficient light is present (**Chap. 12**, **CC14**). With only minor exceptions, the photosynthetic organisms of the oceans are small (less than about 2 mm), as are most of the animals that feed on them (**Chap. 12**). Very few of these organisms live out a full natural life span, because most are consumed by larger animals. The resulting fecal material is excreted in the form of **fecal pellets**. Fecal pellets, although small, are much larger than the microscopic organisms of which they are made up. Therefore, they tend to sink toward the seafloor relatively rapidly, as do bodies of larger animals that have not been ingested by another animal (**CC4**). Most life in the deep-ocean waters and on the seafloor is sustained by the rain of this organic-rich **detritus** (**Chap. 13**).

One species, a small **planktonic tunicate** called a “giant larvacean,” may contribute a substantial proportion of the total detritus that rains into the deep. The giant larvacean secretes an intricate netlike structure to capture its food. The larvacean lives inside the structure, which, consequently, has been called a “house.” When the web becomes clogged with particles, which happens about once a day, the larvacean simply releases the house and secretes a new one. The abandoned houses are large enough that they sink fairly rapidly, and this source may then contribute a substantial proportion of the detritus that reaches the seafloor, at least in some areas.

As detritus falls through the water column, it is either utilized by animals or decomposed by **bacteria, archaea** and **fungi**, and decomposition continues on the seafloor. In all but a few locations where the rain of detritus is extraordinarily heavy or where there is insufficient oxygen to support bacterial decomposition, the organic matter in detritus is essentially completely decomposed. Consequently, little organic matter is incorporated in the accumulating bottom sediment in most areas. In contrast, many marine species have hard parts that are not decomposed as they fall through the water column, and these materials constitute the overwhelming majority of inputs of biogenous particles to ocean sediment.

The **hard parts**, shells, or skeletons of marine organisms are either **calcareous** or **siliceous**. Calcareous organisms have hard parts of calcium carbonate ($CaCO_3$); siliceous organisms have hard parts of silica in the form of opal, which has the same basic composition as glass (SiO_2). Ocean surface waters are populated by very large numbers of calcareous and siliceous organisms, most of which are smaller than a few millimeters. Hence, a continuous rain of both calcareous and siliceous particles falls from the surface layers. This material does not rain uniformly on the sediment surface because marine productivity and types of organisms vary by region, and because calcium carbonate and silica dissolve in seawater at slow but variable rates.

The two major factors that determine the accumulation rate of biogenous sediment are the rate of production of biological particles in the overlying water column and the rate of decomposition or dissolution of these particles as they fall to the seafloor. The percentage of biogenous material in the sediment is determined by these two factors plus the rate of accumulation of other particles.

Regional Variations of Biogenous Particle Production

Chapters 12 and 13 examine the factors that determine primary productivity rates and the types of organisms that inhabit various regions of the oceans. The distributions of productivity, biomass, and dominant organism type (e.g., calcareous or siliceous) in the ecosystem are major factors in determining the nature of sediments that accumulate in any given area.

In high latitudes and areas of coastal upwelling (**Chap. 13**), siliceous **diatoms** are the dominant photosynthetic organisms (**Fig. 6-6**). Diatoms are among the largest of the phytoplankton (up to about 2 mm) and yield relatively large siliceous particles. At lower latitudes and in the open ocean, many of the dominant photosynthetic organisms have no hard parts. However, one group of very tiny photosynthetic organisms called **coccolithophores** (**Fig. 6-7a**) may grow in abundance. They are covered by a number of small calcareous plates that contribute extremely fine particles to the sediment. The white chalk cliffs of Dover in England consist primarily of coccolithophore plates (**Fig. 6-7b**).

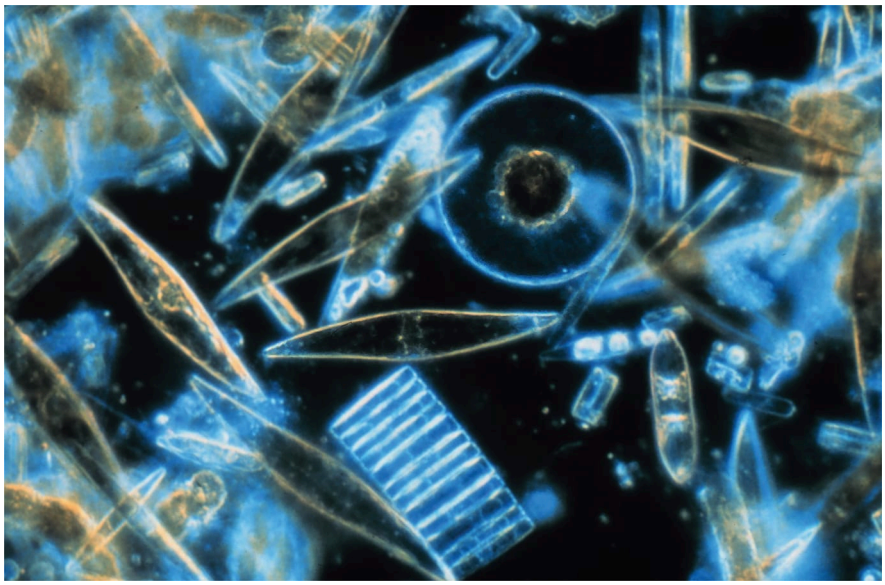


Figure 6-6. Several species of diatoms under a microscope. These are species found living between crystals of annual sea ice in McMurdo Sound, Antarctica. Diatoms are algae that are a preferred food source for many small marine animals and animal larvae. They are photosynthetic, and each one is covered by hard silica frustules.

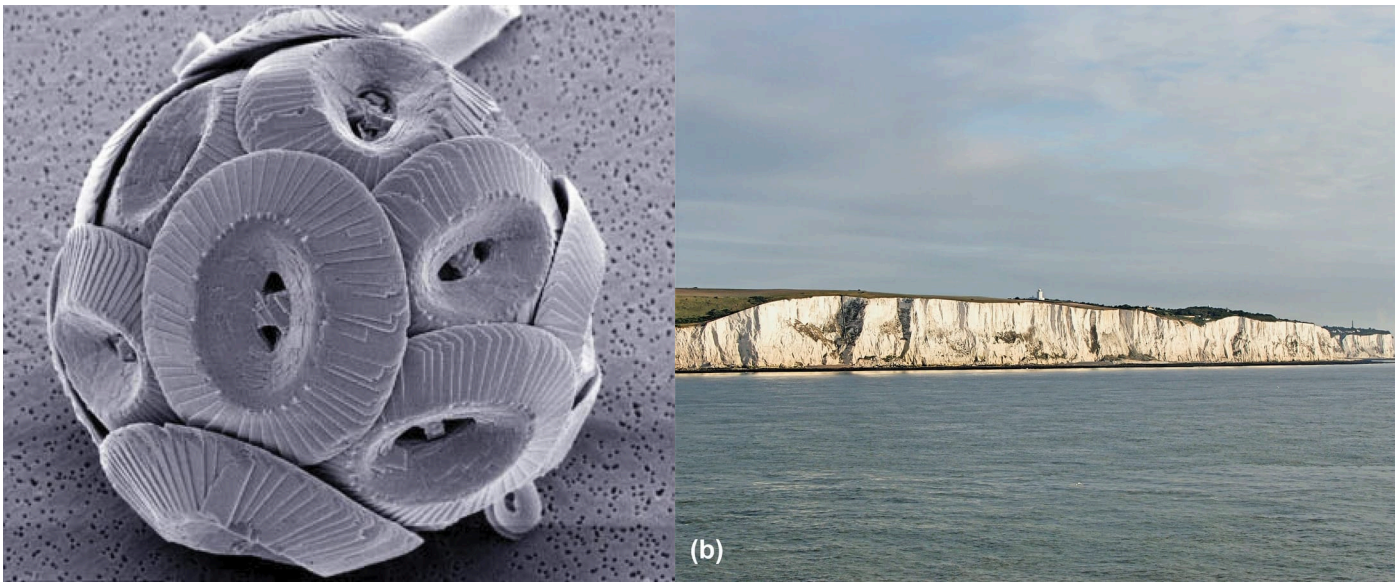


Figure 6-7. (a) Coccolithophore (Coccolithus pelagicus) photographed at approximately 5000 times magnification. Note the covering of intricate calcareous plates. (b) The rocks that make up the White Cliffs of Dover, seen here, are composed primarily of coccolithophore plates.

Zooplankton with calcareous shells include **foraminifera** and **pteropods** (Fig. 6-8), many species of which are present throughout the oceans. Animals with siliceous hard parts are less common, but the microscopic **radiolaria** (Fig. 6-9) have intricate silica shells and are abundant in tropical waters that have high primary productivity.

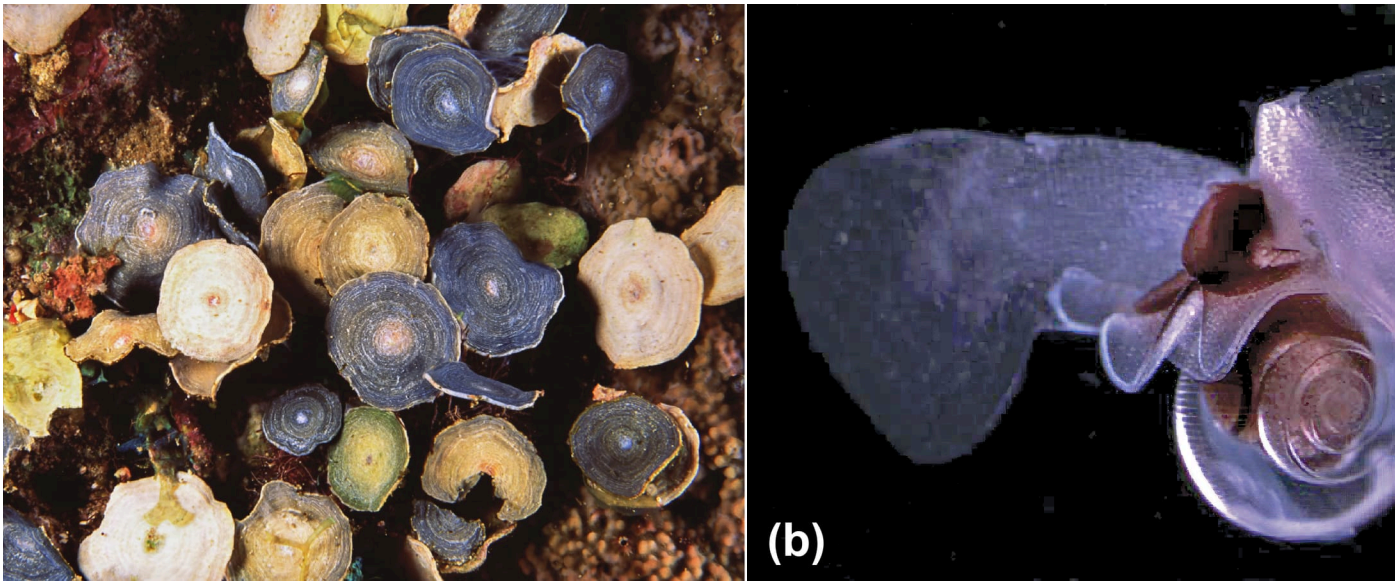


Figure 6-8. Zooplankton with calcareous hard parts (a) Most foraminifera are pelagic (live in the water column). However, the species shown in this photograph (the round platelike objects that are about 1 cm in diameter, Marginopora vertebralis, Papua New Guinea) is benthic and lives on the surface of coral reefs. Several of the individuals shown here have died and lost their green pigmentation, revealing the calcareous hard parts that can survive to become sediment particles. (b) A pteropod, or sea butterfly, *Limacina helicina*. The animal "flaps" its "wings" (modified foot) to propel itself through the water. It is pelagic and eats plankton.

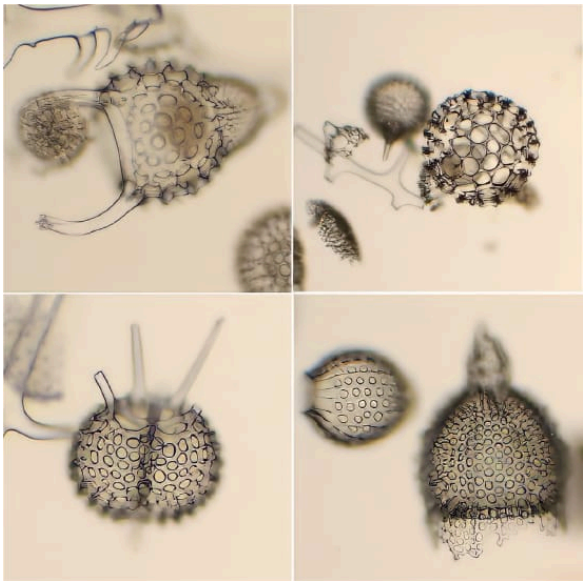


Figure 6-9. Radiolaria (photographed here at about 10 times magnification) are small zooplankton that have intricate silica shells, which make a major contribution to the sediments. They occur in abundance in some tropical waters where primary productivity is high in the surface water layer.

Dissolution of Biogenous Particles

Seawater is undersaturated with silica. Therefore, siliceous particles dissolve as they fall through the water column to the seafloor. The rate of dissolution is very slow at all depths, but it decreases with depth throughout the upper 2 km of the water column (**Fig. 6-10**). Thin siliceous hard parts are almost always totally dissolved and reach the sediment only in areas where siliceous organisms grow in great abundance. For example, many radiolaria have thin, intricate shells that are dissolved relatively easily and reach the sediment in substantial quantities only in highly productive tropical waters where they abound.

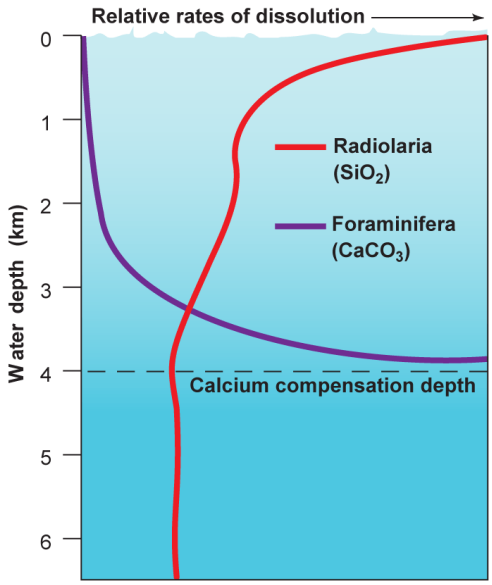


Figure 6-10. Changes in the dissolution rate of siliceous shells (radiolaria) and calcareous shells (foraminifera) with depth. Siliceous shells are more soluble in surface layers; calcareous shells are more soluble at greater depth.

The production of fecal pellets can enhance the accumulation of siliceous and calcareous particles. When packaged in relatively large fecal pellets, such particles fall more quickly to the seafloor and are partially protected from dissolution until the organic matter in the pellet disintegrates or is decomposed.

Many diatoms have a relatively thick siliceous **frustule** (hard part), much of which reaches the seafloor, where it continues to dissolve until buried by other particles. Diatom frustules are present in the sediment in locations where their production rate and the sedimentation rate are high.

Calcareous particles also dissolve in seawater, but their behavior is different and more complex than that of siliceous particles. In the upper water layers, where water temperatures are high and pressures low (**Chap. 8**), seawater is generally supersaturated with calcium carbonate. Consequently, calcareous shells dissolve only very slowly or not at all in near-surface waters (**Fig. 6-10**). The solubility of calcium carbonate increases with pressure and with decreasing temperature. Therefore, solubility tends to increase with depth.

Not all calcareous hard parts are made of the same mineral. Calcium carbonate shell material is of two distinct forms: **calcite** and **aragonite**. Some types of animals, including most foraminifera, have calcite shells, whereas others, such as pteropods, have aragonite shells. Aragonite dissolves much more readily in seawater than calcite. Therefore, pteropod shells are dissolved more completely at shallower depths than foraminiferal shells. Where pteropods are more abundant than foraminifera, shallow-water sediments can consist of predominantly pteropod particles. Sediments at intermediate depths are predominantly foraminiferal. Sediments at greater depths, where the dissolution rate is high, have almost no calcareous component (**Fig. 6-11**).

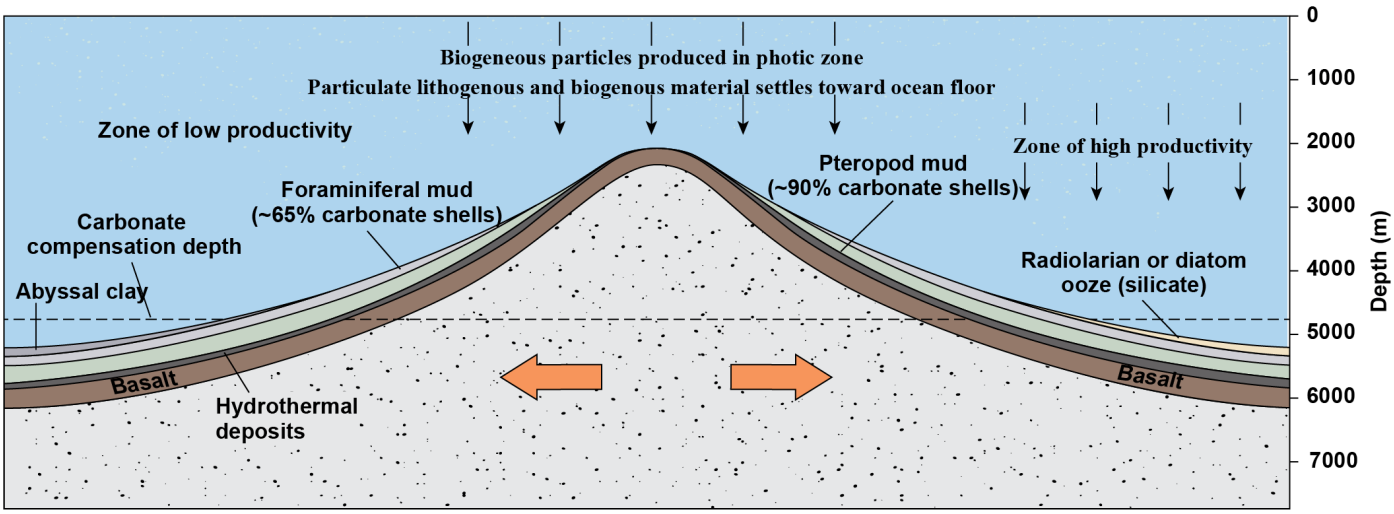


Figure 6-11. Idealized depiction of layered sediment accumulation on an actively spreading oceanic ridge. There is a thin, patchy cover of predominantly hydrothermal sediments at the center of the oceanic ridge. On the flanks of the ridge, which are shallower than the carbonate compensation depth (CCD), calcareous sediments dominated by pteropods and foraminifera are deposited. Pteropods dominate the surface sediments only at depths much shallower than the CCD because their shells are made of a more soluble form of calcium carbonate than those of foraminifera. As the new oceanic crust moves away from the spreading center and sinks isostatically, it is covered by layers of hydrothermal sediments, pteropod and foraminiferal oozes, and then siliceous sediments in sequence as the seafloor descends. The siliceous sediments can be either biogenous, if diatomaceous or radiolarian production in the overlying water column is high, or fine-grained lithogenous clays. In the Pacific Ocean, the lithogenous sediments are primarily deep-sea clays because subduction zones prevent inputs of turbidites. In the Atlantic Ocean, they are generally a mixture of deep-sea clays and turbidites.

Carbonate Compensation Depth

Below a certain depth, seawater is undersaturated with respect to calcium carbonate, and calcareous debris starts to dissolve. Below this level, the degree of undersaturation and hence the dissolution rate of calcium carbonate increases with depth (Fig. 6-10). Eventually, at the **carbonate compensation depth (CCD)**, the dissolution rate is fast enough to dissolve all of the calcium carbonate before it can be incorporated in the sediment.

The CCD depends not only on pressure and temperature, but also on other factors, especially the concentration of dissolved carbon dioxide. An increase in carbon dioxide concentration lowers the pH, making the water more acidic, and thus increases the solubility of calcium carbonate.

Carbon dioxide solubility in seawater increases as temperature decreases. The deepest water layers of the oceans were formed by the sinking of cold water in certain high-latitude regions (Chap. 8). Thus, the deepwater layers contain relatively high concentrations of dissolved carbon dioxide. In addition, deep waters below the warm surface water layer accumulate dissolved carbon dioxide as animals respire, and bacteria, and other decomposers convert organic matter to energy and carbon dioxide (Chap. 12). Therefore, older deep water (water that has been away from the surface longer) has a higher carbon dioxide concentration than newer deep water. The deep water of the Pacific Ocean is older than the deep water of the Atlantic Ocean (Chap. 8). Consequently, calcium carbonate is more soluble, and the CCD is shallower in the Pacific Ocean because of its higher dissolved carbon dioxide concentrations. The CCD is at approximately 4000 m in the Atlantic, at about 2500 m in the South Pacific, and at less than 1000 m in the North Pacific, which has the oldest deep waters with the highest carbon dioxide concentrations.

Anthropogenic Carbon Dioxide and the Carbonate Compensation Depth

The carbonate compensation depth (CCD) does not remain fixed but changes with climate, particularly with changes in the carbon dioxide concentrations of ocean water. An understanding of these changes is necessary to predict the fate of the excess carbon dioxide released to the atmosphere by the burning of fossil fuels (Chap. 7, **CC9**).

Atmospheric carbon dioxide, carbon dioxide dissolved in seawater (primarily as carbonate and bicarbonate ions), and carbon locked up in calcareous sediments or sedimentary rocks are all involved in the same biogeochemical cycle. Some of the excess carbon dioxide released to the atmosphere has been absorbed by ocean water, and some may have been captured in calcareous

sediments. However, the higher carbon dioxide concentrations lower seawater **pH** and will move the CCD to shallower depth, causing more calcium carbonate to dissolve (**Chap. 5**). The effects of this ocean **acidification** on marine organisms in the surface layers of the oceans are a primary concern. However, as more calcium carbonate dissolves, this will result in an even higher concentration of carbonate and, therefore, carbon dioxide in the deep layers of the oceans. Eventually, this will cause more carbon dioxide to be released to the atmosphere when the deep waters are brought to the surface. How fast this will take place is, as yet, unknown. However, because atmospheric carbon dioxide and the CCD have varied substantially throughout the Earth's history, we may be able to find an answer to this and other questions about the effects of ocean acidification in the record preserved in deep-sea sediments. Indeed, existing evidence suggests that release of carbon dioxide to the atmosphere from carbon dioxide-rich deep ocean waters may have caused or contributed to major warming events in Earth's history, such as the interglacial periods of the last **ice age**.

This page titled 6.4: Biogenous Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.5: Hydrogenous Sediments

The input of hydrogenous sediments to the ocean floor is small in comparison with the inputs of biogenous and lithogenous material. Nevertheless, hydrogenous material can be an important component of the sediment, particularly in some areas where the accumulation rate of biogenous and lithogenous material is low.

Seawater is generally undersaturated with most dissolved substances. Consequently, hydrogenous sediments are formed under special conditions where the chemistry of ocean water is altered. Various types of hydrogenous sediment form under different sets of conditions. The types include hydrothermal minerals, manganese nodules, phosphorite nodules and crusts, carbonates, and **evaporites**.

Hydrothermal Minerals

Because the Earth's crust is thin at the **oceanic ridges**, more heat flows through the seafloor from the **mantle** at the ridges than elsewhere. The excess heat drives a series of **hydrothermal vents** that discharge hot water (sometimes several hundred degrees Celsius) into the surrounding cold seawater. The hot water is **anoxic**, devoid of dissolved oxygen, when discharged, and it contains high concentrations of metal sulfides, including iron sulfide and manganese sulfide. These sulfides are soluble in the absence of dissolved oxygen, but they are oxidized to form insoluble hydrous oxides as the vent plumes mix with much larger volumes of the surrounding oxygenated seawater. The hydrous oxides precipitate to form a cloud of fine, metal-rich particles. These particles sink to the seafloor and accumulate as metal-rich sediment primarily in the area surrounding the vent.

In addition to iron and manganese, hydrothermal vent deposits contain high concentrations of other metals, including copper, cobalt, lead, nickel, silver, and zinc. Although some particles accumulate to form metal-rich sediment near the vents, many fine particles formed at the vents are probably transported large distances before settling and provide an input of hydrogenous particles to sediments throughout the deep-ocean basins. This may also be the origin of the material that forms the manganese nodules described later in this chapter.

The mechanism of hydrothermal vent circulation is not fully understood. The excess heat flow at the ridge is believed to drive the mechanism. Water within the rocks and sediments beneath the hydrothermal vents is heated and rises through the vents into the water above (**CC1**). This rising water is replaced by cold seawater that seeps through cracks in the rocks and sediment at other locations nearby and is transported to the vent through the rocks and sediment (**Chap. 15**). As the seawater migrates to the vent through the sediment and rocks, it is heated, its oxygen is depleted by reaction with sulfides, and its pH drops so that it leaches salts and metal sulfides from the sediment and surrounding rock before it is discharged at the vent. The vents are, in essence, parts of convection cells (**CC3**, **Fig. 15-19**).

Hydrothermal vents were not discovered until the late 1970s. They are surrounded by commercially attractive deposits of metal-rich sediment, and they also support unique communities of organisms, many of which are found nowhere else on the Earth (**Chaps. 12, 15**). Hydrothermal vents are present throughout the oceanic ridge system, but they are limited to small areas on the ridge that are often separated by substantial distances. Whether the metal-rich sediment is abundant enough to be commercially exploitable is not known. In addition, it is uncertain whether the unique biological communities could be safeguarded if the minerals were commercially exploited.

Along the axis of the Red Sea, where a new oceanic ridge is forming, hydrothermal vents have discharged huge quantities of hydrothermal minerals. The water in the deep basins of the Red Sea is isolated from and does not mix with open-ocean waters. As a result, hot water discharged by the hydrothermal vents has accumulated in the Red Sea deep basins because, even though the water is hot, it also has very high salinity and thus high density (**CC6**). The high density strongly inhibits vertical mixing of this deep water with Red Sea surface water. Therefore, the hot, high-salinity, deep water is also anoxic, which prevents the minerals discharged by the hydrothermal vents from being oxidized. These factors have caused metal-rich **hydrothermal minerals** to accumulate in large quantities in sediments, and high concentrations of suspended sediments to accumulate in Red Sea deepwater areas. The deposits are so extensive that test mining was performed in 1979, but no further mining has occurred yet, as the minerals are still available at lower cost from mining on land. Nevertheless, these hydrothermal minerals that have collected on the Red Sea floor may eventually represent an economic resource as mineral scarcities grow in the future.

Undersea Volcano Emissions

Undersea volcanic activity is not limited to the oceanic ridges. Volcanoes also erupt under the sea at some hot spots and at some convergent plate boundaries, such as the Mariana Arc. Sulfide-rich water, comparable to the hydrothermal fluids discharged at oceanic ridge vents, is discharged into the water column at some locations on these undersea volcanoes. The sediments in the areas surrounding these discharges sustain populations of biological communities similar to those found surrounding oceanic ridge hydrothermal vents, and sediments in some of these areas are likely to be similarly rich in some metals. There is particular interest in studying these communities and discharges because these undersea volcanoes are located at much shallower depths than most oceanic ridge hydrothermal vents, making them easier to reach for study. Even more importantly, some of these shallower vents are different because they discharge into the upper layers of the ocean, where photosynthesis also occurs. This raises many questions about how these relatively shallow discharges and communities may affect and/or interact with the chemistry and biology of the surrounding waters.

Manganese Nodules

Manganese nodules are dark brown, rounded lumps of rock, many of which are larger than a large potato. Enormous numbers of nodules litter parts of the deep-ocean floor (**Fig. 6-12a**). They are potentially valuable as a mineral resource because they usually consist of about 30% manganese dioxide, about 20% iron oxide, and up to 1–2% of other metals that are much more valuable, including copper, nickel, and cobalt.

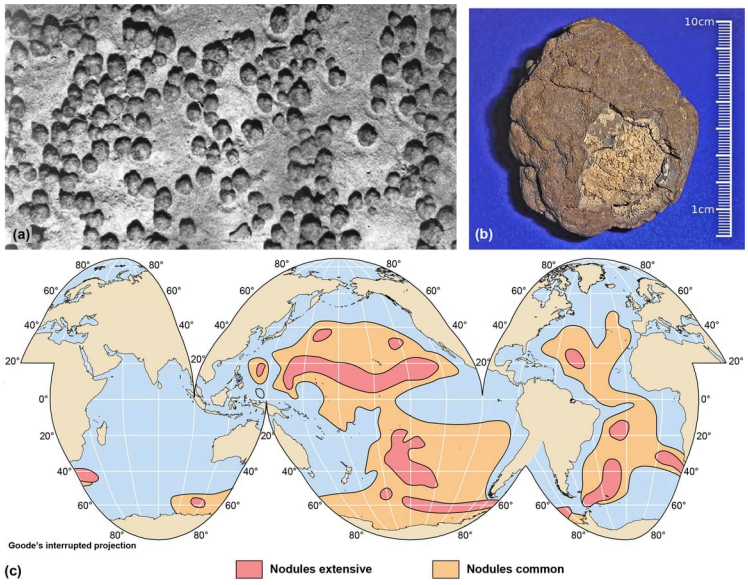


Figure 6-12. (a) Manganese nodules cover certain areas of the deep-ocean floor, such as this area of the northeastern Atlantic Ocean. (b) A typical manganese nodule. The nodule is made up of concentric layers. (c) The areas of the greatest density of manganese nodules are on the abyssal floor of the North Pacific Ocean between about 10°N and 20°N and in the central South Pacific Ocean.

Manganese nodules are probably formed by various mechanisms and from various sources of manganese, iron, and other elements. These chemicals may come from the dissolved ions in river water, from seawater leaching of volcanic materials in the ocean, or from hydrothermal vents on the oceanic ridges. At present, the major source of minerals in the nodules is thought to be oceanic ridge hydrothermal vents. Particles of colloidal size (submicroscopic) are hypothesized to form at the hydrothermal vents, be transported throughout the oceans, and accumulate by adsorption on the surface of the nodule, perhaps aided by microbial action.

Most nodules initially form around a large sediment particle (e.g., a shark's tooth) and are then built up in layers (**Fig. 6-12b**) at the very slow rate of about 1 to 10 mm per million years. Nodules are most common where the sedimentation rate is extremely slow. In areas where sediment accumulates rapidly, nodules are buried by new sediment before they have time to grow. Where nodules do form, occasional disturbance of the nodule by marine organisms is hypothesized to be necessary to prevent the nodule from being buried.

The areas of greatest accumulation of manganese nodules are the deepest parts of the Pacific Ocean (**Fig. 6-12c**), where the

sedimentation rate is very low. The greatest density of manganese nodules is in the region of the Pacific Ocean between about 10° and 20°N latitude (**Fig. 6-12c**).

Phosphorite Nodules and Crusts

Phosphorite is a mineral composed of up to 30% phosphorus. Phosphorite nodules form in limited areas of the continental shelf and continental slope and on some seamounts. Phosphorite nodule formation likely requires low dissolved oxygen concentrations in bottom waters and an abundant supply of phosphorus. These conditions are present in upwelling regions where productivity is high. In such regions, the decomposition of falling detritus depletes oxygen in the bottom waters and supplies relatively large quantities of phosphorus, which is released as phosphate when the organic matter decays.

Phosphorite nodules grow slowly (1 to 10 mm per 1000 years). In contrast to manganese nodules, they do not form concentric layers but instead grow only on the underside, accumulating phosphate released by the decomposition of organic matter. Because very large deposits of phosphates are available on land, phosphorite nodules are unlikely to become commercially valuable.

Carbonates

Many **limestone** rocks lack **fossils**. Some consist of biogenous sediments that have lost their fossils through **diagenetic** changes (discussed later in this chapter), but others consist of hydrogenous sediments formed by direct precipitation of calcium carbonate. Calcium carbonate precipitates from seawater under conditions that were apparently widespread in the oceans at various times in the past, but are now present only in very limited regions, such as the Bahamas.

Calcium carbonate precipitation occurs when the concentration in the water is higher than the calcium carbonate saturation solubility. This is more likely when water temperatures are high and the concentration of dissolved carbon dioxide is low, raising the pH (**Chap. 5**). High temperatures reduce the solubility of carbon dioxide in surface waters, allowing some dissolved carbon dioxide to escape into the atmosphere. Carbon dioxide concentrations in surface waters can also be reduced by high rates of photosynthetic production (**Chap. 12**, **CC14**).

In the present-day oceans, both high temperature and high productivity appear to be necessary to reduce the calcium carbonate solubility enough to cause precipitation. For example, warm water flowing north through the Straits of Florida (**Chap. 8**) is upwelled across the shallow Bahama Banks (**Fig. 6-13**). Over the banks, it is heated further and sustains high primary productivity because it is nutrient-rich and sunlight is intense (**Chap. 12**, **CC14**). This causes the pH to rise, and as a result, calcium carbonate precipitates around suspended particles to form white, rounded grains called “ooids.” Ooids collect in shallow-water sediment in many areas. Where they are especially abundant in relation to other particles, the sediments are called “oolite sands.”

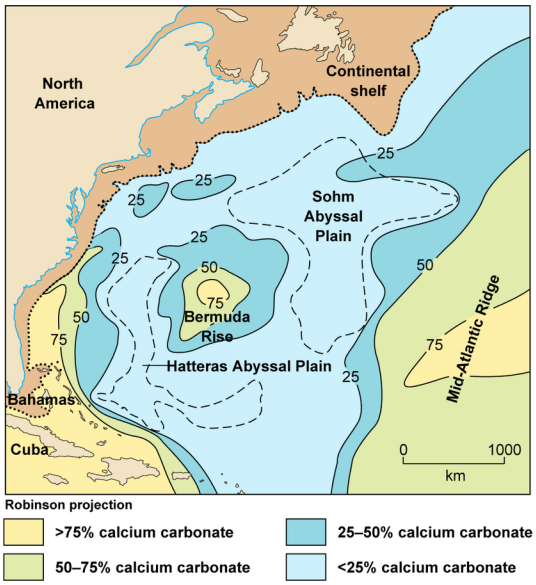


Figure 6-13. Calcium carbonate distribution in the sediments of the western North Atlantic. Values are the percentage of calcium carbonate in the sediment. The high percentages of calcium carbonate in the Bahama Banks sediments are due primarily to the dominance of hydrogenous calcium carbonate. In the shallow parts of the Bermuda Rise and the Mid-Atlantic Ridge, lithogenous sediment accumulation rates are low, allowing biogenous carbonate particles to dominate the sediments in these regions.

Evaporites

In marginal seas in arid climates, the evaporation rate of ocean water exceeds the rate of replacement by rainfall and rivers (**Chap. 7**). Under such conditions, salinity progressively increases until the seawater becomes saturated with salts. As evaporation continues, salts, called “evaporites,” are precipitated in succession: calcium and magnesium carbonates first, then calcium sulfate, and finally sodium chloride. Only a few marginal seas are evaporating in this way today, and even in these areas, salt precipitation occurs only on tidal flats or in shallow embayments around the perimeter of the sea. Areas where evaporites form today include the Dead Sea, Red Sea, and Persian Gulf. Such marginal seas must have been more abundant in the past because salt layers are present at many locations deep within ocean sediments, as well as in previously submerged parts of the continents.

Seismic profiles reveal several extensive layers of buried evaporite sediments, some more than 100 m thick, beneath the floor of the Mediterranean Sea. For such layers to have formed, the entire Mediterranean must have evaporated almost to dryness. This evaporation process apparently occurred several times when global sea level was lower and the connection between the Mediterranean and the Atlantic Ocean was broken at the sill that now lies underwater across the Straits of Gibraltar.

This page titled 6.5: Hydrogenous Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.6: Cosmogenous Sediments

Most meteorites that enter the Earth's atmosphere are destroyed before reaching the surface. As they burn up, the meteorites form particles that rain out onto the land and ocean and are eventually incorporated into sediments. These cosmogenous particles contribute only minor amounts of material to ocean sediments compared to biogenous, lithogenous, and hydrogenous sources. Nevertheless, the total input of meteorite dust to ocean sediments is estimated to be tens of thousands of tonnes each year.

Two types of meteorites hit the Earth: stony and iron-rich. The dust from the meteorites is found in sediment as cosmic spherules, which are spherical particles that show the effects of having been melted as they pass through the atmosphere. Stony meteorites form silicate spherules that are difficult to distinguish from lithogenous sediment grains. Most iron-rich spherules are about 0.2 to 0.3 mm in diameter and are magnetic, which makes them easier to find.

Recent studies have focused on finding the impact craters and debris from large meteorites that were not destroyed in the atmosphere but instead struck the oceans. A particularly large meteorite may have landed in the Caribbean Sea or the Gulf of Mexico 65 million years ago. That impact is discussed later in this chapter.

This page titled 6.6: Cosmogenous Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.7: Sediment Transport, Deposition, and Accumulation

Sediment particles are transported by ocean currents and waves in the same way that dust and sand are blown around by winds. Strong winds and fast water currents both cause particles to be suspended and carried until the wind or current speed diminishes. Particles are then deposited on the ground or ocean floor, but they may be picked up again if the wind or current speed increases. **CC4** explores the relationship between particle size and sinking rate, between current speed and sinking rate, and between current speed and resuspension of deposited particles.

Large particles sink rapidly, and high current speed is necessary to prevent them from being deposited. Once deposited, large particles are not resuspended unless the current speed is considerable. Many large particles are transported to the ocean by rivers during periods of peak river flow that follow flooding rains. When such particles reach the ocean, where currents are slower, they are either deposited or resuspended and transported by waves. In areas where wave energy is limited, the large particles are not resuspended so they are deposited at the river mouth.

Waves can generate orbital water motions that have higher speeds (**orbital velocity**) than ocean currents do. As a result, they can resuspend large particles in waters that are shallow enough for the wave energy to reach the seafloor, and sand-sized particles can be resuspended and transported in the nearshore zone. Waves can transport sand long distances along the coast in shallow water (**Chap. 11**). However, the speed of water motion in waves is reduced with depth below the surface (**Chap. 8**). Sand-sized particles that are transported offshore are deposited when they reach depths at which the wave orbital velocity is no longer high enough to resuspend them. Smaller sand and silt-sized particles that can be resuspended at lower water speeds are deposited in deeper waters than larger particles. Thus, waves tend to sort sand-sized particles in nearshore sediments: larger particles in shallow water and progressively smaller particles with increasing water depth.

Clay-sized particles are generally **cohesive**, but they tend to remain in suspension once resuspended, unless current speed is reduced substantially (**CC4**). The peak water velocity in waves is often sufficient to resuspend most cohesive sediments, and fine particles accumulate in nearshore areas only if these areas are well protected from waves and have slow currents. Such areas include both wetlands and fjords. Fine particles are transported by currents until they reach a low current area, where they are deposited permanently. The finest particles may be transported many thousands of kilometers and for many years before being deposited in the deep oceans, but they are often combined into clumps by electrostatic attraction or packaged into fecal pellets. The larger conglomerated particles are deposited more rapidly.

Aside from an occasional large particle (e.g., shark's tooth or whalebone), deep-ocean sediments remote from coastal sediment inputs are fine-grained because almost all large particles are deposited in nearshore sediments, and most particles introduced directly to the deep ocean are silt and clay-sized. For example, most marine organisms are microscopic and therefore produce small particles. Meteorite dust particles, dust particles carried by winds, and most hydrogenous particles are also small.

Turbidity Currents

In some deep-ocean sediments, particularly those on **abyssal plains** adjacent to continental slopes, layers of coarse-grained sediments (sand and gravel) can be found. Layers of such sediments are separated by layers of fine-grained sediments that normally accumulate on the deep-ocean floor. The coarse-grained sediments are transported downslope to the abyssal plain by **turbidity** currents.

Turbidity currents are similar to avalanches. In an avalanche, snow accumulates on a mountainside until it becomes unstable, breaks loose, and tumbles down the slope as an avalanche. Similarly, turbidity currents occur when sediments accumulate on the continental shelf edge and slope until they become unstable, break loose, mix with the surrounding water, and flow down the continental slope in. Turbidity currents can be triggered by disturbances such as earthquake vibrations or sudden large discharges of sediment by rivers in the same way that avalanches can be triggered by noises or storm winds. Pockets or layers of methane or **methane hydrates** (**Chap. 16**) formed by the decay of organic matter in the sediments may rupture and be released when a turbidity current occurs. This may have the effect of providing a "lubricated" layer on which the turbidity current travels, which may enhance both the size and speed of the turbidity current.

Once a turbidity current has been triggered, the disturbed sediments flow down the continental slope. As it gathers momentum, the turbidity current entrains more sediment and water, just as an avalanche gathers more snow.

Turbidity currents can flow down the continental slope at speeds of 70 $km \cdot h^{-1}$ or more, which are sufficient to suspend and retain large particles in suspension. The speed of the turbidity current is reduced when it reaches the abyssal plain, and the entrained sediment particles are deposited from the suspended sediment cloud. The larger particles are deposited first, followed by progressively smaller particles, as speed and turbulence diminish. The result is the formation of a graded bed of sediments on the abyssal seafloor within which the largest grains are at the bottom and grain size progressively decreases upward (**Fig. 6-14**). Such graded beds, called "**turbidite** layers," can be meters thick. In many locations, a number of them are separated by layers of finer sediment that were deposited slowly over the years, centuries, or millennia between successive turbidity currents.

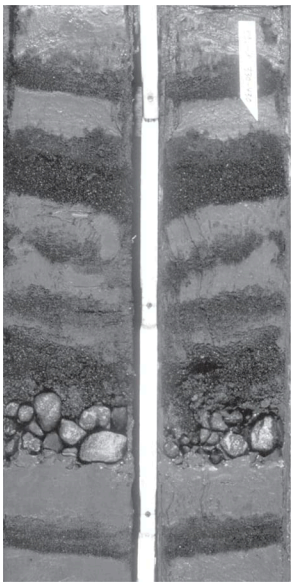


Figure 6-14. There are a number of turbidite layers in these cores. Note the sharp change in grain size at the bottom of each turbidite layer and the gradual reduction in grain size above this boundary. The convex shape of the boundaries between layers is due to distortion of the core during the drilling process. Most turbidite layers would not exhibit the very large particles seen at the lower part of some of the turbidite layers. However, these cores are from the Antarctic continental slope, where large particles are carried onto the continental shelf by glaciers.

Turbidity currents can travel long distances on the abyssal plain before finally depositing all their terrigenous sediment load. Turbidity current deposits are generally thickest at the lower end of **submarine canyons**, where they form **abyssal fans** that decrease in thickness and median grain size in a seaward direction.

Where a deep-ocean trench is present at the base of the continental slope, turbidity currents are intercepted by the trench and do not reach beyond it to the abyssal plain. This is one reason why the topography of the Atlantic Ocean abyssal floor is flatter than that of the Pacific Ocean. The topography of the Atlantic Ocean floor is buried by turbidites that have reached the oceanic ridge since the Atlantic Ocean first began to form. Because most of the Pacific Ocean floor was created after subduction zones formed around this ocean, turbidite deposits in the Pacific are rare.

Turbidity currents must be common events in the oceans, particularly where continental shelves are narrow and terrigenous sediment inputs are high. They are difficult to observe because they last only a few hours at most and occur unpredictably, although some cause tsunamis, marking their occurrence. Nevertheless, their destructive power must be respected. In fact, this destructive power is what led to the first quantitative observation of the speed and geographic extent of turbidity currents. In 1929, an earthquake occurred in the Atlantic Ocean off Nova Scotia. The turbidity current triggered by the earthquake plunged down the continental slope and snapped and buried several undersea telephone cables (**Fig. 6-15**). Because the precise times at which successive cables broke were known, the peak turbidity current speed was later estimated to be at least 70 $km \cdot h^{-1}$. This turbidity current, like many others, moved primarily down a submarine canyon and traveled more than 600 km before slowing and depositing a turbidite layer across the abyssal plain. Because turbidity currents often flow down submarine canyons, it has been suggested that their scouring action maintains or even creates the canyons.

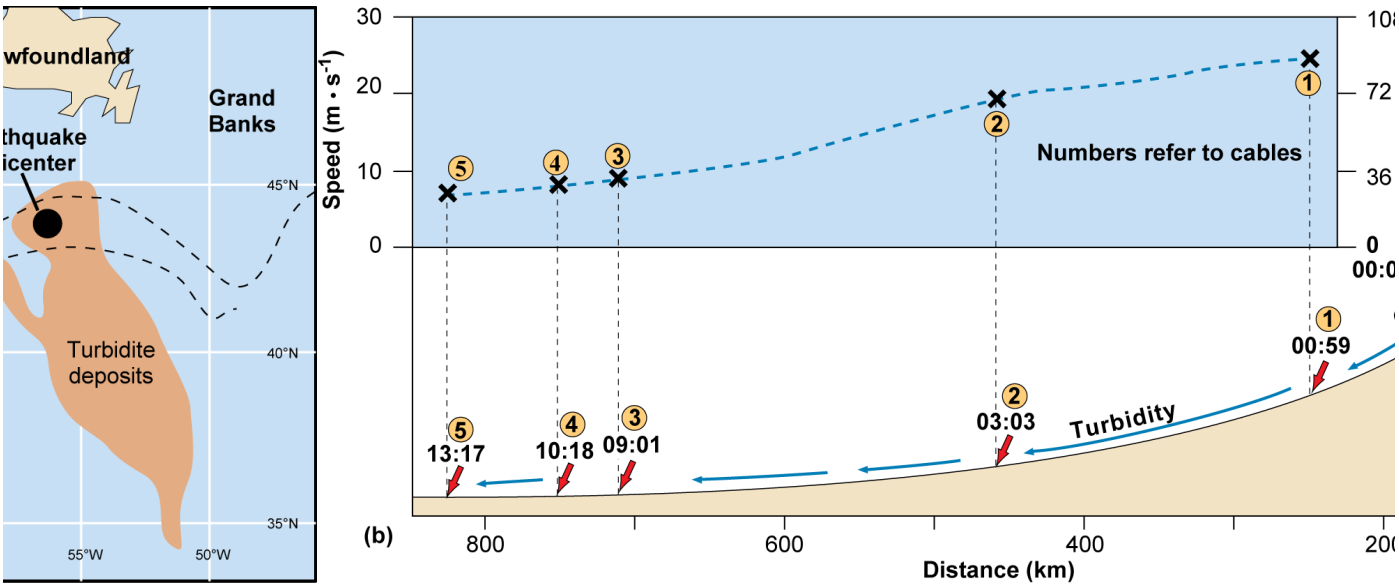


Figure 6-15. Submarine cables in the North Atlantic were severed in succession by an earthquake-triggered turbidity current in 1929. (a) The earthquake epicenter was located on the continental slope. Turbidity currents caused by the earthquake spread turbidite sediments across large areas of the deep-sea floor, extending more than 1000 km from the source. (b) The turbidity currents broke several submarine telephone cables at precisely known times. From these data, it was found that the turbidity currents had reached speeds in excess of 70 km•h^{-1}.

By studying the layers of sediments, scientists have determined that extremely large turbidity currents have occurred in the past and may therefore occur again. For example, about 20,000 years ago, a turbidity current occurred in the western Mediterranean that was estimated to have deposited 500 km^3 of sediments on the deep-sea floor, enough material to cover all of Texas with nearly 2 m of mud and sand. Another slide off the coast of Norway occurred between 30,000 and 50,000 years ago, involved more than twice the volume of sediment as the Mediterranean example, and left a scar on the continental slope that is larger than the state of Maryland.

Turbidity currents carry shallow-water organisms to great depths, where some, particularly microorganisms, may survive and adapt. Their occurrence may also affect the food web. After the 1989 Loma Prieta earthquake in California, dense schools of fish were observed feasting on the sediment-dwelling organisms exposed on the floor and sides of Monterey Canyon, where the sediment slumped and presumably triggered a turbidity current.

Debris fields surrounding some volcanic islands, such as Hawaii, provide evidence that massive slides occur periodically when parts of the volcano that created the island collapse, carrying volcanic rocks far from the island. Such avalanches may cause intense turbidity currents and tsunamis (**Chap. 9**).

Accumulation Rates

All sediments are mixtures of particles from many different sources. The accumulation rate and type of sediment are determined by the relative quantities of particles from each source deposited at each location. For example, **Figure 6-16** schematically illustrates the accumulation of sediments at two hypothetical locations. The input rate of biogenous particles to the sediment is the same at each location, but the input rate of lithogenous sediment is much higher at the first location than at the second. The sediments that accumulate at Site A are predominantly lithogenous, whereas the sediments at Site B are predominantly biogenous. However, the accumulation rate is much faster at A than at B. Because sediments are characterized by their predominant material, the Site A sediment would be called “lithogenous sediment,” and the Site B sediment would be called “biogenous sediment,” or **ooze**. An ooze is a sediment that contains more than 30% biogenous particles by volume. Where one type of organism is responsible for most of the biogenous particles, the ooze can be named after this type: pteropod ooze, radiolarian ooze, diatom ooze, foraminiferal ooze, and so on.

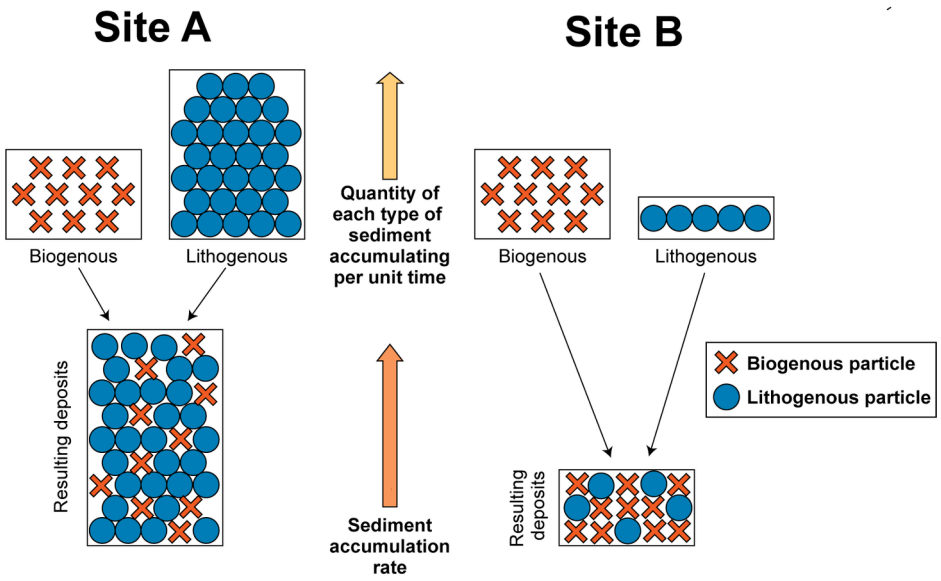


Figure 6-16. The relationships between the accumulation rates of different types of particles, and the accumulation rates and characteristics of the sediments.

Lithogenous particles are the dominant input to ocean sediments. Most lithogenous material is discharged to the oceans from land. Relatively large particles are deposited near river mouths and glaciers and in estuaries and wetlands. Sediment accumulation rates in these nearshore regions range from about 100 cm per 1000 years up to extreme rates such as the 7 m per year found in the delta of the Fraser River in British Columbia, Canada. Somewhat less but still large quantities of lithogenous sediment are transported offshore and are deposited on continental shelves. Such sediment can also reach the deep-ocean floor in areas where the continental shelves are narrow, or when turbidites flow down a continental shelf. Many continental shelves are areas of high biological productivity (**Chap. 13**). Accordingly, sedimentation rates on the continental shelf and slope and within marginal seas, such as the Mediterranean, generally are about 10 to 100 cm per 1000 years.

In the deep oceans, remote from land, lithogenous inputs are much less, and biogenous material, especially calcium carbonate, is dissolved before it can settle and be buried. Therefore, sedimentation rates in the deep-ocean basins are very low, approximately 0.1 cm per 1000 years. Under highly productive areas, on shallow seamounts or on oceanic plateaus remote from land, the increased sedimentation rate of biogenous material input can raise the overall sedimentation rate by about an order of magnitude, to approximately 1 cm per 1000 years.

Although sediments accumulate very slowly on a human timescale, they accumulate to substantial thickness in some places (**Fig. 6-17**). At a sedimentation rate of 0.5 cm per 1000 years, sediments approaching 1 km thick (850 m) can accumulate in 170 million years, which is the approximate age of the oldest (except for some remnants) oceanic crust.

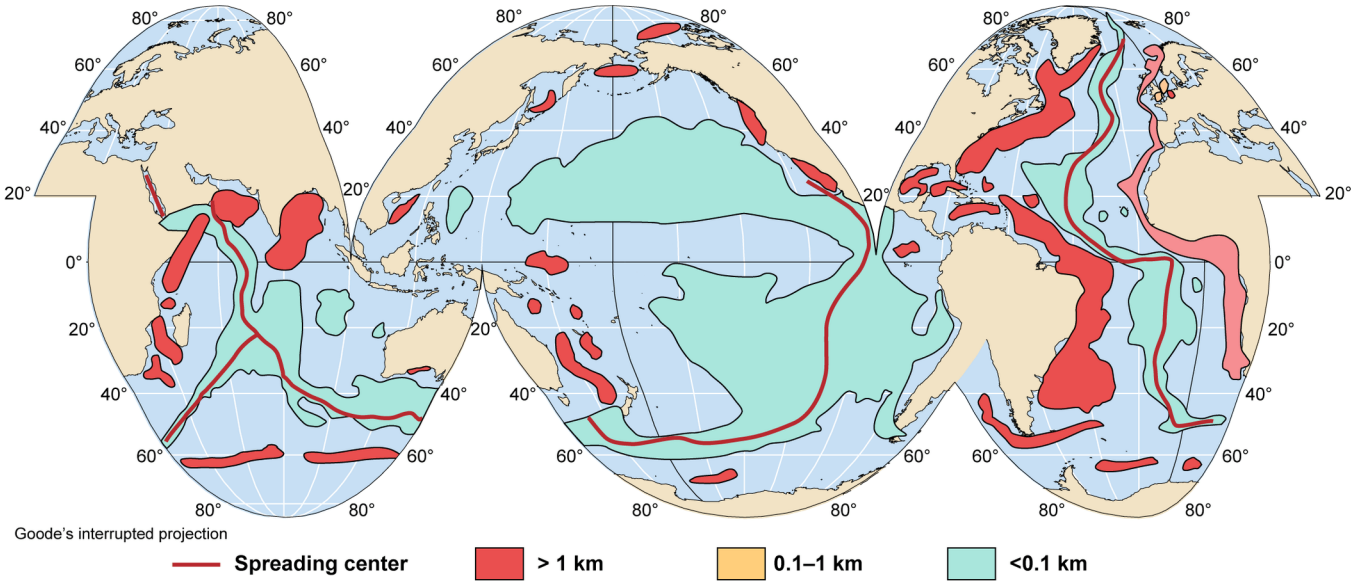


Figure 6-17. Thickness of accumulated sediments on the world's oceanic crust.

This page titled 6.7: Sediment Transport, Deposition, and Accumulation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.8: Continental Margin Sediments

Because most biogenous, hydrogenous, and cosmogenous sediments consist of silt- or clay-sized particles, they do not readily accumulate on continental shelves where currents are comparatively swift. Consequently, continental shelves are covered with lithogenous sediment of larger grain sizes, except in those areas where currents are slow or production of biogenous particles is very high.

On many continental shelves, lithogenous sediments accumulate continuously as rivers and coastal erosion supply new sediment. The particles are sorted by grain size, and grain size generally decreases with distance from shore. However, on some continental shelves, the supply of lithogenous material is very limited. For example, along the Mid- and North Atlantic coasts of the U.S., most river-borne sediment and sediment eroded from the shore is trapped in estuaries and coastal lagoons. Consequently, little lithogenous sediment is transported offshore beyond the longshore drift system (**Chap. 11**). If the shelf is wide and currents are generally swift on the outer part of the shelf, little or no new sediment is supplied to the seafloor. Such areas of the continental shelf are not bare rock, but are covered by generally coarse-grained sediments, known as relict sediments, that were deposited under a different set of ocean conditions (**Fig. 6-18**).

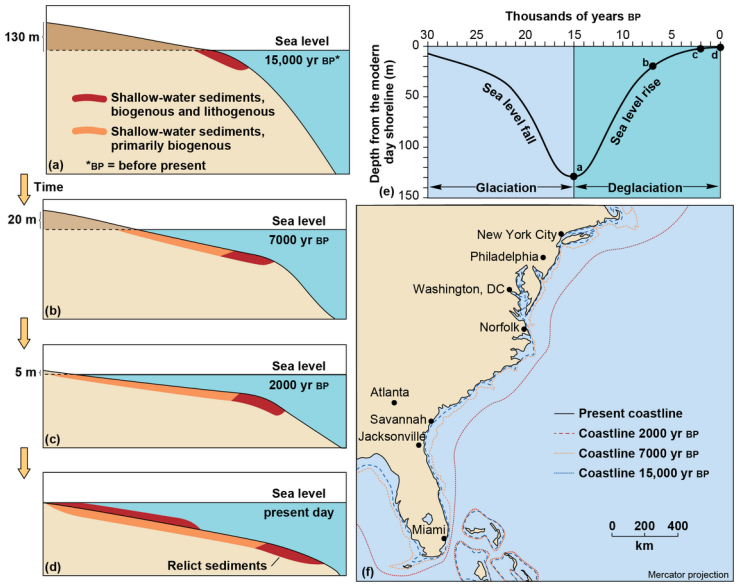


Figure 6-18. Relict sediments. (a) Sediments characteristic of shallow water were deposited on the outer continental shelf when sea lev lower. (b) As sea level rose, the relict sediments were not buried by more recent sediments, probably because of the rapid rise of sea le the trapping of lithogenous sediments in newly submerged river valleys. (c) Shallow-water sediments formed close to the new shoreline migrated inland. (d) As sea level rose more slowly, the sediments close to the now slowly retreating shoreline were buried by terrigenou shallow water biogenous sediments. (e) The history of sea-level change during the past 30,000 years shows the rapid rise between 19 and 4000 years ago and the much reduced rate of rise in the past 5000 years. (f) Locations of the Atlantic coastline of North America years ago and today. Relict sediments are found primarily near the coastline location of 15,000 years ago.

Relict sediments, including those on the outer continental shelf of the U.S. East Coast, contain terrestrial fossils and shells of organisms, such as oysters, that live only in water less than a few meters deep. These materials are too big to have been moved by currents, but are now under 100 m or more of water. The reason is that the relict sediments were deposited several thousand years ago when the sea level was lower, and the coastline was near what is now the edge of the continental shelf.

Sea level determines the distribution and age of relict sediments. Much of the detailed history of recent sea-level change has been learned through studies of the location of relict sediments. Sea level has been rising from its most recent low point about 19,000 years ago to the present day, although not at a uniform rate.

This page titled 6.8: Continental Margin Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.9: Distribution of Surface Sediments

Surface sediments are the materials currently accumulating on top of older sediment. The older sediment buried below the surface layer may have a different character if the deeper layer was deposited under different conditions.

The distribution of **pelagic** surface sediments and their current rate of accumulation are shown in **Figure 6-19**. Continental margin sediments are not shown because they are lithogenous except in areas dominated by coral reefs and in some areas of calcium carbonate precipitation. Each type of sediment and the factors that govern its occurrence are described briefly in the sections that follow.

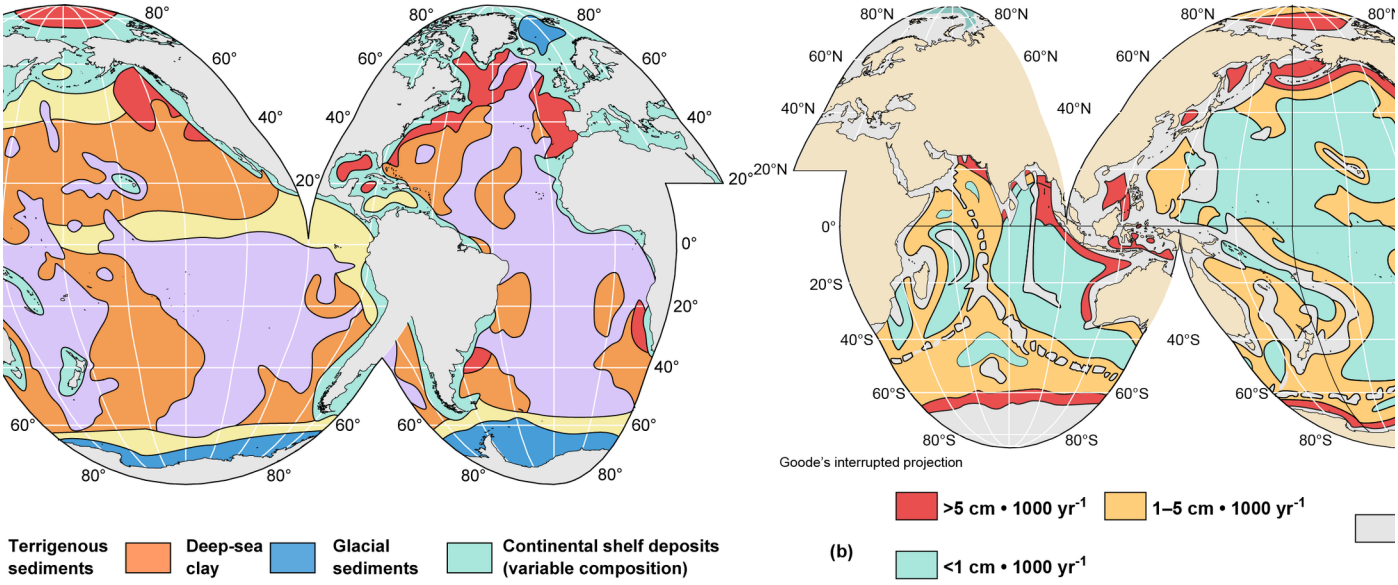


Figure 6-19. Present-day sedimentation. (a) The distribution of present-day surface sediments of the world oceans. (b) Approximate sedimentation rates in the present-day oceans. Note the higher rates of sedimentation near the continents. Surface sediments in these areas are predominantly terrigenous. Sediments are predominantly red clays in areas where the sedimentation rate is lowest. Note the generally slower rate of sediment accumulation on the abyssal floor of the Pacific Ocean compared to that of the Atlantic Ocean.

Radiolarian Oozes

In the Pacific and Indian Oceans, under a region of high productivity that extends in a band across the deep oceans at the equator (**Chap. 12**), surface sediments are fine muds that consist primarily of radiolarian shells. Although both calcareous and siliceous organisms grow in abundance in this upwelling region, calcareous material dissolves and does not accumulate in sediments below the CCD. Radiolaria (**Fig. 6-9**) are prolific in tropical waters, and the rate of input of radiolarian shells to the sediment is much higher there than in other deep-sea mud areas. Because siliceous material dissolves slowly, many radiolarian shells are buried in the surface sediment before they fully dissolve.

The rate of input of fine-grained lithogenous particles that are resistant to dissolution is approximately the same in the equatorial band and in the adjacent areas where deep-sea mud accumulates. However, the rate of accumulation of radiolarian shells, and therefore the overall sedimentation rate, is much higher in the equatorial band. Therefore, the deep-sea lithogenous mud particles are “diluted” to become a minor component in radiolarian oozes.

The Atlantic Ocean has no band of radiolarian ooze because the accumulation rate of lithogenous sediment is much higher there than in the tropical Pacific and Indian Oceans, primarily due to the relatively shallow seafloor and the relatively deep CCD. The radiolarian content of tropical Atlantic sediments is diluted and masked by the greater contributions of calcareous and lithogenous sediment. Hence, the overall sedimentation rate in the tropical Atlantic Ocean exceeds the sedimentation rate in the tropical Pacific and Indian oceans.

Diatom Oozes

Diatoms dominate the siliceous phytoplankton in upwelling areas except in the tropical upwelling zone, where radiolaria dominate. Upwelling and abundant diatom growth occur in a broad band around Antarctica and in the coastal oceans along the west coasts of continents (eastern ocean margins) in subtropical latitudes (**Chaps. 8, 12**). Diatom oozes dominate in these locations if the sedimentation rates of lithogenous and calcareous particles are relatively slow.

Aside from large coastal inputs from Antarctica's glaciers, sedimentation in the Southern Ocean is limited because of the lack of landmasses and river inflows. Hence, the accumulation rate of lithogenous particles is slow in the band of high diatom productivity that surrounds Antarctica, and the sediments are dominated by diatom frustules.

Terrigenous sediment inputs are also limited on the west coasts of North and South America because of the limited drainage areas of coastal rivers and the trapping of terrigenous sediment in the trenches that parallel these coasts. Because the seafloor is deep in the Pacific coast upwelling regions and the CCD is relatively shallow in the Pacific Ocean, calcareous particles are dissolved and do not accumulate in sediments. Diatom frustules therefore dominate the sediments in the north and south subtropical Pacific offshore basins. In contrast, on the eastern ocean boundaries in the Atlantic Ocean (southern Europe and West Africa) and Indian Ocean (Australia), in areas where the seafloor is relatively shallow, calcareous particles accumulate faster than diatomaceous particles, and the sediments are calcareous.

Diatoms are very abundant in Northern Hemisphere cold-water regions (Bering Sea and North Atlantic), just as they are near Antarctica. However, lithogenous sediment inputs are greater in the Northern Hemisphere than in the Southern Ocean, and much of the seafloor is above the CCD. Therefore, diatomaceous particles are diluted by calcareous and lithogenous particles, and diatomaceous oozes accumulate only in limited deep areas of the Bering Sea and northwestern Pacific Ocean.

Calcareous Sediments

In areas where the seafloor is shallower than the CCD, calcareous sediment particles accumulate faster than deep-sea clays. Therefore, where the seafloor is shallower than the CCD and terrigenous inputs are limited, calcareous particles are a major component of the sediment. Calcareous sediments are present on oceanic plateaus and seamounts and on the flanks of oceanic ridges.

Deep-Sea Clays

Few large particles are transported into the deep oceans remote from land unless they are carried by turbidity currents (mostly in the Atlantic Ocean). Most deep-ocean areas far from land sustain relatively poor productivity of marine life (**Chap. 12**). Because of the great depth, the relatively small quantities of biogenous material that fall toward the seafloor are mostly dissolved before reaching, or being buried in, the sediment. This is especially true for calcareous organisms in areas where the seafloor is below the CCD. Therefore, the particles deposited on the deep-sea floor far from land are mainly fine lithogenous quartz grains and clay minerals.

Deep-sea clays (sometimes called "red clays") are reddish or brownish sediments that consist predominantly of very fine-grained lithogenous material. Everywhere but at high latitudes and in a high productivity band extending across the equatorial Pacific and Indian Oceans, they cover the deep-ocean floor that is remote from land. The reddish or brownish color is due to oxidation of iron to form red iron oxide, which we know as rust, during the slow descent of the particles to the seafloor. Oxidation continues on the seafloor until the particles are buried several millimeters deep. The sedimentation rate is very low, so burial is very slow, and the interval during which oxidation occurs is very long.

Siliceous Red Clay Sediments

In the deep basins of the North and South Pacific, South Atlantic, and southern Indian Oceans, transitional areas are present between the central deep basin, whose surface sediments are red clays, and higher latitudes, where surface sediments are diatom oozes. Sediments in these transitional areas are mixtures of deep-sea clay and diatomaceous sediments in varying proportions. The transitional sediments are often classified with deep-sea clays and often called "deep-sea muds."

Ice-Rafted Sediments

In the Arctic Ocean, northern Bering Sea, and a band immediately surrounding Antarctica, the sediment consists largely of lithogenous material carried to the oceans by glaciers. It includes sand, pebbles, and even boulders up to several hundred kilometers offshore from the glaciers. Some lithogenous material is released directly to the ocean when the edge of the ice shelf or glacier melts. Some is transported by icebergs, also called “ice rafts,” that break off the glaciers. The material released and deposited as the ice melts is known as glacial, or “ice-rafted,” debris. Because glaciers can carry huge quantities of eroded rock, the sedimentation rate can be very high. Biogenous particles therefore, are diluted by and mixed with much larger volumes of ice-rafted lithogenous material.

Terrigenous Sediments

Sediments dominated by terrigenous particles are present near the mouths of major rivers and may extend hundreds of kilometers offshore. Such sediments are deposited in the northern Arabian Sea, the Bay of Bengal, the Gulf of Mexico, and the Atlantic Ocean near the mouth of the Amazon River in Brazil. Terrigenous sediments transported by turbidity currents are present at the foot of the continental slope in the North Atlantic. On the North Pacific coast of North America, rivers carry large sediment loads from the glaciers in the mountains of British Columbia and Alaska. Many of the glaciers terminate and melt before reaching the ocean. Ice rafts are few, so coarse sand, pebbles, and boulders are not transported beyond the immediate vicinity of the glacier termination. However, large quantities of glacially ground sand, silt, and mud are transported to the ocean by rivers, and offshore by waves and currents.

Hydrothermal Sediments

The central basin of the Red Sea is the only area known to have hydrothermal mineral deposits as the dominant sediment type. The other locations where such sediments accumulate are small and scattered throughout the oceans, mostly on the oceanic ridges. The extent of such deposits is not known because only a very tiny fraction of the oceanic ridge system has been surveyed by methods that reveal hydrothermal vents and their associated sediments. Vents have been found on oceanic ridges in all oceans and are likely to be present at intervals of a few kilometers to hundreds of kilometers along the entire oceanic ridge system.

This page titled 6.9: Distribution of Surface Sediments is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.10: The Sediment Historical Record

Sediments accumulate continuously by deposition of new layers of particles on top of previously deposited layers. Therefore, individual layers that make up the sediment below the sediment surface were deposited progressively earlier as depth below the sediment surface increases. The type of particles deposited at a specific location and time depends on many factors, including, the proximity to land, whether rivers or glaciers flowed into the adjacent coastal ocean, the type and amount of dust in the atmosphere, and the composition and production rates of organisms in the water column. These factors do not remain constant over geological time because they are affected by plate tectonic movements, climate changes, and changes in volcanic activity, all of which may be related to each other in complex ways. Such changes, singly or in combination, can move coastlines, change seafloor depth, raise mountains, alter rivers, create and melt glaciers, change atmospheric dust composition by creating or reducing deserts or by altering the amounts of volcanic material injected into the atmosphere, and change the temperature, salinity, and nutrient distributions in the oceans. Each of these changes affects the type of sediment that accumulates. Hence, buried sediments may be very different from those currently accumulating.

Buried sediments provide information about the conditions at the time they were buried. Ocean sediments, therefore, can provide a history of plate tectonics and climate changes in that location covering about 170 million years, the approximate age of the oldest remaining oceanic crust in the major oceans. Older fragments of oceanic crust are known to exist, but it is not yet known whether they support undisturbed sediment layers. To read the history of ocean sediment layers, we must be able to determine the date at which a particular layer of sediment was deposited. Then, we must examine the composition of the sediment particles to reconstruct the sedimentation conditions at the time the layer was formed.

Stratigraphy is the study of the Earth's history through investigation of the sediment layers beneath the ocean floor (and on land, where sediments have been compressed and converted into sedimentary rock). Stratigraphic studies of ocean sediments usually involve the examination of sediment cores (**Chap. 3**), which are sampled every 1 or 2 cm (**Fig. 6-20**). Each 1–2 cm section represents the accumulation during a period of thousands or tens of thousands of years, depending on the sedimentation rate.

Figure 6-20. Sediment cores, such as this deep-sea drilling core, are first cut down the middle and then marked every 1 to 2 cm, each 1–2 cm section representing the accumulation of sediments during a period of hundreds or thousands of years in the past.

The stratigraphic record can be complicated to read. Gaps may appear in the record if previously deposited sediment was eroded because of a temporary increase in current speed or a decrease in sea level relative to the continent. Furthermore, adjacent layers may have had substantially different sedimentation rates. Biological activity also may have altered the historical record. Bottom-dwelling organisms rework, or mix, the upper few centimeters or tens of centimeters of sediment as they consume the organic matter, feed on other sediment dwellers, or build burrows. Their churning of the sediments is called bioturbation. Despite these limitations, the stratigraphic record provides vital information about the Earth's history, including previous climate changes. Understanding climate changes will enable us to better understand the factors that control changes in climate, such as those expected to result from the enhancement of the greenhouse effect.

Sediment Age Dating

The age (date of deposition) of the sediment layers is essential for stratigraphic, or sediment layering, studies. While we can date the sediments relative to one another (younger sediments are on top), determining their numerical age is often difficult. Methods for determining sediment ages include radioactive dating (**CC7**), magnetic dating, and biological dating.

Radioactive dating of sediments is difficult because often the assumptions on which the technique is based, including minimal weathering of the grains, are not met. Therefore, researchers commonly use several different methods to be certain of a measured age, including dating of sediments based on their magnetic properties. To explain how magnetic dating is done, we need to understand first that the magnetic poles appear to have been close to the Earth's north and south poles of rotation for billions of years, even though they wander slightly. However, the Earth's magnetic field completely reverses periodically for reasons we do not yet fully understand. During a reversal, the north magnetic pole is located close to the geographic South Pole and the south magnetic pole close to the geographic North Pole. The sequence and ages of these reversals are well documented from studies of terrestrial rocks. In sediments, particles that contain iron or other magnetic materials tend to align with the magnetic field during or soon after sediment deposition. As a result, the magnetic orientation of sediment varies with depth below the sediment surface, changing at each depth in the sediment column that corresponds to a magnetic reversal. Measuring the magnetic properties of the sediment layers and correlating the depth of layers where changes in magnetic field occur with the known dates of magnetic pole reversals, therefore allows us to date the sediments.

Another important dating technique is based on the fossil species found within the sediments. Many of the marine species whose remains make up the biogenous fraction of older sediments have become extinct, have been replaced by new species, or have evolved into substantially different forms. Consequently, the ages of sediments at different locations can often be matched by comparing the species compositions of organisms found in their biogenous fractions. If the sediment at a specific depth at one sampling site contains the fossil remains of all the same species that the sediment at a different depth at another sampling site contains, the sediments are likely to be the same age. Thus, we can date sediments in relation to each other and determine relative sedimentation rates.

Diagenesis

Interpreting the stratigraphic record requires an understanding of **diagenesis**, the physical and chemical changes that occur within sediment after it is buried. In some cases, during the millions of years they lie buried, minerals are altered from one form to another as weathering processes that began on land or in suspended sediment continue. However, the more important diagenetic changes occur in the pore water (interstitial water) that is trapped between the sediment grains.

Chemicals that dissolve in pore waters during diagenesis include silica and calcium carbonate, which continue to dissolve from buried siliceous and calcareous material. Also, during diagenesis, other substances are released by continuing oxidation of sediment organic matter and minerals, and these too may be dissolved in the pore water. Chemicals dissolved from the sediment particles during diagenesis migrate upward by **diffusion** toward (and in some cases into) the overlying water. As sediments accumulate, their weight compacts the underlying sediment. In the process, pore waters are slowly squeezed out and migrate upward through the sediment.

As organic matter decomposes in the sediment, dissolved oxygen in the pore water is consumed, and oxidation continues with oxygen from nitrates and then sulfates (**Chap. 12**). The sulfates are reduced to sulfide, which dramatically changes the solubility of many elements. Iron, manganese, and several other metals form sulfides that are much more soluble than their hydrated oxides, which form in oxygenated water. In sediment with enough organic matter to produce sulfides, iron and manganese are dissolved in pore waters that migrate upward. The metal sulfides migrate with the pore water until it approaches the sediment surface, where it encounters dissolved oxygen diffusing down into the sediment or introduced by bioturbation. The iron and manganese then are oxidized and redeposited on sediment grains. As sediment accumulates, these elements may be dissolved and moved upward continuously, accumulating in a layer near the sediment surface where sulfide and oxygen mix.

In contrast to iron and manganese, the sulfides of many elements (e.g., copper, zinc, and silver) are less soluble than their hydrated oxides. These metals are not dissolved in pore waters that contain sulfide, so they do not migrate up toward the oxygenated surface sediments and overlying seawater. On the contrary, these elements can diffuse with oxygenated seawater downward into the sediment, where they can be converted to sulfides and buried with the accumulating sediment. The complex migrations of elements within pore waters and the associated changes in the nature of the sediment particles must be understood if we are to deduce the nature of the sediment particles at the time of their deposition and thus read the stratigraphic record.

Diagenetic processes are also important to life. For example, the nutrients nitrate, phosphate, and silicate are transported to the sediments in detritus that falls to the sediment surface. The nutrients are released to the pore water as organic matter is oxidized and they then diffuse back into the water column, where they can be reused by living organisms.

Tectonic History in the Sediments

Much of the history of tectonic processes can be revealed by studies of the changes in sediment characteristics with depth below the sediment surface. For example, consider the sediment layers that have accumulated at locations between the oceanic ridges and continents (**Fig. 6-11**). Remember that the seafloor is progressively older with increased distance from the oceanic ridge (**Chap. 4**). However, the older seafloor was once at the oceanic ridge and sank isostatically (**CC2**) as it cooled, so it becomes progressively deeper. Remember also that all sediments are mixtures (**Fig. 6-16**). As a result, sediment deposits can reveal the history of plate tectonic movements. For example, the age of a buried calcareous sediment layer can reveal the times during which the seafloor was at depths shallower than the CCD.

In reality, reading the sediment history is complicated because important factors, such as the CCD level, have undoubtedly varied over the millennia. In addition, the locations at which certain types of organisms are abundant may have changed. Currently, siliceous diatoms abound near the poles, and siliceous radiolaria dominate near the equator. However, the distributions of these organisms might have been very different at times in history when the climate and paleogeography (**Fig. 4-9**) was different from the present.

Climate History in the Sediments

The past 170 million years of the Earth's climate history are preserved in ocean sediments, primarily in biogenous particles. Each marine species that contributes calcareous and siliceous material to ocean sediments has an optimal set of conditions for growth. Although nutrient concentrations, light intensity, and other factors are important, temperature is crucial for most species. Hence, fossils in sediments can be used to determine the geographic variations in ocean water temperatures.

Because different species of foraminifera thrive at different temperatures, the species composition of foraminiferal fossils in buried oozes can be used to study past climate. However, because some species have evolved or become extinct, the temperature tolerances of fossil species often are assumed to be the same as those of modern species. This assumption may not be correct.

Past climate information also can be obtained from measurements of the ratio of oxygen **isotope** concentrations in calcareous sediments. The basis for this method is somewhat complicated. Oxygen has three isotopes: oxygen-16, oxygen-17, and oxygen-18. Almost all ocean water molecules consist of either an oxygen-16 atom combined with two hydrogen atoms, or an oxygen-18 atom combined with two hydrogen atoms (the concentration of oxygen-17 is very small). Oxygen-16 is lighter than oxygen-18. Consequently, water containing oxygen-16 is lighter than water containing oxygen-18. The lighter water molecules evaporate more easily than the heavier molecules. Hence, when seawater evaporates, the water vapor is enriched in oxygen-16. When the Earth's climate cools, more water is precipitated and stored in the expanding glaciers and polar ice sheets. Because this water was evaporated from the oceans, oxygen-16 is transferred preferentially to the ice, the ratio of oxygen-16 to oxygen-18 in ocean water decreases, and sea level falls. Organisms that live in seawater incorporate the oxygen-16 and oxygen-18 of seawater into their calcium carbonate body parts in a ratio that is determined partly by the ratio of these two isotopes in the seawater. Therefore, the ratio of oxygen-16 to oxygen-18 in calcareous sediment can be used to deduce climatic characteristics and sea level at the time of deposition. The ratio is low when sea level is lowered by evaporation and more water is stored on land as ice and snow and high when the climate is warm and sea level is high.

In the same way that water molecules containing oxygen-16 and oxygen-18 evaporate at slightly different rates, molecules containing these oxygen isotopes react at slightly different rates in the chemical processes that produce the calcium carbonate of calcareous organisms. The ratio of oxygen-16 to oxygen-18 in the calcareous parts differs between species, but for a single species, it depends on the ratio in the seawater and the seawater temperature. At any given time, the oxygen isotope ratio of ocean surface water is relatively uniform throughout the oceans. Hence, calcareous remains of the same species deposited in different parts of the ocean at the same time have different ratios of oxygen-16 to oxygen-18 because the water temperatures were different where the particles were deposited. A low ratio of oxygen-16 to oxygen-18 indicates colder water.

Studies of sediments aimed at revealing the Earth's climate history have become more important because of growing concerns about human enhancement of the greenhouse effect. An understanding of past climates can help us understand future climate change due to both natural causes and human activities. Stratigraphic studies are becoming ever more sophisticated and are now beginning to reveal details of ancient ocean current systems and ocean chemistry, including carbon dioxide concentrations.

Such information continually improves our understanding of the complicated feedback mechanisms between the atmosphere and oceans (**Chap. 7**, **CC9**).

History of Ocean Acidification and Deoxygenation in the Sediments

As we have seen in the previous section, ocean sediments can be used to study the variation of temperatures in Earth's past. Sediments also provide a historical record of changes in ocean water chemistry. For example, past changes in the acidity of the oceans and in the extent of oxygenation of the oceans can be revealed by studies of sediments and the fossils found in sediments. The techniques used to extract acidity and oxygenation information from sediments are complex and beyond the scope of this text. However, the results of such studies provide important information on the likely effects of anthropogenic emissions on ocean ecosystems. This information suggests that acidification and **deoxygenation** may have severe adverse effects comparable to, and perhaps greater than, the adverse effects of climate change.

Evidence of Mass Extinctions

Earth has experienced at least five major mass extinctions of species. All but one of these occurred farther in the past than the record provided by current ocean sediments. These earlier extinctions can be studied by records left in sedimentary rocks that were formed from sediment deposited when each event occurred. Sedimentary rocks are more difficult to study than sediments since they are more likely to have been altered by chemical and physical processes that occurred after they were formed and uplifted by tectonic processes. However, each of the mass extinctions has been investigated and it has been learned that, in at least two of these events, the extinction was the result of climate warming due to rising atmospheric carbon dioxide levels, and associated ocean acidification and deoxygenation. The atmospheric carbon dioxide rise in most or all extinctions was likely due to increased volcanic activity. Volcanic activity can introduce large amounts of dust and gas to the atmosphere, altering global climate and causing extinctions. For example, the 1883 eruption of Krakatau, a relatively small eruption in comparison with some that have occurred in the past, is known to have affected the world climate for a decade. Large increases in volcanic activity are known to have occurred periodically in Earth's history

The rate of increase of carbon dioxide in the atmosphere during Earth's mass extinctions is much slower than what Earth is now experiencing due to anthropogenic releases, but past extinctions were not instantaneous events and instead took place over at least thousands of years. Volcanic activity can release large quantities of sulfur that have a similar but different effect on the atmosphere and oceans than carbon dioxide. Despite these differences, a better understanding of past extinctions can provide us with a better understanding of the relative effects of anthropogenic climate change, ocean acidification, deoxygenation in the future and of whether or not anthropogenic inputs may lead to another mass extinction.

Investigating a Mass Extinction

About 65 million years ago, the last of the dinosaurs became extinct. Fossil and sediment records show that, at the same time, about 70% of all species then on Earth, including more than half of all species of marine animals, became extinct. Although other extinctions have been found at different times in the sedimentary rock record, the dinosaur extinction was among the most dramatic.

For a number of years, there was a major debate about what caused this extinction. It was known that a major increase in volcanic activity occurred at about the time of the extinction event. However, dating of events such as the eruption and extinction is difficult and until recently the volcanic event, called the Deccan Traps, flood basalt eruption in China was thought to have occurred about 300,000 years before the extinction event. Instead, evidence supported the hypothesis that a meteorite impact near Chicxulub, Mexico could have been responsible for the extinction. The studies leading to this meteorite impact conclusion provide a good illustration of how past events can be deduced from sediments.

The area that is now the Yucatán Peninsula of Mexico was underwater 65 million years ago at a depth of about 500 m. Magnetic surveys in this area reveal a circular feature (**Fig. 6-21**), 180 km in diameter, that may be the buried remains of the impact crater of a huge meteorite. Studies of sedimentary deposits from many areas surrounding this circular feature, now called Chicxulub, show an unusual series of layers that correspond to deposition 65 million years ago.

Figure 6-21. The location of the Chicxulub crater, which is thought to be the impact site of a massive meteorite that may have been responsible for climate changes that, in turn, led to the extinction of the dinosaurs. The crater is at least 180 km in diameter, but it may be twice that size.

Sediments below the strange layers are fine-grained biogenous oozes of the type normally deposited at a depth of approximately 500 m. The unusual layers immediately above these normal sediments have characteristics that could be explained by the sequence of events that followed this massive meteorite impact. The deepest and oldest strange layer consists primarily of rounded, coarse grains several millimeters in diameter. These grains were not rounded by water weathering. They are glassy tektites, which form when rocks are melted by meteorite impacts, ejected into the atmosphere where they resolidify, and fall back to the Earth. This layer also contains high concentrations of iridium, an element rare on earth but more common in meteorites. An iridium-rich layer is found in 65-million-year-old sediments from all parts of the oceans but decreasing in concentration with distance from Chicxulub, providing evidence that a major meteorite hit spread dust through the entire Earth's atmosphere at this time and that the impact took place near Chicxulub.

The next higher layer consists of coarse-grained sediments that contain fossilized wood and pinecone fragments. These materials are not normally found in marine sediments. It is hypothesized that this layer was created when the meteorite impact generated a huge tsunami (**Chap. 9**) that smashed onto land, tore rocks and trees loose, and carried them back to sea as it receded. The tsunami may have been so large that it sloshed back and forth across the Gulf of Mexico and around the world for at least several days. It must have been many times higher than the tsunamis that hit Indonesia in 2004 and Japan in 2011, perhaps several hundred meters or more high.

Above the two older extraordinary layers is a series of layers of progressively finer-grained sand mixed with other mineral particles that are not normally found in marine sediments. These progressively finer-grained layers may consist mostly of beach sands and eroded soil particles suspended and transported to the deep water by waves generated by the meteorite impact. As wave energy decreased after the initial meteorite impact and tsunami, progressively smaller particles would have been deposited in the same way that turbidites are. Much of the finer-grained material overlying the graded sequence may have come from the fallout of dust ejected into the atmosphere during the impact.

Above these finer-grained layers, the sediments finally grade into normal biogenous oozes, similar to those that underlie the unusual layers, but with fewer and different species. For example, below the anomalous layers, foraminifera species were large, abundant, and diverse, but above they were almost absent, smaller and of many fewer species.

There are many possible variations of the details of the events that deposited the anomalous sediment layers 65 million years ago. For example, one suggestion is that the tsunami may not have been caused directly by the impact of the meteorite on the ocean. Instead, the impact may have caused a magnitude-11 earthquake (about a million times more powerful than the 1989 Loma Prieta earthquake near San Francisco), which could have sent giant turbidity currents down the continental slope and thus generated massive tsunamis.

Additional details of the Chicxulub impact have emerged as studies of the impact continue. For example, it has been estimated that the object that hit the Earth was about 10 km in diameter, that the impact created a crater more than 180 km in diameter,

and that it must have caused strong earthquakes, and most likely a massive fireball. The meteorite impacted carbonate and sulfur-rich rocks and it has been estimated that 100–150 gigatons of sulfur were released to the atmosphere. This would have caused devastating acid rains, while the dust blown into Earth's upper atmosphere would have blocked enough sunlight to substantially cool Earth's climate for decades or longer. This amount of dust would also drastically reduce the sunlight available for photosynthesis. Photosynthetic organisms are the base of both terrestrial and marine food chains, so the loss of available food would have caused many species populations to crash to, or close to, extinction even if they were not adversely affected by the initial firestorm, earthquakes, and tsunami or the subsequent acid rain.

As we can see, the odd sediment layers in the Gulf of Mexico have begun to reveal a detailed picture of what must have been one of the largest events in Earth's history. This evidence appeared to support a hypothesis that the mass extinctions were caused by the meteorite impact rather than by a major increase in volcanic activity. However, as often happens in science, the simple answer is not always correct. More recent studies of fossils in sediments from before and after the impact, and better dating of sediment layers have now revealed the impact is likely not the single cause of the mass extinction. This more recent data now suggests that a series of massive volcanic eruptions took place over a time period that straddles the extinction, with at least four large pulses of volcanic activity, one of which preceded the extinction. Also, more detailed dating of the disappearance of species in the fossil record showed that a number of species became extinct before the impact, while a comparable number of species became extinct after the impact.

It now appears that this mass extinction, that included an end to the dinosaurs and led to the rise of mammals, was caused by a double disaster. First, there was an episode of increased volcanic activity on the planet, causing rapid climate warming and almost certainly acidification and deoxygenation (as scientific evidence supports likely occurred during other extinctions). Second, during the period of increased volcanic activity, there was the massive Chixulub meteorite impact.

The sediments related to this mass extinction will undoubtedly be studied extensively, and their meaning debated for many years, and this double disaster scenario may change again, but there are two important lessons to learn. First, environmental events do not always have a simple single cause. They are most often a combination of several processes interacting with each other. Second, our tendency to think of natural events on human time scales can be misleading. A mass extinction that takes 100,000 years or longer to take place is a very abrupt change on the timescale that controls Earth's climate and life. We think of the anthropogenic impacts on Earth as having taken place over a long period of centuries, but in geological time, humans have caused changes that are far faster than anything seen on Earth before, or at least since the Earth's formation and early history.

Evidence of Other Impacts

Researchers believe that there may have been a number of other large impacts similar to the Chicxulub event at different times in the past, and they are searching for evidence of the impact craters. A number of locations have been identified as possible impact sites, including an area off the northwestern coast of Australia, where there may have been an impact about 250 million years ago, and an area at the mouth of the Chesapeake Bay, where an impact may have left an 85-km-wide crater about 35 million years ago. Decades of careful research will undoubtedly be needed to demonstrate that meteorite impacts did or did not occur in these areas and others. If confirmed, additional research will be needed to investigate and identify the effects that those impacts may have had on the biosphere.

This page titled 6.10: The Sediment Historical Record is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.11: CHAPTER SUMMARY

Classification of Sediments.

Sediments can be classified by their predominant grain size or by the origin of the majority of their particles. Grain size ranges are usually classified as gravel, sand, silt, and clay, in order of decreasing size.

Lithogenous Sediments.

Lithogenous sediment particles are primarily chemically resistant minerals produced by weathering of continental rocks. They are transported to the oceans by rivers, glaciers, waves, and winds. Lithogenous sediments are deposited in thick layers beyond the mouths of some rivers, but many rivers trap sediments in their estuaries and deltas. Fine-grained dust, especially from deserts, can be transported by winds for thousands of kilometers before settling on the sea surface and eventually sinking to the seafloor. Some volcanic eruptions eject large quantities of ash that reach ocean sediment in a similar way.

Biogenous Sediments.

Biogenous sediments are the remains of marine organisms. In most cases, the organic matter is decomposed before the hard inorganic parts are deposited and buried.

Some marine species, including species of phytoplankton and zooplankton, have calcareous or siliceous hard parts, both of which dissolve in seawater. Most siliceous diatom frustules dissolve very slowly and become included in sediments. The intricate siliceous shells of radiolaria generally dissolve before being buried in sediments in all but tropical waters, where radiolaria are abundant. Because the solubility of calcium carbonate in seawater increases with pressure, the rate of dissolution of calcareous hard parts increases with depth. Calcareous shell material has two forms: calcite and aragonite. The aragonite of pteropods is dissolved much more easily than the calcite of foraminifera. Pteropod oozes are present only in relatively shallow water, whereas foraminiferal oozes predominate at intermediate depths. Below the carbonate compensation depth (CCD), little or no calcareous material survives dissolution to be incorporated in the sediment. The CCD varies between oceans because of differences in dissolved carbon dioxide concentrations, and it also varies historically with climate changes.

Hydrogenous Sediments.

Hydrogenous sediments, which are less abundant than terrigenous (lithogenous) or biogenous sediments, are deposited by the precipitation of minerals from seawater. They include manganese nodules that lie on the sediments in areas of the deep ocean, phosphorite nodules and crusts on continental shelves beneath oxygen-deficient water, calcium carbonate deposits in a few shallow areas where water temperature is high and dissolved carbon dioxide concentration is low, and evaporites (salt deposits) in coastal waters with limited water exchange with the open ocean, low rainfall, and high evaporation rate. Hydrogenous sediments also include metal-rich sediments accumulated by the precipitation of minerals from water discharged by hydrothermal vents.

Cosmogenous Sediments.

Cosmogenous sediment particles are fragments of meteorites or tektites created by meteorite impacts. They constitute only a tiny fraction of the sediments. Many are found as cosmic spherules.

Sediment Transport and Deposition.

Large particles are deposited quickly, unless current speeds are high, and they are difficult to resuspend. Large particles collect close to river mouths, glaciers, and wave-eroded shores. Because orbital velocity in waves is higher than ocean current speeds, large particles can be transported in the nearshore zone but cannot be transported far from shore. Smaller particles may be carried long distances before being deposited.

Turbidity currents that resemble avalanches carry large quantities of sediment down continental slopes and onto the abyssal plains. Graded beds of turbidites are present in deep-ocean sediments, and they are common on the abyssal floor near the continents with passive margins.

Sediment Accumulation Rates.

Sediments are mixtures of particles of different origins. Sediment characteristics are determined by the relative accumulation rate of each type of particle. Sediment accumulation rates range from about 0.1 cm per 1000 years in the deep oceans to more than 1 m per year near the mouths of some rivers.

Continental Margin Sediments.

Because fine-grained sediments are transported off the continental shelf by currents, shelf sediments are generally sand- and silt-sized lithogenous particles. Relict sediments, deposited when sea level was lower and containing remains of terrestrial and shallow-water organisms, are present in some areas on the continental shelf.

Distribution of Surface Sediments.

Deep-sea clays are present far from the continents on the abyssal plains. Radiolarian oozes occupy a band of high primary productivity that follows the equator, but not in the Atlantic, where the radiolarian shells are diluted with lithogenous sediment. Diatom oozes are present in high-latitude areas and in other upwelling areas where primary productivity is high and lithogenous inputs are low. Sediments dominated by terrigenous particles are deposited near the mouths of major rivers. Calcareous sediments dominate where the seafloor is shallower than the CCD and lithogenous inputs are low. Hydrothermal sediments are deposited in some areas on oceanic ridges.

The Sediment Historical Record.

Sediments accumulate continuously, providing a record of the Earth's history. The stratigraphic record can be complicated because sediment layers may be disturbed or chemically and physically altered by diagenesis. Stratigraphic studies of ocean sediments have revealed much of the Earth's climate and tectonic history during the past 170 million years. For example, such studies have provided information about the impact of a large meteor that may have contributed to the extinction of the dinosaurs and other species 65 million years ago.

6.11.1: Key Terms

6.11.2: Chapter Questions

6.11.3: References

This page titled 6.11: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.11.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- abyssal fan (plain)
- adsorption
- biogenous
- biogeochemical cycle
- bioturbation
- calcareous
- carbonate
- compensation
- coastal plain
- coccolithophores
- cohesive
- colloidal
- continental shelf
- (margin, slope)
- cosmogenous
- decomposer
- delta
- deposition
- depth (CCD)
- detritus runoff
- diagenesis
- diatom shore
- diffusion
- electrostatic
- erosion
- evaporite
- excreted
- fecal pellets
- fjord
- foraminifera
- fossil
- frustule
- glacier (glaciation)
- graded bed
- grain size

- hard parts
- hydrogenous
- hydrothermal vent
- lagoon
- leaching
- limestore
- lithogenous
- manganese nodules
- marginal sea
- mollusk
- ooze
- pelagic
- phytoplankton
- pore water

- pteropod
- radiolaria
- relict sediment
- residence time
- resuspended
- runoff
- sedimentation rate
- (shoreline)
- siliceous
- slump
- sorted
- steady state
- submarine canyon
- suspended sediment
- terrigenous
- turbidite
- turbidity current
- turbulence
- weathered
- wetland
- zooplankton

This page titled 6.11.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.11.2: Chapter Questions

Study Questions

1. Sediments are deposits of particles of many different sizes from many different sources. Why can we classify sediments by their grain size?
2. List the four principal types of sedimentary particles classified by their origin. Can we classify sediments from all locations according to these four groups? Explain why or why not.
3. A sample of the surface sediments on the continental shelf a kilometer offshore contains large amounts of organic matter compared to other continental shelf sediments. What can you conclude or hypothesize about the characteristics of this region?
4. What conditions are necessary for deposits of metal rich sediments to be formed at hydrothermal vents?
5. Why would you not expect to find turbidites near the East Pacific Rise?
6. Why do we find relict sediments that have only a small amount of overlying sediment on the continental shelf off the northeastern United States, but not off the coast of Louisiana?
7. Describe the characteristics of surface sediments on the deep-sea floor. Why do they have these characteristics and how do they vary with location?
8. If you drilled into the sediments of the deep-sea floor just beyond the continental rise of the northeastern United States, what layers of sediment would you be likely to encounter with depth?
9. A 30-cm-long sediment core has a high concentration of organic matter throughout its length, but oxygen is not depleted in the pore waters. What might explain this?

Critical Thinking Questions

1. Fifty centimeters deep in a 10-m-long sediment core taken from the center of an abyssal plain in the Pacific Ocean, a thin layer of sediment is found that has a much larger grain size than the sediment either below or above it.
   (a) Without knowing anything more about the material, give the possible explanations for this layer.
   (b) How could you determine which one of these alternative explanations was most likely?
2. If you were dating different levels within a sediment core and found a layer that was older than both the layer above it and the layer below it, how would you explain this?
3. Offshore oil drilling now takes place in very deep waters of the continental slope and may eventually take place at abyssal depths. In many cases, plans call for oil and gas produced from these wells to be collected through short pipelines from several wells in a small area and then shipped ashore on tankers.
   (a) Why might it not be a good idea in some deep locations to build pipelines on the seafloor to transport the oil ashore (ignoring economic reasons)?
   (b) On what parts of the continental slope, if any, do you think pipelines should not be built?
4. Some sedimentary rocks found in the interior of continents are composed of very fine-grained silicate mineral particles with no calcium carbonate and low metal concentrations. Other continental sedimentary rocks are composed of coarse sand grains with little calcium carbonate, and yet others are composed of fine-grained material that is principally calcium carbonate. Describe the characteristics of the locations at which each of these three rock types were originally accumulated as sediments.

This page titled 6.11.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

6.11.3: References

Figure 6-1a – Segar, D., and E. Stamman Segar, 1997: Green Sand Beach, Hawai'i.

Figure 6-1b – Segar, D., and E. Stamman Segar, 1997: Sand grains of Green Sand Beach.

Figure 6-2 – Adapted from Gross, M. G. and E. Gross, 1996, *Oceanography, a view of earth*, 7th Edition, Prentice Hall, 462 pp.

Figure 6-3 – Adapted from Milliman, J. D., and R. H. Meade, 1983: World-wide delivery of river sediment to the oceans. *J. Geol.*, **91**, 1–21, Source.

– Updated data from Milliman, J. D., and J. P. M. Syvitski, 1992: Geomorphic/Tectonic control of sediment discharge to the ocean: The importance of small mountainous rivers. *J. Geol.*, **100**, 525–544, Source.

Figure 6-4 – Wing-Chi Poon, 2006: Cavell Glacier with Crevices and Annual Rings. Accessed 26 June 2025, Source, CC BY-SA 2.5.

Figure 6-5 – Allen, J., 2004: Dust over Canary Islands. NASA Earth Observatory Team, accessed 26 June 2025, Source.

Figure 6-6 – Taylor, G. T., 1983: Assorted diatoms living between crystals of annual sea ice in McMurdo Sound, Antarctica. NSF Polar Programs, accessed 26 Jun 2025, Source.

Figure 6-7 – Young, J. R., 2000: Coccolithus pelagicus PM NF 4864 209-73. London Natural History Museum, accessed 26 June 2025, Source, CC0-1.0.

Figure 6-7b – Pkuczynski, 2005: White Cliffs of Dover, England. Accessed 26 June 2025, Source, 4.0 International, 3.0 Unported, 2.5 Generic, 2.0 Generic, 1.0 Generic.

Figure 6-8a – Segar, D., and E. Stamman Segar, 1997: Pelagic foraminifera.

Figure 6-8b – Hopcroft, R. 2002: A pelagic pteropod collected during one of the net tows. NOAA Arctic Exploration 2002, accessed 26 June 2025, Source.

Figure 6-9 – Garcia, L. F., 2007: Radiolaria varios. Accessed 26 June 2025, Source, CC BY-SA 2.5 ES.

Figure 6-10 – Adapted from Berger, W. H., 1975: Dissolution of deep-sea carbonates: an introduction. *Dissolution of Deep-Sea Carbonates,* W. V. Sliter, A. W. H. Bé and W. H. Berger, Eds. Cushman Found. Foram. Res., Spec. Publ., 13: 7–10.

Figure 6-11 – Adapted from Thurman, H.V., 1994: *Introductory Oceanography, 7 Edition*. Macmillan Publishing Company, 276 pp.

Figure 6-12a – USGS, 2000: Manganese nodules. Accessed 26 June 2025, Source, Public Domain.

Figure 6-12b – Beao, 2010: Replaced coin with ruler. USGS, accessed 26 June 2025, Source, Public Domain.

Figure 6-12c – Adapted from D.S. Cronan, 1977: Chapter 2 Deep-Sea Nodules: Distribution and Geochemistry. *Marine Manganese Deposits*, G.P. Glasby, Ed., Elsevier Science Publishing, 11–44, Source.

Figure 6-13 – Adapted from Biscaye, P. E., V. Kolla, and K. K. Turekian, 1976: Distribution of calcium carbonate in surface sediments of the Atlantic Ocean. *J. Geophys. Res.*, **81**, 2595–2603, Source.

Figure 6-15 – Adapted from Heezen, B. C., and W. M. Ewing, 1952, Turbidity currents and submarine slumps, and the 1929 Grand Banks [Newfoundland] Earthquake. *Amer. J. Sci.*, **205**, 949–973, Source.

Figure 6-17 – Adapted from Kennet, J. P., 1982: *Marine Geology*, Prentice Hall Inc., 813 pp.

Figure 6-18e – Adapted from Emery, K. O., 1969: The continental shelves. *Sci. Amer.,* **221**, 106–22, Source.

Figure 6-19a – Adapted from Davies, T. A., and D. S. Gorsline, 1976: Oceanic sediments and sedimentary processes. *Chemical*

oceanography Vol. 5, J. P. Riley and R. Chester, Eds.), 1–80.

Figure 6-19b – Adapted from Heezen, B. C., and C. D. Hollister, 1971: *The Face of the Deep*. Oxford University Press, 672 pp.

Figure 6-20 – With permission from Crawford, W, 2001: Studying sediment cores. IODP/TAMU.

This page titled 6.11.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7: OCEAN-ATMOSPHERE INTERACTIONS

Critical Concepts

- **CC1 Density and Layering in Fluids**: Fluids, including the oceans and atmosphere, are arranged in layers sorted by their density. Air density can be reduced by increasing its temperature, and by increasing its concentration of water vapor causing the air to rise. This is the principal source of energy for Earth's weather.
- **CC3 Convection and Convection Cells**: Evaporation or warming at the sea surface decreases the density of the surface air mass and causes it to rise. The rising air mass cools by adiabatic expansion, and eventually loses water vapor by condensation, which increases temperature as latent heat is released. As air continues to rise, adiabatic expansion and radiative heat loss then cool the air and increase its density so that it sinks. These processes form the atmospheric convection cells that control Earth's climate.
- **CC4 Particle Size, Sinking, Deposition, and Resuspension**: Suspended particles (either in ocean water or in the atmosphere) sink at rates primarily determined by particle size: large particles sink faster than small particles. This applies to water droplets in the atmosphere. The smallest droplets sink very slowly and form the clouds, while larger droplets fall as rain.
- **CC5 Transfer and Storage of Heat by Water**: Water's high heat capacity allows large amounts of heat to be stored in the oceans and released to the atmosphere without much change of ocean water temperature. Water's high latent heat of vaporization allows large amounts of heat to be transferred to the atmosphere in water vapor and then transported elsewhere. Water's high latent heat of fusion allows ice to act as a heat buffer reducing climate extremes in high latitude regions.
- **CC8 Residence Time**: The residence time is the average length of time that molecules of contaminants, such as chlorofluorocarbons, spend in the atmosphere before being decomposed or removed in precipitation or dust. Long residence time allows such contaminants to diffuse upwards into the ozone layer in the upper atmosphere.
- **CC9 The Global Greenhouse Effect**: The oceans and atmosphere are both important in studies of the greenhouse effect, as heat, carbon dioxide and other greenhouse gases are exchanged between atmosphere and oceans at the sea surface. The oceans store large amounts of heat, and larger quantities of carbon dioxide both in solution and as carbonates.
- **CC10 Modeling**: The complex interactions between the oceans and atmosphere that control Earth's climate and affect the fate of greenhouse gases can best be studied by using mathematical models, many of which are extremely complex and require massive computing resources.
- **CC11 Chaos**: The nonlinear nature of ocean-atmosphere interactions makes at least part of this system behave in sometimes unpredictable ways and makes it possible for climate and ecological changes to occur in rapid, unpredictable jumps from one set of conditions to a completely different set of conditions.
- **CC12 The Coriolis Effect**: Water and air masses move freely over the Earth and ocean surface while objects on the Earth's surface, including the solid Earth itself, are constrained to move with the Earth in its rotation. This causes moving water or air masses to appear to follow curving paths across the Earth's surface. The apparent deflection, called the Coriolis Effect, is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The effect is at a maximum at the poles, and is reduced at lower latitudes, becoming zero at the equator
- **CC13 Geostrophic Flow:** Air and water masses flowing on horizontal pressure gradients are deflected by the Coriolis Effect until they flow across the gradient such that the pressure gradient force and Coriolis Effect are balanced, a condition called geostrophic flow. This causes ocean currents and winds to flow around high and low pressure regions in near circular paths. The familiar rotating cloud formations of weather systems seen on satellite images are formed by air masses flowing geostrophically.

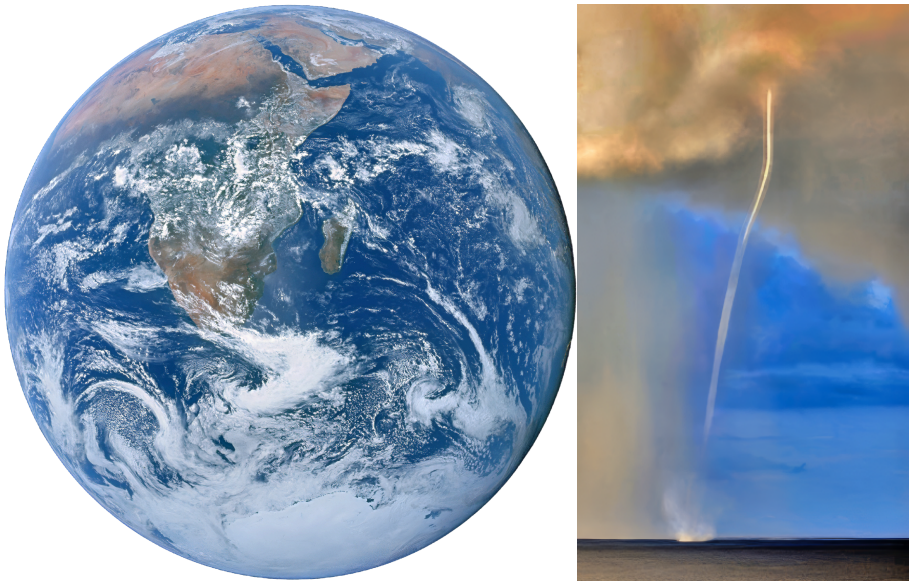


Figure 7-1. Atmospheric phenomena, including thunderstorms, waterspouts, hurricanes, and global winds and cloud patterns, are all driven in-part by heat energy released by ocean surface waters. (a) Apollo 17 took the first picture of Earth from space in 1972. The resulting picture, called the "blue marble" shows the swirling clouds and patterns of the land and oceans. (b) A waterspout, or tornado, is shown as it touches down in a strong thunderstorm near Grand Cayman Island in the Caribbean Sea.

The Earth's **climate** (temperature, rainfall, etc., averaged over several decades of prior data) and **weather** (temperature, rainfall, etc., occurring at a specific time) are determined by the distribution of heat and water vapor in the atmosphere. The oceans play an important role in controlling this distribution.

Ocean-atmosphere interactions are also important because ocean **currents** are generated by winds as well as **density** differences between **water masses**. Seawater density is determined primarily by changes in temperature and **salinity** that occur at the ocean surface (**Chap. 5**). Surface water temperature is controlled by solar heating and radiative cooling, and salinity is altered by evaporation and precipitation.

The ocean contains many times more heat energy than the atmosphere does. This is because water is denser than air, and the total mass of ocean water is more than 200 times the total mass of air in the atmosphere. In addition, water has a much higher **heat capacity** per unit mass than air (or the rocks and soil of the land). Just the top few meters of ocean water have the same heat capacity as the entire atmosphere.

The oceans contain more than 97% of the world's water. The ocean surface covers more than 70% of the Earth's surface. The transfers of heat and water vapor between the ocean and the atmosphere that occur at the ocean surface are the main driving forces that determine the world's climate.

Studies of the complex interactions between the oceans and atmosphere have been important since oceanographic science began. However, such studies have received more intensive attention because of potential climate changes as a result of the enhanced greenhouse effect (**CC9**). We cannot understand such climate changes without understanding ocean-atmosphere interactions.

7.1: The Atmosphere

7.2: Water Vapor in the Atmosphere

7.3: Water and Heat Budgets

7.4: Climatic Winds

7.5: Climate and Ocean Surface Water Properties

7.6: Interannual Climate Variations

7.7: Land-Ocean-Atmosphere Interactions

7.8: Global Climate Zones
7.9: Weather Systems
7.10: Local Weather Effects
7.11: Chapter Summary
7.11.1: Key Terms
7.11.2: Chapter Questions
7.11.3: References

This page titled 7: Ocean-Atmosphere Interactions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.1: The Atmosphere

Gases are highly compressible. Even small changes in pressure produce measurable changes in volume. As pressure increases, the gas molecules are forced closer together and the volume is reduced. In the atmosphere, pressure decreases rapidly with altitude (Fig. 7-2). Consequently, the density of the atmosphere decreases progressively with increasing height above the Earth's surface. As a result, the atmosphere is **stratified** throughout its more than 100-km height (**CC1**).

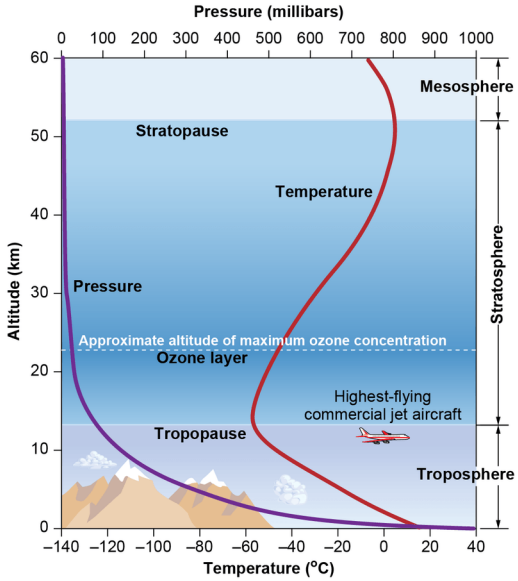


Figure 7-2. Structure of the atmosphere. The atmospheric circulation that determines the Earth's climate and weather takes place within the troposphere. Movements of gases between the troposphere and the stratosphere are limited and slow. Ozone is present throughout the stratosphere, but its concentration peaks at an altitude of about 22 km.

Temperature and vapor pressure (water vapor concentration) also affect the density of air, but much less than the changes in pressure with altitude. The atmosphere is steeply stratified near the Earth's surface because of the rapid pressure change with altitude. As a result, density driven vertical motions of air masses in the **troposphere** are limited to only a few kilometers above the Earth's surface.

Atmospheric Structure

The atmosphere is separated vertically into three distinct zones: troposphere, stratosphere, and mesosphere. Beyond the mesosphere is a zone called the thermosphere which is a transition zone between the atmosphere and outer space. The troposphere lies between the Earth's surface and an altitude of about 16 to 18 km near the equator and less than 10 km over the poles (jet airplanes fly at an altitude of about 10 to 12 km). Vertical movements of air masses are caused by density changes and occur mainly in the troposphere. Although temperature generally decreases with altitude within the troposphere, density also decreases because the reduction in atmospheric pressure with altitude more than offsets the effect of lowered temperature (Fig. 7-3). This chapter pertains primarily to the troposphere because the atmospheric motions created by heat transfer from oceans to atmosphere are restricted to this zone.

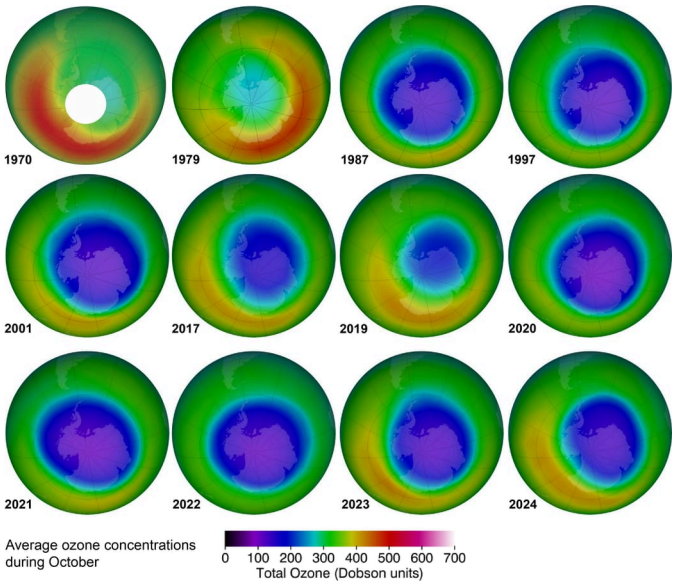


Figure 7-3. Average total amount of ozone above the Earth's surface over Antarctica during the month of October, as measured by a series of satellite instruments. The small circular area that has no color directly over the pole in the 1970 data is an area where no data were obtained. Nevertheless, it is clear that there is a broad region of very low ozone levels over Antarctica during the 1990s that was not present during the 1970s. This is the Antarctic ozone hole. The extent of the ozone hole has declined in recent years due to a global ban on the use of CFCs but other factors are likely involved. For example, the 2022 Hunga Tonga volcanic eruption may have contributed to an increase in ozone depletion in that year.

Between the top of the troposphere and an altitude of approximately 50 km, is the stratosphere. The air in the stratosphere is "thin" (that is, molecules are much farther apart than they are at ground level, because pressure is low), vertical air movements are primarily very slow **diffusion**, and temperature increases with altitude (**Fig. 7-3**). Within the stratosphere is a region called the **ozone layer**, where oxygen gas (O_2) is partially converted to ozone (O_3) by reactions driven by the sun's radiated energy.

Ozone Depletion

The concentration of ozone in the Earth's ozone layer decreased progressively during the 1980s and 1990s, particularly in the region around the South Pole, where the depletion was so great that it has become known as the "ozone hole" (**Fig. 7-3**). The decrease varies seasonally for both poles and is greatest during the spring, which is in October for the Southern Hemisphere.

The ozone layer absorbs much of the sun's ultraviolet light, so depletion of the ozone layer could have severe adverse consequences for people and the **environment**. Ultraviolet light causes sunburn, eye cataracts, and skin cancers in humans; inhibits the growth of **phytoplankton** and possibly land plants; and may have harmful effects on other **species**.

The effects of ozone depletion on the environment are largely unknown. Studies have shown that the growth rate of phytoplankton in the Southern Ocean around Antarctica is significantly reduced (by 6 to 10%) in areas under the ozone hole. All life in the Antarctic, including whales, penguins, and seals, is ultimately dependent on phytoplankton for food (**Chaps. 12**). Consequently, besides the direct effects of the increased ultraviolet radiation, ozone depletion may adversely affect species by reducing their food supply.

There is still some uncertainty about the causes of ozone depletion, but the accepted view is that the primary cause is gases called **chlorofluorocarbons** (**CFCs**) released to the atmosphere. CFCs are synthetic chemicals used in many industrial applications. Until close to the end of the last century, most spray cans used CFCs to produce the aerosol and almost all refrigerators and air conditioners used substantial quantities of a mixture of CFCs called "freon."

CFCs are highly resistant to decomposition and may remain in the atmosphere for decades. The long **residence time** (**CC8**) allows the CFCs to diffuse slowly upward until they reach the ozone layer. There they are decomposed by ultraviolet light, and their chlorine is released. The chlorine atoms take part in a complicated series of chemical reactions with ozone molecules whereby the ozone molecule is destroyed but the chlorine atom is not. Because it is not destroyed, the chlorine atom can destroy another ozone molecule and then another. In fact, one chlorine atom, on average, can destroy thousands of ozone molecules.

Ozone depletion in polar regions is greatest in spring because stratospheric clouds of supercold ice crystals are formed at that time. These clouds remove nitrogen compounds from the ozone layer. Thus, in spring, chlorine that would normally react with the nitrogen compounds reacts with the ozone instead.

CFC use as an aerosol gas in spray cans was banned and eliminated virtually worldwide in the early 1980s. Later, a global international treaty, the Montreal Protocol of 1987, was established that called for total elimination of CFC manufacture and replacement of these compounds by other chemicals by the year 2000. Although these measures will probably eliminate ozone depletion eventually, most scientists believe that the depletion will continue until at least the middle of this century, if not longer, until the CFCs already released have diffused slowly upward through the atmosphere and been decomposed. Also, the compounds used to replace CFCs destroy ozone in much the same manner as CFCs but far less effectively. The number of refrigeration units has rapidly increased and continues to increase as the standard of living rises globally. This has substantially delayed recovery of the ozone layer. Indeed, ozone depletion continues to be observed in both Northern and Southern Hemispheres although it appears to be decreasing slowly. However, other factors may interfere. For example, the 2022 Antarctic ozone hole was one of the largest on record, possibly partially due to the effects of the Hunga Tonga volcanic eruption in January of that year. However, the year to year variability in the magnitude of the ozone hole is large so Hunga Tonga contribution is uncertain.

Depletion of ozone in the ozone layer in the stratosphere should not be confused with the problem of elevated ozone concentrations in smog in the troposphere (lower atmosphere). Ozone is released into the troposphere by various human activities and is created by photochemical reactions of other gases in the troposphere. Ozone is one of the principal harmful components of smog. However, this ozone does not contribute to the ozone layer because it does not last long enough in the atmosphere to be transported up through the troposphere and stratosphere to the ozone layer.

This page titled 7.1: The Atmosphere is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.2: Water Vapor in the Atmosphere

In the troposphere, air masses move continuously, both vertically and horizontally, and these movements control the Earth's weather and climate and the winds that create ocean waves and ocean surface currents. The movements of air masses are caused primarily by changes in air mass density that occur as water vapor is added or removed.

Water Vapor and Air Density

Water vapor in the atmosphere behaves in much the same way as sugar in water. Sugar dissolves in water, but the maximum quantity of sugar that can be dissolved is limited. More sugar can be dissolved at higher temperatures. Similarly, water vapor can be "dissolved" in air in limited quantities, and the amount of water vapor that can be present in air at saturation is greater at higher temperatures. The maximum amount of water that can remain in the vapor phase in air is expressed by the water vapor **saturation pressure**. The variation of water vapor saturation pressure with temperature is shown in **Figure 7-4**. The saturation pressure of water vapor at average ocean surface temperatures (approximately 20°C) is many times greater than the saturation pressure of water vapor in air near the top of the troposphere (approximately –40°C to –60°C).

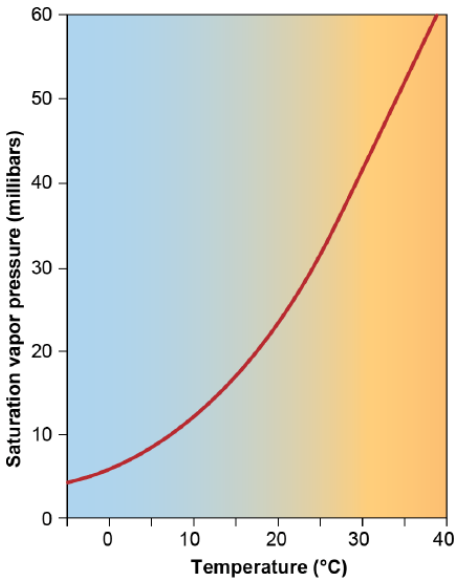


Figure 7-4. Variation of the water vapor saturation pressure with temperature in air. The water vapor saturation pressure is lower at lower temperatures. This is the reason that water condenses into rain as warm, moist air rises and cools.

When water evaporates from the sea surface, water molecules displace (move aside) oxygen and nitrogen molecules in the atmosphere. The displaced molecules are mixed and distributed in the surrounding air mass by random motions of the gas molecules. This is important because the atmospheric pressure and, therefore, density of the air mass would be increased if this were not the case. In contrast, water molecules (H_2O, molecular mass 18) are lighter than either the nitrogen (N_2, molecular mass 28) or the oxygen (O_2, molecular mass 32) molecules that they displace. Hence, light molecules displace heavier molecules, and the density of air is, therefore, reduced when water vapor is added (provided there is no change in pressure). As a result, moist air is less dense than dry air at the same temperature and pressure. Therefore, an air mass to which water vapor has been added tends to rise until it reaches its equilibrium density level in the density-stratified atmosphere (**CC1**). Conversely, if water vapor is removed, air increases in density and tends to sink.

Water vapor is continuously added to the atmosphere by evaporation of water from the ocean surface and removed by condensation and precipitation. The rate of evaporation varies with location, as explained later in this chapter. Where evaporation of ocean surface water is particularly rapid, the density of the air mass at the ocean surface is reduced, and the air tends to rise. In many parts of the oceans where the surface water is warmer than the overlying air, **conduction** and radiation of heat from the ocean surface also reduce the density of the air immediately above it.

Water Vapor, Convection, and Condensation

If a gas expands without any external source of heat, its temperature decreases in a process called **adiabatic expansion**. As a result, when air rises in the atmosphere, the pressure decreases, and the air is cooled by adiabatic expansion. Because water vapor saturation pressure is highly temperature-dependent, cooling the air also results in a decrease in the water vapor saturation pressure. When the air has risen sufficiently that the actual vapor pressure becomes greater than the saturation vapor pressure, the air is supersaturated. In this condition, excess water vapor condenses into liquid water droplets (if above freezing). Because water molecules are typically too far apart to condense spontaneously, condensation in the atmosphere almost always occurs on cloud condensation nuclei (CCN).

In many cases, the water will condense to form large numbers of extremely small water droplets—droplets that are too small to fall out of the sky (**CC4**) or to combine with each other to form larger droplets. These water droplets form the clouds of the atmosphere. When cloud droplets are in the right environment for significant growth, precipitation can eventually form.

Water vapor contributes to vertical movements of and within air parcels, not only because of its low molecular weight, but also because of water's **latent heat of vaporization** (Chap. 5). When warm, moist air rises through the atmosphere and cools, and water vapor condenses, the water's latent heat of vaporization is released. The heat warms the air mass and thus reduces its density, causing the air parcels to rise. Release of latent heat is the major driving force for the spectacular rising plumes of moist air that form cumulonimbus (thunderstorm) clouds, which can reach the upper parts of the troposphere. Latent heat released to the atmosphere is also the main energy source for **hurricanes** and other storms.

Atmospheric **convection cells** are created by evaporation or warming at the sea surface that decreases the density of the air mass and causes it to rise. The rising air mass cools by adiabatic expansion, and eventually water vapor is lost by condensation. As the water vapor condenses, temperature is increased by the release of latent heat and the air tends to rise farther. Adiabatic expansion and radiative heat loss then cool the air parcel and increase its density so that it eventually sinks (**CC3**). The location, size, and intensity of convection cells determine the location and intensity of rainfall and snowfall, as well as the distribution of atmospheric heat energy. Convection cells exist at a range of spatial scales. We will discuss how global-scale cells form and how they influence climate and weather at different latitudes in the next several sections of this chapter.

This page titled 7.2: Water Vapor in the Atmosphere is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.3: Water and Heat Budgets

Averaged over the entire planet and over time periods of months or years, the amounts of heat and water that enter the atmosphere from all sources must equal the amounts removed from the atmosphere by all routes. If total inputs of heat or water did not equal total outputs, the average atmospheric temperature and/or the amount of water vapor in the atmosphere would progressively increase or decrease from year to year. In fact, the Earth's climate normally remains essentially unchanged for centuries.

In the Earth's past, very small imbalances between heat gained and heat lost have slightly changed the total amount of heat in the atmosphere and, thus, the Earth's climate. Today, **anthropogenic** releases of greenhouse gases, such as carbon dioxide, may disrupt, or already have disrupted, the balance between heat gained and heat lost that has remained stable over the past several centuries (**CC9**).

Because the inputs and outputs of heat and water to and from the atmosphere must balance, we can construct global heat and water budgets.

Water Budget

The global water budget is relatively simple (**Fig. 7-5**). Water evaporates from the oceans, lakes, rivers, soils, and vegetation on land. The amount of water evaporated from land and ocean annually is enough to cover the entire globe with water almost 1 m deep. The evaporated water eventually condenses and falls as rain or snow over both land and ocean. Approximately 93% of the water that evaporates into the atmosphere comes from the oceans. However, only about 71% of the total global precipitation (rain and snow) falls on the oceans. Thus, more water evaporates from the ocean than reenters it as rain, and more rain falls on the land than is evaporated from it. This pattern is fortunate because the excess precipitation on land is the source of the Earth's freshwater supply. Excess water flows back into the oceans through rivers, streams, and groundwater.

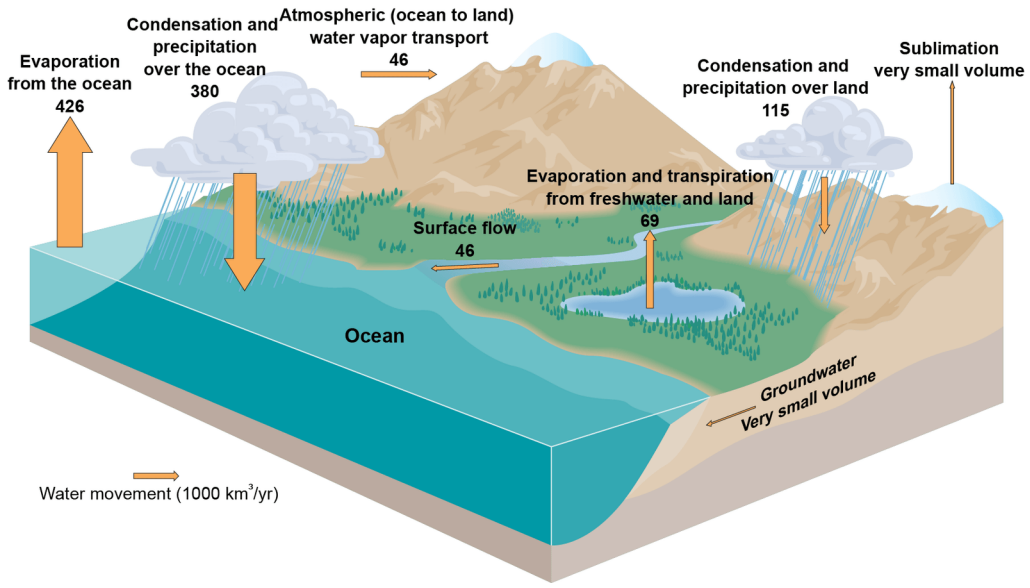


Figure 7-5. The Earth's water budget. The values represent the estimated volumes of water transferred annually between the various reservoirs by the indicated routes.

Neither the rate of water evaporation nor the amount of rainfall is the same everywhere on the oceans or on land. Clearly, deserts and tropical rainforests receive different amounts of rainfall and sustain different amounts of evaporation. The distribution of evaporation and rainfall is discussed later in this chapter.

Heat Budget

The Earth's heat budget (**Fig. 7-6**) is more complicated than its water budget. The source of heat energy for the atmosphere is the sun. Solar radiation energy is partially absorbed by water vapor, dust, and clouds in the atmosphere and converted to heat energy. Part of the sun's energy is also reflected back, or backscattered, to space by clouds. Solar energy that reaches the Earth's surface is partially reflected and partially absorbed and converted to heat energy by the land and ocean waters. The Earth, ocean, atmosphere, and clouds also radiate heat energy. In fact, all bodies radiate heat. The peak wavelength radiated increases as the temperature of the body decreases. Because the Earth is much cooler than the sun, the wavelengths radiated by the Earth, ocean, and clouds (infrared) are much longer than the wavelengths emitted by the sun (visible and ultraviolet). The difference in wavelength and the differential absorption by atmospheric gases at different wavelengths are the basis for the greenhouse effect (**CC9**).

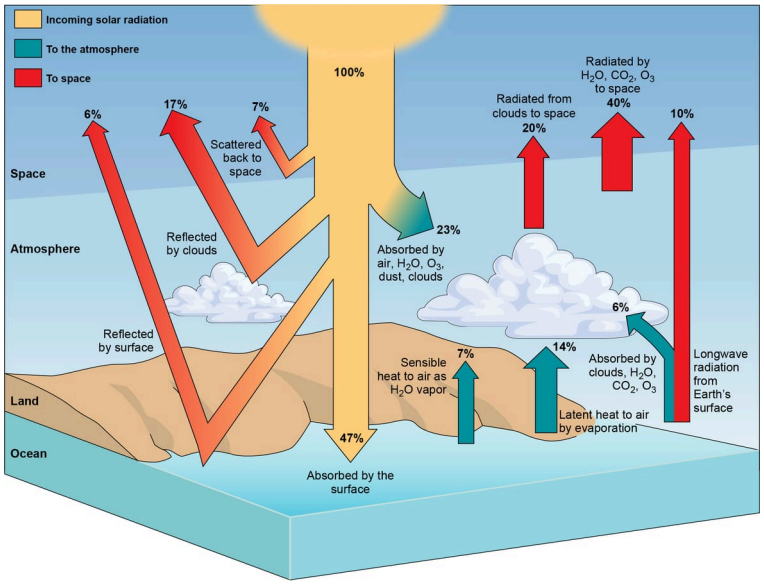


Figure 7-6. The Earth's heat budget. Solar radiation, which is concentrated in the visible light wavelengths, is reflected, scattered, and absorbed by the gases and clouds of the atmosphere and by the ocean and land surface. All of the absorbed energy is radiated back to space, primarily as infrared radiation. The diagram shows the estimated percentages of the incoming solar radiation that are absorbed, reflected, and reradiated to space by the various components of the Earth-ocean-atmosphere system. Each of these may change in an unpredictable way as the greenhouse effect is enhanced by anthropogenic inputs to the atmosphere (**CC9**, **CC10**, **CC11**).

About 20 to 25% of the solar radiation reaching the Earth is absorbed by atmospheric gases and clouds and converted to heat energy. Almost 50% of the solar radiation is absorbed by the ocean water and the land, then transferred to the atmosphere by the conduction of **sensible heat**, as latent heat of vaporization, and by radiation. The net transfer averaged over the global oceans is from ocean to atmosphere. Heat transfer by conduction is relatively small, but the transfer of latent heat from ocean to atmosphere equals about one-quarter of the total solar energy that reaches the Earth. Thus, the oceans effectively capture a major portion of the sun's radiated energy and transfer much of it to the atmosphere as latent heat of vaporization. This mechanism moderates climate differences between Earth's tropical and polar regions.

Latitudinal Imbalance in the Earth's Radiation

At the equator, the sun passes directly overhead at noon on the spring and autumnal **equinoxes** (**Fig. 7-7a**). The sun's radiated energy therefore passes vertically through the atmosphere, and the **angle of incidence** at the Earth's surface is 90°. At the poles on the equinoxes, the sun remains on the horizon all day, and its radiation must pass through the atmosphere parallel to the Earth's surface.

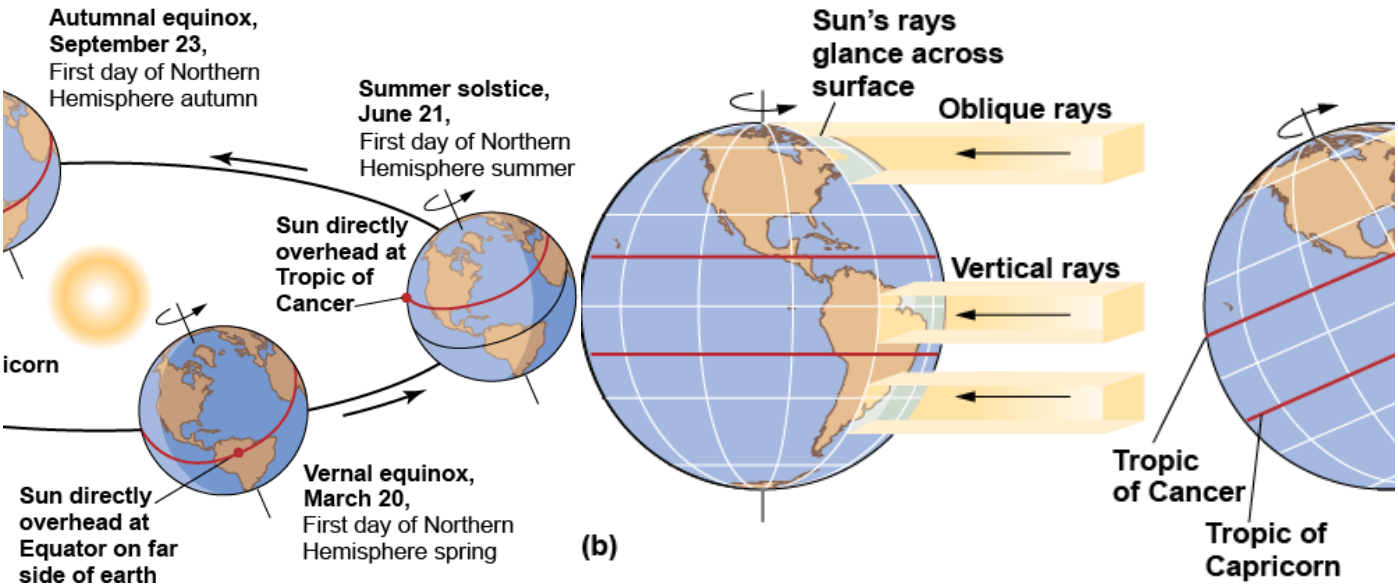


Figure 7-7. (a) The Earth, spinning on its tilted axis, moves around the sun once a year. In northern summer (on the right), the Earth is tilted such that the sun is directly overhead at a latitude north of the equator at a point that moves daily around the Earth. Similarly, in northern winter, this point is south of the equator. At the spring and autumnal equinoxes, the sun is overhead at the equator throughout the day as the Earth spins. (b) The angle of incidence of solar radiation received at the Earth's surface also changes with the seasons, but generally, the area over which a given amount of the sun's energy is spread is less near the equator and increases toward the poles. The globe on the left shows the Earth at an equinox. The globe on the right shows the Earth at the winter (northern) solstice.

Because the poles are only about 6000 km farther from the sun than the equator is, and because the mean distance from the sun to the Earth is 148,000,000 km, the intensity of the sun's radiated energy is only about 0.01% lower at the poles than at the equator. However, solar energy is spread over a progressively larger area with increasing **latitude** (Fig. 7-7b). Consequently, on the equinoxes, the solar radiation received by an area of the Earth's surface at 30°N or 30°S is only 86% of that received by an equal area at the equator. The proportion at 60°N or 60°S is only 50%, and at the poles it is zero.

Nearer the poles, solar radiation must also travel a longer path through the atmosphere because of the lower angle to the Earth's surface. More of the solar radiation is absorbed as it passes through the atmosphere because of the increased path length, but this effect causes a relatively small difference in solar intensity at the Earth's surface.

The intensity of the Earth's long-wavelength radiation increases with temperature. However, the temperature difference between equatorial and polar regions produces a latitudinal variation in the Earth's longwave radiation intensity that is not as great as the latitudinal variation of solar insolation of Earth's surface (Fig. 7-8).

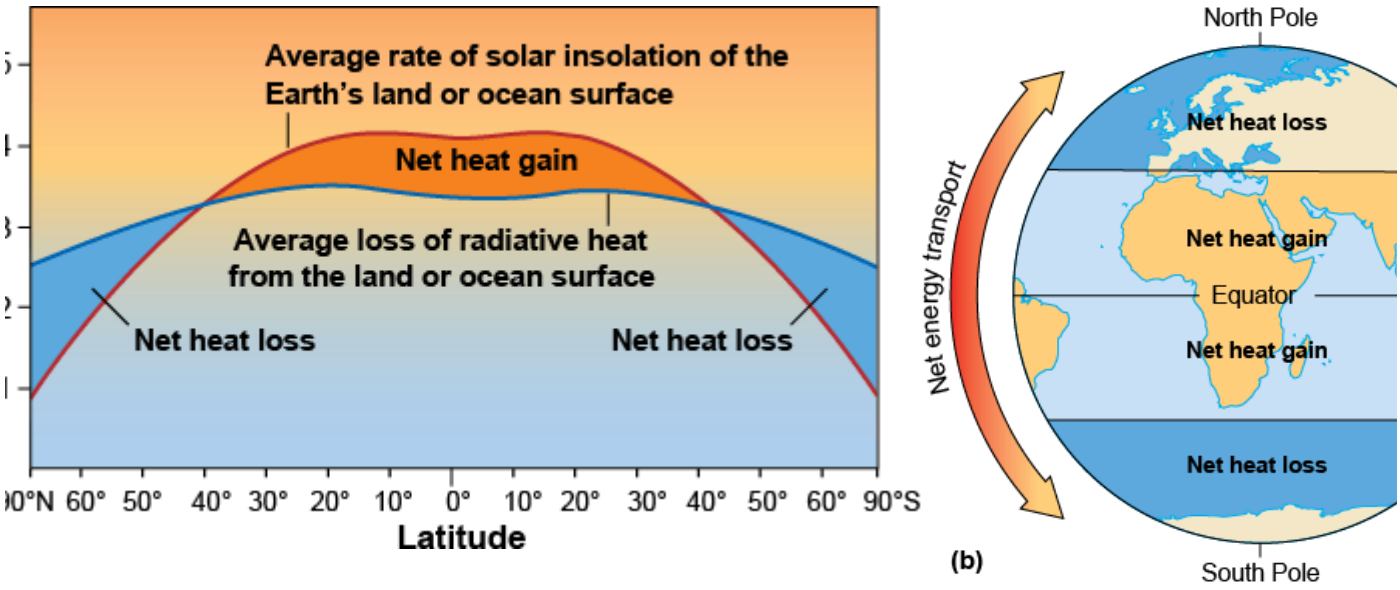


Figure 7-8. Variation in radiative heat loss and gain at the Earth's surface with latitude. There is a net gain between the equator and about 35° to 40°, and a net loss at higher latitudes.

The total solar radiation absorbed by the Earth must equal the total long-wavelength radiation lost by the Earth when averaged over the whole Earth, but this balance does not occur at all latitudes. There is a large difference between the rates at which solar heat is absorbed at the poles and at the equator. There is a much smaller difference between the rates at which long-wavelength radiation is lost at the poles and at the equator (**Fig. 7-8**). Consequently, there must be mechanisms for transferring heat energy from low latitudes to high latitudes. If heat were not transferred, the polar regions would be much colder and the equatorial regions much warmer. Planets and moons that have no (or a limited) mechanism of heat transfer between latitudes have dramatically greater differences in surface temperature between their polar and equatorial regions than does the Earth. Hence, latitudinal heat transfer is one of the critical mechanisms that maintains the habitability of the Earth's entire surface.

Heat is transported latitudinally in two ways. First, latent heat of vaporization enters the atmosphere as water is evaporated in warm tropical regions and some of the heat is transported to higher latitudes by atmospheric convection cells. At higher latitudes, cooling causes water vapor to condense in the form of rain or snow, thus releasing latent heat to the atmosphere. Second, ocean water warmed by the sun in low latitudes is transported by ocean currents to higher latitudes, where it transfers heat to the atmosphere by conduction and evaporation. Near the equator, ocean currents transport more heat to higher latitudes than does atmospheric transport, whereas near the poles atmospheric transport of heat to higher latitudes is greater than ocean current transport.

This page titled 7.3: Water and Heat Budgets is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.4: Climatic Winds

Winds are extremely variable from one day to the next at any location. The day-to-day variations are part of what we call “weather.” Weather events may include calm, thunderstorms, hurricanes, fog, and the other phenomena that a weather forecaster tries to predict hours or days before they happen. If we average the day-to-day variation over relatively large areas and over periods of several days or weeks, we find patterns of average winds, temperatures, rainfall, and so on that are relatively repeatable from one year to the next in each area. These long-term average patterns are called “climate.” This chapter is concerned primarily with climatic winds, temperatures, and rainfall, although some aspects of various weather events are also examined.

Climatic Winds on a Nonrotating Earth

On a nonrotating Earth with no continents or **Coriolis effect** (**CC12**), wind patterns would be simple (**Fig. 7-9**). More heat is transferred to the atmosphere from the oceans than is absorbed by the atmosphere directly from the sun (**Fig. 7-6**). Solar heating of the oceans is greatest at the equator. Thus, ocean water is warmer and the rate of heat transfer by evaporation of water from ocean to atmosphere is greater than at higher latitudes. The equatorial air mass would rise through the troposphere, creating a low-pressure zone and horizontal **pressure gradient** in the atmosphere at sea level. The rising air would be replaced by air flowing into the region from higher latitudes. Once it reached its equilibrium density level (**CC1**), the rising air would spread toward the poles. As the air mass moved north or south in the upper troposphere, it would progressively cool and release its water vapor as rain or snow.

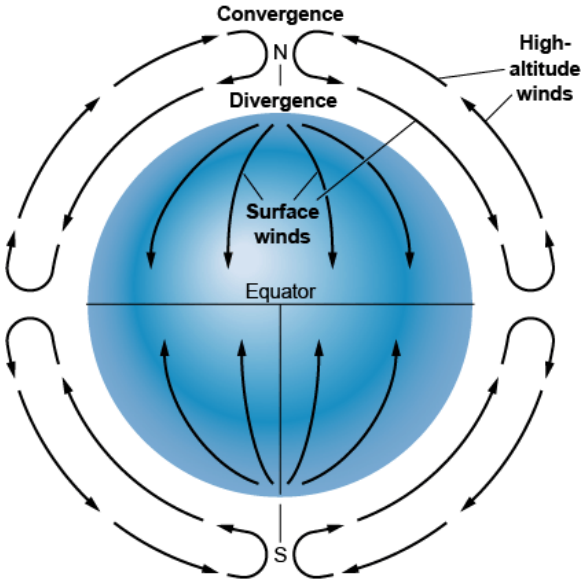


Figure 7-9. Atmospheric convection cells and winds on a hypothetical water-covered, nonrotating Earth.

Once the air mass reached the pole, it would have sufficiently cooled and lost water vapor by condensation or deposition to become dense enough to sink and create a high-pressure zone in the atmosphere at sea level. The air would then move toward the equator along the horizontal atmospheric pressure gradient between the polar high-pressure zone and the equatorial low-pressure zone. As the air moved back toward the equator, it would be warmed and would gain water vapor from evaporation of ocean surface water. This is a simple convection cell motion (**CC3**). There would be a **convergence** at sea level at the equator, and **divergence** at sea level at the poles. Surface winds would blow from the poles toward the equator in each hemisphere. This simple two-convection-cell pattern (**Fig. 7-8**) illustrates the basic principles of atmospheric circulation, but it does not exist on the Earth, because we have so far not accounted for the fact that the planet rotates and moving air is deflected by the Coriolis Effect. The presence of land is also not yet accounted for, but we can best understand the complications that land introduces if we first look at a hypothetical water-covered Earth that is rotating.

Climatic Winds and the Coriolis Effect

The deflection of air movements by the Coriolis effect, caused by Earth's rotation, results in an atmospheric circulation system that consists of three **convection** cells arranged latitudinally in each hemisphere (**Fig. 7-10**). To some extent, this real-world system has the same net result as the simple two-convection cell system: air rises at the equator and sinks at the poles, and heat is transported from the equator to the poles.

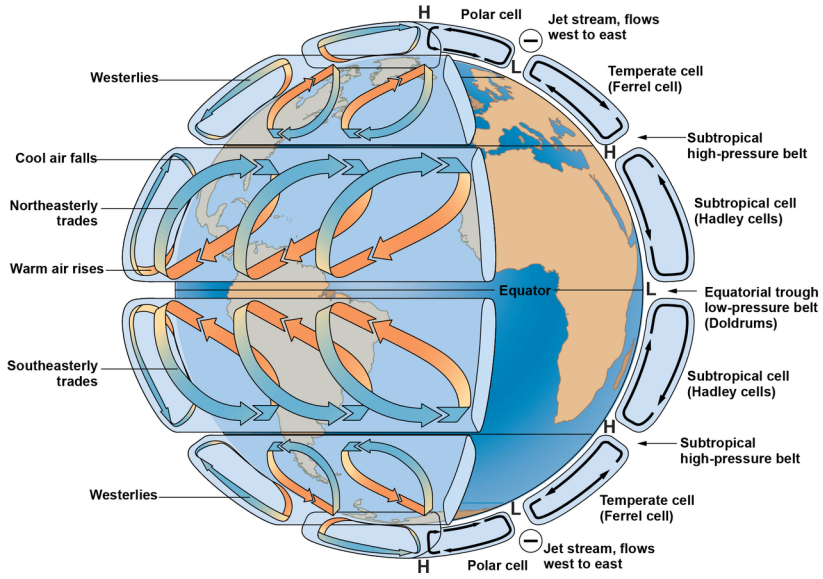


Figure 7-10. Atmospheric circulation cells and winds on a hypothetical water-covered, rotating Earth are arranged in latitudinal bands. There are three cells in each hemisphere. Upwelling occurs at the equator and the polar fronts, and downwelling at the poles and in mid latitudes. Note how the Coriolis effect deflects moving air masses: air masses moving away from the equator are deflected to the east, and air masses moving toward the equator are deflected to the west. Atmospheric convection cells are arranged in this general pattern on the Earth, but they are modified substantially by the influence of the land masses.

To understand how a three-cell per hemisphere circulation pattern develops, consider the movements of an air mass initially located at sea level on or near the equator. The air mass is heated by solar radiation and has a high water vapor content because of the relatively high evaporation rate from the ocean's surface. It rises through the troposphere to its density equilibrium level and spreads out toward the poles, just as it would on a nonrotating Earth. As the air mass moves away from the equator, where there is no deflection by the Coriolis effect, it is not deflected significantly until it reaches 5° to 10°N or S (**Fig. 7-10**). As it continues to move toward the poles, the air mass is deflected increasingly toward the east because the Coriolis Effect increases with latitude.

At about 30°N or 30°S, the now eastward-moving air has risen, expanded, cooled, and lost most of its water vapor. The cooler, dry air sinks to the surface, where some air moves toward the pole and some moves away from it, forming a high-pressure zone and atmospheric divergence at ground level. Air that moves toward the pole in this region enters the "Ferrel cell" circulation (**Fig. 7-10**). Air that moves back toward the equator is deflected toward the west. The deflection decreases as the air mass moves nearer the equator. Consequently, surface winds in the subtropical regions blow predominantly from the northeast in the Northern Hemisphere and from the southeast in the Southern Hemisphere (**Fig. 7-10**). These are the **trade winds**, and the convection cell air movements that occur between the equator and about 30°N and 30°S are called "Hadley cells."

Surface winds in the Ferrel cells (**Fig. 7-10**) are also deflected by the Coriolis effect, but to the east because they are moving away from the equator. Therefore, this is a region of persistent surface winds from the southwest in the Northern Hemisphere and from the northwest in the Southern Hemisphere, commonly called the **westerlies**. Winds are named for the direction from which they are blowing from.

Although the predominant winds are from the east in subpolar regions, air movements in the polar cells are complex because of the stronger Coriolis deflection and the presence of the **jet streams**. Jet streams are swiftly moving west-to-east air currents high in the troposphere. They are located over the boundaries between the Ferrel and polar cells, called the "polar **fronts**" (or "Antarctic front" in the Southern Hemisphere). The jet streams undergo complex **meanders** associated with movements of the polar fronts. Because storms often form along the jet stream, and because meanders in the jet stream can bring cold polar air to lower latitudes and warm air to higher latitudes, the orientation and trajectory of the jet stream are routinely reported in many

weather forecasts.

In any one region, the predominant wind direction and its persistence, the average extent of cloud cover, and the average rainfall are determined largely by the region's location with respect to the atmospheric convection cells (**Fig. 7-10**). In the equatorial region known as the Doldrums, where trade winds converge and atmospheric upwelling occurs, sea-level winds are light and variable. In addition, because the rising moist air is cooled, cloud cover is persistent and rainfall is high. This region is called the **intertropical convergence zone** because it is where two wind systems converge. Air rising from the sea surface through the troposphere causes low atmospheric pressure at sea level in the intertropical convergence zone. Air also rises through the troposphere at the polar front. However, unlike the intertropical convergence zone that migrates seasonally but otherwise remains generally invariable, the polar front oscillates in wavelike motions, called "Rossby waves" (**Chap. 9**), which have timescales of days or weeks. These waves can spawn massive storms called **extratropical cyclones**.

In the atmospheric **downwelling** (or subsidence) regions between the Hadley cells and Ferrel cells and at the poles, atmospheric pressure is high, winds are light, skies are usually cloudless, and rainfall is low. The **subtropical high**-pressure zones between the Hadley and Ferrel cells are called the "horse latitudes." Sailing ships were often becalmed for long periods in these regions, and the horses on board were killed to conserve water. Because of the high-pressure zone at the South Pole, much of Antarctica is a desert with very low annual precipitation. In fact, there are dry desert valleys in Antarctica with no snow cover, and the interior of Antarctica receives only about 50 mm of precipitation per year (desert climates are defined as having less than 254 mm of rain per year). The massive amounts of ice and snow in the thick ice sheet that covers other parts of Antarctica took millennia to accumulate, and the ice sheet exists only because Antarctic temperatures are too low for snow to melt during the summer.

Under the centers of the atmospheric cells (between convergences and divergences), winds are generally persistent, particularly in the trade wind zones. Rainfall and cloud cover are variable, and the zones are affected by periodic strong storms, especially in the higher-latitude areas of the westerly wind zones, where storms form on the polar front. Westerly surface winds under the Ferrel cells can be particularly strong, especially in the Southern Hemisphere, where few continents disturb the circulation. Sailors call the area of strong westerly winds at about 40°S the "roaring forties."

The atmospheric convection cell system described here is a simplified view of the climatic wind patterns on the Earth's surface. Complications of this simplified pattern occur because the interactions between land and atmosphere differ from the interactions between ocean and atmosphere. The pattern of high- and low-pressure zones and winds described by the six-cell convection system model generally holds true over the oceans but is much modified over the large land masses of the continents. The strong Coriolis Effect and shallow troposphere of the polar cell regions favor horizontal transport of air masses by weather systems and poor development of vertical convection cell motion. Complications in the six-cell system also occur because of the seasonal movement of the Earth around the sun and certain longer-term oscillations of the ocean-atmosphere interaction, such as **El Niño**, which is described later in this chapter. Finally, local wind patterns are highly variable because weather events operate at much smaller geographic scales than do the global climatic wind patterns discussed in this section.

Seasonal Variations

Because the Earth's axis is tilted in relation to the plane of the Earth's orbit around the sun, the latitude at which the sun is directly overhead at noon changes progressively during the year. At the Northern Hemisphere summer **solstice** (June 20 or 21), the sun is directly overhead at 23.5°N, the Tropic of Cancer. At the autumnal equinox (September 22 or 23) and the spring equinox (March 20 or 21), the sun is directly overhead at the equator. At the Northern Hemisphere winter solstice (December 21 or 22), the sun is directly overhead at 23.5°S, the Tropic of Capricorn (**Fig. 7-8a**).

As the Earth moves around the sun, the latitude of greatest solar intensity migrates north and south, resulting in a corresponding northward or southward displacement of the atmospheric convection cells. The displacement of atmospheric convection cells causes seasonal climate variations. The seasonal movement of convection cells is shown by a seasonal shift in wind bands (**Fig. 7-11**) and in surface-level atmospheric high- and low-pressure zones (**Fig. 7-12**).

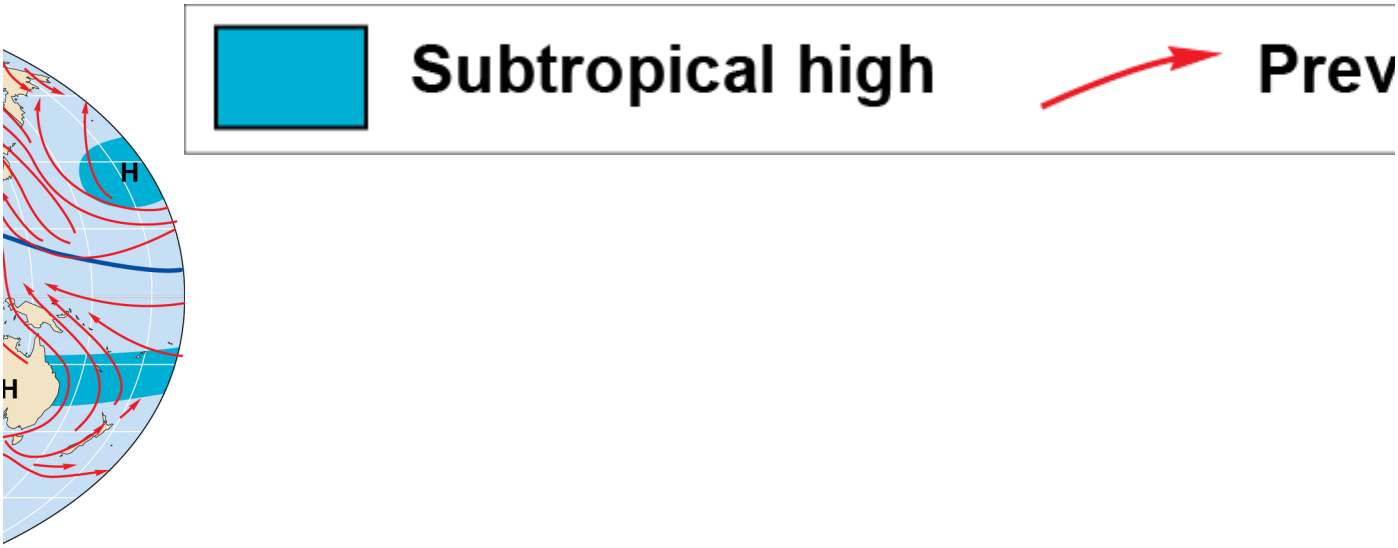


Figure 7-11. Climatic winds over the ocean surface in (a) January and (b) July. Winds generally blow from higher pressure toward lower pressure. Note the seasonal migration of the subtropical high-pressure zones, the intertropical convergence zone, and the associated trade winds. The migration is more pronounced over the major continental landmasses than over the oceans. Over the Indian Ocean, the seasonal migration is enhanced by the strong monsoonal interaction between Asia and the Indian Ocean.

Figure 7-12. Mean atmospheric pressures at sea level in (a) January and (b) July. The greater complexity of the geographical distribution and seasonal change of atmospheric pressure in the Northern Hemisphere is due to the concentration of the landmasses in this hemisphere. The ocean moderates regional coastal climates, leading to larger seasonal changes in pressure over large landmasses such as Asia.

Because the latitude at which tropospheric air temperature is highest changes seasonally, the convection cells also migrate seasonally. Tropical tropospheric air temperature is controlled primarily by radiation and evaporative heat inputs from the oceans and land. Because water has a high heat capacity, a great amount of heat energy is necessary to warm the ocean surface waters (**CC5**). Therefore, water temperature, evaporation rate, and air temperature do not increase in step with changes in solar intensity as the latitude of greatest solar intensity shifts north and south. Several weeks of increased (or decreased) solar heating must elapse before these parameters change enough to cause the atmospheric convection cells to move. Consequently, there is a time lag between the sun's seasonal movement and the movement of the atmospheric convection cells. The time lag is why many northern countries have their warmest weather in August and their coldest weather in February, approximately 2 months after the summer and winter solstices occur in June and December, respectively.

Heat captured from solar radiation is transferred to the atmosphere more rapidly by land than by oceans, primarily because of the high heat capacity of water. As a result, the time lag between seasonal changes in solar radiation and air temperature is

shorter over the continents than over the oceans. Because the continents are concentrated in the Northern Hemisphere, seasonal migration of convection cells has a shorter time lag north of the equator than south of the equator. Furthermore, the center of the latitudinal wind bands, the intertropical convergence zone, migrates farther north of the equator in the Northern Hemisphere summer (especially over the continents) than it does south of the equator during the Southern Hemisphere summer (**Fig. 7-12**).

Seasonal changes in climate are greater in areas that are under an atmospheric convergence or divergence zone in one season and under the middle of a convection cell in another. Seasonal climate changes are also greater in higher latitudes than near the equator. The reason is that the total amount of solar heat reaching the Earth's surface per day depends on both the sun's angle and the amount of time during which the sun is above the horizon on a given day. Both the sun's intensity and the day length vary more between summer and winter at high latitudes than at low latitudes.

Monsoons

In the Northern Hemisphere summer, the intertropical convergence zone moves farther north over the landmasses of Asia and Africa than it does over the oceans. Warm, moist tropical air moves north onto Asia from the Indian and western Pacific Oceans in the trade wind zone under the southern Hadley cell (**Fig. 7-13a**). Thus, in summer the southern Hadley cell trade winds extend into the Northern Hemisphere, where the Coriolis deflection is to the right. These winds, called monsoon winds, are deflected by the Coriolis Effect in the Northern Hemisphere, so they blow from the southwest, whereas trade winds in the Southern Hemisphere blow from the southeast.

Figure 7-13. Indian Ocean monsoon winds reverse seasonally. (a) In summer, an intense low-pressure zone forms over Asia as the land is heated by the sun and loses this heat to the atmosphere more quickly than does the ocean. The heated air rises to form the low-pressure zone at a latitude where, if the Earth were covered by ocean, the Hadley and Ferrel cells would intersect and atmospheric pressure would be high. The low-pressure zone causes the atmospheric convection cells to shift to the north and bring warm, wet, monsoon winds from the southwest to parts of Asia. A similar, but less pronounced effect occurs over North Africa. (b) In winter, the Asian landmass loses heat more quickly than does the ocean, and an intense high-pressure zone is formed by the sinking cold, dry air mass. As this air mass spreads to the south, it shifts the convection cells south, which creates the northeast monsoons and brings dry weather to much of Asia.

The warm, moist air carried by monsoon winds causes storms and torrential rains in India and Southeast Asia. This densely populated region depends on the rains to support agriculture. However, the area is plagued by alternating floods and droughts. When monsoons are strong, devastating floods occur. When monsoons are weak or absent for a year or two, drought and famine occur. The nature of the multiyear, or "interannual," climate variations that cause prolonged famine or drought is just beginning to be understood. We discuss some of these variations later in this chapter.

In the northern winter, the intertropical convergence zone moves south over the Indian and Pacific Oceans and Africa. In addition, the divergence between the Hadley and Ferrel cells moves far to the south over Asia in winter for the same reason that the intertropical convergence zone moves far north in summer. Thus, in winter this downwelling zone lies partly south of the Himalayan Mountains. The cool, dry downwelling air flows south over India, producing dry winters (**Fig. 7-13b**). Similar, but weaker and more complicated, seasonal monsoon wind reversals occur throughout East Asia, tropical Africa, and northern Australia (**Fig. 7-13**).

Sailors have taken advantage of the monsoons for thousands of years. Coastal traders sail on winter trade winds from India and Southeast Asia to East Africa and Madagascar and returned on the summer monsoon winds.

This page titled 7.4: Climatic Winds is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.5: Climate and Ocean Surface Water Properties

Ocean surface water salinity and temperature are controlled by solar radiation, transfer of heat and water between atmosphere and oceans, ocean currents, vertical mixing, and locally, river runoff.

Ocean Surface Temperatures

Figure 7-8 shows the generalized latitudinal distribution of solar radiation that reaches the Earth's surface. Ocean surface water temperatures generally decrease from the equator toward the poles (**Figs. 7-14, 7-18**), and thus, they reflect the distribution of solar radiation.

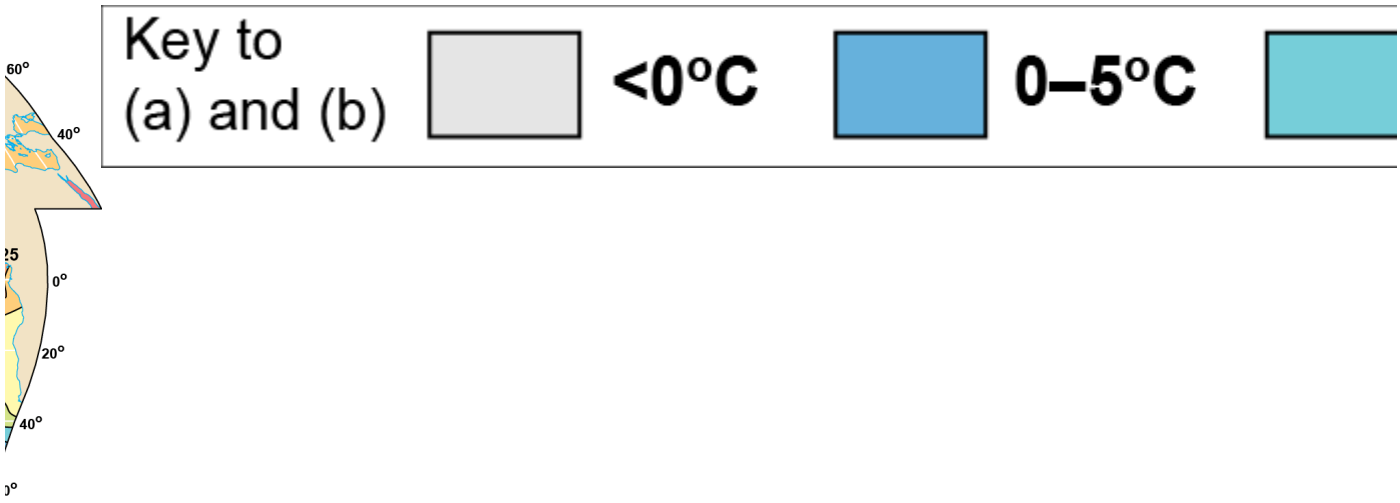


Figure 7-14. Average surface water temperatures in the world oceans in (a) February and (b) August.

On an Earth without ocean currents or winds, ocean surface water temperatures would decrease uniformly with latitude, and the **isotherms** in **Figure 7-14** would be horizontal lines parallel to lines of latitude. Although such a basic pattern is evident in **Figure 7-14**, complications are also readily apparent.

First, a band of low-temperature water lies along the equator on the eastern side of the Pacific Ocean. The band is a surface water divergence where cold, deep water is continuously upwelled (**Chap. 8**). The relatively low surface temperature at the equator in the Atlantic Ocean may be a result of similar, but less well-developed, upwelling. In the Indian Ocean, upwelling at the equator is inhibited by the monsoon winds.

The second complication evident in **Figure 7-14** is that isotherms in mid latitudes are inclined toward higher latitudes on the western sides of the oceans. Such a pattern is especially evident in the North Pacific and North Atlantic Oceans. It is a result of warm surface ocean currents that flow poleward from the subtropics on the western sides of the ocean basins and then across the ocean to the east. The currents are part of the **subtropical** gyres discussed in **Chapter 8**. In the North Atlantic Ocean the current is the Gulf Stream, and in the North Pacific Ocean it is the Kuroshio Current.

Ocean surface water temperature responds to seasonal changes in the sun's angle. High temperatures typical of tropical and subtropical latitudes extend to higher latitudes in summer (**Fig. 7-14b**). However, because of the heat-buffering action of ice (**Chap. 5**), surface water temperatures do not vary significantly with the seasons in the Arctic Ocean or the ocean surrounding Antarctica.

Ocean Surface Salinity

Differences in the salinity of ocean surface waters are caused by variations in the rate of ocean surface water evaporation, by variations in the rate of freshwater input from rainfall and land runoff, and, in limited areas, by upwelling and downwelling. Because geographic variations in evaporation and precipitation rates are somewhat complex, the distribution of ocean surface water salinity (**Fig. 7-15**) is somewhat more complicated than that of surface temperatures. Salinity is generally highest in surface waters in subtropical regions that are remote from land. It is generally lower at the equator, in polar and subpolar regions, and near continents. Surface salinity is higher in the subtropical Atlantic Ocean than in the subtropical regions of the other oceans.

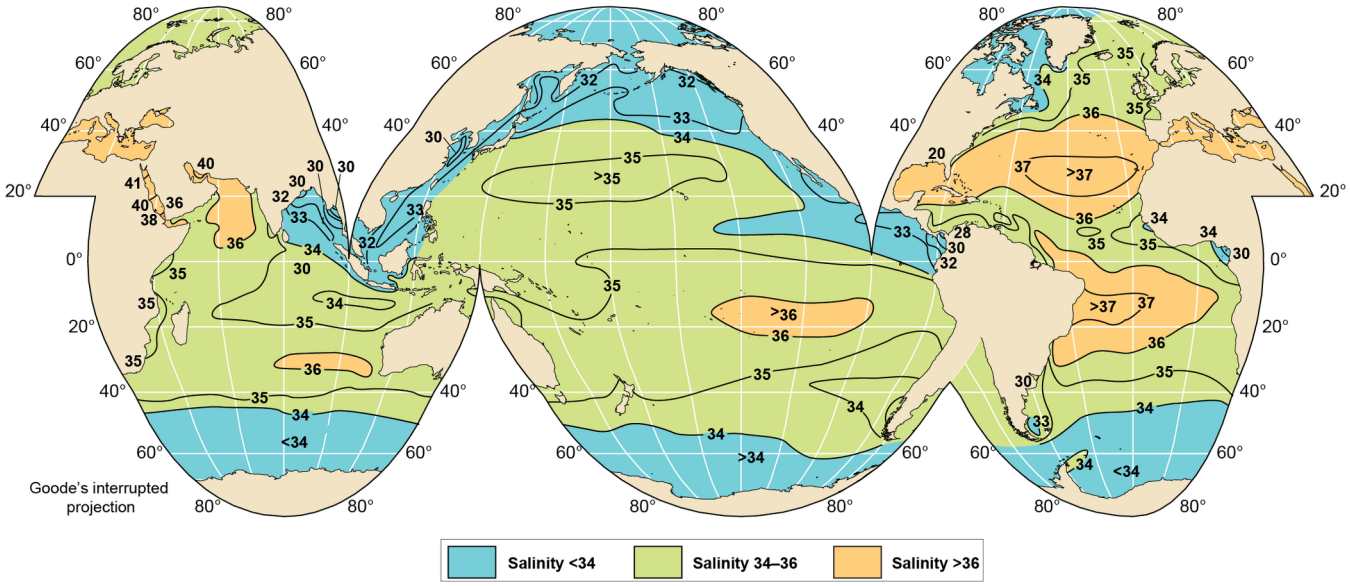


Figure 7-15. Surface water salinity in the world oceans.

The distribution of salinity is related to variations in evaporation and precipitation with latitude (**Fig. 7-16**). Precipitation is high near the equator and at mid latitudes around 40° to 60°N and S, which are the atmospheric convergence regions between the Northern and Southern Hemisphere Hadley cells and between the polar and Ferrel cells (**Figs. 7-10, 7-11**). In these regions, warm, moist air converges and rises through the atmosphere. As the air rises, its water condenses to form clouds and precipitation. Precipitation is lower in subtropical latitudes (the horse latitudes) and near the poles because these regions are under the centers of the atmospheric convection cells or at divergences.

Evaporation varies with latitude in a different way than precipitation does (**Fig. 7-16**). The evaporation rate is higher at warmer ocean surface water temperatures. Hence, evaporation generally decreases progressively from equator to pole. However, the evaporation rate is lower in equatorial latitudes than in subequatorial regions for two main reasons. First, persistent cloud cover reduces the solar intensity in the equatorial region. Second, the equatorial region is persistently calm, whereas the trade wind regions have higher winds and hence greater evaporation. The higher rate of evaporation is due to both increased airflow over the water and the increased surface area of the water caused by waves, particularly breaking waves, which create water droplets and bubbles with relatively large surface areas.

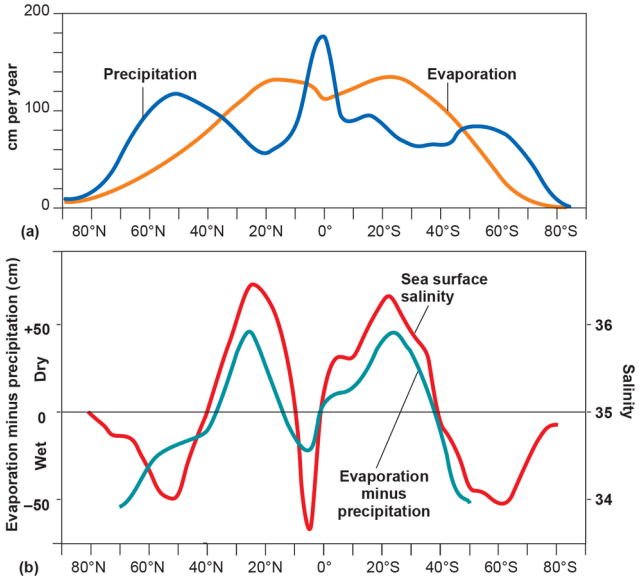


Figure 7-16. Precipitation and evaporation as a function of latitude. (a) Precipitation is high at the atmospheric upwelling zones at the equator and at 50° to 60°N and S, somewhat lower in mid latitudes, and very low at high latitudes. Evaporation generally increases from the poles toward the equator, but it is slightly lower at the equator because of the extensive cloud cover. Thus, there is a net excess of evaporation over precipitation in mid latitudes, whereas there is a net excess of precipitation within a few degrees north and south of the equator and at latitudes above approximately 40° in both hemispheres. (b) The salinity of surface ocean waters varies according to the difference between evaporation and precipitation. It is high where evaporation exceeds precipitation, and low where precipitation exceeds evaporation.

Figure 7-16b shows that precipitation exceeds evaporation across the equatorial oceans and at latitudes above about 40°N and 40°S. Therefore, surface salinity tends to be lower in these regions. In contrast, evaporation exceeds precipitation in subtropical regions. Consequently, surface salinity tends to be higher in these regions (**Fig. 7-16b**). The pattern of higher salinity in subtropical regions and lower salinity in equatorial regions and at high latitudes is observed in surface waters of the central parts of each ocean (**Fig. 7-15**). The distribution is more complex in some regions near continents because of large quantities of freshwater runoff.

The distribution of precipitation with latitude is substantially altered by the presence of landmasses (**Fig. 7-17**). The higher surface salinity in the central North and South Atlantic oceans (**Fig. 7-15**) is the result of interactions of the mountain chains of North and South America with the prevailing winds, and the outflow of high-salinity water from the Mediterranean (**Chap. 8**). In the westerly wind zones, moist air masses that move from the Pacific Ocean onto the American continents lose their moisture as rainfall on the west side of the Andes in South America and on the west side of the Rocky Mountains, the Sierra Nevada, and the coastal mountain ranges in North America. Consequently, the precipitation runs off into the Pacific Ocean. In contrast, the Northern Hemisphere trade winds in the Atlantic Ocean blow across the relatively narrow, low-altitude neck of Central America and therefore carry their moisture all the way to the Pacific. The tongue of low-salinity water that extends westward into the Pacific Ocean from Central America (**Fig. 7-15**) is evidence of this transport.

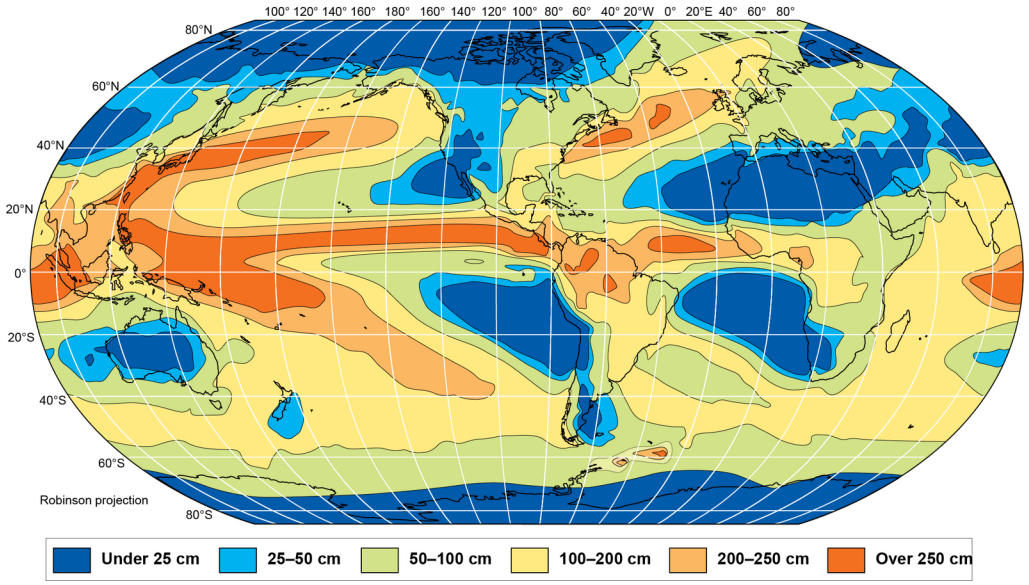


Figure 7-17. Distribution of mean annual atmospheric precipitation. Precipitation is high in the tropical atmospheric upwelling zone at the equator, generally low at mid latitudes in the atmospheric downwelling zones between the Hadley and Ferrel cells, higher in the atmospheric upwelling zones between the Ferrel and polar cells, and low again in the polar atmospheric downwelling zones. This pattern is more evident in the Southern Hemisphere because it is less modified by the presence of landmasses than it is in the Northern Hemisphere.

The Southern Hemisphere trade winds in the Atlantic Ocean carry their moisture predominantly into the Amazon and Orinoco river basins. Both of these major rivers discharge into the Atlantic Ocean equatorial region, where salinity is generally low.

Extremely high and extremely low salinities are most common in **marginal seas**, which have limited water exchange with the open oceans. Hence, if evaporation exceeds precipitation, salinity increases in such seas, and the high-salinity water is not mixed effectively with the lower-salinity water of the open ocean. Surface water salinity is high in the Mediterranean and the Arabian Gulf, and it is especially high in the Red Sea. All of these regional seas are in arid regions (**Fig. 7-17**) with low river runoff and high evaporation rates. Similarly, if precipitation and river runoff exceed evaporation, the salinity of a marginal sea is low. The best examples are the Baltic Sea and the marginal seas of Southeast Asia. In the Baltic, evaporation is low, and rainfall and runoff are moderately high (**Fig. 7-17**). In the marginal seas of Southeast Asia, rivers fed by monsoon rains discharge huge volumes of freshwater, and high rainfall rates overcome a relatively high rate of evaporation.

This page titled 7.5: Climate and Ocean Surface Water Properties is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.6: Interannual Climate Variations

In addition to seasonal variations, there are year-to-year and multiyear oscillations of climate that are related to ocean-atmosphere interactions. Oscillations have been identified in the tropical Pacific Ocean, the North Pacific Ocean, the Indian Ocean, the North Atlantic Ocean, and the Arctic Ocean. These may be linked with each other in ways we do not yet understand. One phase of the oscillation in the tropical Pacific Ocean, called El Niño, is associated with climate changes affecting much of the globe, and this oscillation has been extensively studied.

El Niño and the Southern Oscillation

El Niño (Spanish for "the boy") was named because it was first observed around Christmas time off the coast of Peru. The coastal and equatorial waters off Peru are normally regions where cold water from below the ocean's surface layers is upwelled. **Chapter 8** explains how winds cause this upwelling. The upwelled water contains high concentrations of **nutrients** that are essential to phytoplankton growth (**Chap. 12**). Upwelling areas therefore have high primary productivity rates, and massive **schools** of anchovy feed on the abundant phytoplankton that grow in the Peruvian upwelling region (**Chaps. 12, 13**). Until they were **overfished** (**Chap. 16**), the anchovy populations were so large that they yielded about 20% of the total world fish catch.

Symptoms of El Niño

The first sign of El Niño on the Peruvian coast is a temperature change in ocean surface water. Much warmer water (5°C or more warmer) displaces the normally cold surface water (**Fig. 7-18**). The less dense warmer water causes stable stratification of the water column and inhibits upwelling (**Chap. 8, CC1**), thus reducing phytoplankton populations that were nourished by the cold upwelled water. As the phytoplankton decline, the anchovy population that feeds on them plummets, and the fishery collapses. El Niño recurs generally every 3 to 5 years and usually lasts several months until the situation reverts to neutral. El Niño events have been documented in historical records as far back as 1726.

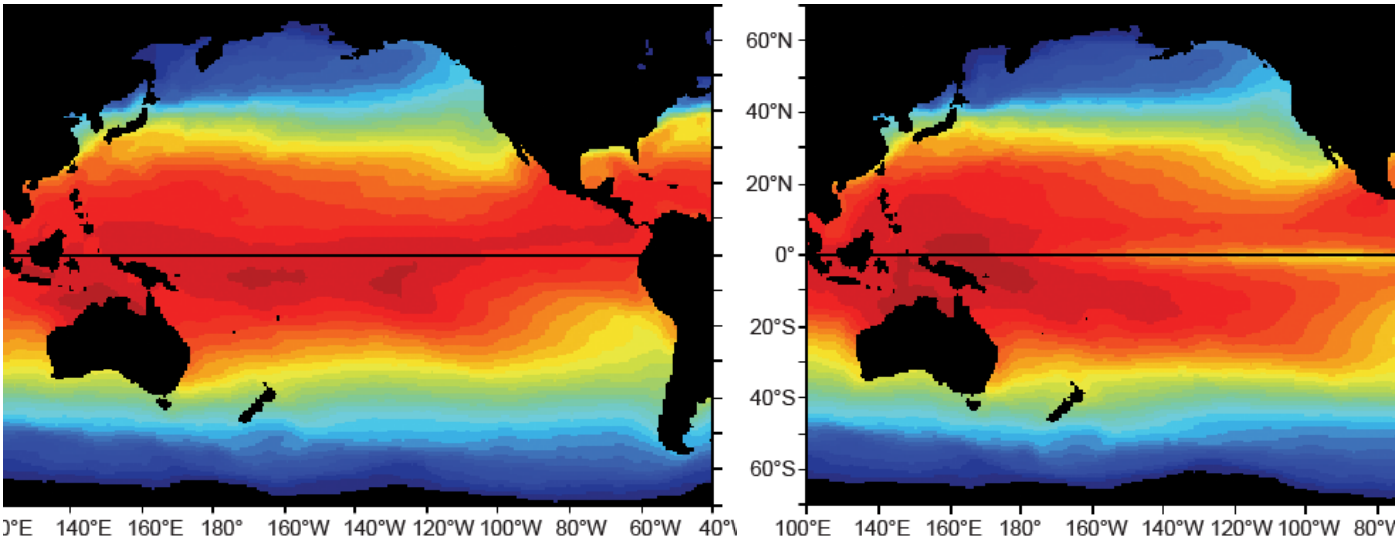


Figure 18. Sea surface temperature in the Pacific Ocean, as measured by satellite-mounted sensors. (a) During non-El Niño (neutral) conditions, there is a strong band of upwelling along the equator and in the Peruvian coastal zone, which can be seen in this December 1993 image by the yellow color that indicates lower surface water temperature in these areas. (b) During a strong El Niño (December 1997), there is no upwelling along the equator, and the Peruvian coastal upwelling is greatly reduced. (c) During La Niña conditions (December 1998), the upwelling area along the equator is substantially larger, and ocean water temperature around Indonesia is higher than normal because warm surface water is driven to the west.

El Niño is not an isolated event that occurs only off the coast of Peru. In fact, it is a complex series of cyclical changes in both ocean and atmosphere that affects much of the Earth's surface. The series of changes has been called the "El Niño/Southern

Oscillation" (**ENSO**). The name reflects the fact that the changes in ocean and atmosphere involve a very large region of the Pacific, especially the region immediately south of the equator. The sequence of ocean-atmosphere interactions during an ENSO, now relatively well known, is described in the next section.

Between El Niños, a broad band of upwelling extends across the equatorial Pacific Ocean, and strong coastal upwelling occurs off the coast of Peru. The upwelling regions and their causes are discussed in **Chapter 8**. The areas of upwelling are shown by colder surface waters (**Fig. 7-18a**). Trade winds move warm surface water to the west, where it accumulates near Indonesia (**Fig. 7-18a**). An atmospheric low-pressure area is well developed in this region because of the strong evaporation from the ocean surface and the high water temperature.

The low-pressure zone normally brings plentiful rains to Indonesia and the surrounding region. In contrast, a well-developed high-pressure zone off Peru maintains low rainfall over that country, which normally has an arid climate.

Trade winds push warm water to the west, and this warm water accumulates near Indonesia, elevating sea level by as much as a few tens of centimeters in extreme cases. For reasons not yet fully understood, this situation generally begins to change from November through April. The Indonesian atmospheric low-pressure zone and the Peruvian high-pressure zone both weaken and, in some years, actually reverse. During such periods, trade winds diminish and may reverse direction for several days. Trade winds do weaken in most years and return to normal by April, but in some years the weakening trade winds are followed by an El Niño.

El Niño/Southern Oscillation Sequence

ENSO follows a relatively well-known sequence that occurs after the Indonesian low-pressure and Peruvian high-pressure systems weaken. The sequence begins as the strength of the southeast trade winds lessens in response to the weakening Peruvian high-pressure zone. The southeast trade winds are the cause of the Peruvian upwelling (**Chap. 8**). Therefore, the upwelling is reduced or may be stopped if the trade winds abate entirely or reverse direction.

Under neutral conditions, before an El Niño is initiated, the sea surface near Indonesia is elevated (a few tens of centimeters) in relation to the region near Peru (**Fig. 7-19c**). The difference is normally partially compensated for (the sea surface slope is balanced) by higher atmospheric pressure near Peru (**Chap. 8**, **CC13**). However, the atmospheric pressure difference depends on the relative strength of the Indonesian low-pressure and Peruvian high-pressure zones. Because both the high- and low-pressure zones are weakened at this stage of ENSO, the horizontal pressure change within the tropical Pacific atmosphere is too small to balance the sea surface slope and surface water flows toward the east to lower the slope.

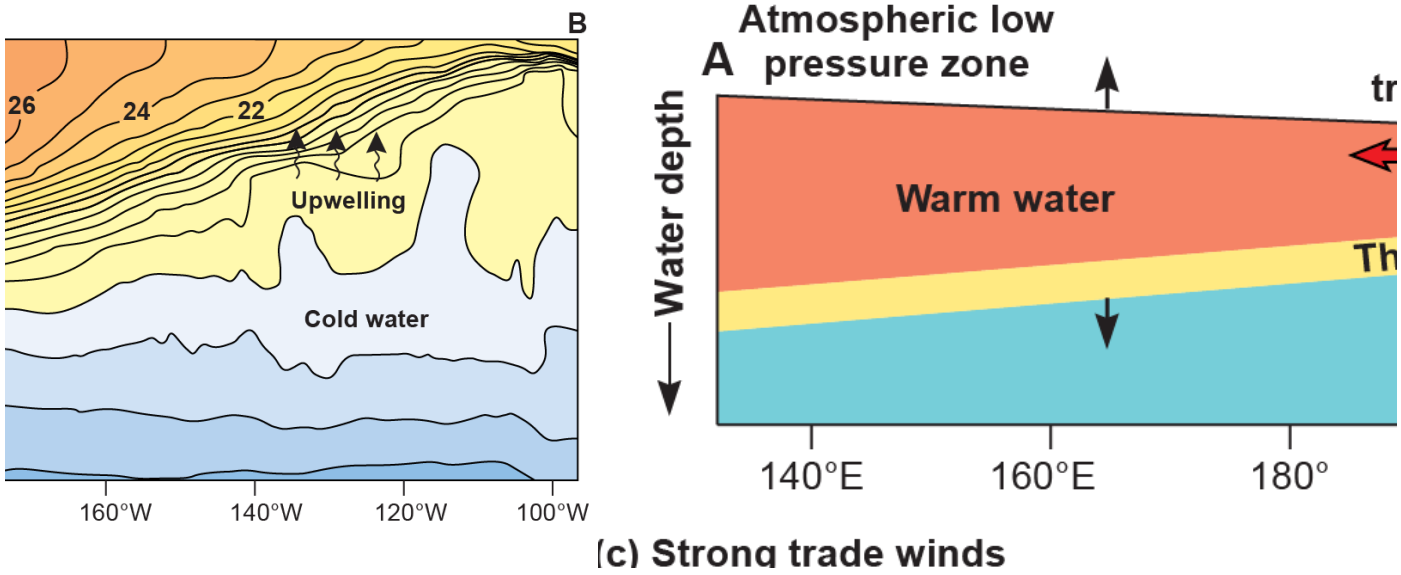


Figure 7-19. One possible model for the development of El Niño. (a) The area of the Pacific Ocean where El Niño develops. (b) When trade winds are strong, during neutral and La Niña phases, warm surface water is pushed westward and piles up on the western boundary of the Pacific Ocean near Indonesia. The surface layer is thinned near Peru, and upwelling occurs. (c,d) Vertically exaggerated cross sections show a simplified model of the interaction of trade winds with sea-level and thermocline depth in the two extreme modes of ENSO. The cross-section in part (c) shows the effect of strong trade winds on the sea surface and surface water layer. This is the normal, or in extreme cases, the La Niña phase. (d) When the trade winds diminish, surface layer water is transported eastward toward Peru, resulting in the El Niño phase. During an El Niño, the surface layer thickens near Peru, and upwelling ceases.

In response, warm surface water from the western tropical Pacific flows eastward toward Peru (**Fig. 7-19d**). The flow has characteristics of an extremely long-wavelength wave that travels directly from west to east along the equator, where there is no Coriolis effect (**CC12**). Therefore, the wave is free to flow across the entire ocean without deflection. This is one of the unique characteristics of ENSO. The warm surface water flows eastward across the Pacific Ocean over the cold upwelled water that is normally found in this region (**Fig. 7-18b**). The warmer, less dense surface water arrives near Peru and establishes a steep **pycnocline** in December, about 9 months after the event has started. Because steep pycnoclines are effective barriers to vertical mixing and water movements, the Peruvian coastal and tropical Pacific upwelling is further inhibited.

El Niño usually persists for about 3 months, but in extreme cases it may last 15 months or more. It ends when the reestablished trade winds again begin to drive the warm surface water of the tropical region to the west, thus restarting the upwelling process near Peru.

The oscillation set in motion by El Niño sometimes appears to “overshoot” as the system recovers. The trade winds become stronger, the water near Indonesia becomes warmer, and upwelling is stronger and water temperature lower near Peru (**Fig. 7-18c**). In addition, the low-pressure zone near Indonesia and the high-pressure zone near Peru become especially well developed. This situation is known as La Niña (“the girl”), in contrast to El Niño (“the boy”).

Effects of El Niño

The climatic effects of El Niño are felt far beyond the tropical Pacific. It appears that the extended area of warmer-than-usual tropical waters in the Pacific enhances the transfer of heat energy toward the poles. This change is manifested in many ways as the atmospheric convection cell system responds to the stimulus. El Niño differs in strength and other characteristics that make each El Niño’s effects somewhat different, but the same general pattern of effects is always present (**Fig. 7-20**).

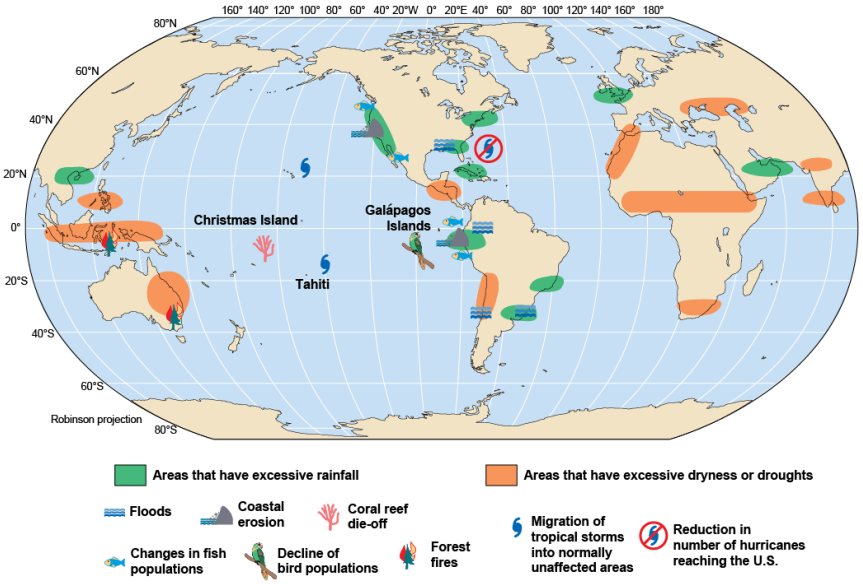


Figure 7-20. Strong El Niño events have worldwide effects that, in different areas, include droughts, excessive rainfall and floods, coastal erosion, tropical storms in areas where they do not normally occur, and various changes in terrestrial and marine ecosystems, some of which can be disastrous.

In the tropical Pacific, El Niño's effects are sometimes devastating. Reduction of the warm-water pool and weakening of the atmospheric low near Indonesia brings droughts to this normally high-rainfall region. In contrast, Peru has heavy rains and coastal flooding as sea level rises. The sea level rises because the surface water layer is warmer and less dense than normal. Consequently, the surface layer is "thicker" than the normally higher-density and cold surface water layer. Offshore from Peru, the marine **food web** collapses from lack of nutrients normally supplied by upwelled waters. In extreme cases, the result can be massive die-offs of marine organisms, which decay, strip oxygen from the water, and produce foul-smelling sulfides (**Chap. 12**). The principal fishery, anchovy, is not the only biological resource affected. In severe El Niños, lack of food decimates the huge colonies of seabirds that live on islands near Peru, and causes penguins and **marine mammals** to undertake unusual migrations and probably suffer population losses.

Changes in atmospheric circulation caused by El Niño also alter ocean water temperatures in areas far from the tropical Pacific. In the particularly strong 1982–1983 El Niño, subtropical fishes were caught as far north as the Gulf of Alaska. This El Niño caused droughts in Australia, India, Indonesia, Central America, west-central South America, Africa, and Central Europe. Excessive rains in many cases plagued parts of Southeast Asia, California, the east coast of the U.S., and parts of South America, Britain, France, and the Arabian Peninsula (**Fig. 7-20**). La Niña, in contrast, typically brings flooding rains to India, Thailand, and Indonesia, and drought to Peru and northern Chile.

Modeling El Niño

We have records of El Niños occurring regularly for centuries. La Niñas also may have occurred regularly, but they have less dramatic effects and were not recorded until scientific studies began. However, there is considerable variability from one decade to the next. In some decades, the cycle was relatively inactive; in others, it was pronounced. Since 1950, El Niño has occurred quite regularly and was followed regularly by La Niña throughout most of the period (**Fig. 7-21**). During the twentieth century, the strongest El Niño occurred in 1982-83 and two consecutive El Niños occurred without an intervening La Niña several times. In the first quarter of the 21st century, there were 7 El Niño episodes. Four of these El Niños were strong, and the 2015-16 El Niño almost tied the 1997-98 episode's record for the strongest El Niño in recorded history. Because El Niño can have worldwide impacts on ecosystems and fisheries, and can cause widespread droughts and floods, major scientific efforts have been made to develop mathematical models (**CC10**) that can predict the occurrence of the phenomenon ahead of time. A number of models have been developed, each calibrated with historical data. As models have developed, they have become increasingly sophisticated, requiring increasingly detailed atmospheric and oceanic data. Gathering these data is one of the most intensive and comprehensive observational programs ever undertaken in the Earth sciences.

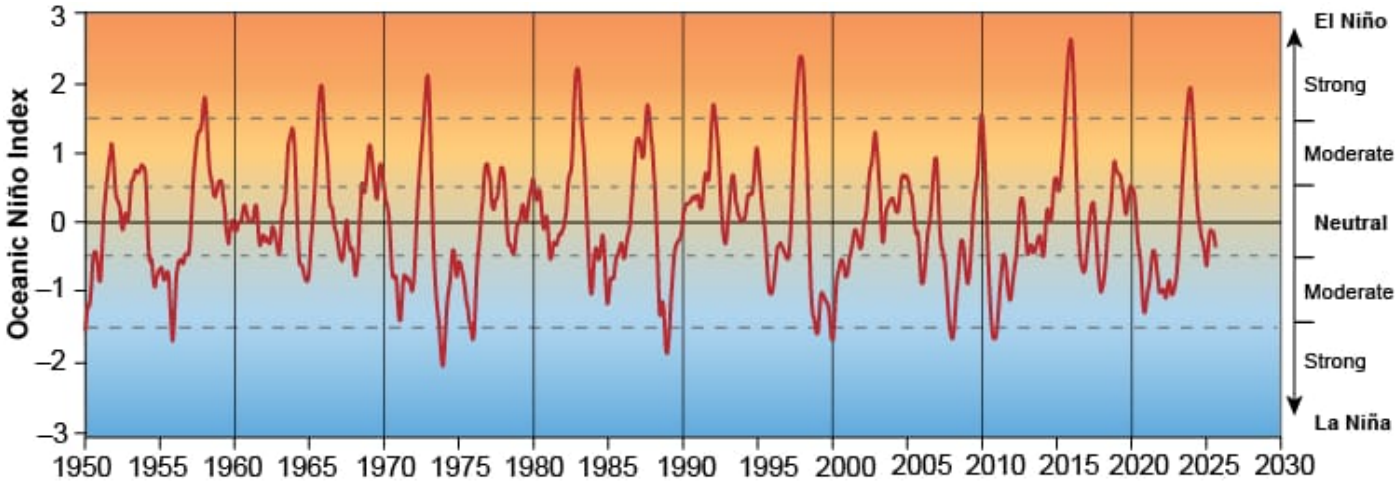


Figure 7-21. The NOAA Relative Oceanic Niño Index since 1950. The index is based on the average surface water temperature in a defined region of the central tropical Pacific Ocean compared to the surface water temperature of global tropical oceans. El Niño (when the index rises above 0.5) and La Niña (when the index falls below −0.5) have occurred regularly about every 2–7 years.

By 1997, several models were sufficiently developed to make predictions, and all of them forecast that a modest El Niño would occur in the winter of 1997-98. The model predictions were only partly correct, as the El Niño was not a modest one. It turned out to be the most intense event on record, costing an estimated 23,000 lives and $33 billion (1998 dollars) in damage worldwide. What went wrong?

Fortunately, this El Niño was extensively observed by satellite sensors, moored instrument arrays, autonomous floats, research vessels, and aircraft. When the data were analyzed, the cause of the model failures appeared to be a burst of westerly winds in the western Pacific that occurred just as the eastward warm-water flow of the El Niño phenomenon was starting. These winds, unanticipated in the models, appear to have been timed perfectly to increase the flow of warm water across the ocean and, thus, intensify the El Niño.

Further research suggested that similar bursts of westerly winds also may have occurred just as other unusually intense El Niños had started. It has been hypothesized that these westerly winds are part of another ocean-atmosphere oscillation, called the Madden-Julian Oscillation (MJO). About every 30 days, this oscillation causes bursts of winds that are associated with an eastward-moving patch of tropical clouds. These clouds extend high into the troposphere and are easily observed from satellites. If this hypothesis is correct, the most intense El Niños may turn out to be formed only when the timing of this burst and the ENSO sequence are exactly right. The MJO cannot yet be reliably forecasted well in advance, and it may never be if the system is **chaotic** (**CC10, CC11**), which is very likely. Unfortunately, this would mean that the intensity of an El Niño could not be accurately forecast until it had already started.

El Niño Modoki

Since the early 1990s, scientists have noted a new type of El Niño that has been occurring with greater frequency and has been getting progressively stronger. In an El Niño Modoki (Japanese for “similar but different”), the maximum ocean warming is found in the central-equatorial, rather than eastern, Pacific. Such central Pacific El Niño events were observed in 1991-92, 1994-95, 2002-03, 2004-05, and 2009-10. Many of the global climate effects of an El Niño Modoki are similar to those that occur in a classic El Niño, but others are significantly different. For example, during a regular El Niño, there are generally fewer Atlantic hurricanes that make landfall in the U.S., whereas in an El Niño Modoki, there are more such hurricanes. Since detailed instrument and satellite studies of ENSO have covered only several decades, researchers are not yet sure whether El Niño Modoki is evidence of human-induced global climate change in the ENSO pattern or whether it is simply a natural long-term variability of ENSO. Some model studies have hypothesized that global warming due to human-produced greenhouse gases could shift the warming center of El Niños from the eastern to the central Pacific, further increasing the frequency of El Niño Modoki events in the future, but recent El Niños do not yet appear to show this trend.

Other Oscillations

ENSO is the best-known ocean-atmosphere oscillation, and it is believed to be the most influential in causing the interannual

variations of the Earth's climate. A number of other oscillations are known, and they too have an influence on climate, but primarily just in specific regions. The MJO, discussed in the previous section, is one such oscillation, but there are several others that have longer periods and greater effects on regional climates. These include relatively well-known oscillations in the North Atlantic and North Pacific Oceans: the North Atlantic Oscillation (NAO) and the Pacific Decadal Oscillation (PDO), respectively. More recently, oscillations have been observed in the Arctic Ocean and in the Indian Ocean. The details of what is known about these oscillations and their effects on regional climates are beyond the scope of this text. However, the following paragraphs provide a very brief overview of each.

The North Atlantic Oscillation is a periodic change in the relative strengths of the atmospheric low-pressure region centered near Iceland and the subtropical high-pressure zone near the Azores. This pressure difference drives the wintertime winds and storms that cross from west to east across the North Atlantic. When the gradient between the two zones weakens, more powerful storms create harsher winters in Europe. The shift can affect marine and terrestrial ecology, food production, energy consumption, and other economic factors in the regions surrounding the North Atlantic. The periodicity of NAO is highly variable, but it ranges from several years to a decade or more. The Arctic Oscillation is closely coupled with the NAO. The Arctic high-pressure region tends to be weak when the Icelandic low is strong. The strength of the Arctic high affects climate around the Arctic Ocean, as well as the direction and intensity of ice drift.

Two modes characterize the Pacific Decadal Oscillation. In one mode, the sea surface temperatures in the northwestern Pacific, extending from Japan to the Gulf of Alaska, are relatively warm, and the sea surface temperatures in the region from Canada to California and Hawaii are relatively cool. In the other mode, these relative temperatures are reversed. In the first mode, the jet stream flows high across the Pacific and dips south over the Pacific coast of North America. Pacific storms follow the jet stream onto the continent over Washington, Oregon, or British Columbia, and California is cool and dry. Moist, relatively warm air is transported across the northern half of the United States, resulting in relatively mild, but often wet, winters in this region and in states influenced by warm, moist air masses from the Gulf of Mexico. In the second mode, the jet stream and Pacific storm tracks parallel the Pacific coast into Canada and Alaska. Parts of Alaska and much of California experience a mild winter. However, the jet stream's location allows cold, dry Arctic air to flow over the central United States, pushing the Gulf of Mexico air mass south, and a generally cold winter occurs in the eastern United States. We discuss some of the effects of the PDO in more detail in the next section.

In the Indian Ocean, there is no strong oscillation comparable to ENSO in the Pacific. However, there is evidence of a weaker oscillation that is similar. Although the Indian Ocean oscillation, known as the Indian Ocean Dipole (IOD), is much weaker than ENSO, it exhibits similar warm, neutral, and cold phases. The warm phase is characterized by warmer water on the western side of the ocean and cooler water on the eastern side, while the cold phase has warmer water in the eastern Indian Ocean and cooler water in the west. Modeling has shown that the intense drought of 1998–2002 over a swath of mid latitudes spanning the United States, the Mediterranean, southern Europe, and Southwest and Central Asia was possibly caused by a combination of the strong El Niño of 1997-98, and the cool phase of the IOD oscillation in the Indian Ocean.

Climate Chaos?

The decade following the 1976-77 winter was extremely unusual in the U.S. and elsewhere. For example, in the following eight years, five severe freezes damaged Florida orange groves. Such freezes had occurred, on average, only once every 10 years over the previous 75 years. In addition, waves 6 m or higher tore into the southern California coastline 10 times in the 4 years from 1980 to 1984. Such waves had occurred there only 8 times in the preceding 80 years.

Were all these "unusual" events the result of a sudden shift in climate or just due to normal year-to-year variations of the Earth's climate? Because any one such change could be just a year-to-year variation, researchers examined the records of 40 different environmental variables that reflect climatic conditions around the Pacific region. The variables include wind speeds in the subtropical North Pacific Ocean, the concentration of **chlorophyll** in the central North Pacific Ocean, the salmon catch in Alaska, the sea surface temperature in the northeastern Pacific, and the number of Canada goose nests on the Columbia River. The 40 variables were combined to derive a single statistical index of climate for each year from 1968 to 1984.

The index showed a distinct and abrupt step-like change in value between 1976 and 1977, which is truly remarkable because such a clear signal of changed conditions is seldom, if ever, found even in any single environmental variable. Usually the many complex natural variations would obscure such a pattern. Was this change real? Did it indicate a chaotic "phase shift" between two relatively stable sets of environmental conditions? This study was perhaps the first evidence that chaotic shifts do indeed occur. It later became known that this shift was associated with the Pacific Decadal Oscillation (PDO), named in 1997. In 2000, this study was repeated, but this time using as many as 100 variables: 31 time series of atmospheric or oceanic physical variables, such as indices of the sea surface temperature; and 69 biological time series, such as catch rates of salmon. The analysis was

performed for the period 1965–1995, but because data were not available for all indices for each year, the analysis was done in two blocks—1965–1985 and 1985–1995—using a slightly different suite of indices for each. The results were as startling as those of the early studies. Distinct phase shifts were identified between 1976 and 1977, and between 1988 and 1989 (**Fig. 7-22**). These dates correspond with reversals of the PDO. Further evidence that these phase shifts may involve elements of chaos comes from the observation that some of the individual biological indices that changed in 1997 did not simply reverse in 1989. Indeed, some made step increases in both years, and some made step decreases in both years. These observations indicate that some species did not return to their former numbers in the ecosystem. This indicates that atmospheric oscillation driven shifts may be an agent of long-term change in the success of individual species within ocean and, perhaps, terrestrial ecosystems.

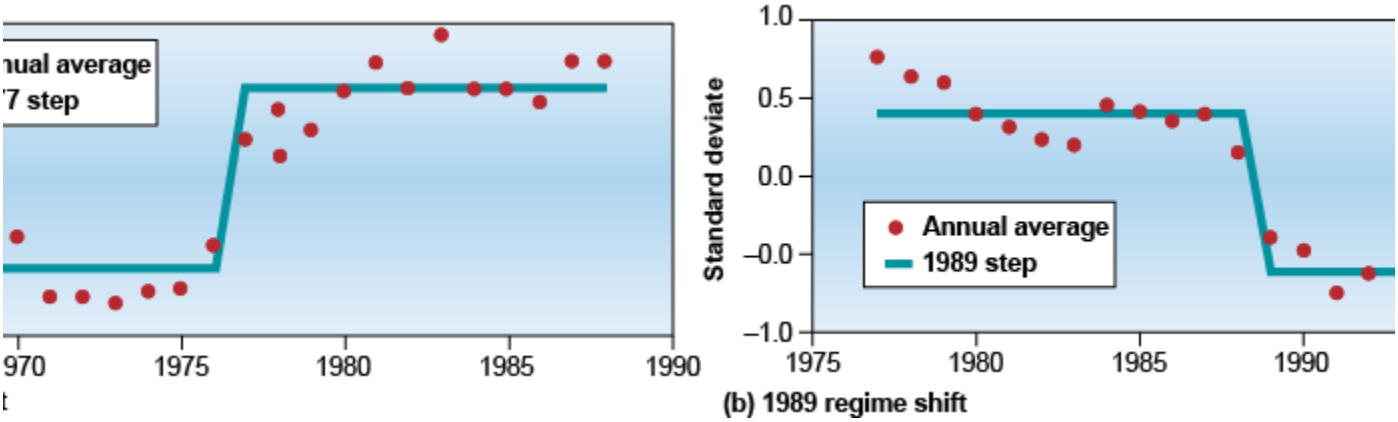


Figure 7-22. Variations in a composite index of environmental variables related to climate in the North Pacific Ocean. Time series were examined for 100 variables—31 climatic and 69 biological—although data were not available for all variables throughout the analysis period. (a) A regime shift occurred in the winter of 1976–1977, as shown by the step change in this function, which includes time series for 91 variables. (b) A regime shift also occurred in 1989, as shown by the step decrease in an index that includes time series for 94 variables. However, this second shift did not restore the system to its pre-1977 state.

If changes in the Earth's ocean-atmosphere climate system and associated ecological effects are chaotic, what does this mean? Chaotic systems often remain stable, varying within a well-defined range of behaviors, until the entire system suddenly shifts to a new stable state and varies within a different range of behaviors (**CC11**). Thus, it may mean that ecosystems are constantly changing in unpredictable ways. It may also mean that, even if no dramatic climate changes have yet occurred in response to human releases of greenhouse gases, there is no guarantee that any future changes will be small or slow. Drastic, sudden climate changes could be triggered if greenhouse gas concentrations in the atmosphere continue to increase, or they may already be inevitable. Decades of research on the Earth's ocean-atmosphere climate system are needed before we can have confidence in any forecasts of future climate in a world in which we have caused an unprecedented, rapid increase in the concentration of greenhouse gases in the atmosphere. Furthermore, such forecasts may never be more reliable than the local weather forecast.

This page titled 7.6: Interannual Climate Variations is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.7: Land-Ocean-Atmosphere Interactions

Landmasses interact with atmospheric circulation in two ways. First, they can physically block or steer air masses moving across the Earth's surface. Air must rise over mountain chains or flow around them. Second, the thermal properties of landmasses and of the ocean are very different because rocks and soil have lower heat capacities than water has. Consequently, land warms more quickly than the oceans under the same solar radiation intensity. Also, land cools and gives off heat to the atmosphere much more quickly than the oceans do.

Interactions of landmasses with the atmosphere and oceans exert substantial control over the climate characteristics of most of the Earth's land and much of its oceans. The effects of the interactions on climate may be extremely localized and limited to one small area, or they may range across an entire continent or ocean. On the local scale, effects include land and sea breezes and the "island effect," which are discussed later in this chapter. On a larger scale, effects include the monsoons and the climate characteristics of the interior of continents. In the following sections of this chapter we discuss how land-ocean-atmosphere interactions can occur on different geographic scales. We review first the global interactions that determine climate, then larger-scale weather events that extend or travel across large regions of a continent or ocean, and finally smaller-cale weather features that occur locally.

This page titled 7.7: Land-Ocean-Atmosphere Interactions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.8: Global Climate Zones

Climate in a given region is defined by several factors, including the average annual air temperature, seasonal range of temperatures, range of temperatures between day and night, extent and persistence of cloud cover, and annual precipitation and its seasonal distribution. Each factor varies with latitude and location in relation to land, mountain chains, and ocean. The factors also vary from year to year. Hence, most climate zones are not separate and distinct, but merge into one another. The demarcation lines drawn between the zones in **Figure 7-23** are therefore only approximations.

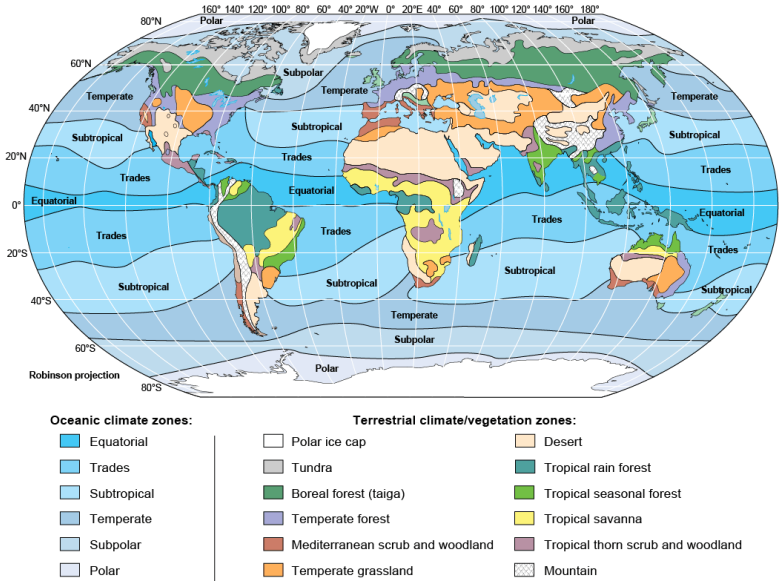


Figure 7-23. The climate zones of the oceans are arranged in bands that generally follow a latitudinal pattern and that are related primarily to ocean surface water temperature (see **Figure 7-14**). Terrestrial climate and vegetation zones follow a more complex pattern that reflects the presence of mountain chains, the prevailing direction of air mass movements, and the proximity to the oceans, all of which affect the distribution of rainfall and seasonal temperature ranges.

Ocean Climate Zones

Climate zones over the oceans are generally latitudinal bands that parallel the average ocean surface temperatures shown in **Figure 7-14**.

Polar ocean zones are covered with ice for much of the year, and ocean surface temperatures remain at or close to freezing all year (**Fig. 7-24a**) because of the heat-buffering effect of ice (**Chap. 5**, **CC5**). These zones have low precipitation and generally light winds (except where land and ocean interact). In subpolar zones, sea ice forms in winter but melts each year. Despite low precipitation amounts in the polar and subpolar zones, ocean surface salinity, particularly in the Arctic Ocean, is low. Surface water salinity is lowered during the continual freeze-and-thaw cycle that creates sea ice. As water freezes, **ice exclusion** causes dissolved salts to be excluded from the ice. The salts are left in the remaining water which raises the salinity and the resulting cold, high-salinity surface water sinks because its density is greater than the water below (**CC1**). When sea ice melts in the spring and summer, the non-salty water released from the melting ice mixes with the upper layer of ocean water and lowers its salinity. In the Arctic Ocean, freshwater runoff from the continents also contributes to low surface-water salinity.

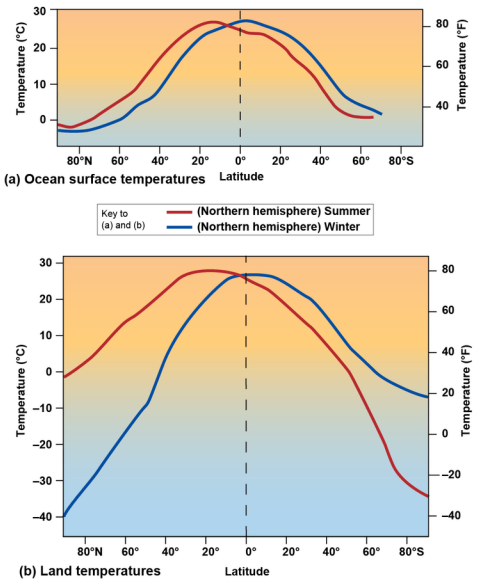


Figure 7-24. Seasonal and latitudinal variation in temperatures at the Earth's surface. (a) Average ocean surface water temperature. (b) Average temperature over land (or ice-covered ocean). Seasonal variation is much greater over land than over ocean, especially at high latitudes.

Temperate ocean zones are located in the zones of strong westerly winds (**Fig. 7-11**). They have high precipitation (**Fig. 7-17**) and are subject to strong storms called "extratropical cyclones." These storms form especially in winter at the atmospheric polar fronts, which are locations where the polar and Ferrel cells meet (**Fig. 7-10**). The storms travel eastward and toward lower latitudes. Extratropical cyclones are described later in this chapter.

Subtropical ocean zones are located at the divergence between the atmospheric Hadley and Ferrel cells (**Fig. 7-10**). In these zones, winds are generally light, skies are usually clear, and precipitation is low. Many of the world's most desirable **beach** vacation areas are on coasts in these zones.

Trade wind zones are generally dry with limited cloud cover, but they are subject to persistent winds. Ocean surface water evaporation is very high, and thus, salinity tends to be high (**Figs. 7-15, 7-16**). In the equatorial region, ocean surface waters are warm (**Fig. 7-14**), evaporation is high (**Fig. 7-16a**), winds are light (**Fig. 7-11**), clouds are persistent, and rainfall is high and continuous throughout the year (**Fig. 7-17**).

Land Climate Zones

Land (terrestrial) climate zones are not arranged into the same type of orderly latitudinal bands that characterize ocean climate zones (**Fig. 7-23**). The reasons for this difference include differences in thermal properties between land and water and the effects of mountains on air mass movements.

The average ocean surface temperature at any latitude changes little between seasons (**Fig. 7-24a**) because of the high heat capacity of water (**Chap. 5**, **CC5**). In contrast, soil and rocks have a much lower heat capacity. Consequently, seasonal changes in solar intensity cause substantial seasonal changes in average land-surface temperature (**Fig. 7-24b**). Seasonal changes are small in equatorial regions and increase with increasing latitude. The reason is that the annual range of daily total solar energy received at the Earth's surface increases with the distance from the equator.

Figure 7-24b shows the temperatures at continental locations far from oceanic influence. Daily and seasonal temperature changes are less in coastal locations than in regions far from the oceans at the same latitude (**Fig. 7-25**). The ocean provides a source of relatively warm air in winter and cool air in summer. Although the coastal air mass moderates climates in all coastal locations, the distance inland to which this effect reaches on a specific coast is determined by the prevailing wind direction and the location and height of mountain ranges.

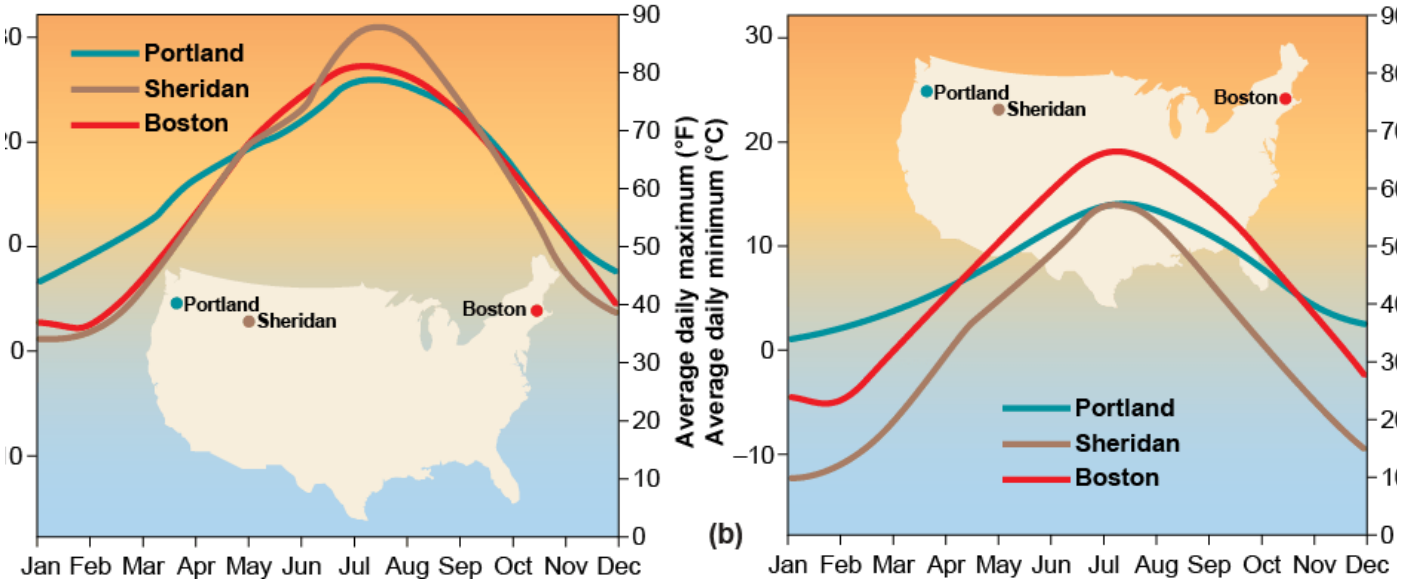


Figure 7-25. Annual variation in monthly average air temperature for Portland, Oregon; Sheridan, Wyoming; and Boston, Massachusetts. (a) Daily high temperatures. (b) Daily low temperatures. These cities are at approximately the same latitude; however, their climates are very different. Portland has a maritime climate with a small annual temperature range; Sheridan has a continental climate with a large annual temperature range; and Boston has a more continental climate than would be expected from its coastal location, because the climatic winds there blow from the continent toward the ocean.

If climatic winds blow onshore, coastal-ocean air can moderate temperatures and enhance precipitation many hundreds of miles beyond the coast. For example, warm coastal air that flows eastward across Europe is unimpeded by mountains. Hence, the moderate marine climate extends farther inland in western Europe than in the western parts of other continents, such as North and South America, where coastal mountains impede the westward flow of warm, moist air from the Pacific Ocean (**Fig. 7-23**). In the latter regions, the mountain effect (described in more detail later in this chapter) removes the principal source of the air mass's heat: its water vapor.

If climatic winds blow offshore, the moderating influence of the ocean is reduced. For example, Boston, Massachusetts, has a wider temperature range than Portland, Oregon (**Fig. 7-25**). Although both are coastal cities at about the same latitude, Boston is in a region of offshore climatic winds, while Portland is in a region of onshore climatic winds.

Coastal land climates are also affected by ocean currents. For example, Scandinavia and Alaska are at about the same latitude, but winters are much more severe in Alaska than in Scandinavia. In winter, most of Scandinavia is in the westerly wind zone and the temperate ocean climate zone, whereas most of Alaska is in the polar easterly wind zone and the subpolar ocean climate zone (**Figs. 7-11, 7-23**). In the North Atlantic Ocean, the Gulf Stream Current carries warm water from Florida north and west across the ocean to the seas around Scandinavia. The warm water provides heat energy through evaporation to moderate the Scandinavian atmosphere in winter. In fact, enough heat energy is provided to cause the westerly wind zone to extend into the region throughout the year (**Fig. 7-11**). A similar warm current, the Kuroshio Current, flows north and west from south of Japan toward Alaska (**Chap. 8**). However, the Aleutian Island arc deflects this warm current and it does not reach the Bering Sea on the west coast of Alaska.

The terrestrial climate zones shown in **Figure 7-23** are defined primarily by seasonal temperature ranges and precipitation rates. In any given location, these parameters are controlled by the interaction of several factors, including ocean water temperatures, climatic winds, distance from the ocean, and the locations of mountain chains. A complete description of the zones is beyond the scope of this text, but the knowledge gained in this chapter can be used to explore the distribution of terrestrial climate zones depicted in **Figure 7-23**.

This page titled 7.8: Global Climate Zones is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.9: Weather Systems

Wherever we live, we know from everyday experience that winds do not always blow in the directions of the climatic winds depicted in **Figure 7-11**. Furthermore, contrary to the average climatic zone characteristics discussed previously, we know that some days are cloudy and rainy, and others are sunny and dry. These short-term variations, called "weather," are caused by temporal and **spatial** variability in the motions of air in the atmosphere that can result from local variations in evaporation rate or in heating or cooling of the atmosphere. They can also result from **friction** as air flows over land or ocean surfaces. We can see this effect by observing how a smooth-flowing stream becomes churned as it interacts with rocks in rapids.

Atmospheric air movements are chaotic, and precise motions of any individual particle of air cannot be predicted (**CC11**). Therefore, we cannot ever hope to forecast weather precisely. Any forecast will be progressively less reliable the further ahead in time it is made. Weather forecasts made more than a few days ahead are unreliable and always will be. At best, future longer-range weather forecasts will be able to predict only trends in average weather that might occur over many days.

A complete discussion of weather systems is beyond the scope of this book. However, the behavior of high- and low-pressure zones, hurricanes, and extratropical cyclones is described briefly in the sections that follow.

High- and Low-Pressure Zones

Atmospheric pressure at the Earth's surface varies at any given location from day to day and over distances of tens or hundreds of kilometers across the surface at any given time. Once created, areas of slightly higher or lower pressure, which may cover hundreds or thousands of square kilometers, can persist for days or weeks. Atmospheric high- or low-pressure zones ("highs" and "lows" in weather media jargon) generate winds because air tends to flow on the pressure gradient outward from a high-pressure zone and inward toward a low-pressure zone. The moving air does not flow directly away from a high-pressure zone or toward a low-pressure zone, because the Coriolis effect deflects the moving air mass (**CC12**).

The Coriolis effect deflects moving air ***cum sole*** (which means "with the sun")—that is, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Thus, in the Northern Hemisphere, air that flows toward a low-pressure zone is deflected to the right. The deflection continues until the pressure gradient balances the Coriolis effect. This balance results in a situation known as **geostrophic** flow, as discussed in **CC13**. A pattern is established in which winds flow counterclockwise around the low-pressure zone (**Fig. 7-26b**). Similarly, clockwise winds are established around a high-pressure zone (**Fig. 7-26a**). In the Southern Hemisphere, winds flow in the opposite direction: clockwise around a low-pressure zone and counterclockwise around a high-pressure zone. Friction between the moving air and the ground or ocean surface reduces the Coriolis deflection. Consequently, winds at, and near, the surface tend to spiral toward the center of low-pressure zones and away from the center of high-pressure zones.

Near-surface winds are influenced by friction and are therefore not true geostrophic winds, but they are in approximate geostrophic balance. As a result, their speed and direction can be reasonably estimated from isobaric charts that show the geographical variations of air pressure (**CC13**). Therefore, atmospheric pressure is among the most important measurements made at weather stations worldwide, and isobaric charts are a major tool for weather forecasting. **Wave height** forecasts also can be made from such charts because the relationships among wave height and wind speed, duration, and distance over which the wind blows (called **fetch**) are known (**Chap. 9**).

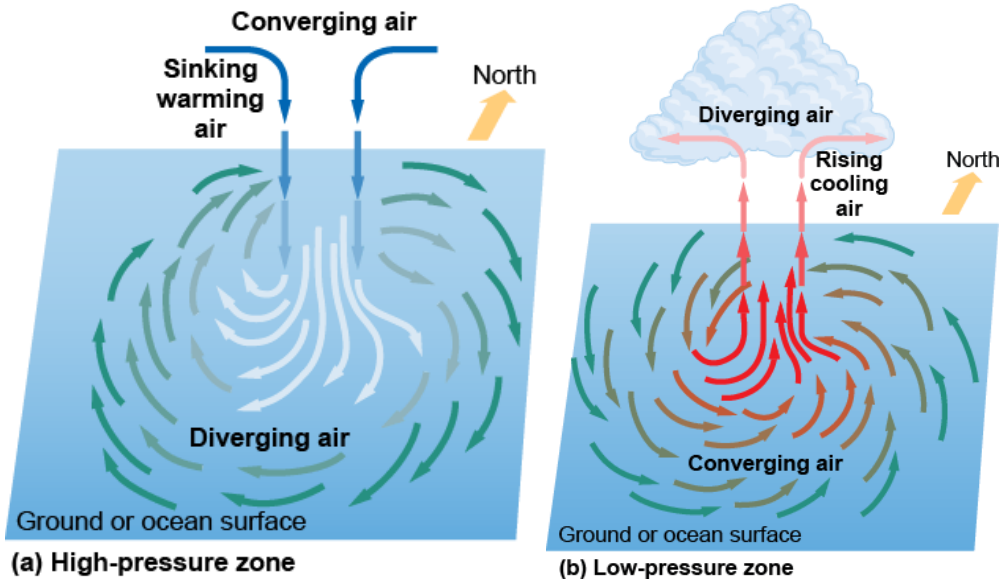


Figure 7-26. High- and low-pressure zones. (a) Cold, dense air sinks, creating a high-pressure zone in the lower atmosphere. The sinking air warms and flows out from the high-pressure zone. As air flows outward, it is deflected by the Coriolis effect, forming winds that flow clockwise around the high-pressure zone (counterclockwise in the Southern Hemisphere). (b) Warm, moist air rises, creating a low-pressure zone in the lower atmosphere. The rising air cools, and water vapor condenses, causing precipitation. The warm rising air at the center of the low-pressure zone is replaced by air drawn toward the low from across the Earth's surface. As air flows inward, it is deflected by the Coriolis effect, forming winds that flow counterclockwise around the low-pressure zone (clockwise in the Southern Hemisphere).

High- and low-pressure zones continuously form and re-form in the atmosphere and vary substantially in size and persistence. Low-pressure zones can be formed over the ocean where slightly higher ocean surface temperatures and/or higher evaporation rates occur. The result is an air mass in the low-pressure zone that has a temperature and water vapor content that is elevated compared to surrounding air. As increasing temperature and water content lead to a decrease in density, this air mass rises as a plume (**Fig. 7-26b**).

Tropical Cyclones

Tropical cyclones are among the most destructive forces of nature. These intense storms form mostly during local summer or fall in both the Northern and Southern Hemispheres. Tropical cyclones develop over the warm ocean waters of tropical regions where a low-pressure zone develops that draws air inward across the ocean surface. If conditions are favorable, the low-pressure zone deepens (that is, the pressure drops further), and a tropical cyclone may form. In the Indian and western Pacific Oceans, tropical cyclones are called "cyclones" north of the equator and "typhoons" south of the equator. They are also called "baguios" around the Philippines and "willy willys" near Australia (**Fig. 7-29**). They are known as "hurricanes" only in the Atlantic and eastern Pacific Oceans, but they are the same phenomenon regardless of the name.

Winds in the developing low-pressure system are deflected by the Coriolis effect (**Fig. 7-26b**) until they flow around the low-pressure zone. Additionally, a centripetal ("center-seeking") force helps confine motion to a curved path. As the air flows in toward and around the low-pressure zone across the ocean surface, it is warmed, gains moisture, and tends to rise. The winds thus blow toward the center, or "eye," of the tropical cyclone in a rising helical pattern (**Fig. 7-27**). Because winds do not reach the eye to alleviate the low pressure, the low-pressure zone continues to intensify. The horizontal pressure gradient increases, and the wind speeds increase (**CC13**).

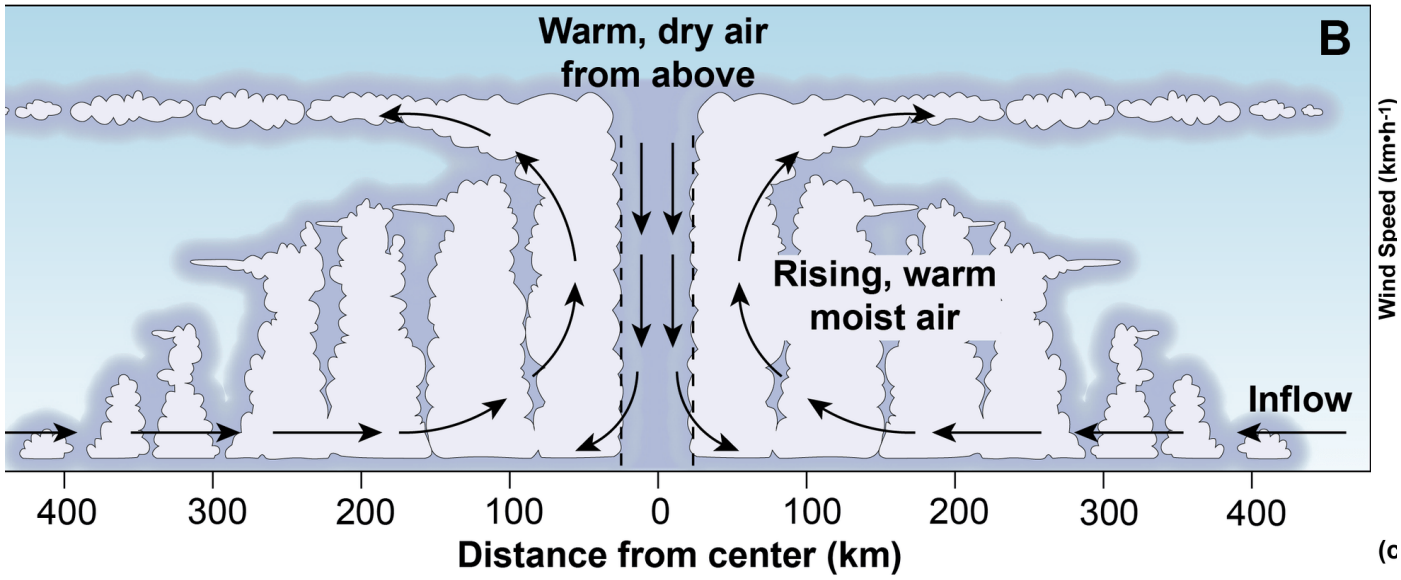


Figure 7-27. (a) Tropical cyclones are rotating storms in which air masses are drawn toward a low-pressure zone and rise in a helical pattern. This process forms spiral rain bands and a cloudless eye. (b) A vertical section of the hurricane across the line from A to B in (a) shows how the warm, wet air rises and spreads out at an altitude of more than 10,000 m. Dry air descends in the center of the storm and is heated adiabatically to form the eye. (c) A plot of wind speed and atmospheric pressure at the surface in the southwest-to-northeast cross section (A to B). Winds are strongest in the northeast quadrant of a hurricane, about 100 km from the eye of a typical storm, but the heaviest rainfall takes place closer to the eye. Storms may vary in size, and the maximum wind speed is higher when the atmospheric pressure in the eye is lower.

A well-formed tropical cyclone has an eye, usually about 20 to 25 km in diameter, within which winds are light and there are few clouds (**Fig. 7-27**). The eye forms when the low-pressure zone deepens enough to draw warm, relatively dry air downward into the eye, as winds are deflected from the eye in their rising helical pattern (**Fig. 7-27**). The eye is surrounded by a wall of clouds (the eye wall) where winds are moderate, but intense thunderstorms are created as the rotating air mass rises. Outside the eye wall is a region of intense winds that blow across the ocean surface (**Fig. 7-27**). The winds are counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere.

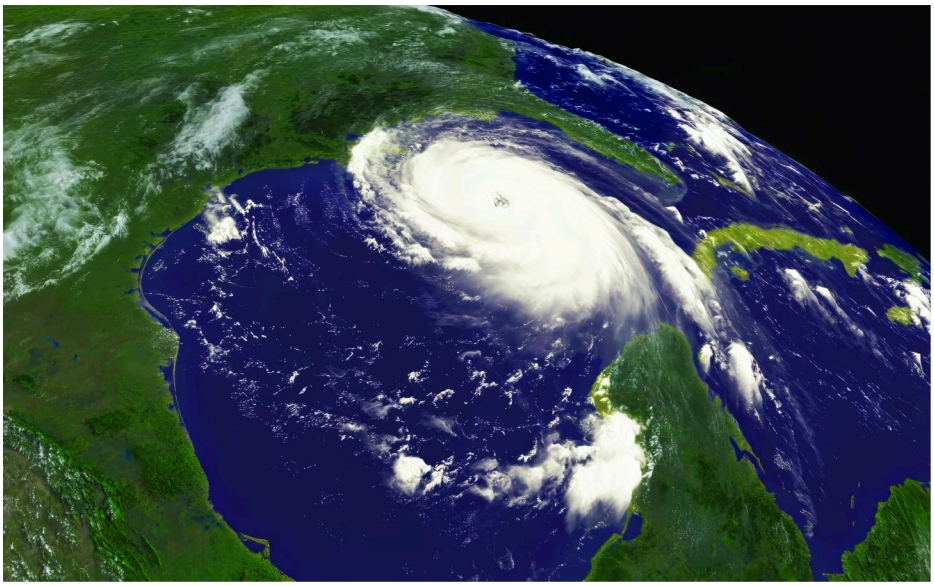


Figure 7-28. This satellite image of Hurricane Katrina (2005) was obtained as the hurricane moved across the Gulf of Mexico, just a few hours before it struck land, costing many lives and causing billions of dollars of damage in Louisiana and Mississippi. In the image, the outer bands of cloud are just reaching the coast near New Orleans while the trailing edge clouds extend as far as the Yucatán Peninsula in Mexico, so this monster storm was about 1000 km in diameter.

Tropical cyclones drift to the west in the prevailing trade wind direction at about 10 to 30 $km \bullet h^{-1}$, and sometimes faster. Their

westward paths are deflected by the Coriolis Effect, and they generally turn away from the equator. Over warm ocean water, hurricanes maintain or gain intensity as high temperatures and high winds encourage evaporation and feed more latent heat energy into the storm. When a hurricane moves over cold water or land, it quickly loses energy and releases its moisture as intense rainfall (10 to 20 billion tonnes of rainfall per day from an average hurricane).

Storm surge, freshwater flooding, wind, tornadoes, and surf and rip currents are all significant tropical cyclone hazards, some occurring far from the storm's landfall. The potentially most destructive and life-threatening effect of a hurricane has historically been the **storm surge**, or wave that pushes ahead of the storm. Storm surges cause flooding of low-lying coastal areas. However, improved early warning and evacuation procedures have reduced storm surge deaths, and drowning due to inland flooding and landslides caused by wind and rainfall from the tropical cyclone are now the largest cause of cyclone-related deaths in most areas.

The deadliest hurricane in U.S. history was the Galveston hurricane of 1900, that killed 8000 people, followed by the 1928 Okeechobee hurricane, that killed 2500. However, since 1961, satellite imagery and progressively improved forecasting have provided early warnings of approaching storms, which has dramatically reduced loss of lives.

In 1969, Hurricane Camille took more than 250 lives, and most of the victims drowned in the storm surge, which crested at nearly 7.5 m. Camille also caused almost $1.5 billion in damage along the Gulf Coast. Hurricane Hugo, which hit South Carolina in 1988, and Hurricane Andrew, which hit South Florida in 1992, caused much greater damage than Camille but took few lives. Hurricane Andrew is estimated to have caused more than $30 billion in damage. Hurricane Katrina came ashore in Louisiana and Mississippi in 2005, causing more than $125 billion in damage and about 1500 deaths. In 2017, Hurricane Harvey hit Houston, Texas and killed at least 70, followed weeks later by Hurricane Irma that killed 134 as it traveled across Puerto Rico, Cuba, the Bahamas, and South Florida. This was the first year on record when two strong (category 4 on a scale of 1–5) hurricanes made landfall in the continental U.S., and the total damages are estimated to be as much as $200 billion. Hurricane Harvey (2017) caused more than 100 deaths and Hurricane Helene (2024) caused over 250 deaths in the U.S. and both caused significant damage in inland areas. In the case of Helene, the majority of fatalities were caused by excessive rainfall well inland in the mountainous region of western North Carolina. The cost of damages in this region are estimated at almost $80 billion.

Tropical cyclones form in all tropical oceans, but they are rare in the South Atlantic and eastern South Pacific (**Fig. 7-29**). Before March of 2004, only two tropical cyclones had ever been recorded in the South Atlantic Ocean, and neither had reached hurricane strength. However, in that month, Hurricane Catarina formed east of Brazil. The storm caused at least two deaths and substantial property damage when it came ashore on the southern coast of Brazil. Tropical cyclone damage and deaths are generally greater in the Islands of the western Pacific and mainland Asia than those in the U.S., as the frequency of these cyclones in the Pacific and Indian Oceans is higher than that in the Atlantic. The most vulnerable locations include the Philippines, which lie in the region with the highest frequency of these storms, and the densely populated low-lying areas of Bangladesh, India, Southern China, and mainland Southeast Asia, where evacuation of the large populations remains difficult and storm surge remains a major threat.

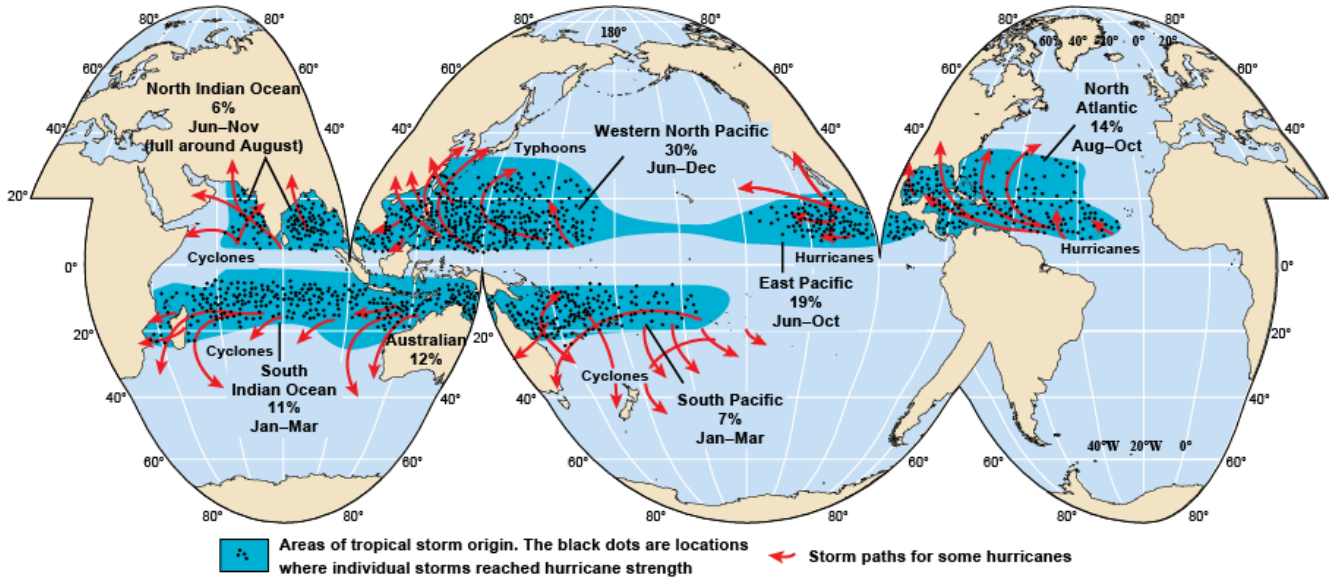


Figure 7-29. Tropical cyclones develop where the surface ocean water temperature is greater than 26°C, except in the equatorial region between 5°N and 5°S, where the Coriolis Effect is too weak to generate these storms. Surface waters are too cold for strong tropical storm formation in the eastern Atlantic Ocean and south of the equator in the eastern Pacific Ocean. Once formed, tropical cyclones travel generally east to west, curving toward higher latitudes as they are steered by the Coriolis Effect and by other weather systems. Tropical cyclones are most frequent in the western North Pacific Ocean, where they are called “typhoons” or “baguios.”

Tropical cyclones form only over surface waters with temperatures above about 26°C and do not generally form closer to the equator than 5°N or 5°S, because the Coriolis effect is too weak in this equatorial band.

Historical records suggest that the frequency and intensity of hurricanes may vary on a cycle lasting several decades. Climate models predict that hurricane frequency will not change significantly due to climate change in the next several decades, but they will become stronger on average, and the most powerful damaging storms will be more frequent and may maintain strength into progressively higher latitudes. For the U.S. and many other countries, this prediction is important. Coastal development has been intense over the past several decades, particularly in the southeastern U.S. and Gulf Coast communities within the hurricane band. In addition, sea level is rising, exposing more coastline to wave action. In the future, large coastal cities of the northeastern United States could be exposed to the full force of devastating hurricanes. Perhaps worse, future hurricanes might have much higher winds than the approximate maximum of 200 $km \bullet h^{-1}$ observed in the most intense recent hurricanes. Studies of **sediment** deposited by ancient storms have provided evidence of “super” cyclones in Earth’s past with wind speeds probably higher than any ever directly observed and greater than what most buildings can withstand.

Extratropical Cyclones

Powerful **cyclonic** storms form in the westerly wind regions in winter. Such storms form primarily at the polar front, where cold polar air moving toward the equator and the west and warmer mid-latitude air flowing east and toward the pole converge. Along the front between these two air masses, which are moving in opposite directions, waves periodically develop. The waves develop into cyclonic storms as the warm air flows over the top of the cold, denser air, and each moving air mass is deflected by the Coriolis Effect (**Fig. 7-30**).

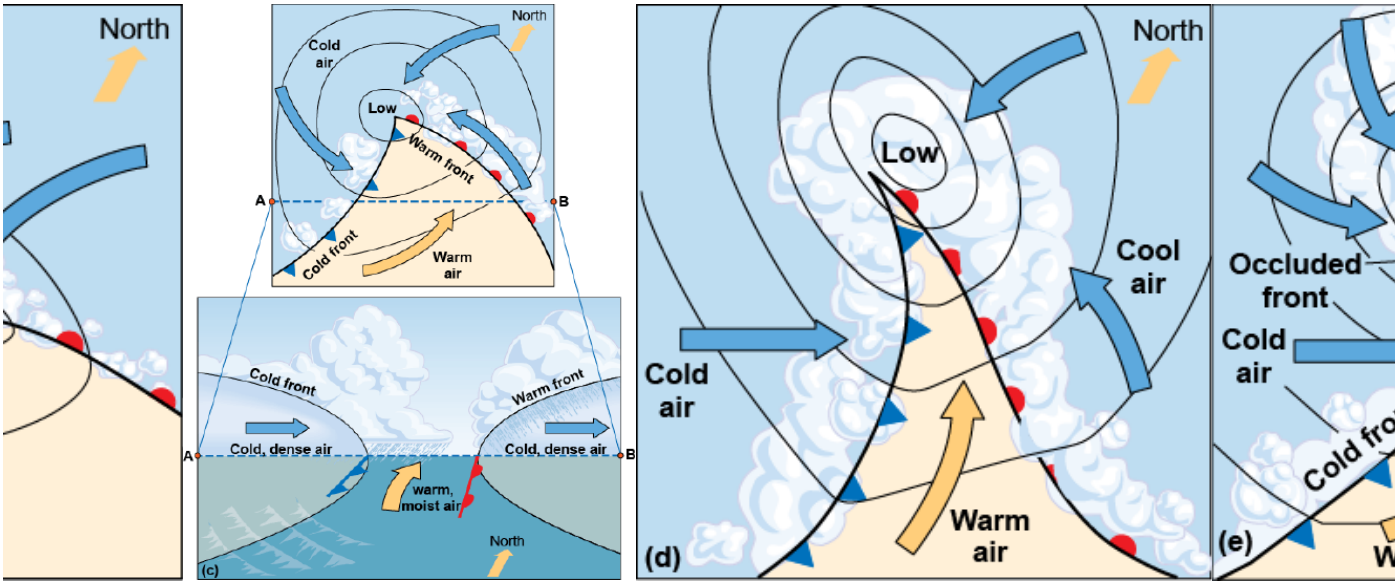


Figure 7-30. Formation of extratropical cyclones in the Northern Hemisphere. All the diagrams except (c) are surface-level map views. (a) The polar front separates cold air on the polar side of the front from warm air at lower latitudes. (b) Extratropical cyclones are formed when a wave develops on this front. (c) A low-pressure zone forms as warm, moist air flows north and rises, initiating counterclockwise rotation of the air masses around it (see Figure 7-26b). Once formed, the low-pressure zone may strengthen if the cold air mass flows over relatively warm ocean surface waters, is warmed, gains water vapor through evaporation, and also begins to rise. The vertical cross section shows that extensive areas of rain clouds form both where the advancing cold air mass forces warm, moist air to rise (cold front), and where warm, moist air is deflected to the right by the Coriolis Effect and rises over the cold air mass (warm front). (d) Eventually, the cold front catches up to the warm front as warm air continues to rise over the cold air mass. (e) This leads to the formation of an occluded front, which is the line where the warm and cold front meet. (f) The polar front reforms, and the extratropical cyclone slowly dissipates when warm air no longer flows into the central low-pressure zone, and atmospheric pressure rises. In the Southern Hemisphere, the rotation is in the opposite direction.

Storms that form at the polar front are called “extratropical cyclones.” They can be extremely large and carry huge amounts of moisture. They move eastward with the prevailing winds. Extratropical cyclones move more slowly than hurricanes and generally have somewhat lower, although still very high, wind speeds. They can be much larger than hurricanes and can cause storm surges. Consequently, they can cause severe and extensive wave **erosion** on exposed **shores**. The massive Pacific storms that bring winter rains, winds, and beach erosion to California and the Pacific Northwest are extratropical cyclones.

Extratropical cyclones can also intensify or deepen by interacting with a warm **western boundary current**. For example, the nor’easters that often strike the Mid-Atlantic coast of the United States in winter are extratropical cyclones that often undergo explosive intensification near the Gulf Stream after passing over the North American land mass.

This page titled 7.9: Weather Systems is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.10: Local Weather Effects

Interactions among ocean, atmosphere, and land are local in geographic scale. These interactions occur on or near the coast and sometimes produce large differences in weather and climate within a coastal region.

Land and Sea Breezes

All parts of the Earth's surface are subjected to a daily cycle of heating and cooling as night follows day. The ocean's surface water absorbs solar radiation during the day without a significant temperature change for three main reasons. First, water has a high heat capacity. Second, much of the solar energy penetrates beneath the water's surface before being absorbed, particularly in low-**turbidity** waters. Third, wind stirs the upper water column, mixing and distributing the heated water. Similarly, ocean surface water cools slowly at night. Consequently, even in locations where daytime solar radiation is intense, the temperature of ocean surface water varies by, at most, a few degrees between day and night. The temperature of the land surface, in contrast, can vary over a much larger range **diurnally**. Anyone who has spent time in the middle of a continent, particularly in the mountains, may have observed the rapid heating and cooling of the land. A rock that is too hot to touch in the middle of a sunny day will feel cold within a few hours after sunset that evening.

Heat is transferred continuously between the land and the atmosphere, and between the ocean and the atmosphere. As a result, and because of its low heat capacity, the air next to the Earth's surface tends to change temperature relatively quickly to match that of the land or ocean surface.

During daylight hours, coastal water and its overlying air mass warm only slowly, whereas the adjacent land and its overlying air mass warm much more rapidly. Where air temperature is higher over coastal land, the air over land becomes warmer and less dense, expands vertically, and thus has reduced surface pressure relative to the cooler marine air (Fig. 7-31a). Cooler air from over the ocean is drawn inland by the resulting horizontal atmospheric pressure gradient to replace the rising heated air. This process creates onshore winds called "sea breezes." The sea breeze blows until the solar radiation is reduced in intensity and the land has cooled to the same temperature as the coastal surface waters. Sea breezes gradually weaken near and after sunset as land cooling reduces the thermal contrast.

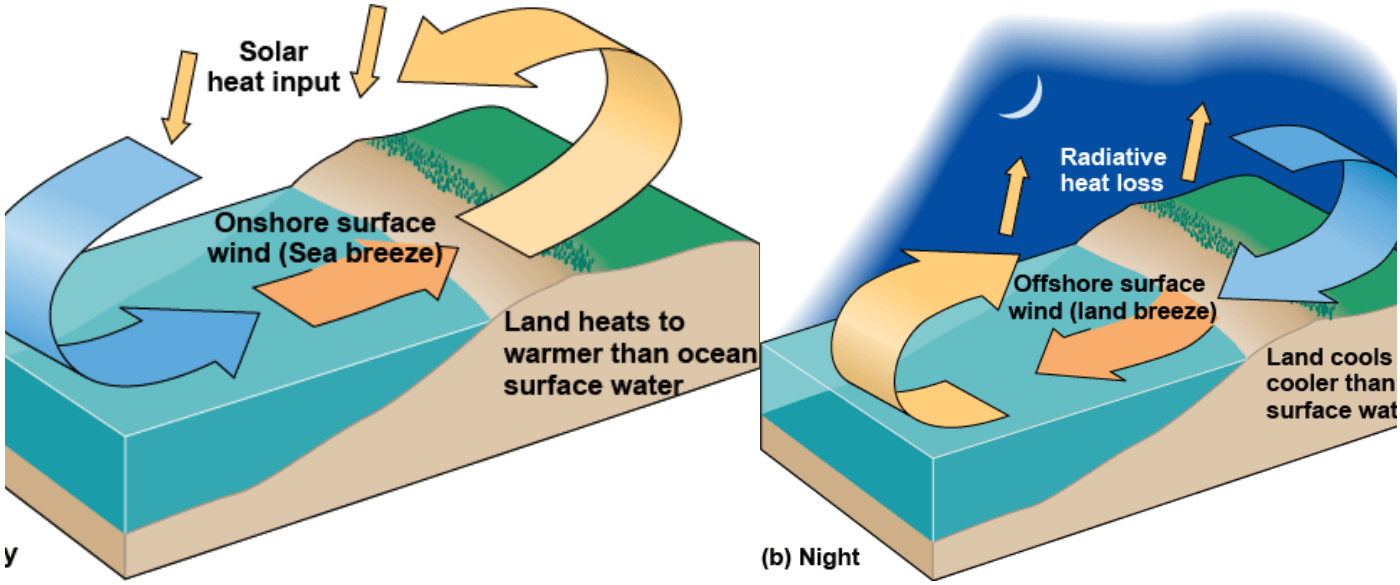


Figure 7-31. (a) Land heated by the sun during the day heats the ground-level air, which rises and is replaced by cooler air drawn in from the ocean. Thus, the surface wind blows from ocean to land and is called a "sea breeze." (b) At night, the land loses its heat quickly through radiation, but the ocean water remains at the same temperature. As the land surface cools to below the sea surface temperature, warmer air rises over the ocean and is replaced by cooler air drawn from over the land. Thus, the surface wind blows from land to ocean and is called a "land breeze."

As night progresses, the land and its overlying air mass cool rapidly until they are colder than the ocean surface water and its

overlying air mass (**Fig. 7-31b**). Air pressure increases over the land as the cooled, denser air mass sinks. This air mass now spreads toward the ocean, where atmospheric pressure is lower. Thus, a land breeze blows offshore during the latter part of the night and the early daylight hours (**Fig. 7-31b**).

Reversing local sea and land breezes are present on almost all coasts, but their intensity varies significantly with location, season, and larger-scale regional weather patterns. Probably for thousands of years, fishers in many countries have used sea breeze and land breeze patterns to their advantage, sailing offshore with the land breezes in the early morning hours and returning with the sea breezes late in the day. Many fishers still follow this pattern, even though they have motorized vessels.

Coastal Fog

The famous fogs of San Francisco and other coastal locations are often related to sea breezes. In many coastal areas, especially where coastal upwelling occurs, surface waters near the coast are colder than those tens of kilometers offshore. When sea breezes develop, warm, moist air over the offshore water is drawn into the sea breeze system. As this air flows shoreward, it passes over cooler coastal water. If it cools sufficiently, some of the water vapor it carries condenses into extremely small water droplets, which we call “fog” (**Fig. 7-32**). Coastal fogs of this type occur along the entire West Coast of the U.S. and along the eastern coast from Maine into Canada. Similar fogs occur on the west coasts of South America and Africa and in many other locations.

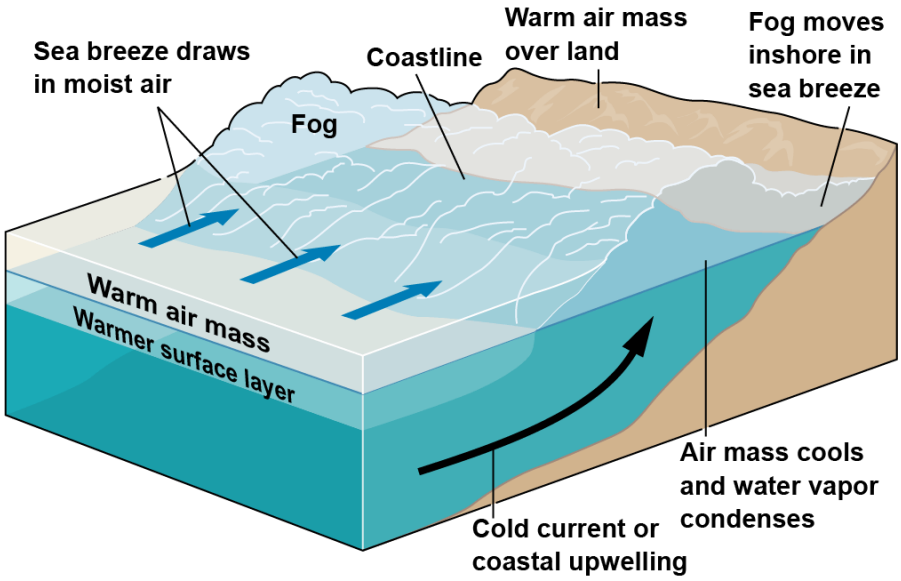


Figure 7-32. Coastal fog develops when moist air is drawn inland by a sea breeze or climatic winds and passes over a cold nearshore water mass. The cold nearshore water can be the result of cold-water currents or, more often, of coastal upwelling that brings cold, deep water to the surface.

The Island and Mountain Effect

When air moving over the Earth’s surface encounters a mountain chain across its path, it is forced to rise up the mountain slope to pass the barrier. As the air mass rises, it cools by adiabatic expansion because the atmospheric pressure decreases with altitude. If the air is moist before it begins to rise and it rises high enough, water vapor will condense to form clouds, which eventually form rain or snow (**Fig. 7-33a**). Once the air has crossed the mountain ridge, it flows down the mountain slope. As it descends and pressure increases, the air warms, and the water vapor saturation pressure increases. Also, the air’s actual water vapor content remains low due to precipitation loss on the windward side. The leeward side of the mountains therefore, receives very little rain.

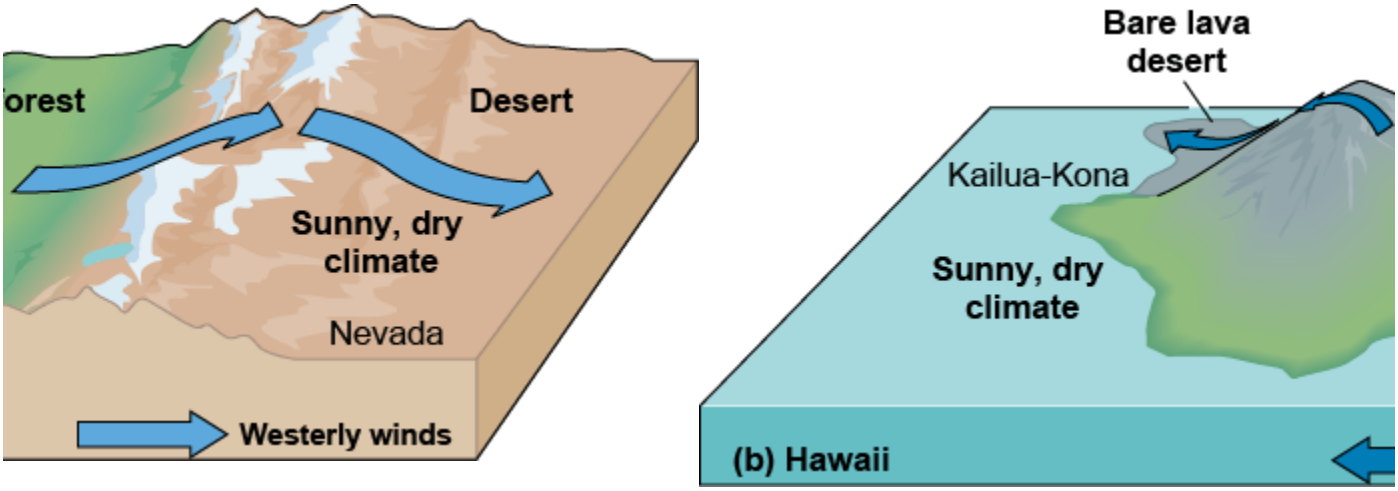


Figure 7-33. The mountain or island effect occurs where a moving mass of moist air, usually from over the ocean, encounters a mountain or mountainous island, rises, cools adiabatically, and loses water as rain or snow. Once over the mountains, the air mass descends, warms, and has a greater capacity for moisture. As a result, rain and snowfall stop. (a) The Sierra Nevada mountains of California intercept moist air moving inland from the ocean. Rain and snowfall are abundant on the California side of the mountains, but the Nevada side is a desert. (b) On the Hilo side of Hawaii, trade winds blowing from the northeast produce clouds, plentiful rainfall, and tropical rainforests, but the opposite side of the island, near Kailua-Kona, is sunny and has low rainfall.

In many coastal locations, the wind blows consistently from the ocean onto land. If a coastal or near-coastal mountain chain of sufficient height is present, the coastal side of the mountain range will have high rainfall, and the landward side will be desert. The best examples are in parts of South America, where the Andes Mountains divide the fertile coastal region of Chile from the desert or near-desert interior of Argentina, and in North America, where the west side of the Sierra Nevada chain is fertile and the east side is desert. Note that the coastal mountains of northern California are too low and have too many gaps to dry the air masses completely as they move off the Pacific toward the Sierra Nevada.

Skiers in the Sierras can use the mountain effect to decide where snow conditions are best. Resorts on the California side usually have more snow, particularly in poor snow years. However, the snow is generally wet and heavy. The resorts on the Nevada side, just over the ridge **crest**, generally get less snow. However, snow that does accumulate falls through dry, undersaturated air and often has a dry, powdery character that is desirable for skiing.

The mountain effect also operates on mountainous islands, where it is called the “island effect.” The island effect is extremely important to the climates of many islands, particularly those in the trade wind zone, where winds blow reliably from one direction almost all year.

The island of Hawaii is a particularly good example. Hawaii is composed of several volcanoes, some of which are over 4000 m high, and it is located in the northeast trade wind zone (**Fig. 7-33b**). Hilo, on the **windward** northeast coast, has rain almost every day. It is within a coastal belt of tropical rainforest, with many continuously flowing streams and waterfalls. Kailua-Kona, on the leeward southwest side of the island, is hot, sunny, and arid, with only occasional showers throughout most of the year. It is surrounded by almost bare fields of solidified **lava**, on which little can grow because of the low rainfall.

Other islands in the Hawaiian chain, and many other tropical Pacific islands, have windward wet climates and leeward dry climates. However, each island is somewhat different. On Oahu, for example, the mountains are relatively low and cut by passes through which clouds carry afternoon showers to Honolulu before they can dissipate. On other islands, the mountains are too low to produce the island effect. Kahoolawe, like most other low islands in the trade wind zone, is extremely arid and gets rainfall only from infrequent storms.

This page titled 7.10: Local Weather Effects is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.11: CHAPTER SUMMARY

Atmosphere and Water Vapor.

Air moves vertically when a change in temperature or in its concentration of water vapor causes its density to change. In the lower atmosphere, vertical movements are generally limited to the troposphere (about 12 km altitude). The addition of water vapor to air decreases its density because lighter water vapor molecules displace heavier nitrogen and oxygen molecules. Atmospheric convection is caused by heating or evaporation at the sea or land surface. Water vapor continuously evaporates from the oceans into the atmosphere. As air rises, its pressure decreases, causing it to expand and cool. The saturation pressure of water vapor in air decreases with decreasing temperature. Hence, as air cools, it becomes supersaturated with water, which condenses to rain or ice crystals.

Water and Heat Budgets.

Enough water is evaporated from land and oceans each year to cover the world about 1 m deep. About 93% of this water comes from the oceans, but nearly 30% of the resulting precipitation falls on land. The excess of precipitation over evaporation on land enters lakes, streams, and rivers and returns to the oceans as runoff.

Almost 25% of the solar radiation reaching the Earth is absorbed and converted to heat in the atmosphere. About 50% is absorbed by oceans and land, and the rest is reflected to space. Of the heat absorbed by the oceans, about half is lost by radiation, and half is transferred to the atmosphere as latent heat of vaporization. Solar energy per unit area received by the Earth is at a maximum at the equator and decreases toward the poles. However, the amount of heat radiated and reflected per unit area varies significantly with latitude and other factors. At the equator, more heat energy is received than is lost to space, whereas at the poles, more heat is lost than is received. Heat is transferred from the tropics to the polar regions by atmospheric and oceanic currents.

Climatic Winds.

Ocean-to-atmosphere heat transfer, particularly through sensible and latent heat fluxes, plays a key role in fueling atmospheric convection, especially over tropical oceans. The atmospheric three-cell per hemisphere system consists of Hadley, Ferrel, and polar cells arranged between the equator and the pole in each hemisphere. This structure is modified, especially over the continents, as a result of land-ocean-atmosphere interactions driven by the far greater heat capacity of ocean waters compared to the land surface. Trade winds in the Hadley cell blow westward and toward the equator, and westerly winds in the Ferrel cell blow eastward and toward the pole. Between cells, winds are generally calm. Rainfall and clouds are heavy at upwelling regions, and winds are light and skies are clear at downwelling regions. The cells shift north and south seasonally as the Earth's angle to the sun changes. Because solar heat is released to the atmosphere relatively slowly by evaporation, the location of the atmospheric cells lags behind the location of greatest solar heating as the cells migrate seasonally.

Climate and Ocean Surface Water Properties.

Ocean surface water temperatures generally decrease with latitude, but currents and upwelling distort the pattern. Surface water salinity is primarily determined by differences between evaporation and precipitation rates, except near continents, where freshwater runoff is high. Salinity is highest in the subtropics and polar regions, where rainfall is low. Salinity is low at mid latitudes, where rainfall is high and evaporation is less than at lower latitudes, and at the equator, where evaporation is reduced by persistent cloud cover and lack of winds. Extreme high salinity occurs in marginal seas where evaporation exceeds precipitation, and extreme low salinity occurs in marginal seas where precipitation exceeds evaporation.

Interannual Climate Variations.

El Niño/Southern Oscillation is a complex, coupled ocean-atmosphere phenomenon that occurs across the equatorial region of the Pacific Ocean. Warm surface water is transported westward by the trade winds until it accumulates near Indonesia. When the winds slow, the water flows back to the east along the equator, where there is no Coriolis Effect. During El Niño, upwelling is

stopped near Peru, with often disastrous effects on marine life. Droughts occur in locations as far away as Central Europe, and severe storms occur in California and other places. When an El Niño ends, the system may overshoot to produce La Niña. However, the strength and severity of La Niña events vary, and in some regions La Niña can have impacts that are equally, or even more, pronounced than those of El Niño.

In addition to ENSO, interannual oscillations of ocean-atmosphere characteristics have been identified in the North Pacific, North Atlantic, Arctic, and Indian Oceans. Each of these oscillations affects regional climates, but the effects are generally smaller and less widespread than those due to ENSO. Studies of the Pacific Decadal Oscillation suggest that it, and probably other such oscillations, and their ecosystem effects are chaotic and may cause unpredictable irreversible ecosystem change.

Global Climate Zones.

Climate zones in the oceans are generally arranged in latitudinal bands. Land climate zones are more complex and depend on proximity to the ocean and the locations of mountain ranges. The ocean moderates coastal climates because surface waters do not change much in temperature either during the day or during the year. The moderating influence of the oceans can extend far into the continents in regions where the winds are generally onshore and no mountain chains block the passage of the coastal air inland.

Weather Systems.

In the free atmosphere, above the influence of surface friction, winds around high- and low-pressure systems often flow approximately geostrophically, or nearly parallel to isobars or height contours due to a balance between the pressure gradient force and the Coriolis force. Near the surface, however, friction disrupts this balance, causing winds to cross isobars toward lower pressure. Winds flow counterclockwise around a low and clockwise around a high in the Northern Hemisphere, and vice versa in the Southern Hemisphere. Tropical cyclones form over warm water in latitudes high enough that the Coriolis effect is significant. Tropical cyclone winds blow in toward the eye in a rising helical pattern. Winds accelerate as heat energy is added by evaporation from the ocean surface. but the hurricane loses strength when it moves over land or cool water, where its energy supply is removed. Extratropical cyclones form at the polar front as warm air flows over cold polar air and is deflected by the Coriolis effect.

Local Weather Effects.

The Earth's surface heats by day from solar radiation and cools at night as the heat is lost. Primarily because of water's high latent heat, the ocean surface water temperature varies little in this daily cycle. Thus, land temperatures vary more. During the day, land next to an ocean heats, then loses some of its heat to the air. The warmed air rises and is replaced by cooler air from over the ocean, creating a sea breeze. At night, the land cools rapidly, which creates a land breeze that flows seaward to displace warmer, less dense air over the ocean. Coastal fogs form when warm, moist air from over the ocean passes over a cold coastal water mass as it enters the sea breeze system.

When moist air encounters mountains, it is forced to rise, expand, and cool. If it rises high enough, water vapor condenses or deposits onto nuclei and causes precipitation. After crossing the mountains, the air mass descends, is compressed, and warms. When it is no longer saturated with water vapor, precipitation ceases. Hence, the windward side of the mountains is wet, and the leeward side is arid.

7.11.1: Key Terms

7.11.2: Chapter Questions

7.11.3: References

This page titled 7.11: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.11.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- adiabatic expansion
- angle of incidence
- anthropogenic
- beach
- chaotic
- chlorofluorocarbons (CFCs)
- chlorophyll
- climate
- conduction
- convection
- convection cells
- convergence
- Coriolis effect
- crest
- cum sole
- currents
- cyclonic
- density
- diffusion
- diurnally
- divergence
- downwelling
- El Niño
- ENSO
- environment
- equinoxes
- erosion
- extratropical cyclones
- fetch
- food web
- friction
- fronts
- geostrophic
- heat capacity
- hurricanes
- ice exclusion
- intertropical convergence zone
- isotherms
- jet streams
- latent heat of vaporization
- latitude
- lava
- marginal seas
- marine mammals
- meanders
- nutrients
- overfished
- ozone layer
- phytoplankton

- pressure gradient
- pycnocline
- residence time
- salinity
- saturation pressure
- schools
- sediment
- sensible heat
- shores
- solstice
- spatial
- species
- storm surge
- subtropical high
- trade winds
- tropical cyclones
- troposphere
- turbidity
- water masses
- wave height
- weather
- westerlies
- windward

This page titled 7.11.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.11.2: Chapter Questions

Study Questions

1. Why do the CFCs released into the lower atmosphere take a very long time to reach the Earth's ozone layer?
2. Why is the amount of water vapor in air so important?
3. What are the reasons for the net heat loss to space from polar regions of the Earth and the net heat gain from the sun in the tropics?
4. Why are there three atmospheric convection cells in the Northern and Southern Hemisphere? Discuss why the cells are not centered on the equator and why they are more complicated in the Northern Hemisphere.
5. Why are persistent cloud cover and rainfall present at atmospheric convergences over the oceans?
6. How do the locations of the world's deserts relate to the locations of the atmospheric convection cells?
7. Given that the summer solstice is in June, why is August usually the hottest month in most of the United States?
8. The Hawaiian Islands lie in the northeast trade wind zone. Why do the northeast sides of the islands have wet climates and the southwest sides dry climates?
9. Antarctica is a desert, despite its ice cap, which is several kilometers thick. Why is it a desert?
10. Why are the cloud patterns that we see in weather satellite images dominated by swirls?
11. Why don't hurricanes form at the equator? Why don't they form over land?

Critical Thinking Questions

1. What causes air masses to move vertically in the atmosphere? Why is the vertical circulation of the lower atmosphere restricted to a layer of only about 12 km, although the entire atmosphere is more than 45 km deep?
2. Would the height of the restricted vertical circulation in the atmosphere be changed if the Earth's average surface atmospheric temperature increased? Why or why not?
3. On a nonrotating Earth, there would be one atmospheric convection cell in each hemisphere, with upwelling at the equator and downwelling at the poles. On the Earth today, we have three cells in each hemisphere because of the Coriolis deflection. Since the Coriolis deflection will be reduced as the Earth's rotation is progressively slowed down, at a lower speed of rotation would you expect a system with two convection cells in each hemisphere to be established eventually? Why or why not?
4. It is known that in Biblical times, parts of the areas that are now desert in Eurasia (Egypt, Israel, Syria, and so on) were fertile regions with plentiful rainfall. What does this tell us about the location of the atmospheric convection cells?
5. If the Earth's direction of rotation on its axis were somehow reversed, would the location of the world's deserts change? Why or why not? If so, which parts of each of the continents would become deserts, and why?
6. If the trade winds were strengthened by global warming, how might the frequency and strength of El Niños be affected? Explain the reasons for your answer.
7. If there were no mountain chains in the western part of the United States, how would the climate in Nevada and the plains states be different? Explain the reasons for your answer.
8. No hurricanes are fully formed on the North African side of the Atlantic Ocean, but many hurricanes are formed in the equivalent region off the Pacific coast of Mexico. Why?
9. Almost no hurricanes form in the South Atlantic Ocean. Why?

This page titled 7.11.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

7.11.3: References

Figure 7-1 – NASA, 2007: Blue Marble – Image of the Earth from Apollo 17. Accessed 21 November 2025, Source.

Figure 7-1b – Crawford, W., 2011: Funnel cloud off the port side bow. IODP/TAMU, accessed 21 November 2025, Source.

Figure 7-3 – NASA Ozone Watch, 2025: October ozone maps of the Southern Hemisphere. Accessed 21 November 2025, Source.

Figure 7-5 – Adapted from Abbott, B. W., and Coauthors, 2019: Human domination of the global water cycle absent from depictions and perceptions. *Nat. Geosci.,* **12**, 533–540, Source.

Figure 7-6 – Adapted from Simmon, R., 2009: Climate and Earth's Energy Budget. NASA Earth Observatory, Accessed 21 November 2025, Source.

Figure 7-12 – Data from NCEP-NCAR Reanalysis, 2025: Monthly mean surface pressure. Accessed 21 November 2025, Source.

Figure 7-14 – Adapted from Sverdrup, H. U., M. W. Johnson and R. H. Fleming, 1942: *The Oceans, their physics, chemistry, and general biology*. Prentice Hall, 1079 pp, Source.

Figure 7-15 – Adapted from Sverdrup, H. U., M. W. Johnson and R. H. Fleming, 1942: *The Oceans, their physics, chemistry, and general biology*. Prentice Hall, 1079 pp, Source.

Figure 7-16 – Data from Sverdrup, H. U., M. W. Johnson and R. H. Fleming, 1942: *The Oceans, their physics, chemistry, and general biology*. Prentice Hall, 1079 pp, Source.

Figure 7-17 – Adapted from NASA IMERG, 2023: Grand Average Precipitation Climatology. Accessed 21 November 2025, Source.

Figure 7-18 – Adapted from NOAA Coral Reef Watch, 2025: Mean Monthly Pacific Ocean Sea Surface Temperature. Accessed 21 November 2025, Source.

Figure 7-21 – Data from NOAA Climate Prediction Center, 2025: Cold and warm episodes by season. Accessed 26 June 2025, Source.

Figure 7-22 – Adapted from Hare, S. R. and N. J. Mantau, 2000: Empirical evidence for North Pacific regime shifts in 1977 and 1989. *Prog. Oceanogr.* **47**, 103–145, Source.

Figure 7-23 – Adapted from D.V. Bogdanov, D. V., 1963: Map of the natural zones of the ocean: Okeanologiya 1961. *Deep-Sea Res. Oceanogr. Abstr.*, **10**, 520–523, Source.

– Adapted from Olson, D.M. and Coauthors, 2001: Terrestrial Ecoregions of the World: A New Map of Life on Earth: A new global map of terrestrial ecoregions provides an innovative tool for conserving biodiversity. *BioScience,* **51**, 933–938, Source.

Figure 7-25 – Data from NWS, 2025: Monthly normal average air temperature for Portland, Oregon; Sheridan, Wyoming; and Boston, Massachusetts. Accessed 21 November 2025, Source.

Figure 7-28 – Science Education Resource Center, 2006: Oblique satellite view of Hurricane Katrina. Carleton College, accessed 21 November 2025, Source.

Figure 7-29 – Adapted from NHC, 2024: Tropical Cyclone Climatology. Accessed 21 November 2025, Source.

Figure 7-30 – Adapted from Scott, R. C., 1996: *Introduction to physical geography*, West Publishing Company, 560 pp.

Doocy, S., A. Dick, A. Daniels, and T. D. Kirsch, 2012: The human impact of tropical cyclones: a historical review of events 1980-2009 and systematic literature review. *PLoS Curr.*, **16**, Source.

Ebbesmeyer, C. C., Cayan, D.R. McLain, F.H. Nichols, D.H. Peterson and K.T. Redmond, 1991: 1976 step in the Pacific climate:

Forty environmental changes between 1968-75 and 1977-1984. *Proc. 7th Ann. Pacific Climate Workshop*, California Dept of Water Resources, Interagency Ecol. Stud. Prog. Report 26.

This page titled 7.11.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8: OCEAN CIRCULATION

Critical Concepts

- **CC1 Density and Layering in Fluids**: Water in the oceans is arranged in layers according to the water density. Many movements of water masses in the oceans, especially the movements of deep water masses, are driven by differences in water density.
- **CC3 Convection and Convection Cells**: Fluids, including ocean water, that are cooled from above, sink when their density is increased. This establishes convection processes that are a primary cause of vertical movements and the mixing of ocean waters. These processes are also important in transporting and distributing heat and carbon dioxide between the atmosphere and oceans and between regions of the globe.
- **CC6 Salinity, Temperature, Pressure, and Water Density**: Sea water density is controlled by temperature, salinity, and, to a lesser, extent pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Movements of water below the ocean surface layer are driven primarily by density differences.
- **CC7 Radioactivity and Age Dating**: Some elements have naturally occurring radioactive (parent) isotopes that decay at precisely known rates to become a different (daughter) isotope. Radioactive isotopes, especially those that were released during the period of atmospheric testing of nuclear weapons, are useful as tracers that can reveal the movements of water masses in the ocean, especially the rates at which deep water masses are formed.
- **CC9 The Global Greenhouse Effect**: The oceans and atmosphere are both important in studies of the greenhouse effect, as heat and carbon dioxide and other greenhouse gases are exchanged between the atmosphere and oceans at the sea surface. The oceans store large amounts of heat and carbon dioxide both in solution and in carbonates.
- **CC10 Modeling**: The complex interactions between the oceans and atmosphere can best be studied by using mathematical models. The motions of water masses within the body of the oceans, especially motions below the surface layer, are also studied extensively using mathematical models because they are extremely difficult to observe directly.
- **CC11 Chaos**: The nonlinear nature of ocean-atmosphere interactions makes at least parts of this system behave in sometimes unpredictable ways. It also makes it possible for changes in ocean circulation to occur in rapid, unpredictable jumps between one set of conditions and a different set of conditions, and these changes can affect climate.
- **CC12 The Coriolis Effect**: Water and air masses move freely over the Earth and ocean surface while objects on the Earth's surface, including the solid Earth itself, are constrained to move with the Earth in its rotation. This causes moving water or air masses to appear to follow curving paths across the Earth's surface. The apparent deflection, called the Coriolis effect, is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The deflection is at a maximum at the poles, is reduced at lower latitudes, and becomes zero at the equator.
- **CC13 Geostrophic Flow**: Air and water masses flowing on horizontal pressure gradients are deflected by the Coriolis effect until they flow across the gradient such that the pressure gradient force and Coriolis effect are balanced, a condition called geostrophic flow. This causes ocean currents and winds to flow around high and low pressure regions (regions of elevated or depressed sea surface height) in near circular paths. The circular gyres that dominate the global circulation of ocean waters are the result of water masses flowing geostrophically.

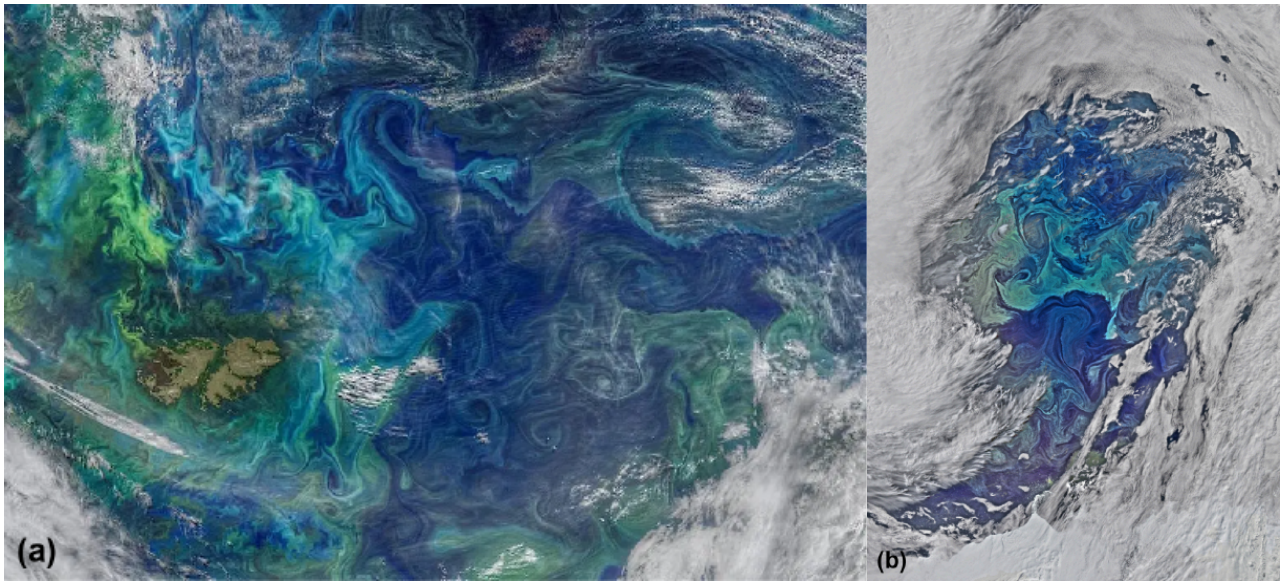


Figure 8-1. These images are scanned from a satellite with a visible/infrared radiometer that senses changes in radiated energy at multiple wavelengths. The data is converted to a false color image that shows elevated chlorophyll concentrations (a proxy for phytoplankton concentration) as the lighter colored blue-green swirls. The radiometers can not see through clouds so most of the bright white parts of the images are cloud cover. The complex swirls and patterns revealed in all these images are caused by eddies and other surface water motions, some of which come and go within a few hours. The eddy motions are too small and too short lived to be observed by oceanographers using ship-based measurements, so their existence was not known until observations were made from space. Eddies exist in all parts of the oceans. (a) South Atlantic Ocean near the Falkland Islands (visible on the left of the image). (b) The Southern Ocean just to the west of Drake Passage (the passage between the tip of South America and the tip of the Antarctic Peninsula).

Throughout the oceans, from surface to seafloor, water is in continuous motion, ranging in scale from movements of individual molecules to oceanwide movements of water by tidal forces and winds. Oceanographers generally group ocean-water movements into three categories: waves, **tides**, and ocean circulation. Waves (**Chap. 9**) and tides (**Chap. 10**) are primarily short-term (lasting from seconds to hours) oscillating motions that move water in orbital paths around a single location.

Seawater properties such as **salinity**, temperature, and chemical concentrations vary with location and depth. However, large volumes of water have nearly uniform properties in many areas and these are called water masses. Water masses are transported across the oceans by **currents**, and vertically up or down by **convection** (**CC3**) or by interaction with seafloor topography. As this circulation occurs, water masses mix with other water masses at their interface (usually a horizontal interface between layers; **CC1**). Mixing is the exchange of water between adjacent water masses, which results in a new water mass with a proportional combination of the properties of the individual water masses before mixing. Currents and mixing are the major processes that transport and distribute heat, dissolved chemicals, **suspended sediments**, and planktonic bacteria, algae, and animals in the oceans.

Chapter 7 describes how ocean surface currents can affect local climates. This chapter examines the physical processes that control present-day ocean circulation, reviews the major features of circulation, and describes some aspects of the relationship between ocean circulation and global climate, both past and present. Subsequent chapters describe how ocean circulation controls and/or affects the distribution and abundance of life in the oceans. Although waves and tides have relatively little influence on large-scale ocean circulation, they contribute substantially to local currents, particularly in coastal waters and **estuaries**, and to vertical mixing, particularly in the upper several hundred meters of the water column (**Chaps. 9, 10**).

8.1: Energy Sources
8.2: Wind-Driven Currents
8.3: Ekman Motion
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents

8.7: High-Latitude Surface Currents
8.8: Upwelling and Downwelling
8.9: Coastal Currents
8.10: Eddies
8.11: Inertial Currents
8.12: Langmuir Circulation
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
8.16: Chapter Summary
8.16.1: Key Terms
8.16.2: Chapter Questions
8.16.3: References
8.17: Box 1- Westward Intensification
8.18: Box 2 - Equatorial Currents

This page titled 8: Ocean Circulation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.1: Energy Sources

Winds are the primary energy source for currents that flow horizontally in the ocean surface layers (less than 100 to 200 m deep). Hence, surface currents are often called “wind-driven currents” or “wind drift currents.” Currents that flow deep in the oceans below the level affected by winds are generated primarily by convection caused by variations in water **density** (**Chap. 5**, **CC3**). Higher-density waters sink and displace (push aside or up) less-dense deeper water. Because density is determined by temperature and salinity (**CC6**), deep-water circulation is called **thermohaline circulation**.

The primary energy source for both wind-driven circulation and thermohaline circulation is the sun. Winds are generated by density variations in the Earth’s atmosphere caused by solar heating and radiative and evaporative cooling (**Chap. 7**). Thermohaline circulation is driven by density differences between ocean water masses caused by temperature and salinity variations. Variations in temperature and salinity are controlled, in turn, primarily by balances between solar heating and radiative cooling, and between solar evaporation and precipitation, respectively (**Chap. 7**).

This page titled 8.1: Energy Sources is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.2: Wind-Driven Currents

On a windy day at the **shore**, we can easily see that winds cause ocean water motions such as waves. Winds also create currents that can transport large volumes of water across the oceans.

Generation of Currents

When winds blow across the ocean surface, energy is transferred from wind to surface water as a result of the **friction** between the wind and the surface. The energy transferred to the ocean surface sets the surface layer of water in motion, generating both waves and currents. The process of energy transfer from winds to waves and currents is complex and depends on many factors, including wind speed, **surface tension** of the water, and roughness of the surface (that is, whether waves are already present, their height, and whether they are breaking). Therefore, the percentage of wind energy that is converted into **kinetic energy** of ocean currents is highly variable. With a steady wind, the speed of the surface current is generally between 1% and 3% of the wind speed, so a wind of 60 km•h^{-1} will generate a surface current of about 1 to 2 km•h^{-1}.

Surface water set in motion by the wind flows horizontally across the water below it. Because of internal friction between the surface water and the water below, wind energy is transferred downward. If we consider the water column to consist of a series of very thin horizontal layers, we can envision the moving surface layer transferring some of its kinetic energy to the next lower layer by friction, setting that layer in motion. The second layer, in turn, transfers some of its resulting kinetic energy to the layer below it, setting that layer in motion (Fig. 8-2a). However, only a fraction of the kinetic energy of each moving layer is transferred to the layer below it. Consequently, the speed of a wind-driven current decreases progressively with depth. Wind-driven currents are restricted to the upper 100–200 m of the oceans and generally to even shallower depths.

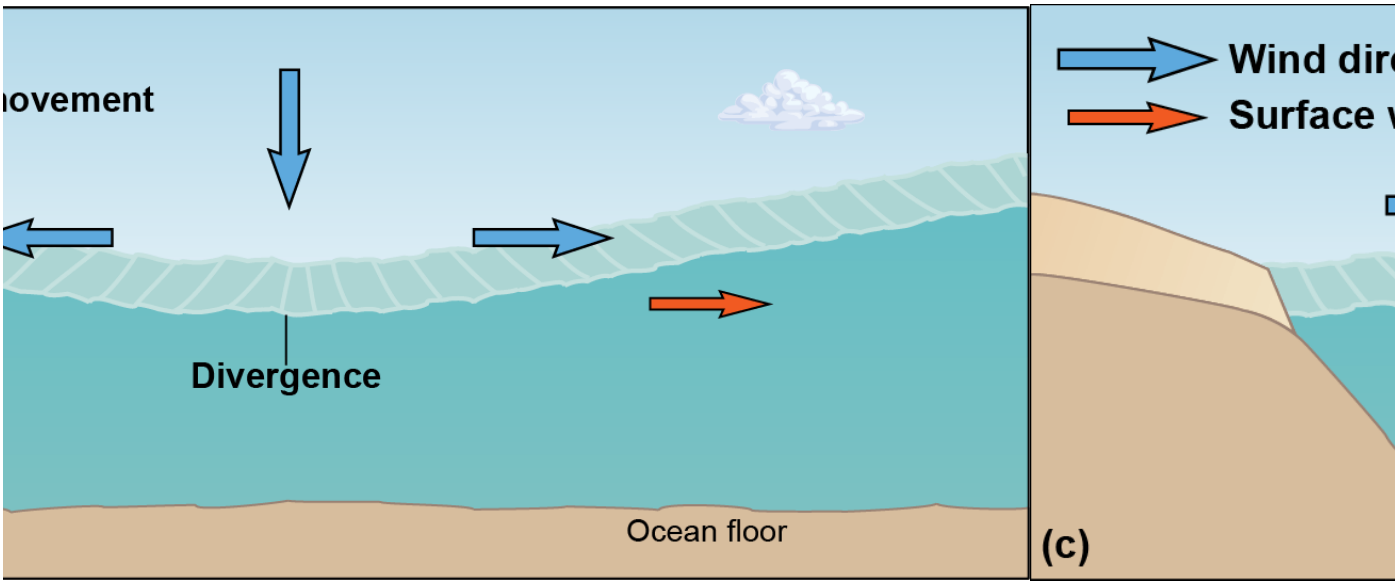


Figure 8-2. (a) Wind energy is converted to water movements, called currents, by friction between the moving air and the water surface. The resulting kinetic energy of the water at the surface is transferred vertically downward into the water column by friction between the water molecules. Friction occurs when the molecules of one layer are moved, as shown in the enlarged diagram of three adjacent molecules. As the upper-layer molecules are moved, the distance between them and the molecules immediately below changes. At point A, the distance between molecules is slightly decreased. With a decrease in distance the repulsive force between the molecules' electron clouds increases faster than the attractive gravitational force between the molecules (partly because the electron clouds are closer to each other than the two centers of mass are). Thus, the molecule below is "pushed" forward and downward. Conversely, the distance between molecules at the point labeled B is slightly increased. Since the repulsive force between the electron clouds is reduced by more than the gravitational force is reduced, the molecule below is "pulled" forward and upward. This process transfers kinetic energy downward through successive layers of molecules. (b) Sloped sea surfaces can be created in the open ocean when winds transport the surface water layer either away from a divergence, as shown here, or toward a convergence that then becomes an elevated area of sea surface. (c) Sloped sea surfaces can also be created at a coastline when winds transport the surface water layer offshore or onshore. The sea surface slopes are greatly exaggerated in these diagrams.

Restoring Forces and Steering Forces

Once current motion has been started, it will continue for some time after the wind stops blowing because the water has momentum. It is like a bicycle that continues to roll forward after the rider stops pedaling. Just as a bicycle slows and eventually stops because of friction in the wheel bearings and between tires and road, ocean currents slow and would eventually stop if winds stopped blowing and did not restart. The energy of a current is dissipated by friction between water layers flowing over each other. However, the frictional force between moving layers of water is small. This is one reason why the currents created by winds can flow for long periods after the winds stop and can also flow into and through regions with little wind, such as the Doldrums (**Figs. 7-10**, **7-11**).

There is a second reason for the continued flow of wind-created currents after the winds stop. Transport of surface water by wind-driven currents while winds are blowing can cause the sea surface to be sloped (**Fig. 8-2b,c**). Sea surface slopes created by wind-driven transport are extremely small, no more than a few centimeters of height difference across hundreds or thousands of kilometers of ocean surface. However, when the surface is sloped, a horizontal **pressure gradient** is formed (**CC13**), aligned in the same direction as the sea surface slope. The water flows from high pressure toward low pressure and tends to restore the ocean surface to a flat horizontal plane.

Once set in motion, the speed and direction of any current are modified by friction and three other factors. First, any body in motion on the Earth, including water moving in currents, is subject to deflection by the **Coriolis effect** (**CC12**). Second, the presence of coasts can block current flow and cause the water to mound up or to be deflected. Third, current speed and direction are affected by the presence of horizontal pressure gradients.

The following sections examine how the interactions between climatic winds and the restoring and steering forces account for the location and characteristics of surface currents in the oceans (**Fig. 8-3**).

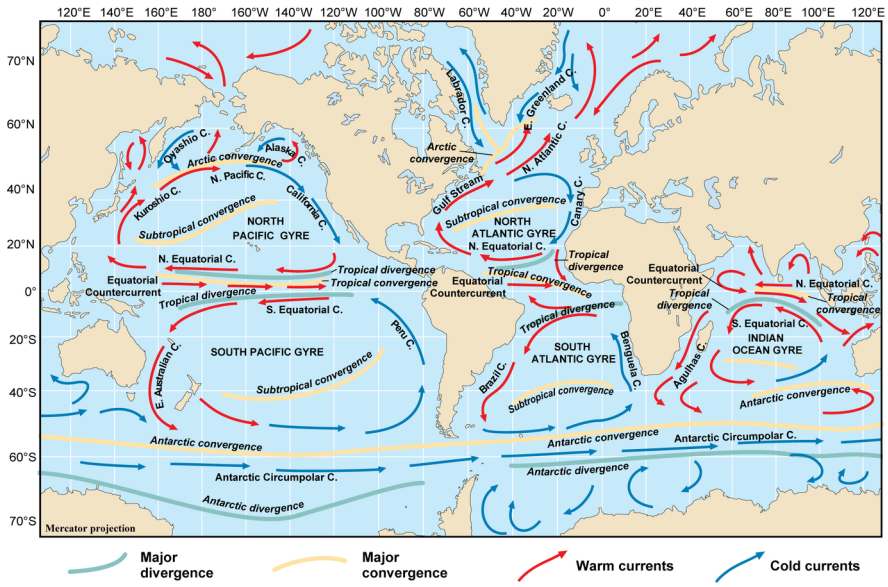


Figure 8-3. The wind-driven surface currents of the oceans and the locations of the major convergences and divergences are related. The surface currents form subtropical gyres both north and south of the equator in each ocean basin, and there are subtropical convergences within these gyres. Convergences are also found at higher latitudes between the subtropical gyres and the Antarctic Circumpolar Current, and between the subtropical gyres and the complicated high-latitude gyres in the Northern Hemisphere. Divergences lie on either side of the equator between the subtropical gyres and surrounding Antarctica.

This page titled 8.2: Wind-Driven Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.3: Ekman Motion

During his Arctic expedition in the 1890s, Fridtjof Nansen noticed that drifting ice moved in a direction about 20° to 40° to the right of the wind direction. The deflection occurs because water set in motion by the wind is subject to the Coriolis effect (**CC12**). The surface current that carries drifting ice does not flow in the direction of the wind but is deflected ***cum sole*** ("with the sun"), in the same way we see the sun move across the sky—to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (**CC12**). Remember the term *cum sole*. It always means "to the right in the Northern Hemisphere and to the left in the Southern Hemisphere."

The Ekman Spiral

As surface water is set in motion by the wind, it successively sets in motion a series of thin layers of water (**Fig. 8-2a**), transferring energy down into the water column. The surface layer is driven by the wind and deflected by the Coriolis effect. Each successively lower layer is driven by the layer above, with speed decreasing in each lower layer, and each layer is also deflected by the Coriolis effect. When it is set in motion by friction with the layer above, each layer's direction of motion is deflected by the Coriolis effect *cum sole* to the direction of the overlying layer. This deflection establishes the Ekman spiral (**Fig. 8-4**), named for the physicist who developed the mathematical relationships to explain Nansen's observations that floating ice does not follow the wind direction.

Surface water set in motion by the wind is deflected at an angle *cum sole* to the wind (**Fig. 8-4**). Ekman's theory predicts this angle to be 45° *cum sole* to the wind direction under ideal conditions, with constant winds and a water column of uniform density. Under normal conditions, the deflection is usually less. As depth increases, current speed is reduced, and the deflection increases. The Ekman spiral can extend to depths of 100 to 200 m, below which the energy transferred downward from layer to layer becomes insufficient to set the water in motion. This depth is a little greater than the depth where the flow is opposite to the surface current direction. This depth is called the "depth of frictional influence." At this depth, the current speed is about 4% of the surface current speed. The water column above the depth of frictional influence is known as the "wind-driven layer."

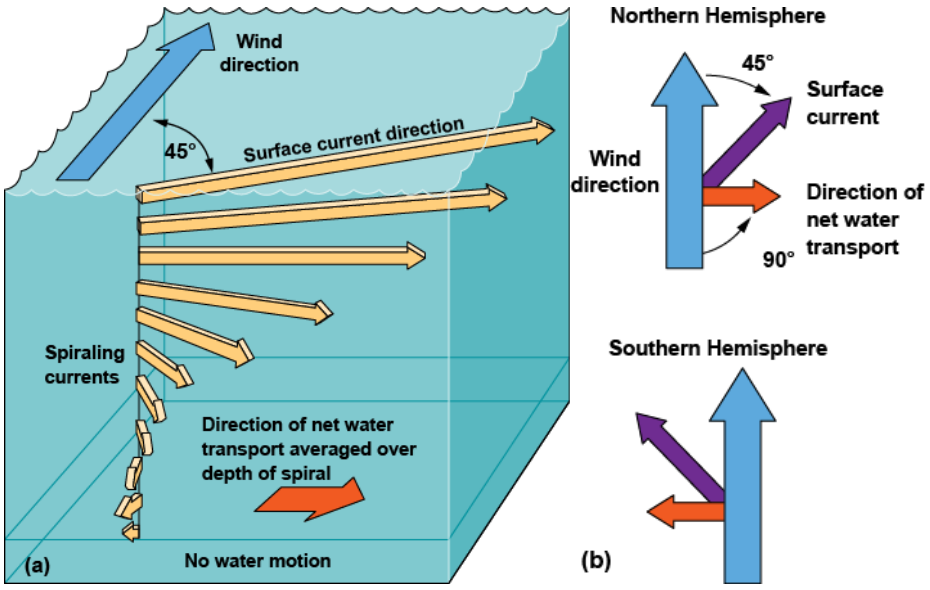


Figure 8-4. The Ekman spiral. (a) In the Northern Hemisphere, the Coriolis deflection causes a surface current to flow at an angle to the right of the wind direction. As the kinetic energy of the wind-driven current is transferred downward in the water column, each layer of water is deflected to the right of the direction of the layer above, producing a spiraling current called the Ekman spiral, in which speed decreases with depth. (b) If the wind blows long enough over deep water, the Ekman spiral motion becomes fully developed. When the spiral is fully developed in the Northern Hemisphere, the surface current direction is 45° to the right of the wind direction, and the net water transport is 90° to the right of the wind direction. In the Southern Hemisphere, the deflection and net water transport are to the left.

The most important feature of the Ekman spiral is that water in the wind-driven layer is transported at an angle *cum sole* to the wind direction and that the deflection increases with depth. Ekman showed that, under ideal conditions, the mean movement

of water summed over all depths within the wind-driven layer is at 90° *cum sole* to the wind. Such movement is called **Ekman transport**.

For the Ekman spiral to be fully established, the water column above the depth of frictional influence must be uniform or nearly uniform in density, water depth must be greater than the depth of frictional influence, and winds must blow at a constant speed for as long as a day or two. Because these conditions are rare, the Ekman spiral is seldom fully established.

Pycnocline and Seafloor Interruption of the Ekman Spiral

In most parts of the ocean, there is a range of depths within which density changes rapidly with depth (Fig. 8-5). Such vertical density gradients are called pycnoclines (**CC1**). In most parts of the open ocean, a permanent pycnocline is present with an upper boundary that is usually at a depth between 100 and 500 m. In some mid-latitude areas, another shallower seasonal pycnocline develops in summer at depths of approximately 10 to 20 m. Where a steep (large density change over a small depth increment) pycnocline occurs within the depth range of the Ekman spiral, it inhibits the downward transfer of energy and momentum. The reason is that water layers of different densities slide over each other with less friction than do water layers of nearly identical density. Water masses above and below a density interface are said to be frictionally decoupled. When a shallow pycnocline is present, the wind-driven layer is restricted to depths less than the pycnocline depth, and the Ekman spiral cannot be fully developed. Pycnoclines and the vertical structure of the ocean water column are discussed in more detail later in this chapter.

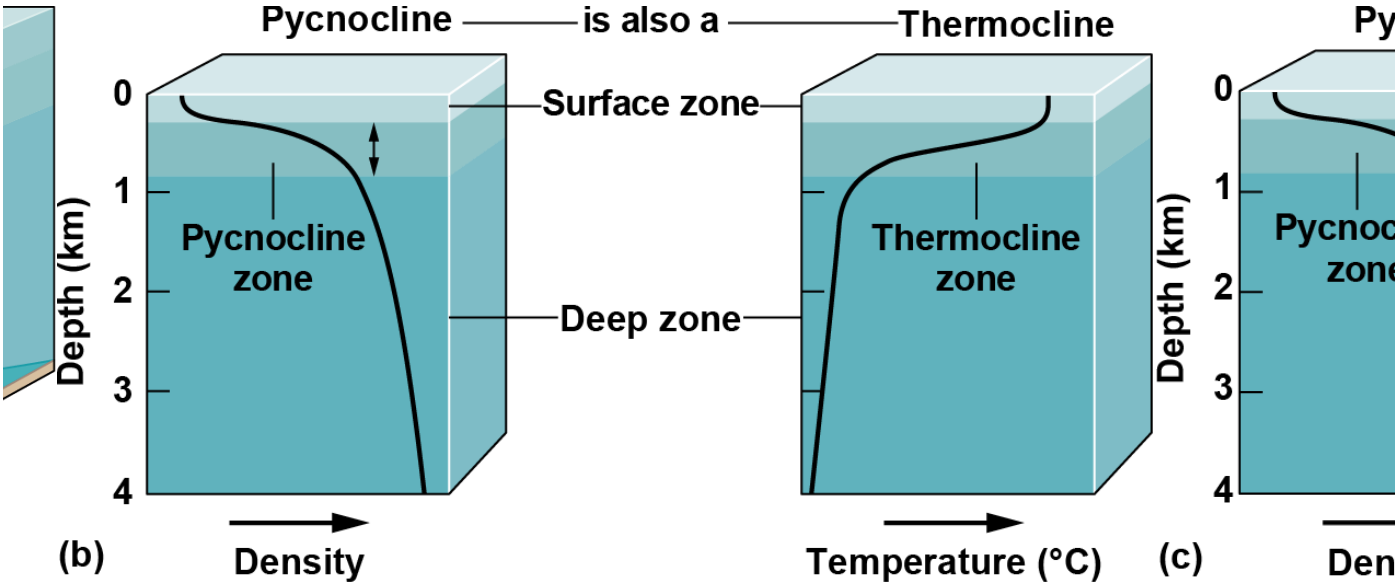


Figure 8-5. Density stratification in the oceans. (a) Throughout most of the open oceans (except in high latitudes where surface waters are cooled), a surface or mixed layer of relatively warm, nutrient-depleted, low-density water overlies a layer called the "pycnocline zone" in which density increases rapidly with depth. Below the pycnocline zone is a deep zone of cold, nutrient-rich water where density increases more slowly with depth. (b) The rapid density change with depth in the pycnocline is due primarily to changes in temperature, and the pycnocline is often called a "thermocline." Shallow seasonal thermoclines can also develop above the main pycnocline and in coastal waters because of solar heating of the surface water. (c) In some other regions, and particularly in shallow water near major freshwater inputs, a pycnocline may develop in which the density change with depth is due to vertical variations in salinity. In this case, the pycnocline is also a "halocline."

The Ekman spiral also cannot be fully established in shallow coastal waters because of added friction between the near-bottom current and the seafloor. Where the Ekman spiral is not fully developed because of a shallow pycnocline or seafloor, the surface current deflection is somewhat less than 45°, and Ekman transport is at an angle of less than 90° to the wind. In water only a few meters deep or less, the wind-driven water transport is not deflected from the wind direction. Instead, the current generally flows in the wind direction, but it is steered by bottom **topography** and flows along the depth **contours**.

Where Ekman transport pushes surface water toward a coast, the water surface is elevated at the **coastline** (Fig. 8-6a). If the Ekman transport is offshore, the sea surface is lowered at the coastline. Similarly, where Ekman transport brings surface water together from different directions, a **convergence** is formed at which the sea surface becomes elevated (Fig. 8-6b). In contrast, where winds transport surface water away in different directions, a **divergence** is formed at which the sea surface is depressed. These elevations or depressions are exaggerated in the diagrams.

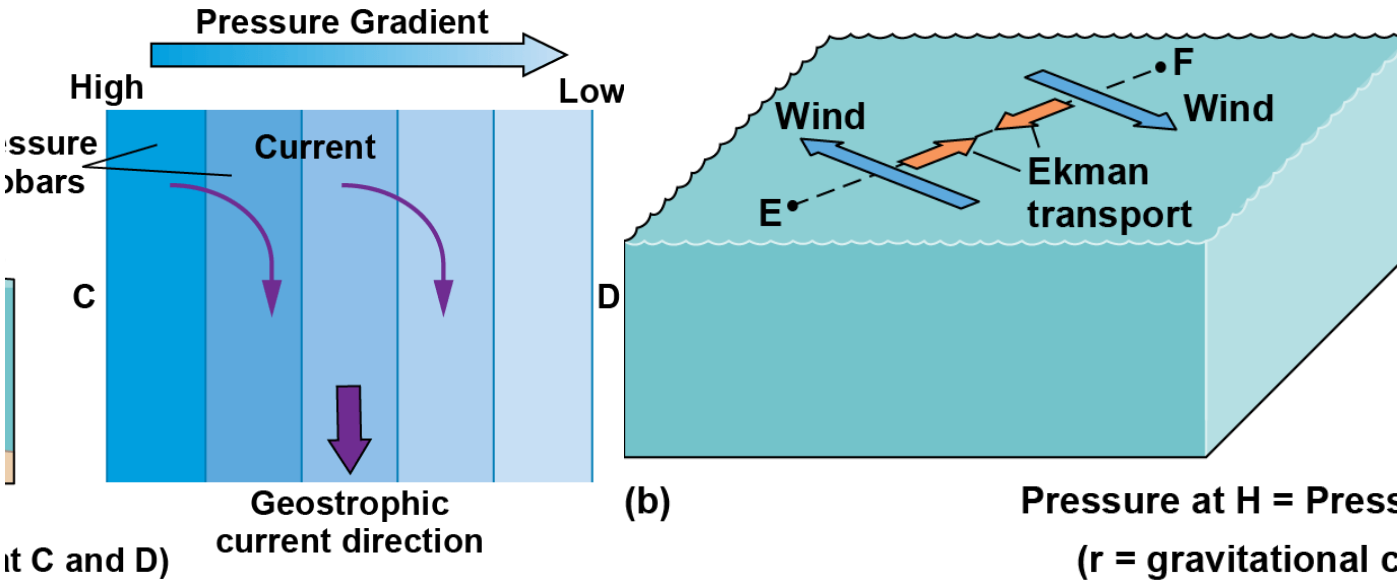


Figure 8-6. Ekman transport, geostrophic currents, and Ekman circulation. (a) Where winds blow surface water shoreward, a horizontal pressure gradient develops that causes water below the surface layer to flow offshore, and water to be downwelled near the coastline. This flow is deflected by the Coriolis effect such that, if the system reaches equilibrium, the lower-layer geostrophic current flows parallel to the coast (across the pressure gradient). (b) Away from the coastline, in locations where Ekman transport of surface waters tends to create a mound of surface water at a convergence, a horizontal pressure gradient develops that causes geostrophic currents below the surface layer to flow across the pressure gradient, and water downwells at the convergence. The circulation caused by the Ekman transport, downwelling, and geostrophic flow in the lower layer is called “Ekman circulation.” Ekman circulation occurs in reverse at divergences or where winds move surface water offshore.

This page titled 8.3: Ekman Motion is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.4: Geostrophic Currents

Ekman transport of surface layer water tends to produce sloping sea surfaces by “piling up” the water in some locations and “removing” surface water from others. **Geostrophic** currents are the result of horizontal pressure gradients in the water column below the surface caused by such variations in sea surface level (**CC13**).

Ekman Transport, Sea Surface Slope, and Pressure Gradient

If winds blew continuously, Ekman transport would cause the sea surface slope to increase continuously unless other water movements caused the sea surface to return to a flat configuration. Although sloping sea surfaces do develop, the slopes are always very small—so small that we perceive the sea surface to be flat even in regions of the most persistent winds. Water movements associated with Ekman transport that tend to restore the flat sea surface configuration are driven by the horizontal pressure gradients created under sloping sea surfaces (**Fig. 8-6**). Horizontal pressure gradients develop in response to the Earth’s gravitational force on the water.

The pressure gradient force tends to move water from the high-pressure region under the elevated part of the sea surface toward the lower-pressure region under the depressed part (**Fig. 8-6**). This direction is opposite to that of the Ekman transport. The flow induced by the pressure gradient tends to reduce the gradient and sea surface slope, restoring a flat sea surface.

Ekman transport occurs only in the wind-driven layer, which usually occupies only the upper few tens of meters of the water column. The mean water flow in the wind-driven layer is toward the highest point of sea surface elevation. However, the pressure gradient created when Ekman transport generates a sloping sea surface extends to depths beyond the wind-driven layer and is not restricted by shallow pycnoclines. As a result, the mean flow below the wind-driven layer is directed away from the area of maximum sea surface elevation (**Fig. 8-6**). In the area of maximum elevation, water is **downwelled** from the surface to below the wind-driven layer to join the pressure gradient flow.

Balance between Pressure Gradient and Coriolis Effect

All fluids, including air and water, are subject to the Coriolis effect when they flow horizontally (**CC12**). In the wind-driven layer, the Coriolis effect causes Ekman transport at approximately 90° *cum sole* to the wind direction. Ekman transport stores wind energy as **potential energy** associated with the elevation of the sea surface in one area relative to another. Water flows in response to the horizontal pressure gradient created by the sea surface elevation. As the water flows, it is accelerated and deflected *cum sole*. If allowed to develop long enough, the resulting currents, called “geostrophic currents” (**CC13**), are deflected to flow parallel to the contours of constant pressure (or along the side of the “hill” parallel to the sea surface height contours), and the Coriolis deflection is balanced by the pressure gradient force.

Geostrophic currents generated under ideal conditions beneath sloping sea surfaces are shown in **Figure 8-6**. These ideal conditions would be established only if winds blew at uniform speed over the entire area represented in the figure for several days or more. Such equilibrium conditions are never fully met in the oceans. Consequently, **Figure 8-6** shows only the general characteristics of geostrophic currents. In reality, the currents are continuously changing as winds vary.

Geostrophic current speed and direction are determined by the magnitude and distribution of the horizontal pressure gradient (**CC13**). The pressure gradient can be determined from the sea surface slope and the water density. For example, the pressure at point C in **Figure 8-6a** is greater than the pressure at the same depth at point D by an amount that is equal to the additional weight per unit area of water column at point C. The additional weight is equal to the water density multiplied by the difference in sea surface height between the locations. Using these simple relationships, oceanographers have been able to estimate some surface current speeds by measuring sea surface height differences very precisely with satellite radar sensors.

Sea surface slopes that are created by wind-driven currents and that sustain geostrophic currents are extremely small. For example, one of the fastest ocean currents is the Florida Current, a part of the Gulf Stream that flows around the tip of Florida from the Gulf of Mexico into the Atlantic Ocean. The Florida Current flows at about 150 $cm \cdot s^{-1}$ (5.4 $km \cdot h^{-1}$). This current flows on a sea surface with an approximate height difference of only 20 cm across about 200 km—a slope of 1 in 1 million. Most sea surface slopes are even smaller, and the geostrophic currents are correspondingly slower.

Dynamic Topography

Measuring very tiny sea surface height differences to determine current speeds would appear to be a very difficult task. However, oceanographers have learned to measure the differences indirectly. They carefully measure the density of seawater throughout the depth of the water column at different locations and calculate the **dynamic height**.

The dynamic height is the height of the water column calculated from a measured density profile of the water column between the surface and a depth below which it is assumed there are no currents. We can see how the calculation of dynamic height works if we remember that density is mass per unit volume. Consider several columns of water, each with a 1-cm^2 surface area, in different locations (**Fig. 8-7**). If the total mass of water in each column were the same, the height of the column with a lower average density would be greater than that of the column with higher average density (**Fig. 8-7**). From several calculated dynamic heights, we can estimate the average sea surface slope between locations. By making such calculations for many different locations, we can draw a detailed map of sea surface height called a "dynamic topography map." Detailed water density distribution data can also be used to identify slopes on pycnoclines, which are then used to estimate geostrophic current speeds and directions in deep-ocean water masses.

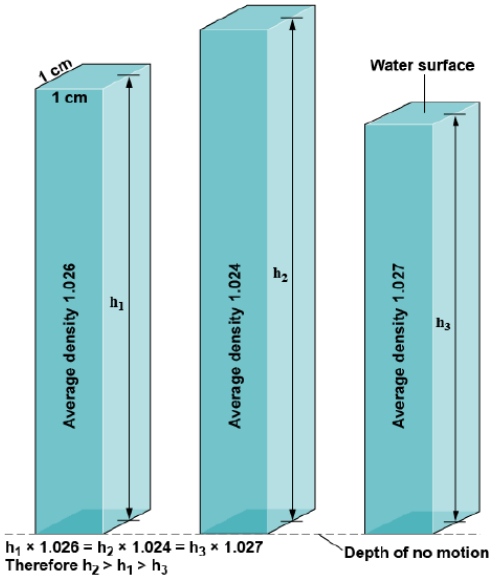


Figure 8-7. Calculations of dynamic height can be used to determine sea surface slopes. The height of the water column above a "depth of no motion" is calculated from measurements of the distribution of density with depth. The depth of no motion is a depth below which it can be reasonably assumed that there are no currents flowing. If the average density of the water column above the depth of no motion is known for several locations, the height of the sea surface at each location and the horizontal pressure gradients between the locations can be calculated.

This procedure for calculating dynamic height or sea surface slope is relatively simple, but it involves assumptions that are usually not fully met. First, we must select a depth of no motion below which we assume there are no currents. The pressure at this depth must be the same at all locations; otherwise, the horizontal pressure difference would cause water to flow horizontally. Because pressure does not vary at the depth of no motion, the mass of water above this depth must be the same at any location.

Until just the past few years, dynamic topography maps were generated from measurements made at many different individual locations at various times, sometimes months or years apart. The necessary measurements of salinity and temperature were made from research vessels, which take many weeks to complete a series of measurements at stations spaced across an ocean. As a result, until relatively recently, dynamic topography maps represented average conditions over many weeks or months, and we knew very little about short-term (up to months) variations in geostrophic currents.

Oceanographers still rely on indirect calculations of dynamic height to investigate geostrophic currents. However, the development of unattended moored arrays, autonomous floats, and satellite sensors now allows simultaneous data to be gathered at many locations. Moored vertical strings (arrays) of instruments at key locations and autonomous floats that profile the water column can collect and send back to shore-based facilities continuous data that can be used to calculate dynamic height. More than 4000 autonomous floats are currently deployed. Satellite radar sensors have now become sensitive enough to measure sea surface heights directly, and with sufficient precision that the data can be used to draw detailed dynamic topography maps of any part of the ocean every few days. Although the satellite data give only the dynamic height at the sea surface, these data,

when combined with the data collected by moored and autonomous instruments, have drastically improved our ability to observe geostrophic currents. As a result, we are just beginning to understand their complexities and variability.

Energy Storage

Geostrophic currents flowing near the surface are generated by winds because they depend on Ekman transport to establish the sea surface slopes under which they flow. However, geostrophic currents continue to flow even if winds abate, because Ekman transport stores wind energy as potential energy by piling up the water. The stored potential energy is converted to momentum to maintain geostrophic currents during intervals when winds are light or cease.

The length of time during which geostrophic currents continue to flow after winds abate depends on the magnitude of the sea surface slope and the area over which the slope occurs. The larger the area of sloping sea surface and the greater the slope, the longer it takes for geostrophic currents to restore the sea surface after winds abate. Where local winds create sea surface slopes over small areas, geostrophic currents restore the sea surface and cease to flow within a few hours after the winds abate. This is often the case for geostrophic currents created by the interaction of storms and coastal water masses.

This page titled 8.4: Geostrophic Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.5: Open-Ocean Surface Currents

Surface currents of the open oceans (**Fig. 8-3**) are wind-driven currents initiated by Ekman transport and maintained as geostrophic currents. Although they typically extend to depths of several hundreds of meters, they may extend as deep as 2000 m in some limited cases. The most obvious features of ocean surface currents are the **gyres** in subtropical **latitudes** of each ocean. Separate **subtropical gyres** are present north and south of the equator in each ocean except in the Indian Ocean, where the Northern Hemisphere gyre would occur at a location occupied by landmasses. Subtropical gyres in the Northern Hemisphere flow clockwise; those in the Southern Hemisphere flow counterclockwise.

At latitudes above the subtropics, ocean surface currents are more complicated. The North Atlantic and North Pacific oceans have high-latitude gyres similar to, but less well formed than, the subtropical gyres. The high-latitude gyres rotate in the opposite direction (counterclockwise) from the northern subtropical gyres. In the Southern Hemisphere, high-latitude gyres are absent because the surface current that flows clockwise from west to east is not interrupted by a landmass, so it flows completely around Antarctica.

Geostrophic Currents on a Water-Covered Earth

If there were no continents, geostrophic surface layer currents would be relatively simple. The climatic winds and **convection cells** that would be present on a rotating, water-covered Earth are shown in **Figure 7-10**. **Figure 8-8a** shows these idealized climatic winds, and the directions in which the wind-driven layer of ocean water would be moved by Ekman transport due to the winds. In the **trade wind** and polar easterly zones, Ekman transport is partly toward the pole and partly toward the west. In the **westerly** wind zones, Ekman transport is toward the equator and to the east. Surface water layer divergences and **upwelling** of subsurface water occur at the equator and at the atmospheric upwelling regions between the polar and Ferrel cells in each hemisphere. Surface water convergences occur at the atmospheric downwelling regions between the Hadley and Ferrel cells in each hemisphere. Thus, Ekman transport would produce a depression of the sea surface at the equator and between the polar and Ferrel cells in each hemisphere, and an elevation of the sea surface between the Hadley and Ferrel cells in each hemisphere (**Fig. 8-8b**).

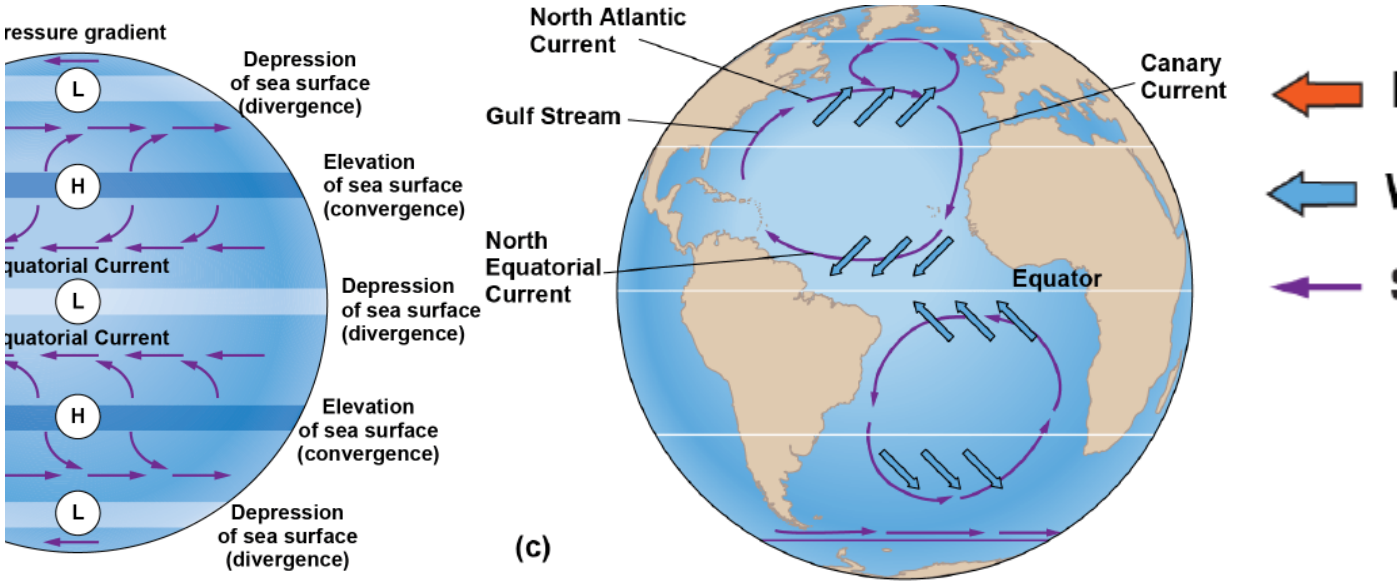


Figure 8-8. Ocean surface layer currents are driven by climatic winds and modified by the presence of landmasses. (a) The climatic winds on an idealized water-covered Earth would cause Ekman transport of surface waters in the directions shown. (b) Those water movements would lead to the formation of convergence and divergence, and geostrophic currents would flow on the pressure gradients created by the Ekman transport of the surface layer. These geostrophic currents would flow around the Earth from east to west, or vice versa, between the latitudes of the convergences and divergences. (c) However, landmasses block the flow of geostrophic currents and divert them to the north or south, resulting in the formation of geostrophic gyres.

The sloping sea surfaces created by climatic winds establish horizontal pressure gradients within the water column and thus

establish geostrophic currents. If an equilibrium were reached, the geostrophic currents would flow around the Earth from east to west or from west to east across the pressure gradients (**Fig. 8-8b**). Currents would flow to the west in the trade wind zones and polar regions and to the east in the westerly wind zones. Geostrophic currents on the real Earth cannot flow in this way because they are interrupted by continents in all but the westerly wind zone around Antarctica. The Antarctic Circumpolar Current does, in fact, flow from west to east around the Earth in the Southern Hemisphere westerly wind zone (**Fig. 8-3**), just as the no-continent model suggests (**Fig. 8-8b**).

Geostrophic Gyres

Around each of the three major oceans, continents provide a western boundaries (the Atlantic coast of North and South America, Pacific coast of Asia and Australia, and Indian Ocean coast of Africa) and eastern boundaries (the Atlantic coast of Europe and Africa, Pacific coast of North and South America, and Indian Ocean coast of Southeast Asia and Australia). Within each ocean, geostrophic east-to-west or west-to-east currents are established at the same latitudes and in the same locations as those in the simple ocean-covered Earth model (**Figs. 8-3, 8-8b**). Where these currents meet a continent, they are blocked and diverted to the north or south.

The westward-flowing currents within the trade wind zones of the Northern and Southern Hemispheres are called the North Equatorial Current and the South Equatorial Current, respectively (**Fig. 8-3**). These Equatorial currents carry large volumes of water from east to west until they meet the western boundary continent. Surface water “piles up” there, and the Equatorial currents are deflected away from the equator. The currents then flow toward higher latitudes as **western boundary currents** until they enter the westerly wind zone, where they join the west-to-east geostrophic flow. When the currents reach the eastern boundary land mass, they are deflected primarily toward the equator. After turning toward the equator, they flow as **eastern boundary currents** until they enter the trade wind zone, rejoin the Equatorial Current, and complete their circuit, or gyre (**Figs. 8-3, 8-8c**).

As western boundary currents flow toward the pole, they are subject to the Coriolis effect. Therefore, if there were no opposing horizontal pressure gradient, they would be deflected to the east. Similarly, as eastern boundary currents flowed toward the equator, they would be deflected to the west (**Fig. 8-9a**). In each case, the deflection would be toward the center of the gyre. In addition, Ekman transport of surface water in the trade wind and westerly wind zones and the Coriolis deflection of the gyre currents, including the boundary currents, all tend to move water toward the center of the gyre. As a result, an equilibrium situation is created in which there is a mounded sea surface at the center of the gyre (**Fig. 8-9b**). Currents that make up the gyre flow around the mound in geostrophic balance between the Coriolis effect and the horizontal pressure gradient that surrounds this central location.

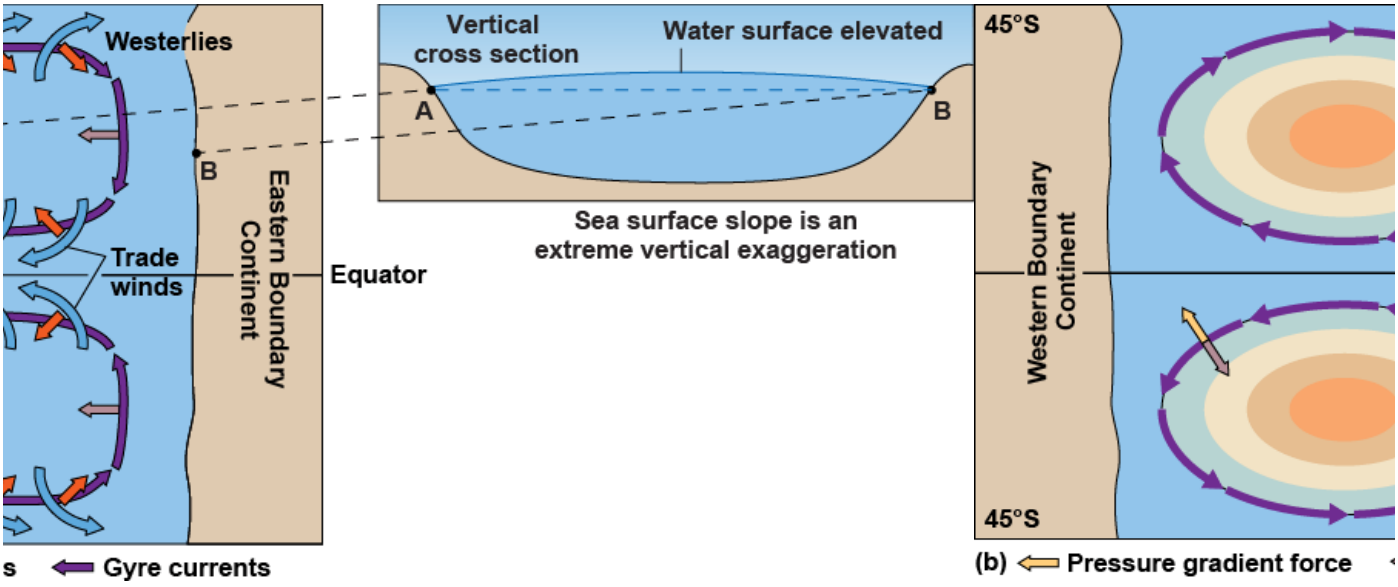


Figure 8-9. Geostrophic currents of the subtropical gyres. (a) Wind-driven Ekman transport due to the trade winds and westerlies, and the Coriolis deflection of the boundary currents, both act to elevate the ocean surface at the center of each subtropical gyre. (b) This elevated sea surface creates a horizontal pressure gradient that tends to move water outward. As water flows on this pressure gradient, it is deflected by the Coriolis effect and flows geostrophically around the mound. At equilibrium, the sea surface elevation does not vary because the tendency of water to be moved to the center of the mound is balanced by the pressure gradient, which generates geostrophic currents. Thus, the wind energy of the trade winds and westerlies is transferred into the geostrophic currents that flow continuously and form the subtropical gyres.

Hence, gyres are geostrophic currents that flow on mounded surfaces. The center of each subtropical gyre is called a "subtropical convergence" (**Fig. 8-3**). Subtropical gyres are essentially the same as the idealized gyres shown in **Figure 8-9**. However, they are modified in shape and location by the configuration of each ocean's coastlines (**Fig. 8-3**).

Geostrophic ocean gyres act like giant flywheels, which use angular momentum to store rotational energy. They spin at an almost constant speed that represents the average wind energy input to the gyre. If winds cease or are abnormally light, the geostrophic current gyre, or flywheel, slows very gradually because turbulence and friction between the gyre currents and the adjacent water masses or seafloor dissipate the energy very slowly. When winds resume, Ekman transport builds the height of the mound, increasing the sea surface slope and speed of the gyre. This process occurs very slowly because massive volumes of water must be moved to elevate the sea surface even a millimeter or two over a mound that is hundreds or thousands of kilometers in diameter. Winds that drive the ocean gyres are variable over periods of days, but very long periods of calm or unusually high winds are rare. Because gyre current speeds change very slowly, they are relatively stable and generally do not reflect normal wind speed variations.

Westward Intensification of Boundary Currents

Western boundary currents include the Gulf Stream and the Brazil, Kuroshio, East Australian, and Agulhas Currents (**Fig. 8-3**). Eastern boundary currents include the California, Peru, Canary, and Benguela Currents. Western boundary currents are narrower, faster, deeper, and warmer than eastern boundary currents (**Table 8-1**). They are generally so deep that they are constrained against the edge of the **continental shelf** and do not extend across the continental shelf to the shore.

Table 8-1. *Boundary Currents*

Properties	Western Boundary Currents	Eastern Boundary Currents
Northern Hemisphere examples	Gulf Stream, Kuroshio Current	California Current, Canary Current
Southern Hemisphere examples	Agulhas Current, Brazil Current	Peru Current, Benguela Current
Width	Narrow (≤100 km)	Broad (≈1000 km)
Depth	Deep (to 2 km)	Shallow (≤500 m)
Speed	Fast (>100 km•day^{-1})	Slow (<50 km•day^{-1})
Volume transport	Large (50×10^{6} m^{3}•s^{-1})	Small (10–15×10^{6} m^{3}•s^{-1})
Boundaries with coastal currents	Sharply defined	Diffuse
Upwelling	Almost none	Frequent

Properties	Western Boundary Currents	Eastern Boundary Currents
Nutrients	Depleted	Enhanced by upwelling
Fishery	Usually poor	Usually good
Water temperature	Warm	Cool

The reasons for the westward intensification of boundary currents are quite complex. However, the major reason is related to the Earth's rotation. Western boundary currents are intensified because the strength of the Coriolis effect varies with latitude.

We can look at westward intensification in a simplified way. The Coriolis effect, which increases from the equator to the poles, is weaker in the latitudes of the trade wind band than in the westerly wind band. Therefore, water moving eastward is deflected more quickly toward the south in the westerly wind zone than water moving westward is deflected toward the north in the trade wind zone (**Fig. 8-10**). Consequently, in the westerly wind zone, geostrophic flow tends to transport surface water toward the center of the gyre over the entire width of the ocean. In contrast, surface water transported westward in the trade wind zone tends to flow with less deflection (that is, in a circle of larger radius; **CC12**) across the ocean (**Fig. 8-10**), is constrained at the equator by the Ekman transport of the trade winds in the other hemisphere, and tends to pile up on the west side of the ocean.

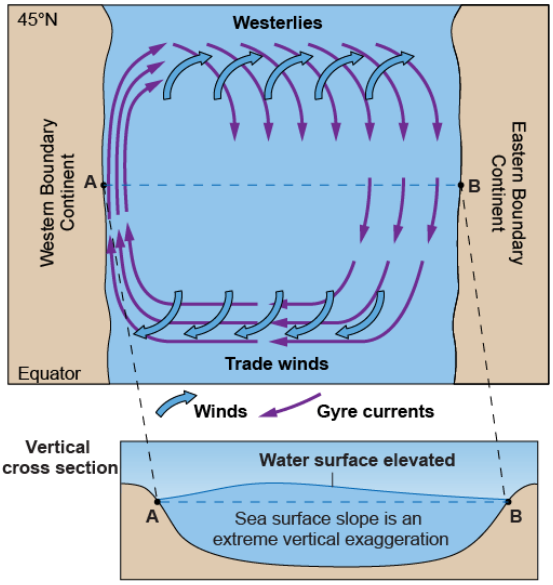


Figure 8-10. The westward-moving geostrophic currents of the subtropical gyres flow near the equator, where the Coriolis deflection is weak, whereas the eastward-moving currents flow where the Coriolis deflection is much greater. Hence, the westward-moving water near the equator flows across the ocean more directly, piling up water on the west side. The greater Coriolis deflection further from the equator tends to turn parts of the eastward-moving gyre currents toward the equator even before it reaches the eastern boundary continent, "spreading" the southerly part of the gyre currents that returns water from the westerly region toward the equator.

As a result, the mound at the center of the subtropical gyre is offset toward the west side of the ocean (**Fig. 8-11**). The western boundary current is laterally compressed against the continent, and the sea surface slope is steeper toward the western boundary than toward the eastern boundary. Steeper sea surface slopes cause faster geostrophic currents. Therefore, western boundary currents are faster than eastern boundary currents. The westward offset of the subtropical gyre mound causes the pycnocline to be depressed (deeper) on the west side of the ocean in relation to the east side (**Fig. 8-11**). Hence, western boundary currents are narrower and deeper than eastern boundary currents.

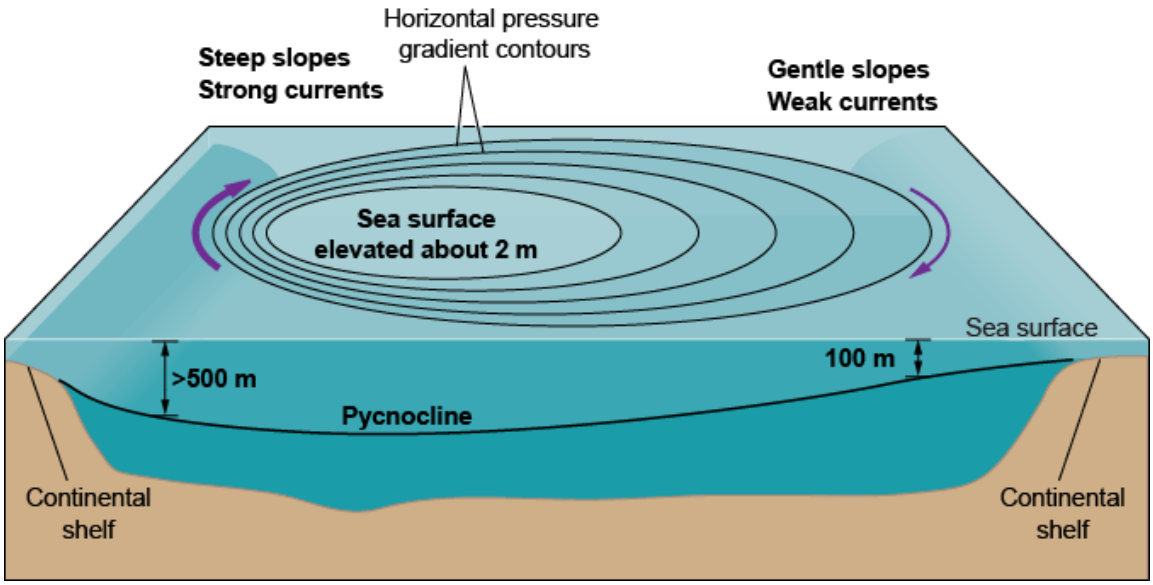


Figure 8-11. In the Northern Hemisphere, the sea surface elevation in the center of the subtropical gyres is displaced to the west. Thus, the horizontal pressure gradients are stronger and the gyre currents are faster on the west side of the ocean than on the east side. In addition, because of this displacement, the mixed layer is thicker on the west side and the deep layer does not extend onto the continental shelf.

This simplified explanation serves reasonably well. However, for a more detailed understanding of westward intensification, it is necessary to examine a difficult concept called “vorticity” (see **Online Box 8B1**).

The characteristic differences between western and eastern boundary currents have major consequences for processes that sustain fisheries off the adjacent coasts (**Chap. 12**). The differences are also an important factor in the transport of heat poleward (**Chap. 7**). Warm western boundary currents are fast-flowing and deep. They have little time to cool and only a small surface area from which to lose heat as they move into higher latitudes. Therefore, heat energy carried away from the equator by western boundary currents is released to the atmosphere primarily when the water enters the westerly wind zone, where the gyre currents flow eastward. Heat transported by western boundary currents moderates the climates of regions into which they flow. For example, the Gulf Stream transports heat north and then east across the Atlantic Ocean, moderating western Europe’s climate.

This page titled 8.5: Open-Ocean Surface Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.6: Equatorial Surface Currents

The Northern and Southern Hemisphere trade wind zones are separated by the Doldrums, where winds are very light and variable (**Figs. 7-10, 7-11**). In the simplified model of subtropical gyre formation discussed in the preceding section, the Northern and Southern Hemisphere trade winds produce Ekman transport of surface water to the northwest and southwest, respectively (**Fig. 8-8**). This pattern would be present if the Doldrums (the atmospheric **intertropical convergence zone**) were exactly over the equator. However, atmospheric circulation interacts in a complex manner with continental landmasses, which are concentrated in the Northern Hemisphere. One result of this interaction is that the Doldrums are displaced somewhat to the north of the equator during most of the year (**Fig. 7-11**).

The displacement of the Doldrums leads to a complex system of west to east and east to west currents that flow on or very close to the equator. These currents and their behavior are important components in the **El Niño** Southern Oscillation (ENSO) described in **Chapter 7**. The most important characteristic of Equatorial Currents is that they are able to flow directly east to west or west to east across entire oceans without being deflected by the Coriolis effect since the Coriolis deflection is zero at the equator and negligibly small within a few degrees north and south of the equator. Equatorial Currents are discussed in more detail in **Online Box 8B2**.

This page titled 8.6: Equatorial Surface Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.7: High-Latitude Surface Currents

In the North Atlantic and North Pacific, secondary gyres occur at higher latitudes than the subtropical gyres. They are called “subpolar gyres” and rotate in the direction opposite that of the adjacent subtropical gyres. Subpolar gyres are wind-driven in the same way as subtropical gyres. However, the prevailing winds of the region, the polar easterlies, are much more variable than trade winds.

Because of this wind variability and the complex shapes of the ocean basins in polar regions, the subpolar gyres are more complex and variable than the subtropical gyres.

In the Southern Hemisphere, subpolar surface currents are different from their Northern Hemisphere counterparts. The primary reason is the lack of continents in the Southern Hemisphere. Because of the wide connection between the oceans, Southern Hemisphere subpolar currents flow all the way around the Antarctic continent (**Fig. 8-3**). At high latitudes, a geostrophic current, the East Wind Drift, flows westward around Antarctica. It flows in response to the pressure gradient caused by the elevation of the sea surface near the Antarctic continent due to southerly Ekman transport in the polar easterly wind belt. Farther from Antarctica, an eastward current, the Antarctic Circumpolar Current (or West Wind Drift), flows geostrophically around Antarctica under a sea surface that slopes up to the north as a result of the Ekman transport of the westerlies (**Fig. 8-3**). The Antarctic Circumpolar Current is combined with, and a part of, the eastward-moving currents of the Southern Hemisphere subtropical gyres in each ocean.

This page titled 8.7: High-Latitude Surface Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.8: Upwelling and Downwelling

Although wind-driven surface water motions are primarily horizontal, winds can also cause vertical water movements. Vertical movements occur because wind-driven Ekman transport drives a layer of surface water up to about 100 m deep across the underlying water layers. Winds moving surface water away from an area cause a divergence at which the surface water is replaced by water upwelling from below (**Fig. 8-12a**). Upwelling is important because colder water upwelled from below the pycnocline has high concentrations of **nutrients**, such as nitrogen and phosphorus compounds. The nutrients are needed to fertilize **phytoplankton**, the microscopic algae and **phototrophic** bacteria and archaea that grow only in near-surface waters and that are the principal source of food supporting all animal life in the oceans. Most surface waters are deficient in nutrients and cannot sustain phytoplankton growth unless nutrients are supplied by upwelling or recycling (**Chap. 12**).

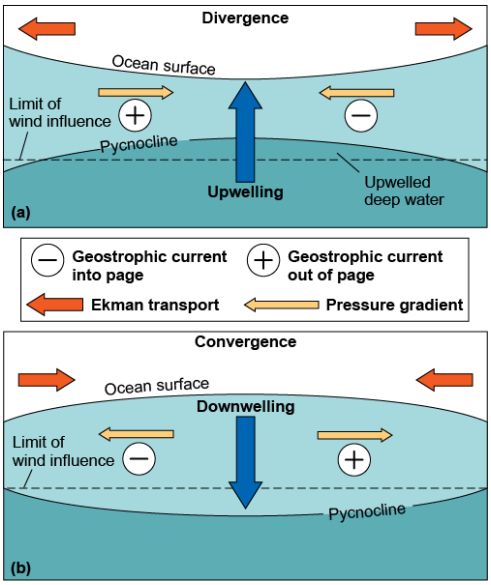


Figure 8-12. Divergences and convergences are areas of upwelling and downwelling in the open oceans. (a) At a divergence, the surface layer is thinned and the pycnocline depth reduced. If wind mixing extends below the pycnocline depth, cold, nutrient-rich, deep water is upwelled and mixed into the surface layer. (b) At a convergence, the surface layer is thickened, and the pycnocline is driven deeper. If the pycnocline is below the maximum depth of wind influence, there is no mixing of cold, nutrient-rich subpycnocline water into surface layers, even when winds are very strong.

At a convergence, winds moving surface waters toward an area elevate the sea surface and create a high-pressure region in the subsurface water, causing downwelling (**Fig. 8-12b**). Surface water is transported downward at the convergence and then outward below the wind-driven layer under the influence of the horizontal pressure gradient. Downwelling thickens the layer of warmer surface water above the pycnocline and tends to isolate cold subpycnocline water, preventing it from mixing upward into the surface layers. Downwelling also occurs in areas where the surface water density is increased by cooling or evaporation sufficiently that this surface water has a higher density than the water below.

Locations of Upwelling and Downwelling

Divergences are upwelling areas where primary productivity is high because of the supply of nutrients from below. Convergences are downwelling areas where productivity is poor. The principal open-ocean divergences include a band that parallels the equator between the North and South Equatorial Currents and a band around Antarctica between the Antarctic Circumpolar Current and the eastward coastal flow around the continent (**Fig. 8-3**). In the equatorial region, upwelling is inhibited by the flow of warm surface water from west to east in the Equatorial Countercurrents (**Online Box 8B2**). Consequently, persistent upwelling of nutrient-rich, cold subpycnocline water at the equatorial divergence is limited primarily to the east side of the Pacific Ocean off the Peruvian coast. Upwelling in this region is inhibited by the movement of warm surface water from west to east during El Niño (**Fig. 7-18b**). The principal open-ocean convergences are in the center of each of the subtropical surface current gyres (**Fig. 8-3**).

Coastal upwelling can occur when winds blow parallel to the coast and cause Ekman transport of the surface layer offshore.

Surface water transported offshore is replaced by deeper water that moves inshore and is upwelled (**Fig. 8-13a**). **Coastal downwelling** occurs when surface water is transported toward shore displaces near-shore surface water downward and forces the deeper water to move offshore (**Fig. 8-13b**). These processes are particularly important to local ecosystems in areas where there is a shallow pycnocline, with warm, nutrient-depleted water at the surface and colder, nutrient-rich water below. In such areas, offshore transport causes coastal upwelling that supplies large quantities of nutrients to support phytoplankton growth (**Chap. 12**).

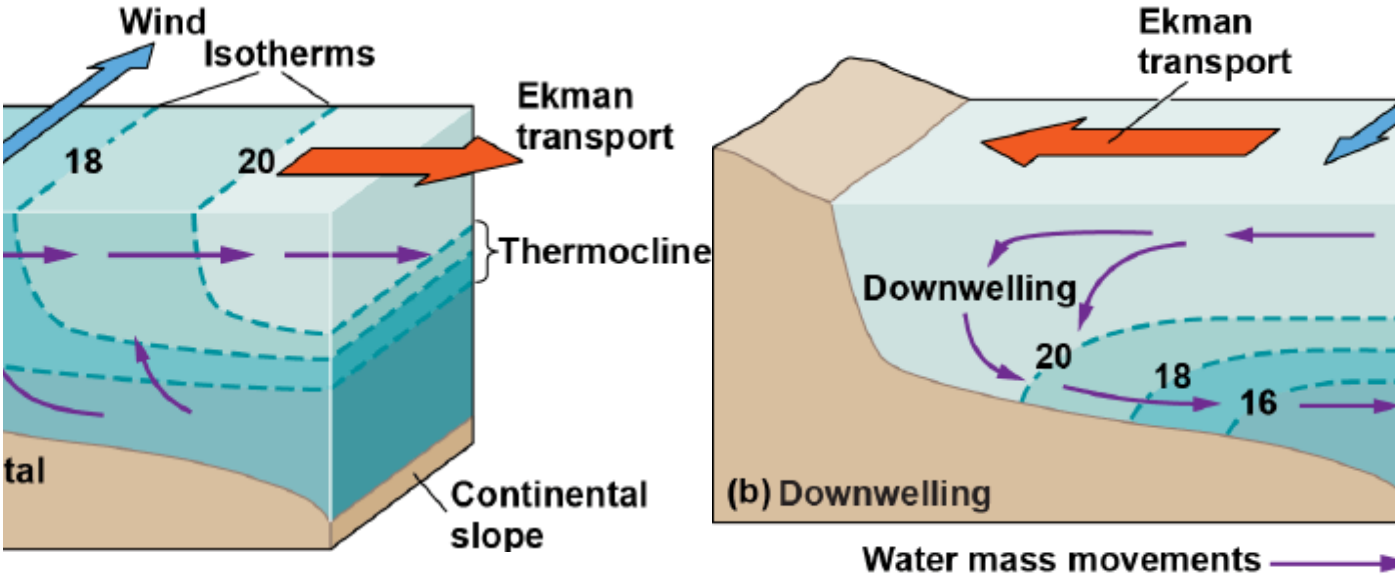


Figure 8-13. Coastal upwelling and downwelling. (a) When winds produce Ekman transport of the surface water layer in an offshore direction, the pycnocline (usually a thermocline) is raised, sometimes all the way to the surface, and cold, nutrient-rich, deep water is upwelled into the surface layer. (b) When winds produce Ekman transport of the surface water layer in an onshore direction, the pycnocline (again usually a thermocline) is depressed, shelf surface water is downwelled, and cold, nutrient-rich, deep water is forced deeper and off the shelf, so it cannot enter the surface layer, even if winds are strong enough to mix coastal water to the depth of the continental shelf break.

Boundary Currents and Upwelling or Downwelling

The characteristics of ocean gyre boundary currents strongly influence the occurrence and persistence of coastal upwelling near the basin's coasts. Western boundary currents are deep and generally compressed against the continental shelf edge. Hence, a deep layer of warm surface water generally lies over the cold, nutrient-rich subpycnocline water and prevents it from extending onto the continental shelf (**Fig. 8-13b**). Because coastal winds normally move aside a surface layer only a few tens of meters deep, offshore Ekman transport in western boundary current regions leads to upwelling of warm, nutrient-poor water from within the deep surface layer (**Fig. 8-13b**). Thus, coastal upwelling of nutrient-rich water is rare in western boundary current regions.

Eastern boundary currents are relatively shallow and wide, and they extend onto the continental shelf. Cold, nutrient-rich, deep water can migrate onto the shelf as a bottom layer below the shallow, warm surface layer (**Fig. 8-13a**). Only moderate coastal winds with offshore Ekman transport are needed to cause cold, nutrient-rich waters to upwell to the surface. Consequently, coastal upwelling is more frequent, widespread, and persistent on the eastern boundaries of the oceans (west coasts of continents) than on the western boundaries of the oceans (east coasts of continents).

This page titled 8.8: Upwelling and Downwelling is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.9: Coastal Currents

Coastal currents are independent of and more variable than the adjacent oceanic gyre boundary currents. The zone in which coastal currents flow varies in width along different coasts. For example, the coastal current zone is wide off the Mid-Atlantic coast of North America because the continental shelf is wide, and the western boundary current (Gulf Stream) flows in deeper water off the edge of the shelf. In contrast, the coastal current zone off the Pacific coast of North America is narrow because the continental shelf is narrow, and the relatively shallow eastern boundary current flows over the narrow shelf almost to the coastline. Variable winds, tides, freshwater inflow from rivers, friction with the seafloor, and steering by coastline irregularities, such as capes, affect currents more in coastal waters than in the open ocean.

Coastal currents generally flow parallel to the coastline in a direction determined by the winds (and by Ekman transport). Coastal winds are more variable in speed and direction than are trade winds and other global winds that drive the ocean gyres. Coastal currents can be established quickly in response to storm winds, but they can also disappear or change direction within a few hours.

The strongest coastal currents occur when strong winds blow in areas with large freshwater inputs from rivers. In such areas, a shallow, low-salinity layer of surface water is formed over a very steep pycnocline. The surface layer slides easily over layers below this pycnocline. Thus, the wind energy is concentrated in the shallow, surface-layer current, rather than being transmitted and distributed throughout a greater depth.

Because they are directed by local wind patterns and interaction with the coastline, coastal currents may flow in the opposite direction from the adjacent ocean gyre boundary currents. For example, coastal currents off the Atlantic coast of North America generally flow to the southwest from Labrador to Cape Hatteras and along the Carolina, Georgia, and Florida coasts (**Fig. 8-14**) in the direction opposite that of the adjacent Gulf Stream.

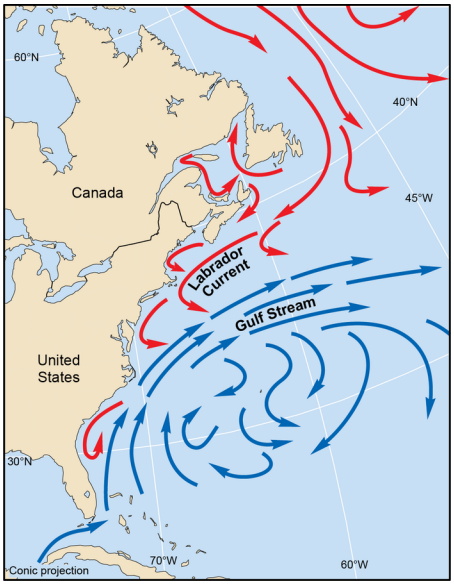


Figure 8-14. Off the Atlantic coast of North America, the coastal currents are cold-water currents that flow to the south on the continental shelf inshore from the northward-flowing Gulf Stream.

The boundary between the warm Gulf Stream and colder coastal water, called a **front**, can be clearly discerned in satellite imagery as a distinct difference in water color. The nearshore water is greenish or brownish, whereas Gulf Stream water is a clear, deep blue. The color difference is due to the greatly reduced concentrations of suspended sediment and phytoplankton in Gulf Stream water (**Chap. 5**).

Off the west coast of North America, the boundary between the water of the California Current (which is part of the North Pacific Ocean subtropical gyre) and the coastal water is not well defined. The main reason is that the California Current is a weak, diffuse eastern boundary current. During winter and spring, a weak and variable coastal current, the Davidson Current, flows north in the opposite direction from the adjacent boundary current. The Davidson Current is driven by the predominant winds of winter and

spring that blow from the southwest. These winds produce Ekman transport of water onshore and a geostrophic flow to the north on the resulting sea surface slope (**Fig. 8-15a**). In summer and fall, prevailing winds change to blow from the north. They cause Ekman transport offshore, a generally southward-flowing coastal current, and upwelling (**Figs. 8-13, 8-15b**). The summer and fall upwelling is responsible for the very high primary productivity and abundant sea life in coastal waters off the west coast of North America.

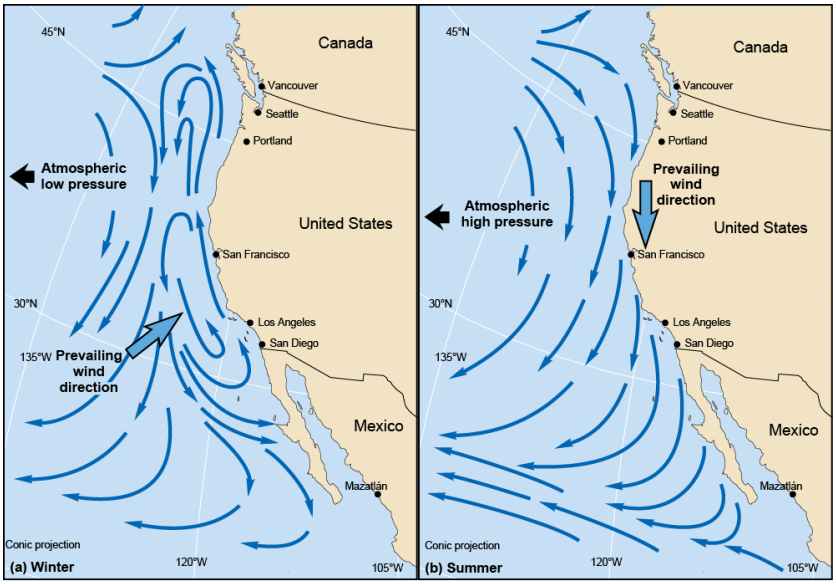


Figure 8-15. Coastal currents reverse seasonally off the Pacific coast of North America. (a) They flow to the north during winter and sometimes spring, when there is an area of atmospheric low pressure offshore and the coastal winds blow mainly from the southwest. (b) In the summer, when there is atmospheric high pressure offshore and coastal winds are mainly from the north, the coastal currents flow to the south, in the same direction as the offshore southward current that is part of the subtropical gyre in the North Pacific Ocean.

This page titled 8.9: Coastal Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.10: Eddies

Chapter 7 describes how winds are arranged in global patterns that determine climate. It also explains that winds are highly variable in any given location. The local variations are part of **weather** systems, which we readily associate with the swirling patterns of clouds and precipitation. The global ocean current systems are analogous to the climatic winds, and the oceans have their own "weather systems," which includes variable swirling motions and meandering fronts, just like atmospheric weather. These swirling motions are called **eddies**.

The principal difference between atmospheric weather and ocean current "weather" is the scale of the eddy motions. Ocean currents generally move more slowly than most atmospheric winds. Consequently, the Coriolis effect tends to make ocean currents flow in curving paths that have a smaller radius than comparable atmospheric eddies (**CC12**). In other words, ocean eddies are generally much smaller than atmospheric eddies. Because ocean eddies are smaller, they are more numerous than atmospheric eddies.

Satellites can detect many atmospheric circulations and patterns that contribute to the weather. However, such **synoptic** measurements cannot be made so easily within the body of the oceans. Therefore, we do not have a detailed understanding of deep ocean current variability.

Satellite Observations of Eddies

Satellite infrared sensors detect ocean surface temperature by measuring the sea surface heat radiation. Satellite optical sensors tuned to different **wavelengths** of light can detect ocean color by measuring **back-scattered** sunlight, creating images that look similar to photographs. Such satellite observations are most effective for observing ocean surface currents where temperature and/or color differ markedly between water masses. Unfortunately, in most areas, the differences are very small. The easiest surface currents to observe by satellite are western boundary currents because the water they carry is warmer than the adjacent coastal waters. Areas near river discharges are also good for satellite observations of currents because the suspended particles in river outflow alter ocean color.

Even in areas where satellite observations are most effective, color or temperature variations of the sea surface are so small that they must be computer-enhanced to be seen by researchers. Computer enhancement of satellite images generally assigns bright false colors to areas that are slightly different in temperature and true color, thus making the demarcations between currents or water masses more prominent. Figure 8-16 is a color-enhanced infrared satellite image of the Gulf Stream region of the Atlantic Ocean.

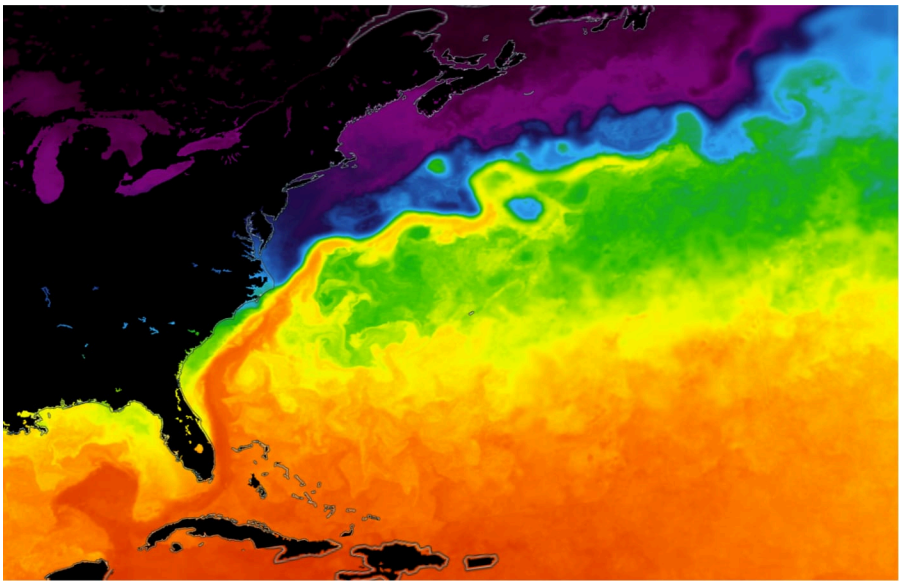


Figure 8-16. This image of the northwestern Atlantic Ocean shows the sea surface temperature derived from NOAA satellite sensors. Temperatures are depicted by the false colors in this image. Shades of red denote the areas of highest temperature, and progressively lower temperatures are depicted by orange, yellow, green, blue, and violet shades. The image shows the Gulf Stream separating the coastal water mass from the Sargasso Sea water mass. The complex meanders of the Gulf Stream and both cold-core and warm-core eddies are easy to see. For example, the isolated blue area surrounded by yellow and green in the offshore water is a cold core eddy, and the green areas isolated in the light blue coastal water are warm core eddies.

Mesoscale Eddies

The image in **Figure 8-16** shows the warm Gulf Stream flowing northward from the tip of Florida along the edge of the continental shelf. The front between this water and the colder coastal water is very sharp (**Fig 8-14**). A similar front can be seen between Gulf Stream water and colder Sargasso Sea water in the interior of the North Atlantic subtropical gyre. The infrared satellite image shows that there is great complexity along these fronts. We can see numerous secondary fronts and **meanders** of the Gulf Stream. We also see isolated, almost circular areas of warm water on the coastal-water side of the Gulf Stream and similar isolated areas of cold water on the Sargasso Sea side called warm-core and cold-core rings.

These Gulf Stream meanders and rings are eddies that continuously form and change shape and location. Meanders travel slowly northward along with the Gulf Stream and can become larger as they move. If a meander becomes tight enough, it can break off the Gulf Stream entirely and form a spinning ring of water. If the meander is to the coastal-water side of the Gulf Stream, a warm-core ring that spins clockwise is formed and isolated (**Fig. 8-17**). If the meander is to the Sargasso Sea side of the Gulf Stream, a cold-core ring that spins counterclockwise is formed as colder coastal water is pinched off inside the meander (**Fig. 8-17**).

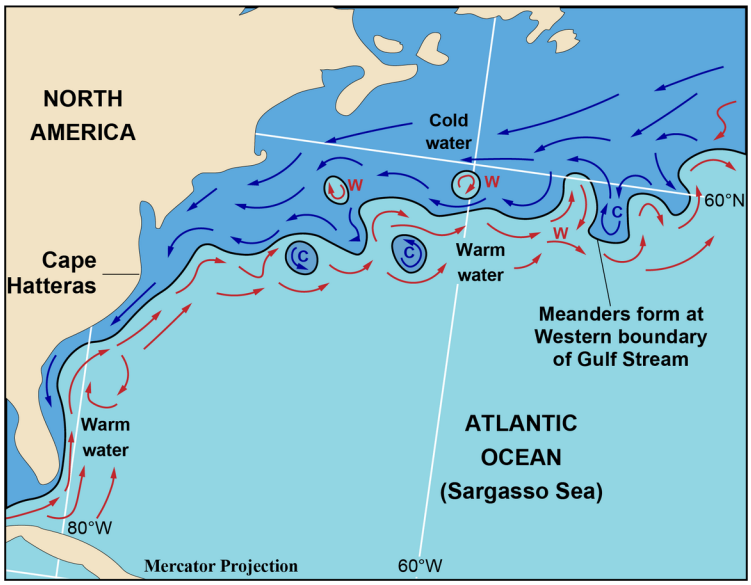


Figure 8-17. Gulf Stream rings. As the Gulf Stream passes north of Cape Hatteras and no longer flows along the edge of the continental shelf, meanders form. The meanders become more extreme as they move north and may eventually be pinched off to form warm- or cold-core rings. The rings generally drift south and dissipate or are reabsorbed by the Gulf Stream. Meander and ring formation is variable and complex, as shown in **Figure 8-16**.

The rings are from 100 to 300 km across and can extend to the seafloor. They are encircled by swiftly flowing currents that move at approximately 90 cm•s^{-1}. Both cold- and warm-core rings drift southeastward and slowly disappear as the core water temperature changes to match that of the surrounding water. Many rings are reattached to, and reabsorbed by, the Gulf Stream. Cold-core rings can maintain their integrity for up to several months depending on their direction of travel, but warm-core rings do not generally last as long, because of the shallower water and narrow shelf they travel over.

Gulf Stream rings transport heat, dissolved substances (including nutrients), and marine organisms that are weak swimmers or that prefer the warmer or colder water. Specific locations within the complex frontal system, with its associated warm- and cold-water rings, are better than others for certain fish **species**. Hence, the location of the Gulf Stream and its rings is important to fishers. This information is also important to ships if they are to minimize fuel costs by taking advantage of, or not fighting, ocean currents. The position of the Gulf Stream front and its rings is now monitored and forecasted. Many oceangoing fishing and other vessels obtain frequent up-to-date satellite images of the Gulf Stream region while at sea, just as they receive weather forecast maps.

Swiftly flowing meanders and rings are also present in other western boundary currents, such as the Kuroshio Current and eddies similar to Gulf Stream rings are present throughout the oceans. The general term for these eddies, including Gulf Stream Rings is “mesoscale eddies.” Most mesoscale eddies are less well defined than Gulf Stream rings, and their current speeds are generally lower. Mesoscale eddies, which in some cases extend all the way to the deep-sea floor, are the ocean equivalent of atmospheric high- and low-pressure zones. They range in diameter from 25 to 200 km, drift a few kilometers per day, and generally have rotating currents of between about 10 cm•s^{-1} and the 90 cm•s^{-1} of the Gulf Stream ring currents. In comparison, atmospheric depressions are about 1000 km across, travel about 1000 km per day, and sustain rotating winds of up to approximately 20 m•s^{-1}.

Mesoscale eddies transport and distribute heat and dissolved substances within the oceans. One area off the tip of South Africa has been well studied because numerous eddies are created at the location where the south-moving Agulhas Current meets the subtropical convergence and turns eastward in the Indian Ocean subtropical gyre. In this region, numerous large eddies are formed (**Fig 8-18**), and some of the eddies form warm core rings that spin off into the South Atlantic Ocean, transferring heat and water from the Indian Ocean into the Atlantic Ocean. These eddies travel across the South Atlantic Ocean and to the north and can persist for 3 to 4 years. This transfer of water and heat between oceans is an essential part of the global ocean circulation called the Meridional Overturning Circulation (MOC), which is discussed later in this chapter.

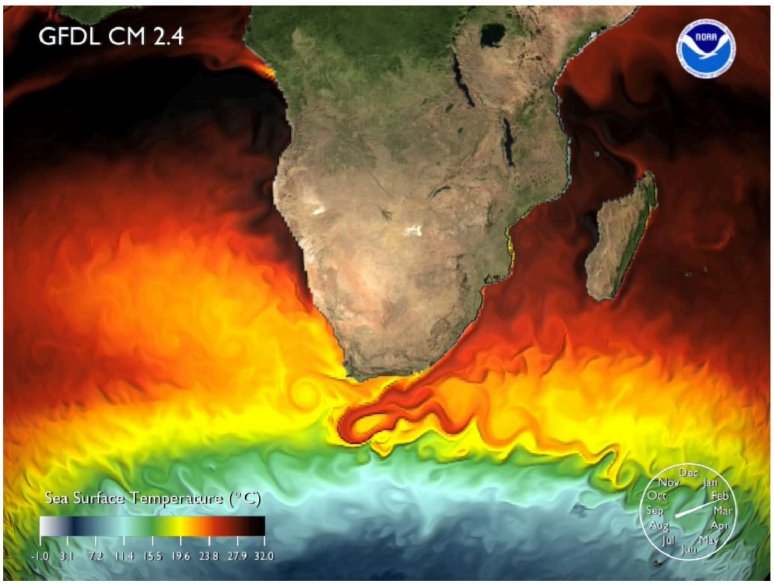


Figure 8-18. Meanders and mesoscale eddies formed at the southern end of the Aghulhas Current. Eddies spin off to the west into the Atlantic Ocean and transport water and heat from the Indian Ocean surface waters to the Atlantic Ocean surface waters. These eddies, similar to Gulf Stream rings, travel to the northwest and persist for several years before being fully mixed with Atlantic Ocean water.

Mesoscale eddies are the subject of considerable ongoing research efforts. However, mesoscale eddies are smaller and slower-moving than atmospheric eddies, and ocean currents are difficult to measure, especially below the surface layer. As a result, the tracking, understanding, and forecasting of ocean "weather," particularly below the surface layers, will likely always be much more difficult than the observation and forecasting of atmospheric weather.

This page titled 8.10: Eddies is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.11: Inertial Currents

Once established by the wind, currents will continue to flow even after the wind stops, until the momentum built up by the wind is dissipated by friction. If there is no other wind or pressure gradient force on the water after the wind stops, the water will flow in a curved path under the influence of the Coriolis effect. If the current is limited to a narrow range of latitude within which the Coriolis effect does not change significantly in magnitude (**CC12**), the water flows in circles. The circular currents are called "inertial currents." Although inertial currents flow even after wind has stopped, they are ultimately wind-driven currents because winds are responsible for starting the water in motion and thus provide the momentum needed to drive the currents.

Inertial currents occur frequently throughout the oceans, but the water rarely moves in perfect circular patterns. In **Figure 8-19**, for example, a circular inertial current is superimposed on a mean drift to the north-northwest. The inertial current and mean wind-driven current were generated by winds blowing at different times, different velocities, and probably different locations. Studies of ocean currents are always complicated by the simultaneous presence of wind-driven and inertial currents, which are caused by different wind events but combine to form the current actually measured at any given place and time. In fact, measured currents are even more complex because geostrophic currents, tidal currents (**Chap. 10**), and wave-induced longshore currents (**Chap. 9**) are also combined with inertial and other wind-driven currents.

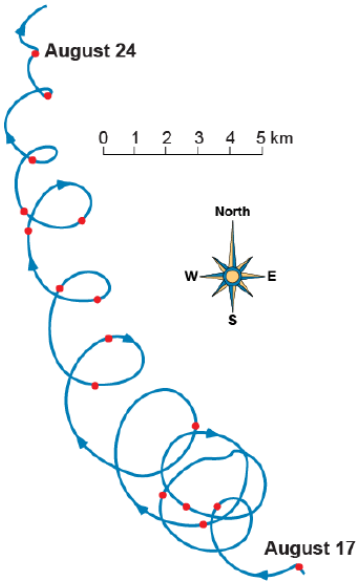


Figure 8-19. Inertial current in the Baltic Sea tracked by a surface float. The inertial motion is superimposed on a mean drift toward the north-northwest. The red dots indicate the float's position every 12 h.

This page titled 8.11: Inertial Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.12: Langmuir Circulation

In addition to currents generated by Ekman transport, winds cause a completely different type of near-surface motion in which the upper few meters of water flow in corkscrew-like motions. These motions, called **Langmuir circulation**, are generated when the wind blows across the sea surface at speeds in excess of a few kilometers per hour.

In Langmuir circulation, water travels in helical vortices aligned in the direction of the wind (**Fig. 8-20**). The helical vortices are stacked side by side and rotate in alternating directions like convection cells (**CC3**). Within each cell, water moves across the surface, sinks at a line of downwelling between two cells, moves back under the surface, and rises at a line of upwelling between two cells, while still moving in the wind direction (**Fig. 8-20a**).

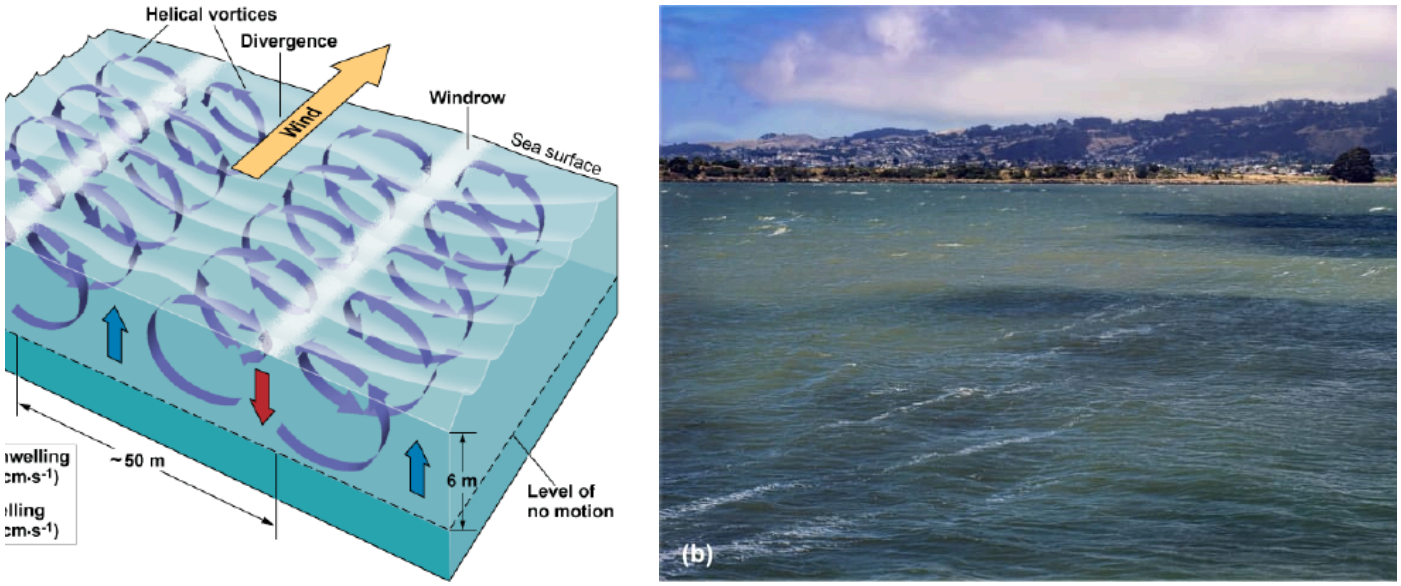


Figure 8-20. Langmuir circulation. (a) Moderate but persistent winds produce corkscrew-like motion of the upper few meters of water. Long linear cells of this helical motion line up alongside each other in the direction of the wind. Adjacent cells rotate in opposite directions, and alternating convergences and divergences form between them. (b) Convergences between Langmuir circulation cells are visible in this photo as the windrows of foam that collect at the convergences when the wind is blowing strongly.

The width and depth of the Langmuir cells are determined by the wind speed and may be limited by shallow water or a shallow pycnocline. Typical cells are a few meters deep and a few meters to about 30 m wide. The top of the downwelling zone between two Langmuir cells is a sea surface convergence. Any oil, debris, or foam that is floating on the surface when Langmuir circulation is established is transported by the circulation to a convergence. The convergence zones are readily observable in lakes or oceans as the linear or nearly linear windrows of floating material that form when strong winds blow (**Fig. 8-20b**). The distance between two of the parallel windrows is equal to twice the width of each Langmuir cell.

Langmuir circulation, together with wave action, is known to be very important in mixing the upper layers of water—a key process in the transport of dissolved gases and heat between the ocean and atmosphere (**Chap. 7**).

This page titled 8.12: Langmuir Circulation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.13: Thermohaline Circulation

Wind-driven currents dominate water motions in the upper layer of the oceans above the pycnocline. Below the pycnocline, currents are driven primarily by density differences between water masses. Density differences cause water masses to sink or rise to the appropriate density level (**CC1**). In areas where water masses sink or rise, the density distribution with depth, and hence the pressure distribution, is different from that in surrounding areas. Thus, horizontal pressure gradients are formed. Once a water mass has moved vertically to reach its density equilibrium level (**CC1**), it flows horizontally in response to the gradient. Because seawater density is determined primarily by temperature and salinity, these water movements are called "thermohaline circulation."

Thermohaline circulation is difficult to study. Our knowledge about this circulation, especially the details of the complex flow patterns in the deepest water layers and the mechanism and geographical distribution of the upward mixing of water that returns deep water to the surface, is far from complete. Most of our information about these comes from studies of density and other characteristics of the deep-ocean water masses and from models (**CC10**). However, the models are themselves limited by the relatively small amount of data that is available to calibrate and test them.

Depth Distribution of Temperature and Salinity

Ocean waters are arranged in a series of horizontal layers of increasing density from the surface to the ocean floor (**CC1**). Throughout most of the oceans, the layers form three principal depth zones: the surface zone (usually called the **mixed layer**), the pycnocline zone, and the deep zone (Fig. 8-5). The surface zone, which is approximately 100 m thick, is usually of uniform or nearly uniform density. However, a seasonal pycnocline is present in the surface zone in many mid-latitude areas (Fig. 8-21). Water in the surface layer above the pycnocline, or above the seasonal pycnocline if it is present, is continuously mixed or stirred by winds. In the permanent pycnocline zone, water density increases rapidly with depth. This zone extends from the bottom of the surface layer, where water temperature is approximately 10°C, to depths that vary by location between about 500 and 1000 m. In the deep layer below the permanent pycnocline, water density increases slowly with depth.

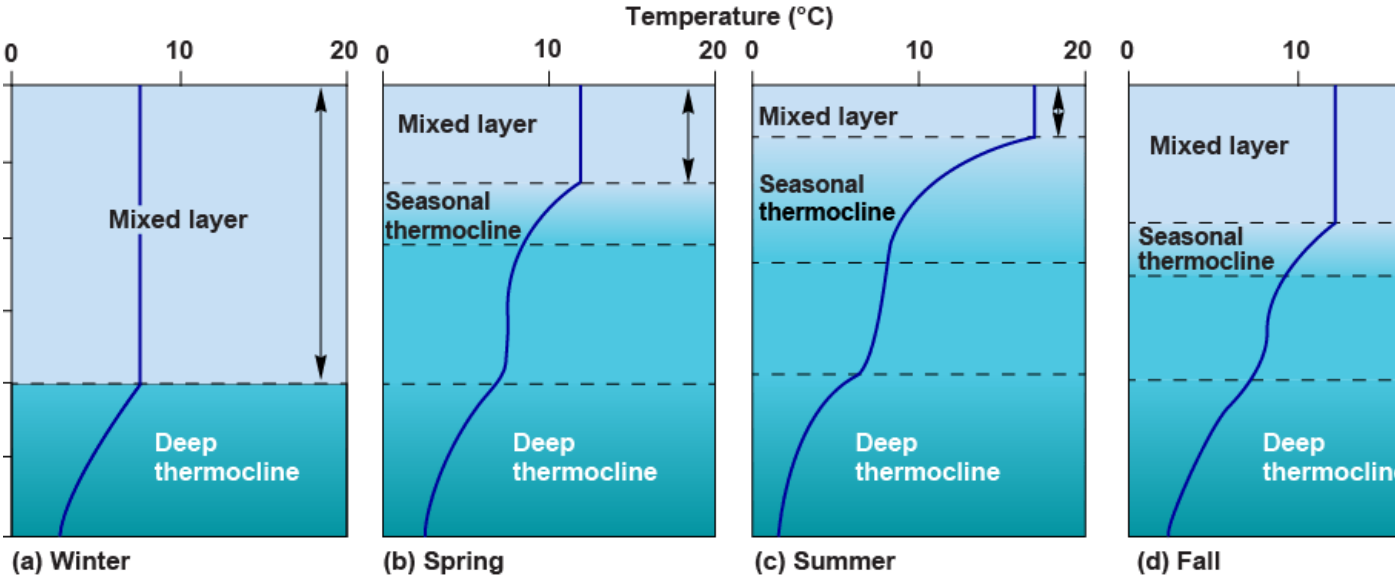


Figure 8-21. In many subtropical and temperate regions, a shallow summer thermocline forms, acting as a barrier to vertical mixing. (a) In winter, the water column is well mixed by the strong winter winds and by convection (due to cooling of surface water) down to the top of the permanent thermocline. (b) In spring, the surface layer is warmed and becomes less dense, forming a shallow thermocline whose depth is determined by wind-mixing depth. (c) As surface water becomes warmer and winds lighter in summer, the seasonal thermocline becomes stronger and shallower. (d) During fall, increased winds increase the depth of the mixed layer, and cooling of the surface water causes convective mixing, which eventually completely destroys the seasonal thermocline.

The marked density increase with depth in the pycnocline zone is due to decreasing temperature, increasing salinity, or a combination of temperature and salinity changes (Fig. 8-5, **CC6**). Where temperature changes cause density to change with

depth, the pycnocline is also a thermocline (**Fig. 8-5b**). Where salinity changes cause density to change with depth, the pycnocline is also a halocline (**Fig. 8-5c**). Haloclines are more important in nearshore waters where freshwater runoff produces surface waters of low salinity.

Figure 8-22 and **Figure 8-23** show the vertical distributions of temperature and salinity, respectively, in the centers of the Atlantic, Pacific, and Indian Oceans. Throughout much of the area between 45°N and 45°S, surface water salinity is actually higher than that of deep waters (**CC6**). The reason is that evaporation exceeds precipitation throughout most of the open oceans in mid- and subtropical latitudes (**Fig. 7-16**). If the water temperature were uniform with depth, the surface waters would be denser than those below and would sink. However, the temperature difference between surface and deep waters is more than enough to offset the density difference due to salinity.

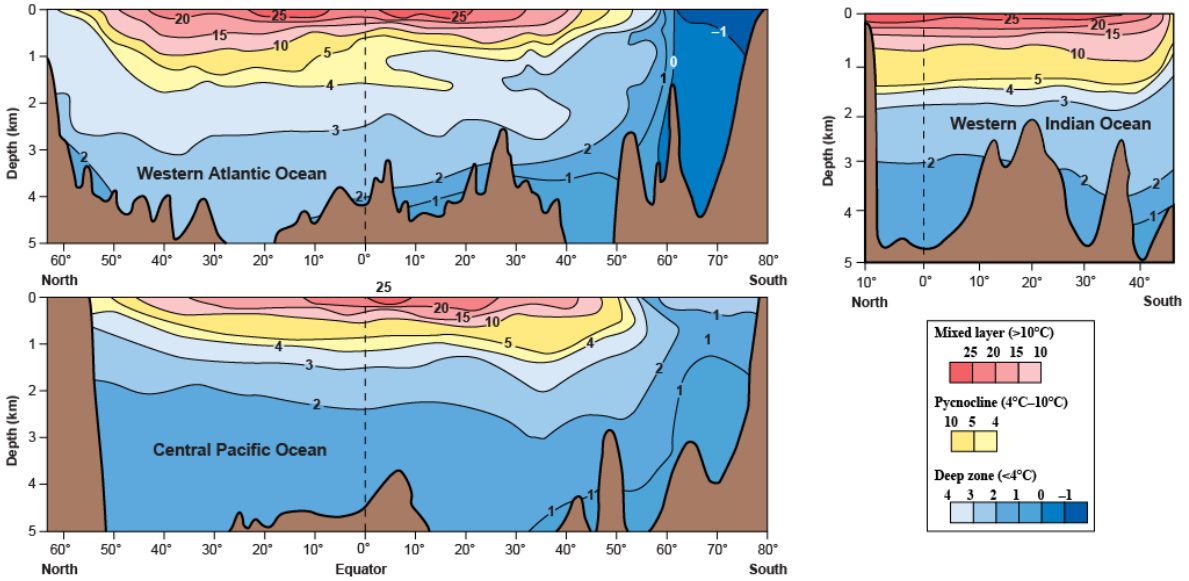


Figure 8-22. North-south profiles showing the distribution of temperature with depth in the open oceans. There are three distinct zones arranged by depth. In the deep zone below the pycnocline, water is cold (below 4°C). Except in high latitudes, there is a warm (>10°C) surface layer that is generally less than 1 km deep. Between the surface and deep zones, the thermocline zone is a region in which temperature decreases rapidly with depth. Around Antarctica, in the North Atlantic Ocean, and to a lesser extent, in the North Pacific Ocean, the warm surface layer and thermocline layer are absent because surface waters in these regions are cooled and sink to form the deep water masses.

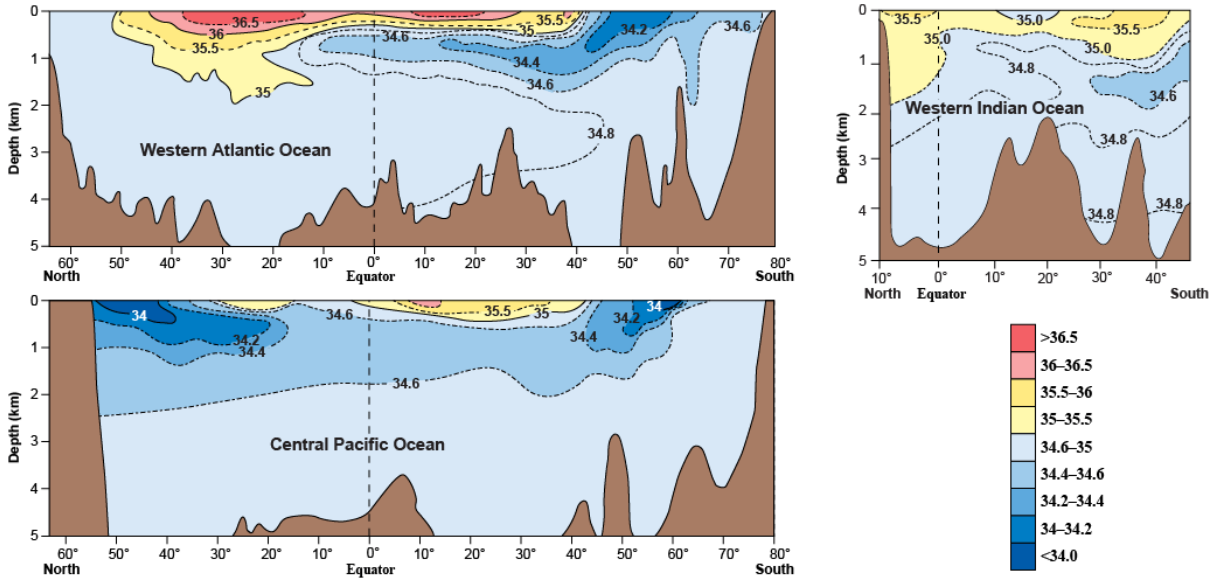


Figure 8-23. North-south profiles showing the distribution of salinity with depth in the deep oceans. This distribution is more complex than the temperature distribution. The surface layer has more variable and generally slightly higher salinity than the deeper water masses, especially in the Atlantic. Salinity is low in the Pacific high latitudes, where precipitation is high, and evaporation is low (dark blue). The influence of the outflow of the warm, high-salinity water mass from the Mediterranean is evident in the tongue of higher-salinity water below 1 km depth between 20°N and 40°N in the Atlantic Ocean.

Within the mixed layer, shallow secondary thermoclines may form during summer in areas where intense solar heating warms surface water, winds are light, and vertical mixing is limited to shallow depths (**Fig. 8-21c**). Seasonal thermoclines are important to biological processes, particularly in coastal waters (**Chap. 13**).

The pycnocline acts as an effective barrier to vertical mixing of water masses. Where density changes rapidly with depth, large amounts of energy are needed to move water either up or down across the density gradient (**CC1**). Consequently, seasonal changes in temperature and salinity caused by changes in solar intensity and rainfall rates (**Chap. 7**) do not generally penetrate below the mixed layer. The pycnocline also acts as a barrier to vertical mixing of dissolved gases and other chemicals. Therefore, deep water is effectively isolated from the mixed layer and atmosphere.

In high latitudes, especially in the northern Atlantic, there is no pycnocline because heat lost from the oceans exceeds heat gained from solar radiation (**Fig. 7-8**). Consequently, surface waters are cooled and their density increases. If cooling is intense enough, the cooled surface water becomes more dense than the water below and sinks. Thus, in high-latitude regions without a pycnocline, deep-ocean water masses are formed as cooled surface waters sink.

The pycnocline layer and deep layer extend throughout each ocean in all but high latitudes. The pycnocline layer is shallower at the equator as a result of upwelling there, and is deeper in some areas, particularly the North Atlantic, and in some areas where complicated tongues of water occur, particularly at depths within and just below the pycnocline zone. The tongues are different water masses because they have their own characteristic temperature and salinity, and they indicate the presence of deep-ocean currents.

Formation of Water Masses

Water density can be increased by decreasing temperature or increasing salinity (**CC6**). Salinity can be increased by evaporation or by ice exclusion (dissolved salts remain in solution in the water left behind as seawater freezes and forms ice crystals). Higher-density water sinks to the depth at which the water below has higher density, and the water above has lower density (**CC1**). It then spreads laterally to form a thin layer extending out from the source area (**Fig. 8-24**).

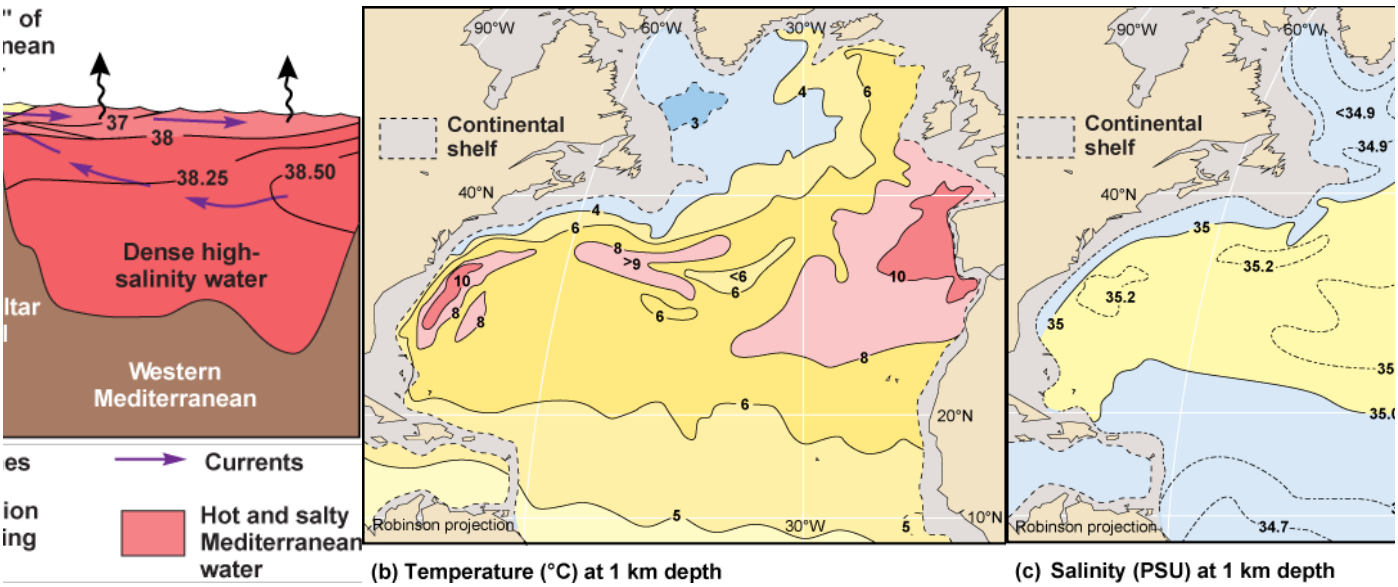


Figure 8-24. (a) The dry, sunny climate in the Mediterranean causes warming and an excess of evaporation over precipitation. The warm, high-salinity water created there spills over the sill at the Straits of Gibraltar into the Atlantic Ocean, where it sinks to its density-equilibrium level at about 1–2 km and spreads out as a distinctive layer. The (b) temperature and (c) salinity characteristics of the Mediterranean water can be detected throughout much of the North Atlantic Ocean.

Water within each such layer is a separate water mass within which salinity, temperature, and therefore density vary only slightly from those of the source water. Water masses can spread horizontally over large areas because the density differences between them inhibit vertical movement and mixing. Less energy is needed for the water mass to flow horizontally than for it to flow or mix vertically because, in a stably **stratified** water column, vertical movements of water must overcome **gravity**. Vertical mixing between water masses of adjacent layers does occur, but the process is generally extremely slow.

Locations of Deep Water-Mass Formation

The densest water masses are created by cooling or freezing of surface waters in only a few locations at high latitudes (**Fig. 8-25**). The densest water is formed by ice exclusion and cooling in the Weddell Sea, a bay on the Antarctic continent opposite the south end of the Atlantic Ocean. This cold, high-salinity water mass, called "Antarctic Bottom Water," sinks to the deep-ocean floor and is transported eastward around Antarctica. As it sinks, it is partially mixed with other water masses and then moves northward along the ocean bottom into each of the three major ocean basins.

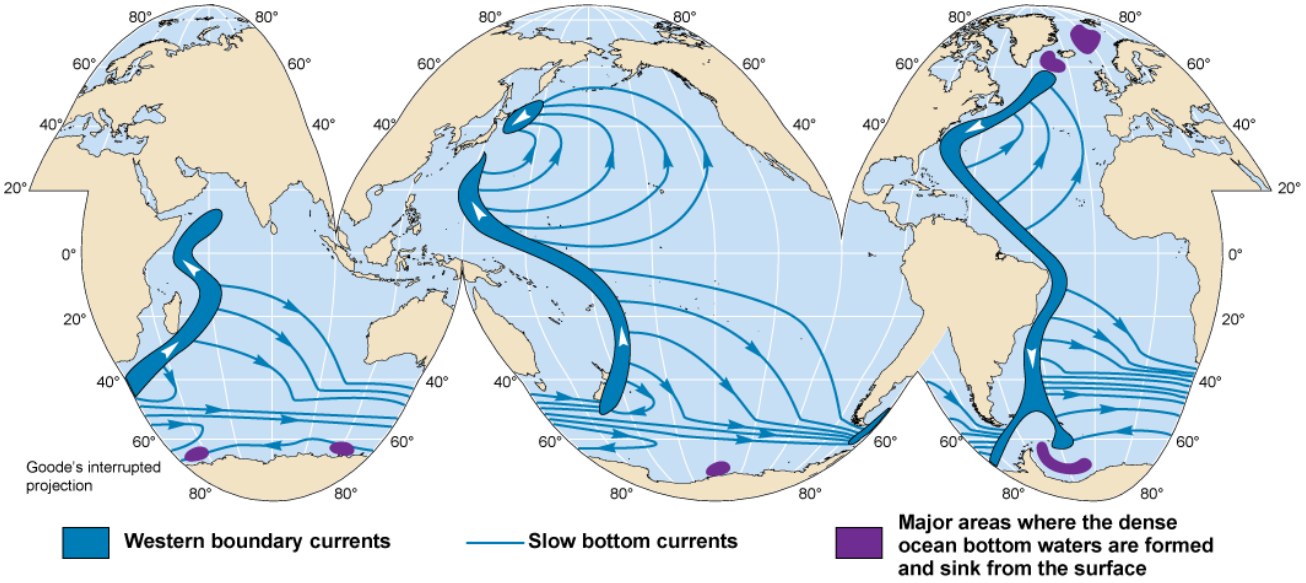


Figure 8-25. The densest water masses that flow along the ocean floor are created by cooling at only a very few locations. Deep water-mass circulation is still not well understood, so this map is only a very general representation of the bottom water flow from each ocean basin through the others. The movement of bottom water within each ocean basin is now known to be far more complex than in this simple depiction. The densest bottom water is formed in the Weddell Sea near Antarctica. It sinks, then flows around Antarctica and northward into each of the oceans. Cold, dense water is also formed near Greenland and in the Norwegian Sea. It flows south until it meets and flows over the more dense Antarctic Bottom Water (Figure 8-29). Bottom currents in each ocean basin are modified by topography, which causes eddy formation. Deep water currents are thought to be weakly intensified on the western boundaries of the oceans (exaggerated in this figure). The western boundary current in the Atlantic Ocean flows southward from the zone of deep water-mass formation near Greenland, and the western boundary currents in the Pacific and Indian oceans flow northward from the Southern Ocean around Antarctica. Return flows of bottom water may exist in some ocean basins but where they exist, they are more complex and diffuse than portrayed here.

No bottom water is formed at the northern end of the Indian or Pacific Oceans. The Indian Ocean does not extend into high latitudes north of the equator. The Pacific Ocean is effectively separated from polar regions by the shallow **sills** between the Aleutian Islands, which mark the southern boundary of the Bering Sea, and by the shallow, narrow Bering Strait that connects the Arctic Ocean and Bering Sea. Cold, dense water formed in the Arctic Ocean sinks and cannot flow into the Pacific Ocean over these shallow areas. In the North Pacific Ocean itself, precipitation rates are high, and low surface salinity prevents the formation of deep water.

The Atlantic Ocean is partially isolated from the Arctic Ocean by shallow ridges between Scotland and Greenland, and the coldest Arctic Ocean deep water cannot readily enter the Atlantic Ocean. However, in the Norwegian and Greenland Sea, intensive cooling forms North Atlantic Deep Water (NADW), which sinks and flows south in vast quantities. It is probably the most voluminous water mass, and its presence can be traced throughout much of the world's oceans. The Atlantic Ocean has water of higher average salinity than any other ocean for two reasons. First, much of the water evaporated in the tropical Atlantic Ocean is transported across the narrow neck of Central America and deposited as precipitation in the Pacific Ocean. Second, large volumes of high-salinity water are introduced from the Mediterranean Sea (which lies in an atmospheric divergence zone with low rainfall and evaporation exceeding precipitation; **Fig. 8-24**).

The higher salinity and temperature of NADW make it readily distinguishable from Antarctic Bottom Water. Because it is less dense than Antarctic Bottom Water and, therefore, southward-moving NADW flows over the northward-moving Antarctic Bottom Water (**Fig. 8-26**). Although the southward flow of NADW is intensified toward the western boundary, detailed studies show that the southward flow takes place in several broad tongues across most of the width of the Atlantic Ocean, and any return gyral flows in the North Atlantic, such as shown in **Figure 8-25**, are minimal at most.

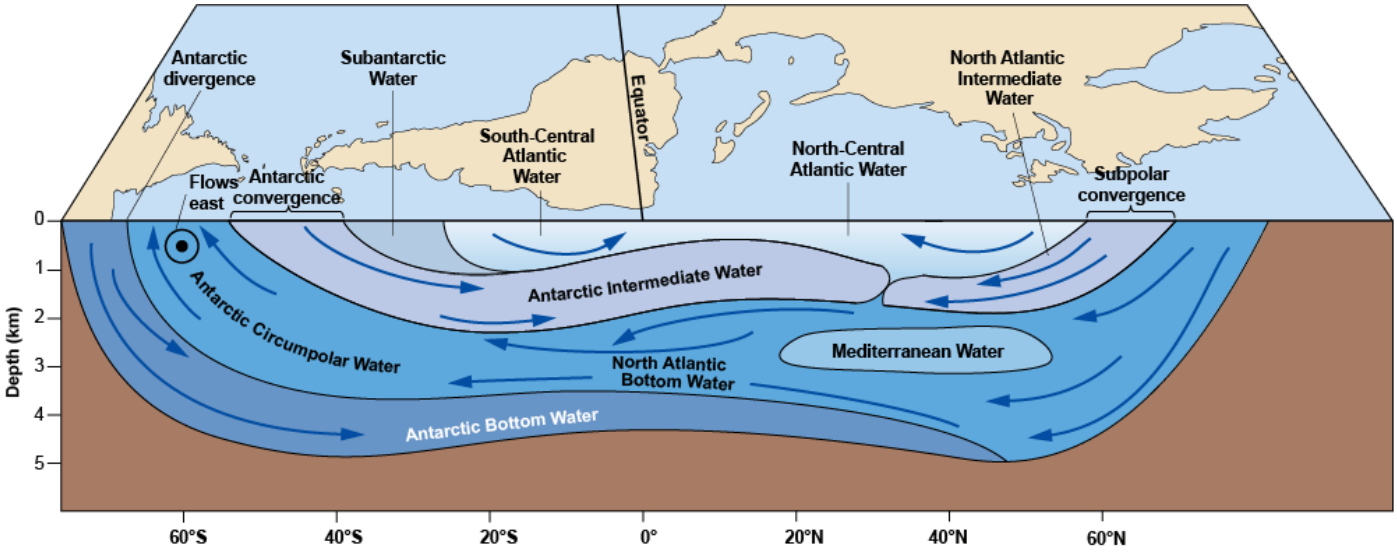


Figure 8-26. A vertical cross-section of the Atlantic Ocean shows the various water masses that form layers at different depths. Antarctic Bottom Water is the densest water mass, and it flows northward from around Antarctica. North Atlantic Bottom Water sinks near Greenland and flows southward over the top of Antarctic Bottom Water until it reaches the surface again in the southern Ocean, where it then mixes with other water masses to form the Antarctic Circumpolar Current. Intermediate-depth water masses are formed and sink at the Antarctic and subpolar convergences. The near-surface layers are more complex. Note the tongue of Mediterranean Water that spreads across the North Atlantic Ocean from the Straits of Gibraltar at about 1 to 3 km depth between 20°N and 55°N.

Most of the Arctic Ocean is covered for much of the year by a floating ice sheet, although the area and thickness of this is rapidly declining (**Fig. 8-27**). Beneath the ice, surface waters have low salinity and are separated from the deep waters by an intense halocline. The low salinity is due partly to river runoff into the Arctic Ocean and partly to ice exclusion. Salt is excluded and transported below the pycnocline during freeze-up in **brines** left by the freezing, but it is not returned in summer when seasonal ice melts. Sea ice contains very little salt, and melting of seasonal ice adds about the same volume of freshwater to the Arctic Ocean surface layer as river runoff does.

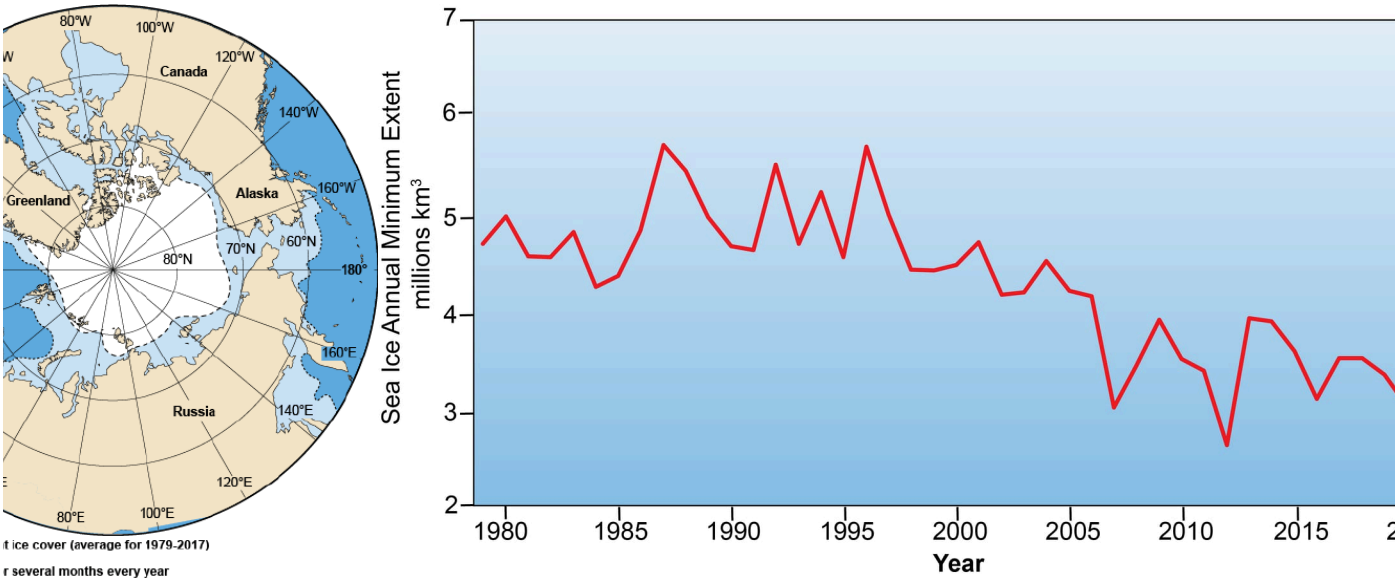


Figure 8-27. (a) Part of the Arctic Ocean is covered by a permanent floating ice pack. Until recent years, only the areas around the edges became ice-free for a few weeks each year. The areas covered by seasonal ice in winter extend far south into the Bering Sea and the Sea of Okhotsk between Russia and Japan, and into the North Atlantic near Greenland and eastern Canada. Most northern European seas are kept ice-free year-round as a result of the warming influence of Gulf Stream water transported into the area. (b) Area of sea ice at its seasonal lowest in September. The area of this permanent ice pack, and of the seasonal ice pack, is becoming smaller from year to year, especially since 2005, and this is believed to be due to the enhanced greenhouse effect. The Arctic Ocean is expected to be ice free in the summer before 2050.

High-salinity brine is also formed by ice exclusion during winter freezing in the continental shelf regions that surround Antarctica. This brine sinks rapidly through the halocline without mixing effectively with the lower-salinity surface water through which it passes. Brine may collect in pools on the continental shelf floor and then drain down the slope. Once in the deep basin below the halocline, the brine eventually mixes with the deep water, most of which enters the Southern Ocean originally from the Atlantic Ocean, and the mixed water mass becomes Antarctic Bottom Water. Most Antarctic Bottom Water is formed in the Weddell Sea section of the Antarctic coast, located south of the southwest corner of the Atlantic Ocean near the tip of South America. Lesser contributions are made at several other locations around Antarctica (**Fig 8-25**).

Deep water-mass movements are affected by the Earth's rotation and the Coriolis effect in the same way that surface currents are. Therefore, deep currents may be intensified along western boundaries of the oceans and may tend to flow in gyre-like motions within the northern and southern ocean basins (**Fig. 8-25**). Eastern boundaries generally have a weak, diffuse flow in the direction opposite to the western boundary flow (**Fig. 8-25**). In addition, deep currents are affected by oceanic ridges and other bottom topography as the flow over and around such topography creates eddy motions.

Water Masses at Intermediate Depths

Intermediate water masses are formed at areas of convergence. For example, sinking of cool, higher-density water at the Antarctic Convergence, and sinking of warm but high-salinity and slightly higher-density water at subtropical convergences, where evaporation exceeds precipitation (**Fig. 8-25**) and form water masses at intermediate depths. Additionally, warm, high-salinity water formed in the Mediterranean Sea, where evaporation exceeds precipitation, is discharged into the Atlantic Ocean, where it sinks to intermediate depths (**Figs. 8-24, 8-26**).

If water continuously sinks to form deep water masses, deep water must be displaced, warmed, and returned to the surface. Cold, deep water is mixed with warmer, less dense water from above by slow vertical mixing between water masses and by more turbulent mixing induced by currents, **internal waves**, and tides. Tides create oscillating currents in the deep waters. Where these currents flow over oceanic ridges or other seafloor topography, eddy formation and turbulent mixing is enhanced, and internal waves can be generated. Vertical mixing is enhanced both at the locations where tidal currents meet rough seafloor topography and at locations where internal waves break as the seafloor shallows on the continental slope or on oceanic ridges and other seamounts. Thus, vertical mixing of deep water upwards back to the surface layers takes place very slowly by **diffusion**, except where seafloor topography interacts with deep currents and internal waves. The locations where such interaction with topography occur are numerous and located primarily at oceanic ridges, seamounts, and on continental slopes. At these locations is known to be vertically mixed with shallower water masses by turbulence at rates far above that in the open ocean, and this also can occur in pulses. This is most likely the dominant mechanism for returning bottom waters back to the surface layers. Water from intermediate depths may also return to the surface by upwelling at divergences such as the Antarctic Divergence (**Fig. 8-26**).

This page titled 8.13: Thermohaline Circulation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.14: Ocean Circulation and Climate

Circulation of water in the oceans can be likened to a giant conveyor belt. Water is cooled near the poles; transported through the deep oceans, where it mixes with other water; returned to the surface far from where it sank; and warmed and transported back to a location where it cools and enters the cycle again.

The ocean circulation system carries heat from one part of the planet to another and transfers dissolved chemicals between surface and deep-water layers. Consequently, changes in circulation can cause and be caused by changes in climate, and they can cause changes in the distribution of dissolved chemicals that directly affect the biological **communities** of the oceans. Ocean circulation also carries some of the excess carbon dioxide from **fossil fuel** burning into the deep oceans (**CC9**). Therefore, deep-ocean circulation research is of great interest, particularly for global climate change studies.

Meridional Overturning Circulation: The Conveyor Belt

The Meridional Overturning Circulation (MOC), which is often referred to as the "conveyor belt circulation," is a closed-loop circulation of water that travels through all of the world's oceans. The general features of the MOC are depicted in **Figure 8-28**. The complete circuit takes an average of about 1000 years, and the amount of water transported is enormous—about 20 times the combined flow of all the world's rivers. The MOC is usually considered to start in the North Atlantic near Greenland and Iceland, where surface water is cooled by the cold air mass that flows from the Canadian Arctic. The cold, high-density water sinks to form North Atlantic Deep Water (NADW), which then flows south through the deep Atlantic Ocean, around Africa, where it flows over and mixes with the deep water formed near Antarctica. The deep water is transported around Antarctica, then north into the Indian and Pacific Oceans. The water mixes progressively toward the surface. The details of where and how this mixing takes place are not yet well known, but it is thought to take place slowly throughout most of the oceans and more quickly through turbulent mixing in some areas where currents and internal waves interact with seafloor topography. Eventually, the deep water mixes with surface waters and enters the mixed-layer current system. Complex exchanges and movements of surface water eventually return surface water to the Atlantic Ocean, replacing the water that originally formed NADW. This system is instrumental in transporting heat from low latitudes to high latitudes of the North Atlantic Ocean.

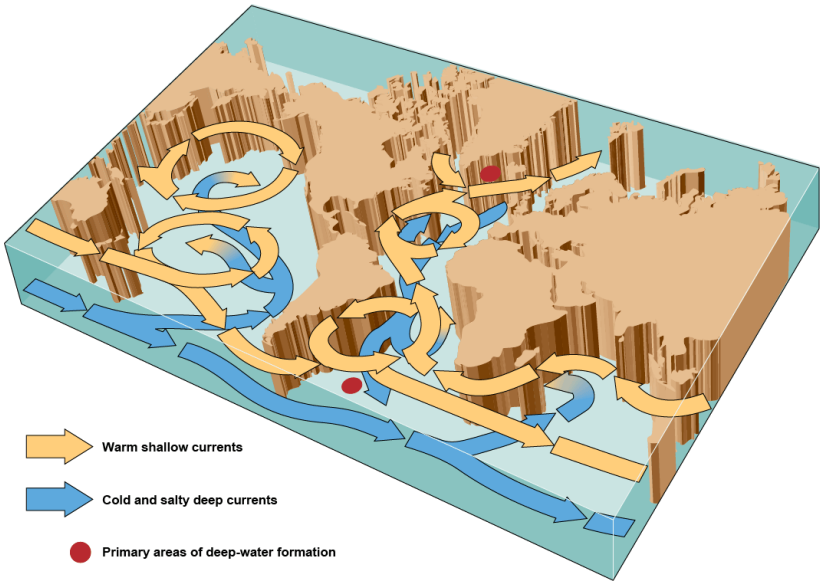


Figure 8-28. A simplified depiction of the Meridional Overturning Circulation (MOC), or "conveyor belt," that starts with the formation of North Atlantic Deep Water in several locations near Greenland.

Surface water in the North Atlantic Ocean is warmer than NADW. As the Gulf Stream flows into the northeastern Atlantic Ocean near Europe, prevailing westerly winds blow over it, transferring heat and moisture into the atmosphere. The westerly wind air mass releases its heat and moisture over Europe, causing Europe's climate to be extremely mild and wet in comparison with the climates of other land areas at the same latitude (**Chap. 7**). Thus, the MOC transfers heat from the Indian Ocean and Pacific Ocean to the North Atlantic region near Europe. At the same time, the MOC transports dissolved nutrients from the Atlantic Ocean

to surface waters of the Indian and Pacific Oceans.

The MOC Climate Switch

Ocean **sediment** records have shown that, during the last **ice age**, which peaked about 18,000 years ago, the MOC generally operated more slowly and weakly than it does today, and varied in strength. The variances may have occurred abruptly at times. Some of those abrupt changes may have almost entirely turned the MOC off or back on again during the past tens of thousands of years. It is believed that such abrupt changes have, at least on some occasions, coincided with very abrupt (on the scale of decades) changes in climate. Evidence suggests that variations in the MOC circulation, at least during the cold, ice-age period, may have been **chaotic** (**CC11**) and that ocean circulation may have switched periodically between a very weak MOC and a strong MOC, which is seen today.

About 15,000 years ago, the last ice age ended, although it may be that the current warm period is just a warm interglacial interval similar to other interglacial periods that have interspersed the ice age in the past. Nevertheless, about 15,000 years ago, the Earth's average temperature increased by about 6°C in as little as 100 years. This abrupt change was accompanied by a 20% increase in atmospheric carbon dioxide concentration and a substantial, but not yet well quantified, increase in the intensity of the MOC.

During the past 13,500 years, the Earth's climate has been warmer than at any other time during the past million or more years. However, this warm period was interrupted by a short cold period, the Younger Dryas period or "Little Ice Age", that occurred between approximately 12,900 years ago and 11,500 years ago. At the beginning of the Younger Dryas period, western Europe's climate cooled dramatically by about 5°C within a matter of decades or less and then returned just as abruptly to its former warm condition at the end of this period. Apparently, the Younger Dryas cold climate occurred when the MOC was severely slowed or stopped. The reason for the temporary slowing or stoppage of the MOC is not known, but might be related to the flow of meltwater from **glaciers**.

During the early postglacial period, about 13,000 years ago, glaciers extended far to the south, and most meltwater from the North American ice sheet probably drained down the Mississippi Valley to the Gulf of Mexico. By 13,000 years ago, as the glaciers continued to melt, a lake called Lake Agassiz covered much of Manitoba, western Ontario, northern Minnesota, eastern North Dakota, and Saskatchewan. At its greatest extent, Lake Agassiz may have covered as much as 440,000 km^2, larger than any currently existing lake in the world. It is believed that the cold, fresh meltwater that was collected in this lake emptied rapidly either through the McKenzie River to the Arctic Ocean or through the Saint Lawrence River to the North Atlantic, or perhaps both. Because it was fresh or very low salinity water, it would have floated on the ocean water to form a stable, low-density surface layer over a large area of the Arctic Ocean and North Atlantic Ocean. The low-salinity surface layer would have acted as a virtual cap, severely restricting the formation of NADW and slowing the MOC. Within several hundred years after glacial meltwater first flooded out of Lake Agassiz in large volumes the freshwater flow to the North Atlantic probably diminished because, by that time, most glaciers had finished melting. At that point, NADW again formed, the MOC resumed, and Europe's climate warmed rapidly.

We are not sure whether a number of additional abrupt climate changes that have occurred during the past 10,000 to 12,000 years coincided with significant changes in the rate of water transport within the MOC, but it appears that at least some such abrupt climate changes did coincide with such changes. We cannot yet determine for certain whether changes in the MOC caused these additional abrupt climate changes, or vice versa.

Recent studies of the MOC suggest that the rate of formation of North Atlantic Deep Water and thus the rate of flow in the MOC exhibits strong short-term variability that makes it very difficult to assess whether there are any long-term trends in its strength. However, some estimates suggest that the MOC may have slowed down, perhaps as much as 30% during the past several decades. This slowdown, if it can be confirmed, may be related to climate change. If so, and if this indicates that the MOC will continue to slow down as the planet warms due to anthropogenic climate change, one possible result could be the abrupt (years to decades) onset of a cold climate period in Europe, similar to the Younger Dryas period, and to other abrupt climate changes elsewhere. However, modeling studies (**CC10**) suggest that climate change is likely to slow the MOC but not to cause it to slow so much that catastrophic changes, including drastic cooling of Europe, will occur within the next century or two.

This page titled 8.14: Ocean Circulation and Climate is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.15: Tracing Water Masses

Although direct measurements have been made of deep-ocean currents, most of what we know about deep-ocean circulation was learned from studies of the horizontal and vertical distribution of water masses. Water masses are most often characterized by their temperature and salinity, but the concentrations of several dissolved constituents are also used to distinguish and trace water masses (**CC6**).

Conservative and Nonconservative Properties

Conservative properties of seawater are properties that are not changed by biological, chemical, or physical processes within the ocean. A conservative property can be changed only by mixing with other water masses that have different values of that property, or by processes that occur at the ocean surface, or where rivers or other sources, such as **hydrothermal vents**, enter the oceans. For example, salinity, a conservative property, can be changed at the ocean surface by evaporation or precipitation, or at river mouths by the introduction of freshwater. Small salinity changes can also occur at hydrothermal vents, where high-salinity waters are emitted. Similarly, temperature is a conservative property until water contacts the atmosphere or seafloor, with which it can exchange heat.

Dissolved constituents of seawater, such as Na^+ and Cl^-, whose concentrations are not significantly affected by biological or chemical uptake or removal from ocean water, are also conservative properties. In contrast, concentrations of constituents such as oxygen, carbon dioxide, phosphate, and many trace metals are substantially altered by biological and chemical processes that remove and release them to solution within the ocean. These are nonconservative properties.

Conservative properties are particularly useful in tracing water masses because they can be used to identify the masses as they are formed, to trace their transport through the oceans, and to determine how they mix with other water masses. When two water masses mix, the value of the conservative property in the mixed water mass is determined by the proportions in which the two water masses have mixed. For example, consider the mixing of two water masses, one with a salinity of 35 and the other with a salinity of 37. If the salinity of the mixed water mass is 36, the mixture must consist of equal volumes of the two water masses (2 volumes of 36 = 1 volume of 37 + 1 volume of 35). If the salinity of the mixture is 35.5, the mixture must consist of three parts water of salinity 35 and 1 part water of salinity 37 (4 volumes of 35.5 = 3 volumes of 35 + 1 volume of 37).

TS Diagrams

We can identify different water masses by plotting temperature against salinity in what are called “TS diagrams” (**Fig. 8-29**). Water samples from the same water mass have the same salinity and temperature and therefore will appear as a single point on a TS diagram. Samples from water masses that have different temperature and salinity appear as separated points. When two water masses are mixed, the temperature and salinity of the mixed water lie along a straight line drawn between the two points representing the original water masses. The location of the TS value for a mixed water sample on the line between the TS values of the two water sources indicates the relative proportions of the two original water masses in the mixture.

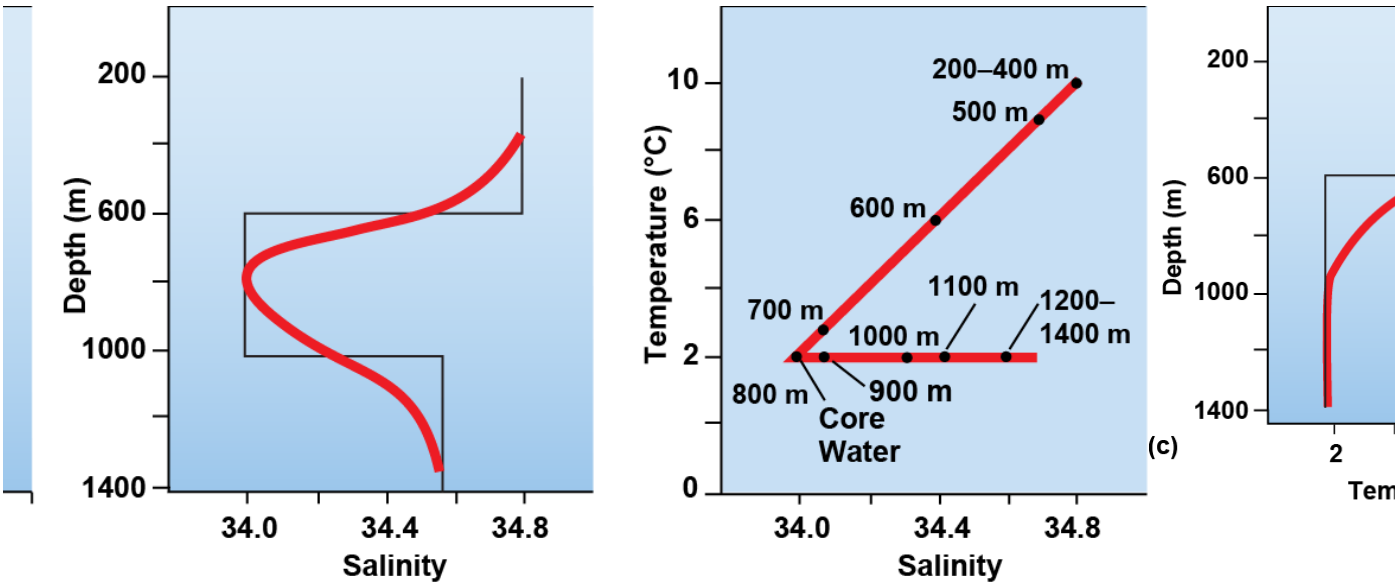


Figure 8-29. TS diagrams are used to trace water masses. In this figure, hypothetical vertical temperature and salinity profiles are used to construct examples of TS diagrams. (a) A simple three-layer system. All TS points plot at one of three locations, representing the three water masses. (b) If the three layers mix, but some of the unmixed mid-depth water mass is still present (at 700 m depth), the TS points plot on two straight lines that connect the three original water-mass characteristic points. (c) Even if all of the mid-depth water mass is mixed with at least some of the other two water masses, a TS plot can be used to identify the temperature and salinity characteristics of the original unmixed water mass. (d) A typical TS diagram for a station in the South Atlantic Ocean.

When three water masses mix in such a way that no unmixed sample of one of the three water masses remains, some of the mixed water samples will have TS values that are no longer on straight lines between the three original TS points (**Fig. 8-29c**). However, in such cases, the temperature and salinity of the original water mass, which now exists only as a part of mixtures, can often be deduced by extrapolation of the straight-line portions of the TS curve (**Fig. 8-29c**).

A TS diagram for water samples taken at different depths at a single station in the South Atlantic Ocean is shown in **Figure 8-29d**. At this station, the principal water masses are warm, saline surface water; North Atlantic Central Water; water that is a mixture of surface water and North Atlantic Deep Water; Antarctic Intermediate Water; North Atlantic Deep Water; and Antarctic Bottom Water. We need not be concerned with how these water masses acquired their different temperature and salinity characteristics. What is important is that, once the source properties are known, the presence of a particular water mass in the area for which the TS diagram is drawn can be determined. Note that we can detect the presence of Antarctic Intermediate Water in the diagram, even though no unmixed Antarctic Intermediate Water is present at the station. We can also calculate from the diagram the percentage of Antarctic Intermediate Water present in the mixed water at any depth. For example, at 800 m, where the proportion of this water mass is greatest, Antarctic Intermediate Water is about 55% of the mixture.

By drawing TS diagrams for many different stations across the oceans, oceanographers can trace movements of individual water masses and investigate mixing processes at different points within the oceans.

Tracers

Salinity and temperature are excellent tracers, but they tell us only which water masses are mixed and where the mixed water is transported. Without other information, these tracers cannot tell us how fast the water masses move and mix. Several different dissolved components of seawater are now used in conjunction with temperature and salinity to trace ocean water masses and to provide information on their rate of movement.

In some applications, nonconservative properties, such as oxygen and carbon dioxide concentrations, can be used as tracers. When water sinks below the pycnocline and is no longer in contact with the atmosphere, the concentrations of these gases are altered by the decomposition of organic particles, which consumes oxygen and releases additional carbon dioxide to solution. Therefore, if we know the oxygen and carbon dioxide concentrations in water masses when they sink below the mixed layer, changes in those concentrations can provide information about the relative ages (since the time when they sank) of water masses in different parts of the oceans.

Human activities have provided a number of useful tracers during the past half century. These tracers include **radionuclides** created by nuclear weapons testing, and certain synthetic organic compounds, particularly **chlorofluorocarbons (CFCs)**. Some radionuclides that do not occur naturally are now found dissolved in ocean water in extremely small concentrations. They are derived primarily from nuclear bomb tests in the atmosphere, which peaked in the 1950s. These radionuclides are good tracers because the time and location of their introduction to the oceans is well known, and they decay at known, constant rates (**CC7**).

Tritium (3H), a **radioactive isotope** of hydrogen, is a particularly useful tracer. Tritium is produced naturally by cosmic rays in the atmosphere. Thus, it occurs naturally in ocean water, but only at extremely small concentrations. In water unaffected by human releases, approximately one of every 1019 atoms of hydrogen is a tritium atom. Releases of tritium during the nuclear bomb testing era of the mid-twentieth century raised the tritium concentration in surface seawater to a concentration of approximately 6 atoms of tritium in 1018 atoms of hydrogen. Tritium reacts quickly with oxygen in the atmosphere to form tritiated water, which enters the oceans in rainfall. Because the tritium is incorporated in the water molecule itself, it is a perfect tracer for water masses. Tritium has a **half-life** of only approximately 12 years, so bomb-produced tritium will continue to be useful as a tracer for only a few decades, until all the tritium introduced by nuclear bomb testing has decayed.

In the Northern Hemisphere, most tritium was produced by nuclear tests in the Soviet Arctic. Therefore, tritium concentrations are particularly high in water flowing into the Atlantic Ocean from the Arctic Ocean. This tritium is now distributed throughout the mixed layer of the oceans and is currently being transported into the deep oceans with sinking cold water masses created in polar regions. The progress of deep water-mass formation has been monitored using tritium as a tracer (**Fig. 8-30**).

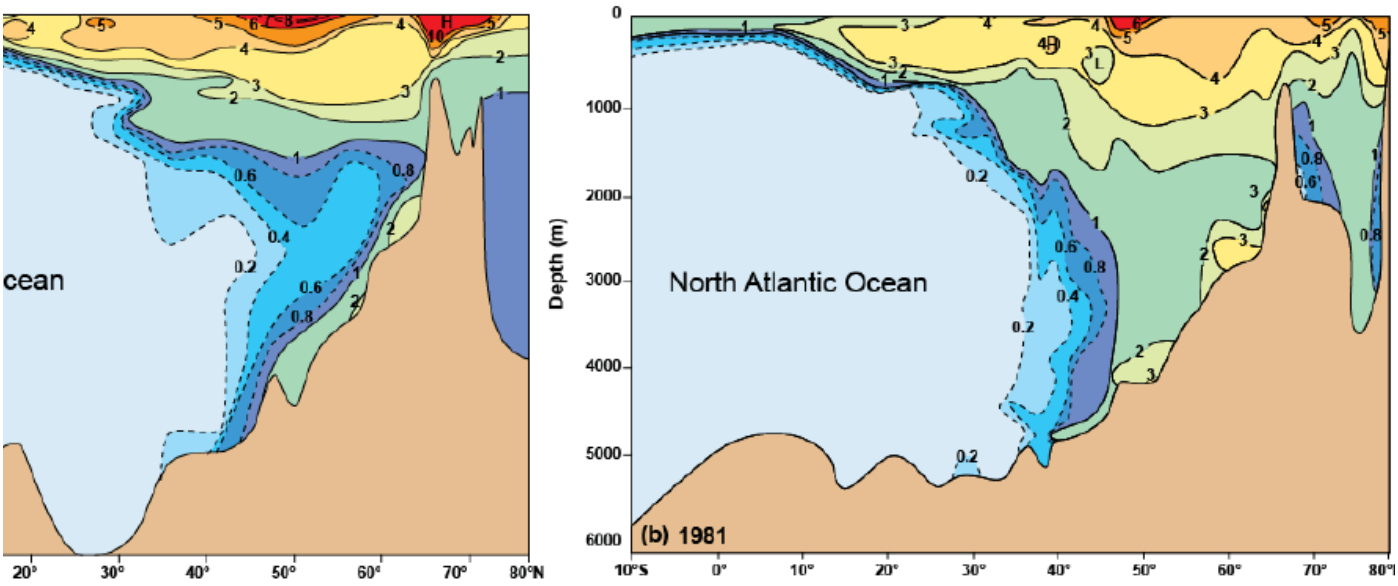


Figure 8-30. (a) Tritium distribution with depth in the western North Atlantic Ocean (1972–1973). Released primarily by nuclear bomb tests in the 1950s, tritium has spread throughout the ocean surface layers and is being transported steadily into the deep oceans with North Atlantic Deep Water, which is formed by the cooling of surface waters in the North Atlantic. (b) Distribution in 1981. The values are corrected for the radioactive decay that occurred during the 9 years between the two surveys. The strong source in the surface layer at 50°N is the outflow from the Arctic Ocean where the Soviet Union did most of its nuclear bomb testing. Comparing the two diagrams shows the progressive movement of tritium into the deep layer as North Atlantic Deep Water continuously forms, sinks, and moves southward.

Chlorofluorocarbons (CFCs), which were used as refrigerants and are thought to be responsible for **ozone** depletion in the atmosphere (**Chap. 7**), are also excellent water-mass tracers (**Fig. 8-31**). These compounds do not occur naturally. They are dissolved at very low concentrations in seawater and are not readily decomposed by biological processes.

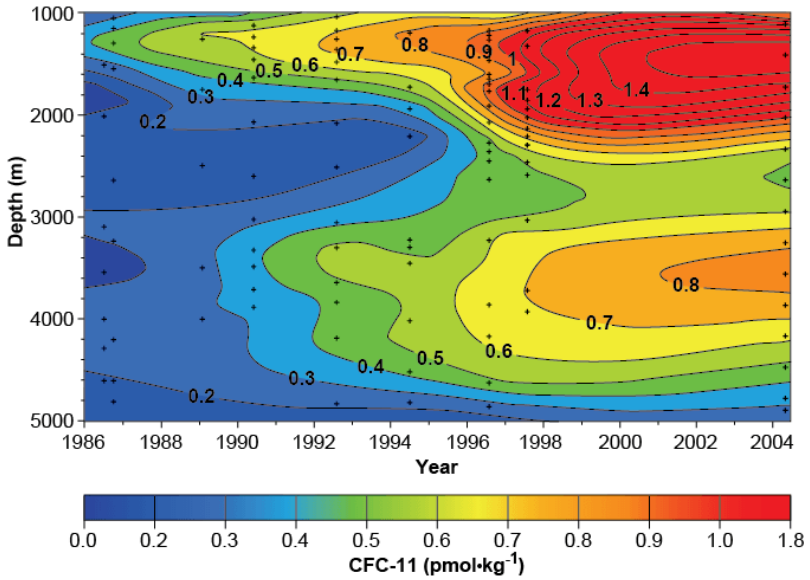


Figure 8-31. The concentrations of CFC-11, a specific chlorofluorocarbon compound, have been measured throughout the water column to the east of Abaco Island in the Bahamas, periodically between 1986 and 2004. This figure plots the vertical distribution of the concentrations of CFC-11 in this area as a function of time. The data show the arrival at this subtropical location of CFC-11 in recently formed deep water masses originating from the high latitude, deep water mass formation areas in the polar and subpolar regions of the North Atlantic (**Figs. 8-26, 8-28**). There are three distinct water layers seen arriving in this area, centered at about 1200 m, 1800 m, and 4000 m. The deepest layer is deep water formed from the outflow of the Arctic Ocean, and the two shallower layers are both deep water formed near Greenland. The 1800 m layer arrived near Abaco later than the shallower layer, and appears to reflect a small reduction in the temperature of deep water formed near Greenland that had been observed since about a decade earlier.

Unfortunately, for most water-mass tracers, **in situ** instrumentation is not sensitive enough, and thus, water samples must be collected for laboratory analysis. Because the cost of ship time required to collect water samples from deep within the water column is very high, sampling cannot be as intensive either spatially or temporally as would be desirable in most instances. Thus, many studies rely on mathematical modeling (**CC10**) using the limited data that are available.

Studying the recent and ongoing processes of deep water mass formation is extremely important to our understanding of the effects of fossil fuel burning on global climate. Estimates suggest that approximately 30% of the carbon dioxide released by burning fossil fuels since the Industrial Revolution in the mid-1800s is now dissolved in ocean water. However, there is still considerable uncertainty about the accuracy of the estimates. Ocean water that sinks below the mixed layer carries carbon dioxide from the atmosphere into the deep oceans, where it is locked away from contact with the atmosphere for hundreds or thousands of years (or longer if it is incorporated in carbonate sediments; **Chap. 6**). Tracer studies will help us to determine how much carbon dioxide from fossil fuels has already been transported safely, at least for now, below the pycnocline and to estimate how quickly additional carbon dioxide will follow it.

This page titled 8.15: Tracing Water Masses is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.16: CHAPTER SUMMARY

Energy Sources.

Winds are the energy source for most currents in the ocean's upper few hundred meters. Deeper currents and most vertical movements of ocean water are caused by thermohaline circulation, which is driven by differences in the density of water masses due to temperature and salinity differences.

Wind-Driven Currents.

Winds blowing across the sea surface transfer kinetic energy to the water through friction and cause wind-driven currents with speeds generally about 1% to 3% of the wind speed. Currents continue after winds abate until their momentum is eventually lost because of internal friction. Currents also continue because winds transport water, cause the sea surface to slope, and create horizontal pressure gradients that drive water motion. Current speed and direction are modified by friction, the Coriolis effect, horizontal pressure gradients, and land and seafloor topography.

Ekman Motion.

Surface water set in motion by winds is deflected *cum sole* relative to the wind direction by the Coriolis effect. Wind energy is transferred down into the water column as each water layer transfers its energy to the layer below. If winds blow consistently for long enough over deep water, an Ekman spiral develops in which the surface current is deflected 45° *cum sole*, current speed decreases with depth, and current direction is progressively deflected *cum sole* with depth. Mean water transport (Ekman transport) in the layer within which the Ekman spiral develops is at 90° *cum sole* to the wind.

Wind-driven currents rarely reach deeper than 100 to 200 m. Where the seafloor is shallower than the Ekman spiral depth, friction reduces the surface water deflection below 45° and the mean transport below 90°. If a pycnocline is present within the Ekman spiral depth, downward transfer of wind energy is inhibited, and wind-driven currents do not occur below the pycnocline.

Geostrophic Currents.

Ekman transport causes surface water to "pile up" in some areas. The resulting sloping sea surface results in a horizontal pressure gradient within the upper several hundred meters. Water flows on such gradients and is deflected by the Coriolis effect. The deflection continues until the flow is along the gradient (parallel to the pressure contours) at a speed where the Coriolis deflection is just matched by the pressure gradient force. This is geostrophic flow.

Ocean Surface Currents.

Trade winds drive surface waters generally west in the Equatorial Currents. Westerly winds drive surface waters generally east in the subtropics. Continents deflect the flow north or south. Thus, current gyres are created between trade wind and westerly wind latitudes in each hemisphere and ocean. These gyres flow clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere. Ekman transport by trade winds and westerlies piles up water in the gyre center, creating horizontal pressure gradients that maintain geostrophic gyre currents even when winds cease. Western boundary currents are faster, deeper, and narrower than eastern boundary currents.

In the high latitudes of the North Atlantic and Pacific Oceans, counterclockwise subpolar gyres are complex, weak, and variable. At high southern latitudes, no continents block the Antarctic Circumpolar Current that flows westward around Antarctica.

Upwelling and Downwelling.

Upwelling occurs where Ekman transport moves surface waters away from a divergence or coast. Downwelling occurs where Ekman transport adds to the surface layer depth at convergences or coasts. Upwelling areas are usually highly productive because

upwelling can bring cold, nutrient-rich water from below into the mixed layer, where nutrients are essential for phytoplankton growth. Divergences and upwelling occur in part of the area between the North and South Pacific Equatorial Currents, around Antarctica, and in many coastal locations. Convergences occur in the centers of ocean gyres and in many coastal regions. Coastal upwelling is more prevalent inshore of eastern boundary currents than inshore of western boundary currents.

Coastal Currents.

Local winds determine coastal current directions and they may be opposite those of the adjacent gyre currents. Coastal currents are variable and are affected by freshwater inflow, friction with the seafloor, steering by seafloor topography and coastline, and tidal currents.

Eddies.

Ocean eddies are similar to, but smaller than, the atmospheric eddies seen on satellite images as swirls of clouds. Satellites can be used to observe some ocean eddies. Gulf Stream rings are mesoscale eddies 100 to 300 km wide that form cold-core (counterclockwise-rotating) or warm-core (clockwise-rotating) rings when a meander of the Gulf Stream breaks off. The rings, which may last several months, may reattach to the Gulf Stream as they drift slowly south. Mesoscale eddies are present throughout the oceans, generally have slower currents than Gulf Stream rings, may extend to the deep-sea floor, and are 25 to 200 km wide.

Inertial Currents.

Once established, wind-driven currents that continue to flow and are deflected into circular paths by the Coriolis effect are called “inertial currents.”

Langmuir Circulation.

Strong winds can set up a corkscrew-like Langmuir circulation aligned in the wind direction. Langmuir cells are a few meters deep and about 30 m wide, and they lie side by side. Foam or floating debris collects in the linear downwelling regions between cells.

Thermohaline Circulation.

When surface water density is increased by cooling or evaporation, the water sinks and spreads horizontally at a depth where the water above has a higher density and the water below has a lower density. Throughout most of the oceans, the water column consists of a uniform-density mixed surface layer (about 100 m deep), a permanent pycnocline zone (extending from the bottom of the mixed layer to a depth of 500 to 1000 m) in which density increases progressively with depth, and a deep zone in which density increases slowly with depth. Most high-latitude regions where surface waters are cooled lack a pycnocline. Secondary, temporary pycnoclines can develop in the mixed layer in summer. Pycnoclines act as a barrier to vertical mixing.

Deep water masses are formed by cooling and ice exclusion at high southern latitudes, in the Arctic Ocean, and at high latitudes in the North Atlantic Ocean. Antarctic Bottom Water is formed primarily in the Weddell Sea and flows north as the deepest and densest water layer. North Atlantic Deep Water is formed in the Greenland and Norwegian Seas and flows south. Water masses at intermediate depths are formed by the sinking of cooled water at the Antarctic Convergence and of warm, high-salinity water produced by evaporation at subtropical convergences and in the Mediterranean Sea. Deep and intermediate-depth water masses are subject to the Coriolis effect, and thus, currents in the deep layer tend to be intensified along western boundaries and tend to flow in gyres. They are also affected by seafloor topography, which generates eddies that enhance vertical mixing.

Ocean Circulation and Climate.

Ocean circulation transfers heat from the tropics toward the poles, moderating mid- and high-latitude climates. North Atlantic Deep Water forms near Greenland and Iceland. It sinks and spreads south in the deep Atlantic and then north into the Pacific and Indian Oceans. It mixes progressively upward to the surface layer, which is warmed and eventually transferred back to the North Atlantic by the Gulf Stream.

This Meridional Overturning Circulation (MOC) has varied in intensity and switched off and on in the past. Periods when it has not operated seem to be associated with colder climates, especially in Europe. The change to a colder climate appears to take place abruptly when the MOC is switched off.

Tracing Water Masses.

Water masses are characterized by their salinity and temperature, which are conservative properties everywhere but at the surface and in some areas of the seafloor. Nonconservative properties such as oxygen concentration, as well as concentrations of human-made radionuclides and persistent organic compounds, are also used as water-mass tracers.

8.16.1: Key Terms

8.16.2: Chapter Questions

8.16.3: References

This page titled 8.16: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.16.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- back-scattered
- brines
- chaotic
- chlorofluorocarbons (CFCs)
- coastal downwelling
- coastal upwelling
- coastline
- communities
- contours
- convection
- convection cells
- convergence
- Coriolis effect
- cum sole
- currents
- density
- diffusion
- divergence
- downwelled
- dynamic height
- eastern boundary currents
- eddies
- Ekman transport
- El Niño
- fossil fuel
- friction
- front
- geostrophic
- gravity
- gyres
- half-life
- hydrothermal vents
- ice age
- in situ

- internal waves
- intertropical convergence zone
- kinetic energy
- Langmuir circulation
- latitudes
- meanders
- mixed layer
- nutrients
- ozone
- phytoplankton
- phytotrophic
- potential energy
- pressure gradient
- radioactive isotope
- radionuclides

- salinity
- sediment
- shore
- species
- stratified
- subtropical gyres
- surface tension
- suspended sediments
- synoptic
- thermohaline circulation
- tides
- topography
- trade wind
- upwelling
- wavelengths
- weather
- westerly
- western boundary currents

This page titled 8.16.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.16.2: Chapter Questions

Study Question

1. What are the major energy sources for the surface currents and the thermohaline circulation of the oceans?
2. Why doesn't a wind-driven current flow in the same direction as the wind that causes it?
3. Wind-driven Ekman transport moves surface water across the oceans. How does this lead to geostrophic currents that continue to flow even if the winds abate? Why don't geostrophic currents flow in the same direction as the Ekman transport?
4. Subtropical gyres are present in each ocean and each hemisphere in the latitudes between the trades and westerlies. Why? Would they exist if the Earth were totally covered by oceans? Why or why not?
5. Why do the subtropical gyres rotate clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere?
6. How are upwelling and downwelling caused in the oceans, and why are they important? Where would you expect to find upwelling and downwelling areas in the oceans? Why?
7. What factors affect coastal currents but not deep-ocean currents?
8. Why are the deepest water masses in the oceans formed at high latitudes? Why are they not formed in the Indian and North Pacific oceans?
9. Why is there no pycnocline in most high-latitude regions of the oceans?
10. Why are temperature and salinity the two most important tracers of water masses?

Critical Thinking Questions

1. You might have heard it said that because of the Coriolis deflection, water going down a bathtub drain will swirl in opposite directions depending on whether the drain is located in the Northern or Southern Hemisphere. Do you think this is true? Explain why or why not.
2. Imagine that the viscosity of water was much higher, so that it flowed more like syrup. (a) Would the Ekman spiral and Ekman transport be different? If so, how would they be different and why? (b) Would there still be geostrophic currents? If so, how would they be different?
3. If seawater had a maximum density at 4°C, how would the deep water layers be different from those of today? Where would deep water masses be formed?
4. If we were to build a solid causeway across the Pacific Ocean from Seattle to Tokyo, how would the surface currents in the North Pacific Ocean be affected, and how might this affect the climate of Japan, Alaska, and California?
5. If the Earth were covered in water except for the North American continent, would there be geostrophic gyres? If so, where? What would the current pattern be in the Southern Hemisphere?
6. Discuss why you would expect the Gulf Stream to slow down and Europe to be colder if the rate of formation of North Atlantic Deep Water was reduced.
7. Why is there no geostrophic gyre in the Mediterranean Sea?
8. Global climate may change in the next few decades, with the average temperature of the ocean surface waters increasing by as much as 2–3°C. (a) Hypothesize what might happen to the thermohaline circulation of the oceans as a result. (b) Speculate on what would be the effects on the intensity, geographic distribution, and frequency of upwelling of water from below the permanent thermocline. Explain your reasons for these speculative changes.
9. From what you have learned about ocean circulation, what do you think would be likely to happen to the atmospheric concentrations of carbon dioxide if the carbonate compensation depth (CCD) were to migrate to shallower depths? How long would you expect it to take for these changes to occur? Explain why.
10. If the hypothesized future global warming does indeed occur, the Arctic ice pack may melt completely. (a) How would this affect the bottom water mass of the Arctic Ocean? (b) How might it also affect the characteristics of bottom water masses in the other oceans? (c) What other changes in ocean circulation might be caused? (d) Would these changes be significant during a normal human's lifetime, or would they take thousands of years to occur? Explain the reasons for your answer.

This page titled 8.16.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.16.3: References

Figure 8-1a – NASA GSFC, 2017: Eddies in the Southern Ocean. Accessed 9 December 2025, Source.

Figure 8-1b – NASA GSFC, 2021: Taking to the Air and Sea to Study Ocean Eddies. Accessed 9 December 2025, Source.

Figure 8-16 – NASA JPL, 2015: Sea Surface Temperature (L4, MUR) on 18 April 2025. Accessed 9 December 2025, Source.

Figure 8-8 – NOAA/GFDL, 2006: CM 2.4 Sea Surface Temperature Simulation.

Figure 8-19 – Data from Neumann, G., 1968: *Volume 4: Ocean Currents*. Elsevier Science Publishing, 132 pp.

Figure 8-20b – Segar, D., and E. Stamman Segar, 1997: Convergences between Langmuir circulation cells.

Figure 8-22 – Tchernia, P., 1980: *Descriptive regional oceanography*. Pergamon Press Ltd., 253 pp.

– Dietrich, G., 1980, *General Oceanography*. John Wiley & Sons Inc., 626 pp.

Figure 8-23 – Tchernia, P., 1980: *Descriptive regional oceanography*. Pergamon Press Ltd., 253 pp.

– Dietrich, G., 1980, *General Oceanography*. John Wiley & Sons Inc., 626 pp.

Figure 8-24a – Adapted from Wüst, G., 1961: On the vertical circulation of the Mediterranean Sea. *J. Geophys. Res.*, **66**, 3261–3271, Source.

Figure 8-24b,c – Adapted from Pickard, G. L., and W. J. Emery, 1982: *Descriptive Physical Oceanography*. Pergamon Press Ltd., 265 pp.

Figure 8-25 – Stommel, H., 1958: The abyssal circulation. *Deep-Sea Res.* **5**, 80–82, Source.

Figure 8-27a – NOAA NSIDC, 2022: Sea Ice Extent Maps. Accessed 9 December 2025, Source.

Figure 8-28 – Adapted from NOAA, 2005: The Global Conveyor Belt. Accessed 9 December 2025, Source.

Figure 8-30 – Östlund, H. G., and C. G. H. Rooth, 1990: The North Atlantic tritium and radiocarbon transients 1972–1983. *J. Geophys. Res.*, **95**, 20147–20165, Source, with permission

Figure 8-31 – Adapted from Molinari, R. L., R. A. Fine, W. D. Wilson, R. G. Curry, J. Abell, M. S. McCartney, 1998: The arrival of recently formed Labrador sea water in the Deep Western Boundary Current at 26.5°N. *Geophys. Res. Lett.*, **25**, 2249–2252, Source.

This page titled 8.16.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.17: Box 1- Westward Intensification

Imagine a motionless wheel suspended horizontally above the Earth's surface at the equator (**Fig. 8B1-1**). To an observer on the Earth, the wheel does not appear to rotate as the Earth rotates. Now imagine that we transport this still motionless wheel to the North Pole and again suspend it horizontally above the Earth. To an observer on Earth, the stationary wheel appears to spin at the same rate as the Earth. This effect also applies to spinning gyres of water in the oceans. The rate of apparent rotation increases with distance from the equator and is called planetary vorticity. Western boundary currents undergo an increase in their apparent rate of rotation as they flow poleward, whereas eastern boundary currents undergo a decrease in their apparent rate of rotation as they flow toward the equator. This is the underlying reason for the westward intensification, but it requires some explanation.

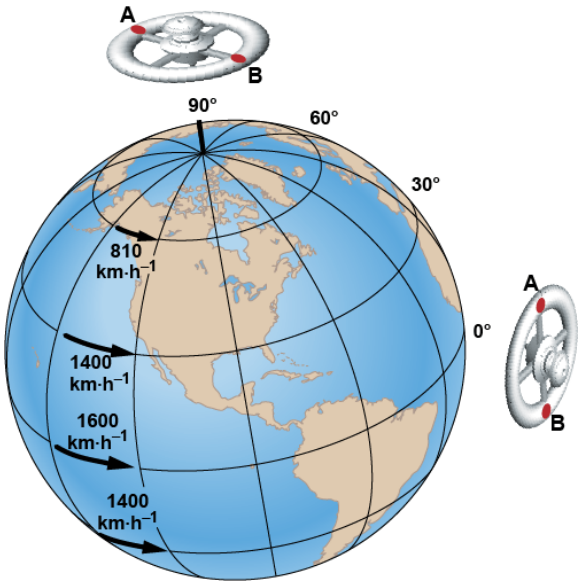


Figure 8B-1. Vorticity (apparent rotation) increases with increasing latitude. Imagine a rotating Earth with a circular space station orbiting above a specific location on the equator but not rotating around its own axis. An observer looking up from the equator would see that the nonrotating space station stays in the same position in the sky and does not rotate. Now imagine the space station being positioned directly over the North Pole, but still not rotating about its own axis. To an observer looking up from the pole, the nonrotating space station would appear to rotate once every 24 h. It is more difficult to explain what an observer would see if the space station were moved from the equator to mid-latitudes. However, the observer would see it rotate, and the rotation rate would increase progressively faster with increasing latitude until, at the poles, it reached one apparent rotation per 24 h.

In a spinning fluid, the angular momentum must be conserved. It is not necessary to fully understand this, but you can see an application of this principle when an ice-skater spins. With the arms close to the body, the skater spins quickly. If the arms are extended, the spin becomes much slower, but momentum is not lost. When the arms are once more brought in, the spin becomes faster again. Of course, friction with the ice does eventually steal momentum, slowing down the spin. Angular momentum and vorticity are related, and vorticity is conserved in much the same way as angular momentum. However, the vorticity that must be conserved is a combination of the planetary vorticity and the relative vorticity. The relative vorticity is the vorticity due to the wind-driven motion of the water in the gyre. The math can be a little complicated, so the process is simplified and summarized in **Figure 8B1-2**.

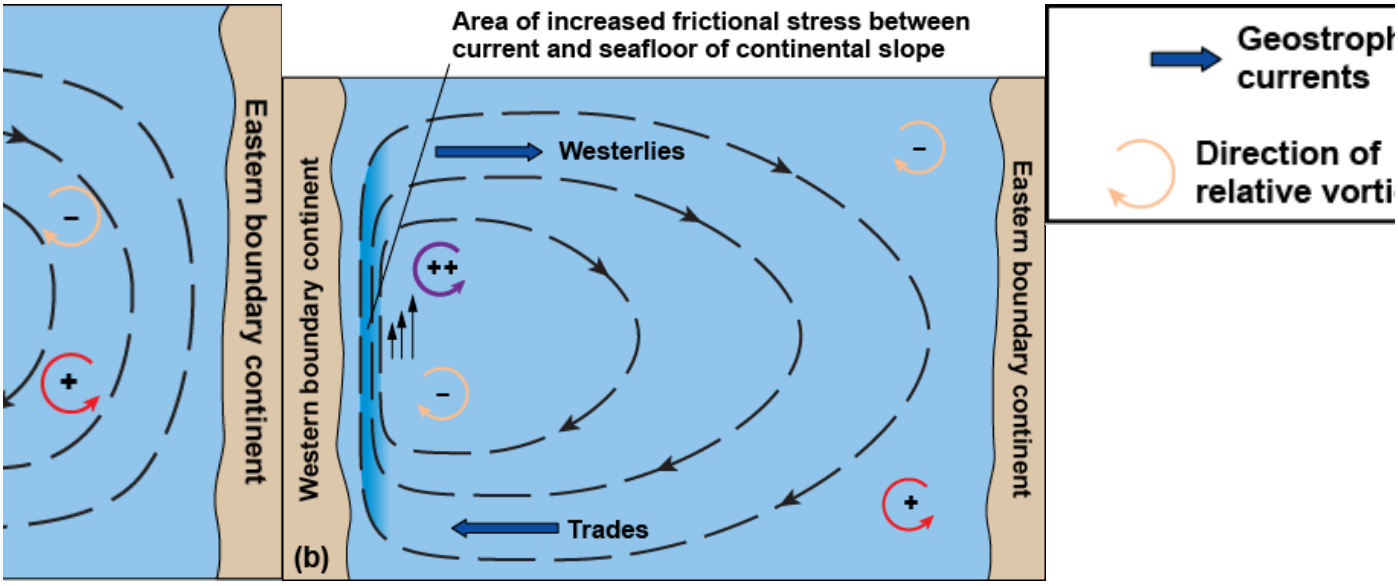


Figure 8B1-2. In the Northern Hemisphere, winds generate clockwise-rotating gyres. This rotation creates clockwise vorticity (or spin) relative to the Earth, and this is called relative vorticity. In addition, vorticity, called planetary vorticity, is imparted by the variations in the magnitude of the Coriolis effect. On the eastward side of the gyre, water flowing southward is subject to a decreasing Coriolis effect, which induces an anticlockwise planetary vorticity (for simplicity, a reduced tendency to turn to the right or spin). On the westward side of the gyre, the northward-flowing water is subject to increasing Coriolis effect, which induces clockwise planetary vorticity. However, the total vorticity, or spin, must be the same for all parts of the gyre. If it were not, then the total mass transport rate of one part of the gyre would speed up while another slowed down, which is not possible. (a) If the gyre were symmetrical, the relative vorticity on the eastward side would be balanced by the planetary vorticity, as they are opposite. On the westward side, planetary and relative vorticity are in the same direction and cannot be balanced. (b) If the gyre is offset (intensified) to the west, friction is increased between the flowing water and the seafloor of the continental slope against which the gyre current is concentrated. The friction provides a source of counterclockwise vorticity to balance the clockwise relative and planetary vorticities.

Figure 8B1-2a represents a symmetrical gyre with no westward intensification. On the east side of the gyre, the planetary vorticity is counterclockwise, and the relative vorticity is clockwise. Thus, the decrease in planetary vorticity as water moves south can be compensated for by an increase in the relative vorticity (speeding up the wind-driven rotation of the gyre, which is in fact what happens as the current flows farther south into the trade wind belt). However, on the west side, both the planetary vorticity and the relative vorticity are counterclockwise. The increase of planetary vorticity could be partly offset if the relative vorticity decreased, but the magnitude of the planetary vorticity at the north end of the western boundary current is much larger than the relative vorticity, so the clockwise motion would be stopped unless there was a strong source of positive (counterclockwise) relative vorticity. This can be supplied by a strong frictional stress that opposes the northward flow of the western boundary current. A strong frictional stress, in turn, can be established if the current is constrained to flow against the land boundary (**Fig. 8B1-2b**), which generates friction at the interface between the current and the seafloor. If the current flows fast and is narrow and deep, the area of contact with the seafloor (along the edge of the continental shelf) is maximized, and, therefore, so is the shear “frictional” stress. As a result, the westward intensification occurs in all ocean current gyres and is caused by the variation in magnitude of the Coriolis effect with latitude.

This page titled 8.17: Box 1- Westward Intensification is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

8.18: Box 2 - Equatorial Currents

Ekman Transport in Equatorial Region

Because the atmospheric intertropical convergence zone is displaced north of the equator, part of the southeast trade wind zone extends into the Northern Hemisphere. Consequently, although the Coriolis effect is weak near the equator, trade winds are deflected to the right after they pass north of the equator. Once the Southern Hemisphere trade winds have crossed the equator into the Northern Hemisphere, they generate Ekman transport that is deflected to the right, away from the equator, in a generally northeast direction toward the Doldrums (**Fig. 8B2-1**). This pattern contrasts with the Ekman transport to the southwest, or away from the Doldrums, that was assumed in the simple model (**Fig. 8-8**). Ekman transport is directed to the southwest in the major portion of the southeast trade wind belt that lies south of the equator (**Fig. 8B2-1a**).

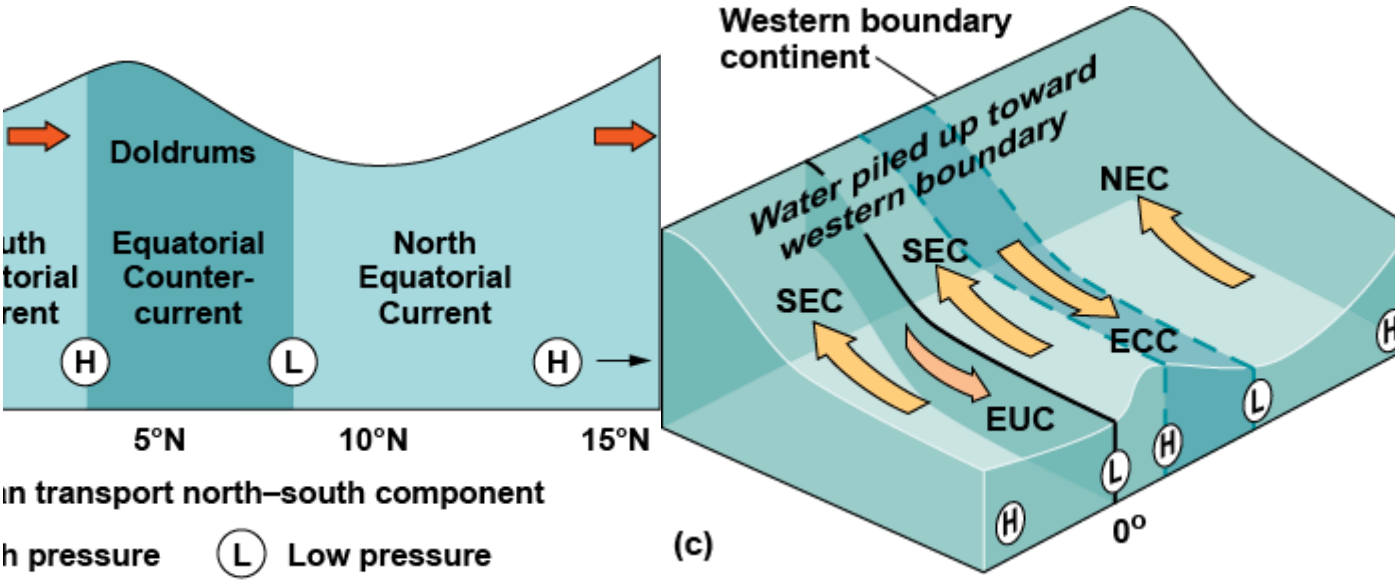


Figure 8B2-1. Currents of the Pacific Ocean equatorial region. (a) The intertropical convergence is displaced north of the equator, especially in the winter, such that the Southern Hemisphere southeast trade winds extend north of the equator. Because Ekman transport is to the right of the wind direction in the Northern Hemisphere, the pattern of Ekman transport just north of the equator is complex. (b) This pattern of Ekman transport produces complex latitudinal variation of sea surface elevation in the equatorial region. (c) The complex sea surface elevation pattern sets up an eastward-flowing Equatorial Undercurrent between the South Equatorial Current and the Equatorial Countercurrent. (d) The Equatorial Undercurrent is a fast current—much faster than the North and South Equatorial Currents or the Equatorial Countercurrent. The Equatorial Undercurrent flows between the surface and a depth of about 300 m, and its maximum velocity occurs at about 100 m.

When the intertropical convergence zone is displaced north of the equator, the sea surface slopes up in a northward direction toward the center of the Northern Hemisphere subtropical gyres throughout the northeast trade wind zone (**Fig. 8B2-1b**). In addition, the sea surface slopes up in a southward direction toward the center of the Southern Hemisphere subtropical gyre throughout the portion of the southeast trade wind zone in the Southern Hemisphere. These slopes are the same as the slopes developed in the simple model of the subtropical gyres shown in **Figure 8-8** and **Figure 8-9**. However, the sea surface slopes upward toward the north in the portion of the southeast trade wind zone that lies between the equator and the Doldrums (**Fig. 8B2-1b**). In this region, Ekman transport is northeast, to the right of the wind, toward the Doldrums, where the sea surface level is undisturbed by local winds.

Geostrophic Currents in the Equatorial Region

As a result of Ekman transport processes, a small “hill” of water is present in the Northern Hemisphere, and its crest is at the intersection of the Doldrums with the southeast trade wind zone (**Fig. 8B2-1b**). In the pressure gradient south of the crest, geostrophic flow is to the west, in the same direction as the geostrophic flow immediately south of the equator. North of the crest, in the pressure gradient within the Doldrums, geostrophic flow is to the east. This is the Equatorial Countercurrent. The arrangement of sea surface slope, pressure gradients, and geostrophic currents in the equatorial region is shown in **Figure**

8B2-1c. The characteristics of each of the currents are described in Table 8B2-1.

Table 8B2-1. *Boundary Currents*

Properties	Western Boundary Currents	Eastern Boundary Currents
Northern Hemisphere examples	Gulf Stream Kuroshio Current	California Current Canary Current
Southern Hemisphere examples	Agulhas Current Brazil Current	Peru Current Benguela Current
Width	Narrow (≤100 km)	Broad (≈1000 km)
Depth	Deep (to 2 km)	Shallow (≤500 m)
Speed	Fast (>100 km•day^{-1})	Slow (<50 km•day^{-1})
Volume transport	Large (50•106 m^{3}•s^{-1})	Small (10–15• ×10^{6})
Boundaries with coastal currents	Sharply defined	Diffuse
Upwelling	Almost none	Frequent
Nutrients	Depleted	Enhanced by upwelling
Fishery	Usually poor	Usually good
Water temperature	Warm	Cool

The equatorial currents of the subtropical gyres and the westward component of Ekman transport in the trade wind zones move very large volumes of surface layer water toward the west. Consequently, the sea surface slopes upward toward the western boundary. Much of the water that flows westward is either diverted north or south in the subtropical gyres or transported back toward the east in the Equatorial Countercurrents. However, water also flows back to the east in a current called the Equatorial Undercurrent.

Equatorial Undercurrents flow at the equator, are about 400 km wide, and have a vertical thickness of about 200 m (Fig. 8B2-1d). In the Pacific, the core of the current rises from a depth of about 200 m in the west to about 40 m in the east. In the Atlantic, the core is at about 100 m depth. Equatorial Undercurrents are very swift, with speeds comparable to those of the Gulf Stream. The Equatorial Undercurrent flows on the west-to-east pressure gradient created by the “piling up” of surface water toward the western boundary. However, the fact that it flows directly down the pressure gradient distinguishes it from other geostrophic currents,

which all flow parallel to the contours of constant pressure (across the pressure gradient) to balance the pressure gradient and the Coriolis effect (**CC12, CC13**).

The Equatorial Undercurrent can flow down the pressure gradient because it flows precisely west to east at the equator, where the Coriolis deflection is zero. If the current deviates slightly to the north, the Coriolis deflection (to the right in the Northern Hemisphere), although weak, turns the current back to the south. If it deviates slightly to the south, the Coriolis deflection (to the left in the Southern Hemisphere) turns it back to the north. Thus, the Equatorial Undercurrent flows directly across the oceans for thousands of kilometers without deviating from its west-to-east direction. The Pacific Ocean Equatorial Undercurrent is particularly important for the critical role it plays in El Niño (**Chap. 7**).

References

Figure 8B2-1 – Adapted from Knauss, J. A., 1961: The Cromwell Current. *Sci. Am.,* **204**, 105–116, Source.

This page titled 8.18: Box 2 - Equatorial Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9: WAVES

Critical Concepts

- **CC5 Transfer and Storage of Heat by Water:** Water's high heat capacity allows large amounts of heat to be stored in the oceans with little change in temperature. Thus, when waves break and release their energy as heat, they cause very little change in the temperature of the water.
- **CC9 The Global Greenhouse Effect**: Major climate and climate related changes may be an inevitable result of our burning fossil fuels. The burning of fossil fuels releases carbon dioxide and other gases into the atmosphere, where they accumulate and act like the glass of a greenhouse retaining more of the sun's heat. One of these climate related changes is the possibility that average wave heights in the oceans will increase, which would add to existing safety issues for ocean vessels and increase shoreline erosion rates.
- **CC12 The Coriolis Effect:** Water masses move freely over the Earth surface while the solid Earth itself is constrained to move with the Earth's rotation. This causes moving water masses, including some long period waves, to appear to follow curving paths across the Earth's surface. The apparent deflection, called the Coriolis effect, is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. It is at a maximum at the poles, reduces at lower latitudes, and becomes zero at the equator
- **CC13 Geostrophic Flow**: Water and air masses flowing on horizontal pressure gradients are deflected by the Coriolis effect until they flow across the gradient such that the pressure gradient force and Coriolis effect are balanced, a condition called geostrophic flow. The interaction of geostrophic currents and the change of the magnitude of the Coriolis effect with latitude are the cause of Rossby waves that flow east to west across the oceans and atmosphere. The meanders in the jet stream seen in many weather maps are Rossby waves.

Figure 9-1. A large ferry boat rests inland amidst destroyed houses in Miyako, Iwate, Japan. The houses were destroyed and the ferryboat and other boats carried inland by a tsunami that followed a 9.0 magnitude earthquake on 11 March 2011. The earthquake and tsunami killed about 27,000 people, most of whom survived the earthquake only to be overcome by the tsunami.

We are very much aware of continuous wave motion anytime we sail or visit the **shore**. Even on the calmest day, the water still laps gently onto the **beach** or rocks. If we stay long enough, we also notice that in most places the water level rises and falls on the shore as the hours pass. This slow up-and-down oscillation of the ocean surface is the incoming and outgoing **tide**. Because the high point of the wave (high tide) is separated by hours from the low point of the wave (low tide), we do not think of tides as waves. However, tides, which are discussed in **Chapter 12**, are a form of wave motion.

For all who visit or live near the shore, ocean waves are important. Waves provide enjoyment for people who visit the shore, and, as described in **Chapter 11**, waves are responsible for building and maintaining the sand beaches that are essential to many of our recreational activities. Unfortunately, waves also drown the unwary, **erode** beaches and **coastlines**, and damage piers and

other structures. Waves and tides are also important because they affect the navigation and safety of the myriad commercial and recreational vessels that ply the world oceans. This chapter describes what ocean waves are, how they are formed, and how they behave. Knowledge of the behavior of waves can enhance our enjoyment of activities such as swimming, surfing, **scuba** diving, sailing, or just standing by the water's edge watching the **surf**. It can also protect us from the dangers that waves often present.

9.1: Complexity of Ocean Waves

9.2: Progressive Waves

9.3: What is Wave Motion?

9.4: Making Waves

9.5: Sea Development and Wave Height

9.6: Wave Dissipation

9.7: Deep-Water Waves

9.8: Waves in Shallow Water

9.9: Tsunamis

9.10: Internal Waves

9.11: Rossby and Kelvin Waves

9.12: Standing Waves

9.13: Chapter Summary

9.13.1: Key Terms

9.13.2: Chapter Questions

9.13.3: References

This page titled 9: Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.1: Complexity of Ocean Waves

Ocean waves are complicated motions of the water surface. The surface of the water oscillates up and down, sometimes only a few centimeters and sometimes several meters, but the oscillations are never exactly regular. Successive waves are of different **wave heights**, and the **wave period** also varies. Sometimes there may seem to be a pattern, especially on days with little or no wind. On such days the waves appear to be less complicated, and the intervals between successive waves are usually longer. However, on the open sea during a storm, waves appear to break at random times and places. In addition, waves that roll in toward shore do not all break at exactly the same distance from shore, and each breaking wave looks a little different from the preceding one.

Although they are complex, ocean waves are not difficult to understand. The reason for the complexity is that what we see at any one place and time is a combination of many different waves. The combination of two or more simple waves of different periods or of two waves traveling in different directions is called **wave interference**, and this process produces complex waveforms. We can see how waves interact and combine in a lake or even a bathtub. When we throw a stone into a lake, the waves (usually called “ripples” if they are small) radiate outward in neat, regular, circular patterns from the point of impact. When we throw two stones into a lake a few meters apart from each other, the waves radiate in regular, circular patterns from each point of impact. Where the waves from one stone meet the waves from the other stone, a complicated wave pattern develops on the water surface as the two sets of waves interact.

If we watch carefully, we may be able to see that the two sets of waves are not really altered when they meet. Each wave simply passes through the other set, maintaining its original form and direction. The complicated surface pattern arises where the two sets of waves cross each other because the vertical displacement of the surface is the sum of the displacements of each of the two sets of waves at that point in space and time. We will return to the concept of wave interference later. For now, the important point to understand is that even the most complicated seas are the sum of a number of different simple waves of different heights and periods that may come from different directions.

This page titled 9.1: Complexity of Ocean Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.2: Progressive Waves

Waves that we create by throwing a rock into a lake and waves that we see on the ocean move freely on the water surface. Therefore, they are called **progressive waves** (Fig. 9-2). Another type of wave, called a **standing wave**, behaves differently, as discussed later in this chapter.

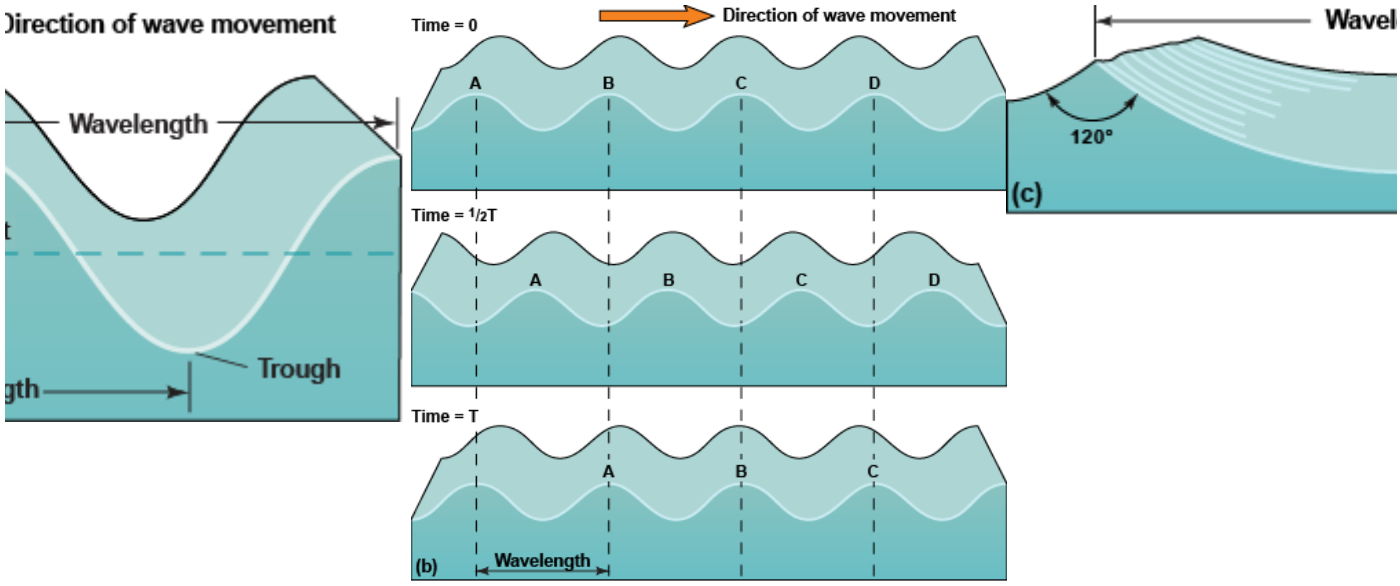


Figure 9-2. (a) All waves can be characterized by the presence of a crest and a trough, and by their wavelength (L) and wave height. (b) Progressive waves travel over the ocean surface at a speed that can be characterized by their period (T) or their frequency (f), which is the inverse of the period. Most progressive waves are smooth sine waves with rounded crests and troughs. (c) As waves are built by the wind, they develop a pointed crest and become trochoidal in shape.

If we look carefully at the ripples caused by tossing a rock into a lake, we see that a series of waves, not just a single wave, moves out from the impact point. Each of the individual waves has a rounded top where the water surface is elevated (Fig. 9-2a). The point of highest elevation of the wave is the wave **crest**. Between each successive pair of crests is a rounded depression of the water surface, or **trough**. Most progressive waves, whether in a lake or in the deep ocean, also have this general shape.

Waves also have several other important characteristics (Fig. 9-2a). The distance between two adjacent crests (or troughs) is called the **wavelength**, represented as *L*. Wave height (*H*) is the vertical distance between crest and trough, and wave **amplitude** is equal to *H/2*, the vertical distance between the crest or trough and the mean water level. Wave period (*T*) is the time the wave takes to move a distance equal to one wavelength. It is equivalent to the time that elapses between the arrival of two successive crests (or troughs) at a point on the surface. Wave period is measured as the number of seconds per wave. Wave **frequency** (*f*) is the number of wave crests (or troughs) that pass a point on the ocean surface in a given time. It is measured as the number of waves (or fractions of a wave) per second. Wave period (*T*), usually measured in seconds, and wave frequency (*f*) are related by a simple equation:

$$T = \frac{1}{f} \tag{9.2.1}$$

or

$$f = \frac{1}{T} \tag{9.2.2}$$

We can estimate frequency without a stopwatch by counting the number of waves that pass in several minutes and dividing by the elapsed time (in seconds).

Two other important characteristics of progressive waves, **wave speed** and **wave steepness**, can be determined from the wavelength, wave height, and period or frequency. Wave speed is often called “celerity” (abbreviated *C*) because there is essentially no net forward movement (or speed) of the water as the wave passes. As explained later in this chapter, only the wave energy and waveform, and not the water, move forward with the wave. Wave speed (*C*) is the ratio of the wavelength to the wave’s period and can be calculated using the following equations:

$$C = \frac{L}{T} \tag{9.2.3}$$

or

$$C = f \times L \tag{9.2.4}$$

Wave steepness is equal to the ratio of wave height to wavelength (*H/L*).

This page titled 9.2: Progressive Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.3: What is Wave Motion?

When a wave moves across the ocean surface without breaking, there is almost no net forward motion of the water itself. As we can easily see at the shore, objects such as logs and surfers floating outside the **surf zone** are not carried inshore with the waves. As waves pass, the floating objects ride up and down on each wave but remain at almost the same location indefinitely. Surfers must paddle forward to "catch" a wave before they can ride it inshore.

If water moved forward with the wave, the world would be an entirely different place. If you have walked or swum out through the surf on a beach, you have experienced what this would be like. When a wave breaks on the beach, the foaming water in the wave crest does move forward faster than the water under the wave is moving back away from shore. Walking or swimming through the surf is difficult because each wave tends to knock you down and carry you back toward the beach until the trough of the wave arrives and carries you seaward as water drains down the beach. However, once you have passed through the surf zone, the same waves that knocked you down in the surf no longer push you toward the beach. Imagine the difficulty ships would have if water on the open ocean were transported in the direction of the waves, in the manner that water in the surf zone is.

Wave Energy

A wave has both **kinetic energy** and **potential energy**. Kinetic energy is possessed by water molecules that are moving in the wave, and potential energy, by water molecules that have been displaced vertically against **gravity** and **surface tension**.

At the wavelengths of most ocean waves, the total energy (*E*) per unit area of a wave is approximately

$$E = 0.125(grH^2) \tag{9.3.1}$$

where *r* is the absolute **density** of water (in g•cm^{-3}), *g* is the acceleration due to gravity (9.8 m•s^{-2}), and *H* is the wave height (in m). *E* is measured in **joules** per square meter (J•m^{-2}). Wave energy does increase with wavelength, but this factor is only important for waves of very long or very short wavelength.

Because water density changes very little in the open oceans (**Chap. 5**) and *g* is a constant, the total energy of a wave depends primarily on its height. The total energy of a wave is multiplied by a factor of 4 if the wave height is doubled.

Restoring Forces

Instead of moving forward with the wave, each water molecule within a wave moves in an orbital path. In deep water, the orbital path is circular (**Fig. 9-3**). Only the waveform and energy associated with the wave move forward. For the waveform to move forward, a restoring force must exist that tends to return the ocean surface to its original flat configuration after the water is initially displaced.

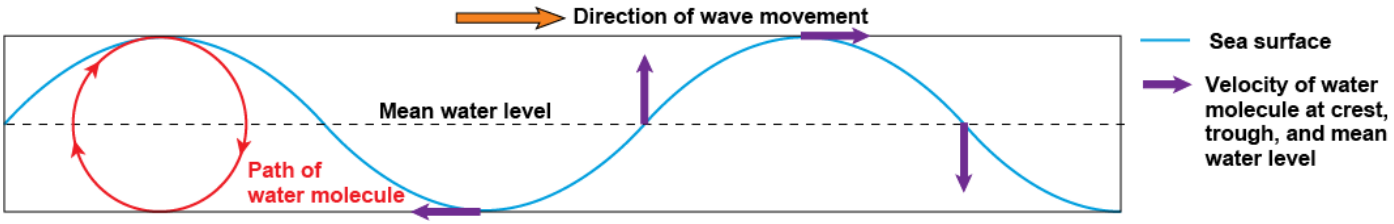


Figure 9-3. In a progressive wave, water at the crest is moving forward but has no vertical velocity. After the crest passes, the forward velocity slows, and vertical velocity increases until, halfway between crest and trough, the velocity is entirely vertical. As the wave continues forward, the vertical velocity is slowed, and the water gains velocity in the backward direction. At the trough, the vertical velocity is zero, and the backward velocity is at a maximum. The waveform moves forward, but each particle of water moves in a circular orbit whose diameter equals the wave height (for water at the surface) and returns to its starting point after each wave. You can see this for yourself if you push down on a water bed. The wave travels across the bed, but the plastic "surface layer" returns to its original location after the wave has passed.

The principal restoring forces acting on ocean waves are surface tension and gravity. Surface tension pulls the surface equally in all directions, contracting the surface to its minimum area—a flat plane (**Chap. 5**). A trampoline provides a good analogy for

surface tension. The trampoline surface is depressed and stretched when someone lands on it, but its “surface tension” causes it to snap back to its normal flat configuration, launching the trampoliner into the air.

Gravity acts on water molecules within a wave and causes a **pressure gradient** to develop beneath the sloping surface of the waveform (**Fig. 9-4**). The water flows in response to the pressure gradients and tends to flatten the ocean surface. In simpler terms, gravity causes the water to move downwards from the high parts of the wave to fill the depressions and restore the surface to a flat configuration.

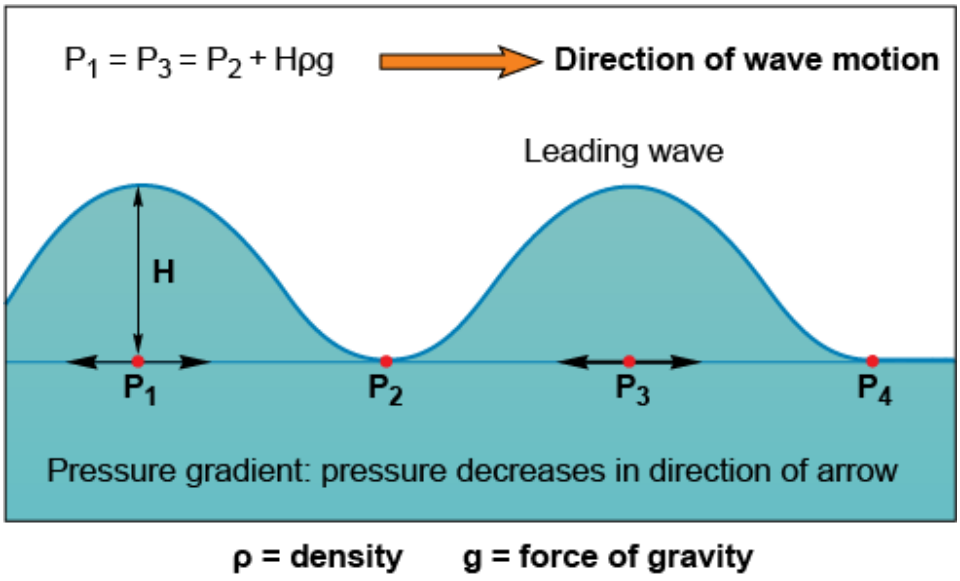


Figure 9-4. The horizontal pressure gradients under a wave illustrate how waves move. The pressure difference between locations directly beneath the crest and locations at the same depth beneath the trough tends to move water away from each crest toward the troughs. Between two crests, the pressure gradients are symmetrical and tend to move water toward the troughs. However, the water under each crest has kinetic energy because it is moving forward with the wave motion, and this energy is transferred forward from wave to wave with the wave motion. In front of the leading wave, the pressure gradient tends to move the water forward, displacing the still water surface in front, transferring some of the kinetic energy into potential energy, and beginning the wave motion. Some energy is transferred backwards from the last wave of a group of waves, in much the same way, to form a new wave at the back of the train.

Although interactions of restoring forces with water molecules in the wave are somewhat complicated, the principle is straightforward. Consider a water molecule located at the high point of an elevation of the ocean surface. The water molecule has potential energy because it is elevated above the mean water level. The restoring forces accelerate the molecule downward, and potential energy is converted to kinetic energy.

When the molecule reaches the mean surface level, its initial potential energy has been converted to kinetic energy and its motion is vertically downward. Because it has kinetic energy, it continues its downward motion but is slowed by the restoring forces. As it slows, kinetic energy is converted to potential energy. The molecule will continue in motion until all of its kinetic energy is converted to potential energy, at which point the molecule is at the same distance below the mean surface level as its starting point was above that level.

This process explains why water moves up and down as potential energy is converted to kinetic energy and back. Why, then, do water molecules in a progressive wave move in a circular path and not simply oscillate vertically up and down, falling from crest to trough and then flowing back up to the crest to repeat the cycle? In fact, this simple vertical oscillation can occur at some locations where a vertical barrier exists at the exact location of a wave crest or trough. Where this happens, the wave is not a progressive wave but instead is a standing wave. Standing waves (discussed later in the chapter) behave differently from progressive waves.

In contrast, if no barrier is present, the leading wave of any series of waves is adjacent to an undisturbed water surface. The horizontal pressure gradient under the leading wave moves water molecules toward the undisturbed surface (**Fig. 9-4**). This movement causes the undisturbed water surface at the leading edge of the wave to be displaced. Thus, the restoring force (gravity) causes a disturbing force that transfers energy to undisturbed water ahead of the leading wave, wave energy is translated forward, and the wave is progressive.

The two restoring forces, gravity and surface tension, act on all waves. Gravity is the principal restoring force for most waves within

the range of wavelengths normally present in the oceans, and such waves are often called **gravity waves**. In contrast, for waves with very short periods (<0.1 s) that can have only very small wave height (for reasons discussed later), gravity is less important and surface tension is the principal restoring force. These small waves are called **capillary waves** because *capillarity* is another term for surface tension. For waves with very long periods, such as the tides, an additional restoring force, the **Coriolis effect** (**CC12**), which is not actually a "force," is important (**Fig. 9-5**).

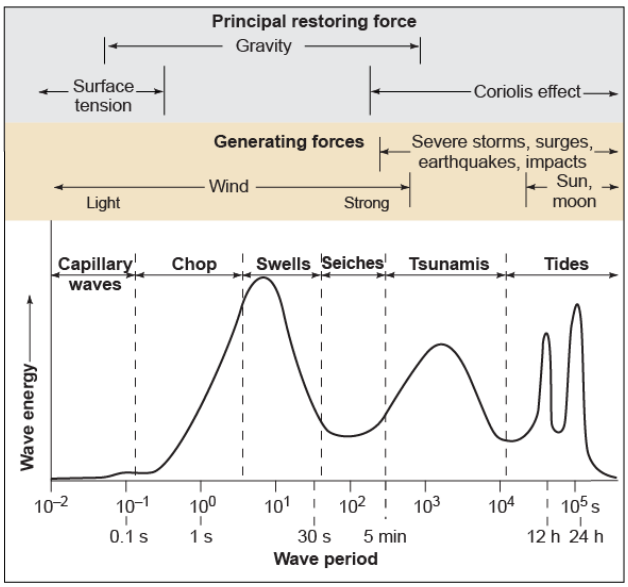


Figure 9-5. It is estimated that most of the energy associated with waves in the world's oceans is contained in wind waves that have periods of between 0.2 and 30 s. However, there is also considerable energy associated with tides (periods of 12 h and above) and, despite their infrequent occurrence, considerable energy is also associated with tsunamis and severe storm waves (periods between about 1 min and 1 h). The smallest waves, capillary waves, are created by winds and restored principally by surface tension. Winds are responsible for creating most waves with periods less than about 15 min, and gravity is responsible for restoring all except the shortest-period capillary waves within this range of periods. Longer waves are formed by earthquakes and tidal forces, and the Coriolis effect is the principal restoring force for the longest waves.

This page titled 9.3: What is Wave Motion? is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.4: Making Waves

Once they are created, waves must have a restoring force to sustain the wave motion, but waves must be formed initially by a displacing force. Small waves can be created by small displacing forces acting over short periods of time. Large waves can be created only by large displacing forces or lesser displacing forces exerted continuously for extended periods of time.

Most ocean waves are caused by the interaction of winds with the water surface, but waves are also created by impacts on ocean water, by rapid displacement of ocean water, by gravitational attraction between the Earth, moon, and sun, and by the passage of vessels or marine animals through the ocean surface. The ocean waves created by these various processes vary from waves with a period of less than one-tenth of a second to waves with a period of more than a day (**Fig. 9-5**).

Impact and Displacement Waves

Waves can also be created by impact or displacement. For example waves can be created by earthquakes or volcanic explosions that cause the seafloor to move abruptly, and by meteorite impacts on the ocean surface. Rapid displacements of ocean water can be caused by seafloor **slumps** (landslides) and **turbidity currents**, and by collapses of coastal cliffs. Such impacts can create waves with very long periods (typically 10 to 30 min) called **tsunamis** (or sometimes “seismic sea waves”). Tsunamis are occasional events and contribute only a fraction of the total wave energy of the oceans (**Fig. 9-5**), but they can be extremely destructive.

Impact and displacement waves can also be generated by comparatively minor events, such as whales and dolphins jumping out of the water and ships moving through the water leaving wakes. Waves created by these minor impacts are insignificant in number and intensity, in comparison with wind-generated waves. However, in sheltered harbors ship wakes may be the principal source of wave energy and erosion of the **shoreline**.

The gravitational attraction between the Earth, moon, and sun is the force that creates waves that we know as tides (**Chap. 10**). Tide waves have periods of predominantly about 12 or 24 h and contribute a significant amount of the total wave energy of the oceans (**Fig. 9-5**).

Wind Waves

As we know from experience, winds are highly variable, or gusty, and may change speed and direction from second to second. Air moving over a land or ocean surface becomes **turbulent** when more than a gentle breeze blows. Small variations of wind speed and pressure that occur as the wind blows across a smooth ocean surface are believed to lead to the creation of capillary waves, which then grow larger if the wind continues to blow. Capillary waves have a maximum wavelength of 1.73 cm, rounded crests, and V-shaped troughs.

Winds generate waves on lakes and other fresh water bodies as well as in the oceans. While ocean waves are considered here, the principles of wind wave generation discussed in this section apply equally to all water bodies, including bodies such as lakes, estuaries, and rivers.

The mechanism by which wind generates capillary waves is not fully understood, but two factors seem to be important. First, because of turbulence in the atmosphere, atmospheric pressure is slightly increased at some locations at the sea surface compared to locations a few millimeters or centimeters away. In these areas with slightly higher air pressure, the sea surface tends to be slightly depressed in relation to the adjacent ocean surface, where the pressure is lower. Thus, when winds blow across a perfectly flat sea surface, tiny areas of elevated or depressed sea surface are formed. Second, when winds blow over the ocean surface, a **shear stress** (**friction**) develops between the air and water because of the velocity difference across the air-sea interface.

The formation of capillary waves increases the roughness of the surface, which in turn increases the shear stress between wind and ocean surface and this enhances the ability of the wind to build larger waves. Capillary waves also alter the sea surface, so that some areas are tilted slightly up toward the wind (**windward** side) and others (**leeward** side) are tilted slightly away from the wind. The wind pushes harder on the back of the wave than on the front, further accelerating and building the wave (**Fig. 9-6**).

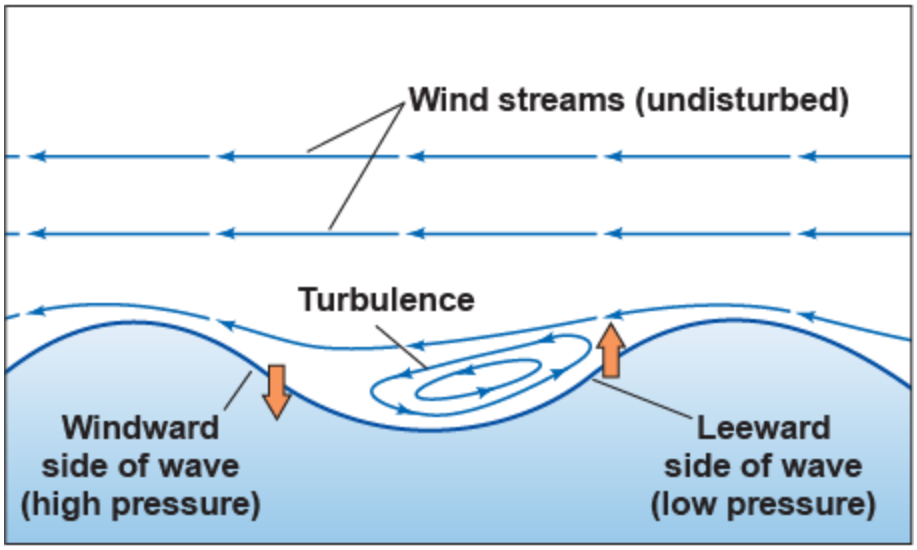


Figure 9-6. When the wind speed is great enough to form waves, the waves interfere with the wind flow immediately above the surface of the water. The wind pushes harder on the elevated, windward surface of the wave, and the air pressure is slightly lower on the leeward side of the wave than on the windward side. These forces combine to increase the height of the wave.

Wind-wave interaction actually is more complicated than this simple description suggests. For example, a wave generates turbulence as the wind flows across the water's surface just as an aircraft wing alters the airflow to provide the lift necessary for a plane to fly. Once a wave is formed, the wind tends to build the wave height, primarily by uplifting the leeward side of the wave. In addition, differences in wave speeds cause small capillary waves to combine into the longer-period, higher waves called gravity waves. In this way, as the wind continues to blow across the sea surface, wind energy accumulates, and longer-period waves form. These longer period waves comprise most of the ocean's wind wave energy (**Fig. 9-5**).

This page titled 9.4: Making Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.5: Sea Development and Wave Height

As a wave's height is built up by winds, its shape is modified from that of capillary waves. The shape first becomes close to the smooth sine wave form shown in **Figure 9-2a,b**. Subsequently, the crest and trough are modified progressively as the wave builds, and the wave shape becomes trochoidal (pointed crests and rounded troughs; **Fig. 9-2c**). As wind energy is absorbed by waves, their height, speed, period, and wavelength are increased.

Winds never blow uniformly over the water because they are highly variable in both time and space. Consequently, in the area where the waves are formed by the wind, waves of many different heights, wavelengths, and even directions of travel are present at the same time. Mariners refer to this confused state as a "**sea**." In a "calm," there are no significant waves, and the sea surface is flat. In a "**swell**", waves are generally smooth, mostly of the same wavelength and from the same direction. How a sea becomes a swell is explained later in this chapter when we consider **wave dispersion**.

Waves sometimes break in deep water just as they do near the shore, but waves breaking in deep water are often called "whitecaps." Waves break because they have become too steep. When the steepness of a **deep-water wave** (*H/L*) reaches 1:7 (wave height equals one-seventh of wavelength), the wave becomes unstable and the crest tumbles down the forward slope of the wave, creating a breaking wave. If winds are very strong, the tops of the largest waves may be blown off by the wind.

Waves break when they have reached their maximum possible height for the wind speed. They also break when seas are developing and the waves cannot be modified quickly enough to longer wavelengths and greater heights to absorb the energy introduced by the winds. As a wave becomes oversteepened and breaks, some of its energy is dissipated by turbulence, which releases heat. Consequently, this excess energy cannot be used to increase the wave height.

Factors Affecting Maximum Wave Heights

A sea is said to be fully developed when the amount of wave energy lost to turbulence in breaking waves is equal to the difference between the total energy input from winds and the amount of wind energy needed to maintain the sea. Waves of many different heights and periods are present in a fully developed sea, but storms with stronger winds produce waves with longer maximum periods than are produced by lower wind speeds (**Fig. 9-7a,b**). The maximum height of waves created by any specific storm or series of storms depends on the wind speed and the length of time the wind blows. The maximum height also depends on the wind **fetch**, which is the uninterrupted distance over which the wind blows (**Fig. 9-7c**). In shallow water (depth less than one-half of the wavelength), water depth is a limiting factor.

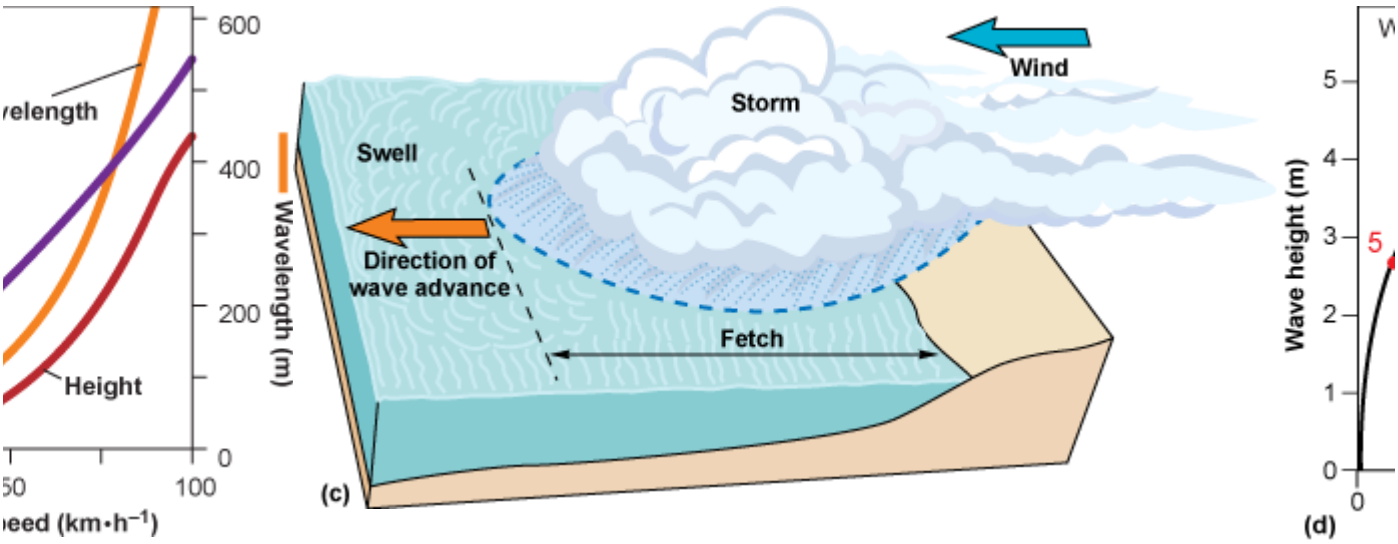


Figure 9-7. Relationships between wind speed, fetch, the length of time over which the wind blows, and the characteristics of the waves that are formed. (a) Strong (higher-speed) winds produce storms with greater total wave energy, and this energy is stored in waves of longer periods. (b) Strong winds also produce waves that have longer wavelengths and greater wave heights, provided that the winds blow long enough and have sufficient fetch. (c) The fetch is the distance over which the winds blow across the water to create waves. Fetch can be limited by the presence of a coastline or, where the storm is completely over water, by the dimensions of the storm itself. (d) The maximum wave height that can be reached is limited by the fetch, and this maximum wave height is reached only if the winds blow for a sufficiently long time.

Figure 9-7d shows the effect of wind duration and fetch on maximum wave height. As the wind begins to blow, waves accumulate energy from the wind. Wave height initially increases rapidly, then at a progressively slower rate. However, if the fetch is small, the maximum wave height is limited because waves relatively quickly travel out of the area where the wind is blowing. We can see the effects of a small fetch in lakes or in harbors protected from ocean waves. The fetch is restricted in such areas because the waves encounter a shoreline. No matter how strong winds are or how long they blow, only small waves can be created in harbors and all but the largest lakes.

The highest ocean waves are created where winds are strong and blow persistently over long fetches. High waves can also be generated when a series of strong storms passes in the same direction across the same fetch over a period of several days or longer. In this situation, some of the shorter-wavelength waves created by one storm do not have time to travel out of the area before the next storm arrives. Each successive storm simply builds on the height of waves remaining from preceding storms.

Maximum Observed Wave Heights

Persistent winds and storm tracks are arranged in bands with an east-west orientation (**Chap. 7**). Therefore, calm regions alternate with stormy regions from north to south in the oceans. The calm regions limit the fetch that can occur in a north or south direction in the ocean basins, but that does not mean that waves travel only east or west. First, the **trade winds** and **westerly** winds in the east-west wind bands do not blow directly east or west. Second, storms produce rotating winds and, thus, winds and waves of all directions. Third, if waves encounter a landmass or **reef**, the direction of wave travel may be altered as discussed later in this chapter.

Not surprisingly, high waves are most common in the Pacific Ocean because the Atlantic and Indian Oceans are much narrower and provide a more limited fetch. The longest fetches are within a band of westerly winds stretching around Antarctica.

A giant wave at least 34 m high was measured by the U.S. Navy vessel *Ramapo* on 7 February 1933 in the tropical North Pacific Ocean. In 1998, a wave estimated at nearly 43 m high was measured during the Sydney to Hobart yacht race in the South Pacific Ocean. Although these were the largest waves reliably reported, higher waves undoubtedly occur from time to time. It is not surprising that such extremely high waves have not been measured, since any sailor whose vessel is faced with a 30-m or higher wave has many concerns other than measuring wave height. Most vessels would survive the ride through even the highest wave unless the wave was extremely steep-sided and breaking. A 30-m-high steep-sided wave breaking over the deck of even the largest ocean vessel may be the last wave the vessel ever encounters. Indeed, such large waves may have caused the

disappearance of many ships in the past.

So far, the highest waves reliably reported in the Atlantic and Indian Oceans were about 15 m high. However, in September of 2004, the center of **Hurricane** Ivan passed directly over six tide/wave gauges mounted at depths of 60 to 90 m on the Gulf of America (Golfo de México) **continental shelf**. The highest wave measured by these gauges was 27.7 m. The area of highest winds within Ivan did not pass directly over any of the gauges, and it was calculated that the highest wave generated by this storm may have exceeded 40 m. Satellites such as the TOPEX/Poseidon satellite can measure ocean surface heights to an accuracy of within less than 5 cm. Unfortunately, these sensors cannot yet measure the heights of individual waves because the width of their measurement beams is large compared to the distance between ocean wind waves.

Effects of Currents on Wave Height

Steep-sided waves can be created when waves travel in the direction opposite that of a strong ocean **current**. The opposition of current and wave motion shortens the wavelength of the waves and increases their steepness, particularly on the front face of the wave. This effect is illustrated in **Figure 9-8**.

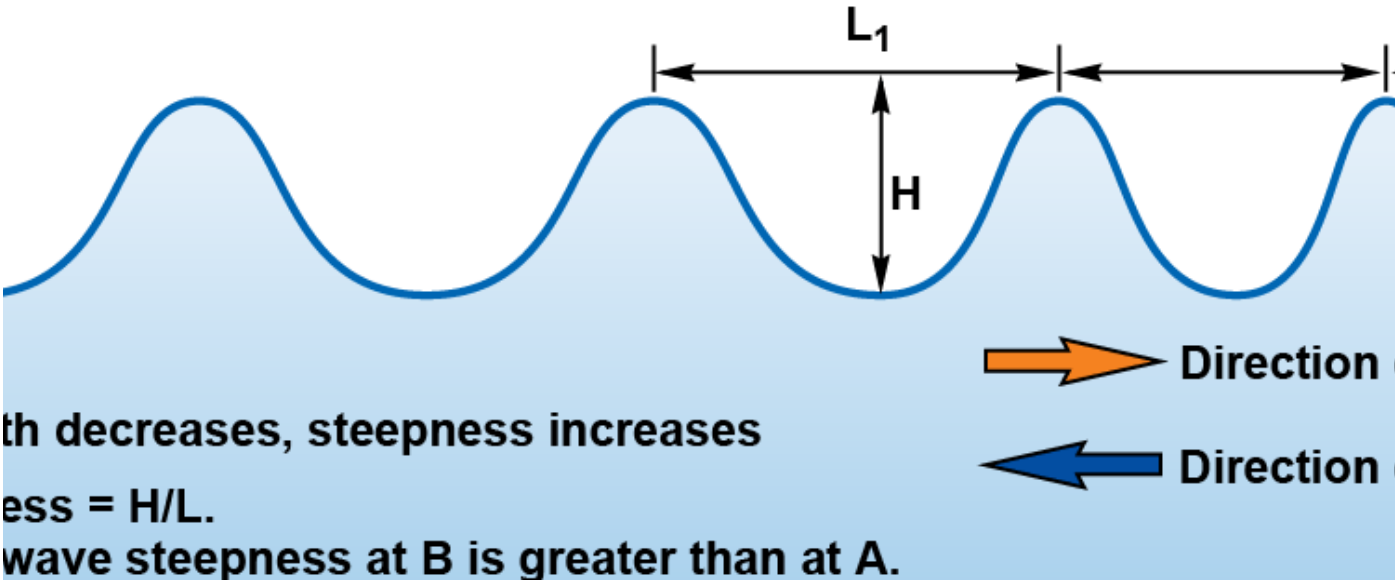


Figure 9-8. (a) Waves from around Antarctica travel northward and meet the strong, southward-flowing Agulhas Current. (b) As each wave encounters the current, its forward speed is decreased because the current moves surface water southward, and water particles in each wave are a little farther south after each wave orbit, instead of in their original position. As each wave slows, the following wave that has not yet encountered the current is not slowed, so its crest becomes closer to the crest of the wave preceding it. Hence, wavelength decreases and steepness increases as the figure shows. (c) The tanker World Glory, sinking after breaking its back in waves steepened by the Agulhas Current.

In extreme cases, waves can have almost vertical faces. Long-wavelength waves travel faster than vessels. When a vessel encounters a large wave with a nearly vertical face, the vessel may be unable to climb up and over the wave rapidly enough to avoid what is often a fatal plunge under the wave.

Such steep waves occur often in some areas, including where the swift Agulhas Current flows south along the Indian Ocean coast of southern Africa (**Fig. 9-8a**, **Chap. 8**). Large waves built by Antarctic storms travel north into this area, which is a major shipping lane for traffic between the Indian and Atlantic Oceans and is heavily traveled by oil supertankers. Several ships have been severely damaged or sunk there due to the shortened and steepened waves. In 1968, for example, the tanker World Glory broke in two, spilled its entire cargo, consisting of 14 million gallons of oil, and sank after encountering such a wave (**Fig. 9-8c**).

Tankers are constructed with most of the hull weight concentrated in the engine room and crew quarters at the stern, and the anchor-handling and other equipment in the bow. In the middle of the ship, oil is stored in separate compartments held together relatively weakly by the hull. A tanker can break in the middle if it becomes suspended with its center section on the crest of a high, steep-sided wave or suspended between two waves with the bow section on one wave and the stern on the other. Sophisticated satellite-based radar and other sensors now provide accurate advanced wave height forecasts in areas such as the Agulhas Current, so vessels are now able to avoid this region when dangerous waves are likely to be present.

Are Wave Heights Increasing?

Higher waves possess higher energy than lower waves and can create more damage when they impact a shoreline. Therefore, it is important to know whether wave heights are increasing due to higher wind speeds caused by a changing **climate**. However, measuring wave heights is not as simple as just recording the maximum wave height at one location over a period of time. Generally, oceanographers measure a parameter called “significant wave height,” which is a measure of the average height of approximately the highest one-third of all waves within a given area at a specific time. Methods of measuring wave heights have varied in the past several decades, so there are very few long time-series records of wave heights measured by a consistent method. This lack of consistent methodology makes conclusions about long-term trends in wave height difficult.

Wave heights in the North Atlantic Ocean were measured by researchers on a lightship moored off the southwest tip of England during 1960–1985. These measurements showed not only that the maximum wave heights vary widely between years, but also that the maximum wave heights apparently increased from about 12 m to about 15 m, about 25%, over this 25-year period. Since 1985, similar increases in wave heights have been observed in other parts of the Atlantic and Pacific Oceans by other techniques, including measurements of significant wave height by satellite-based sensors, calculation of significant wave heights from historical wind data records, and seismological data. Seismological data can be used because the microscale fluctuations in seismic records are related to the impact of waves on the coastline.

Although wave-height studies have shown that wave heights are increasing in some areas of the Atlantic and Pacific Oceans, they have also shown that wave heights have not increased significantly in other areas of these same oceans. At present, we do not know whether a long-term trend toward higher maximum wave heights is real. If it is real, we do not know whether it is related to climate change, whether the trend will continue, or in which areas this change is occurring. We are also unsure to what extent opposite trends or a lack of change may have occurred in other oceans.

However, we do know that any increase in maximum wave heights must be related to increased wind speeds, storm frequencies, or both. Small changes in ocean surface temperature may cause changes in the ocean-atmosphere interactions that create winds (**Chap. 7**). Hence, if wind speeds or storm frequencies have increased, the changes may be related to enhancement of the **greenhouse effect** caused by human release of greenhouse gases (**CC9**). Alternatively, the change in wave heights may be related to a natural long-term climatic cycle. Long-term climatic changes or cycles with periods of 30 years or more are known to occur. For example, historical records for Atlantic hurricanes suggest that they were fewer and weaker during the 1970s through the 1990s than in the preceding decades or in the current millennium (**Chap. 7**).

Considerably more information will be needed to fully identify the trends in wave heights and to determine whether these trends are part of a natural climatic cycle, are related entirely to the greenhouse effect (**CC9**), or are partially natural and partially related to the greenhouse effect. These issues are important because any long-term trend of increasing wave heights will pose problems for ships and increase coastal erosion.

This page titled 9.5: Sea Development and Wave Height is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.6: Wave Dissipation

When the force that creates a wave is removed, the wave continues to move across the ocean surface. For example, waves created by a rock thrown into a lake continue long after the rock's impact, and large waves may crash ashore on days when there is no wind to create them. The reason is that wave energy dissipates (is lost from the wave) very slowly.

Energy is lost from a wave because of viscosity (internal friction) between water molecules moving within the wave and air resistance as the wave moves across the surface, but this energy loss is very small. Frictional loss in a flowing fluid depends on the fluid's viscosity. The viscosity of water is not high (**Chap. 5**), so frictional loss is a very small fraction of the total wave energy in all but very short-period (<0.1 s) capillary waves. For capillary waves, viscosity is important as a means of dissipating wave energy because such waves contain only a very small amount of energy. In fact, the energy in capillary waves can be totally dissipated by viscosity in a minute or less after the wind stops. Capillary waves, therefore, disappear quickly when there is no wind, but larger waves do not.

Wave steepness diminishes once the wind ceases. Hence, in the open ocean, waves do not break and lose energy through turbulence unless winds blow and add more energy to the sea. Minor exceptions occur when waves encounter winds that blow, or currents that flow, against the wave's direction of travel. Waves then slow, and wave steepness increases (**Fig. 9-8b**).

The waves created by a rock thrown into a lake spread in a circular pattern away from the point of impact. This process does not decrease the total energy of each wave. However, as the wave moves out from the impact point, it spreads over an ever-increasing area (**Fig. 9-9**), and the wave height is reduced. Storm waves also spread out from where they are generated, but some storms generate their largest waves across a broad storm front. The resulting waves travel out from the storm area in the same general direction, and wave energy tends to be concentrated in that direction (**Fig. 9-9b**).

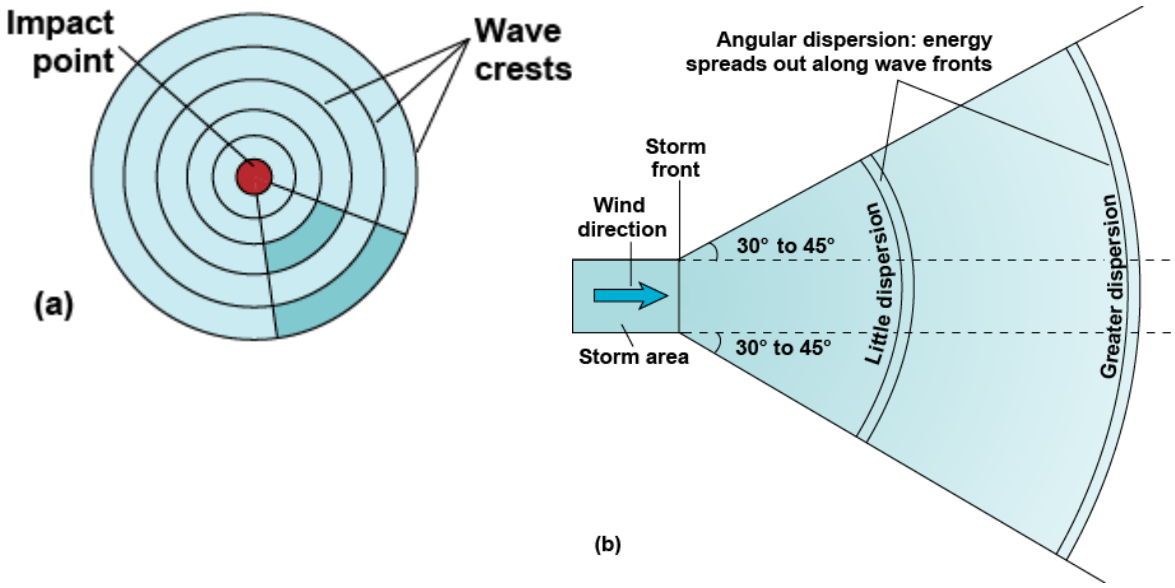


Figure 9-9. (a) Waves spread out uniformly in all directions from an impact point, and energy is spread along the increasing length of each wave as it moves away from this point. (b) Waves generated along a storm front travel forward in the direction of the winds in the storm, but they also spread out in a fan shape. This disperses (spreads) the wave energy along a progressively lengthening wave crest with distance from the storm.

Because waves lose very little energy as they travel across the ocean surface, most waves travel until they meet a shore, where their energy is converted to heat or used to erode the shore, transport sand along the coast, or create **longshore currents** (**Chap. 11**). Although breaking waves release a considerable amount of heat energy, this process does not cause a measurable change in water temperature, because water has a high heat capacity (**Chap. 5, CC5**). In addition, heat from breaking waves is dissipated in very large volumes of water.

This page titled 9.6: Wave Dissipation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.7: Deep-Water Waves

Individual water molecules in waves in deep water (depth >*L*/2) move in vertical circular orbits oriented in the direction of wave travel (**Fig. 9-10**). The forward motion at the top of the orbit is very slightly greater than the backward motion at the bottom of the orbit, hence there is a very small net forward movement of water in the wave, although this forward movement of water is so small that it is negligible (**Fig. 9-10b**). Water motion within a wave is not restricted to water depths between crest and trough. Scuba divers who have dived where long-period swell waves are present know that the surging wave motion can be felt 10 m or more below the surface, even when the surface swell waves are less than a meter high.

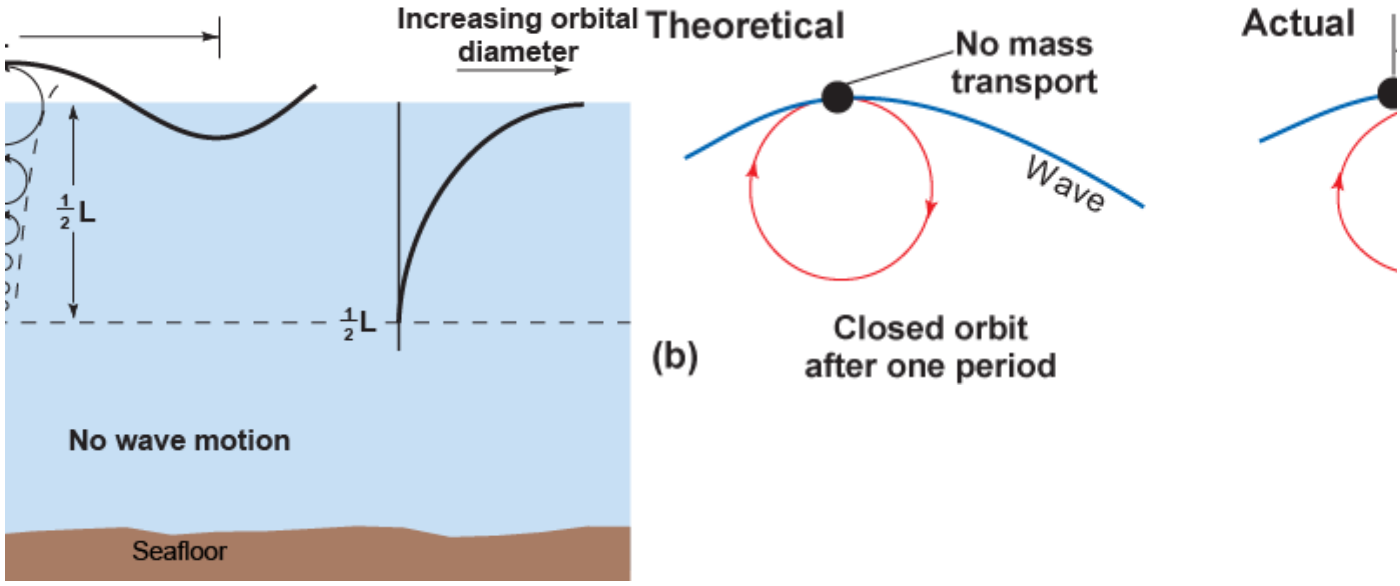


Figure 9-10. Orbital motion of water particles beneath waves. (a) The orbits in which individual particles move within a wave decrease in diameter exponentially with increasing depth. The wave motion becomes essentially zero at a depth equal to one-half of the wavelength. (b) The particle orbits are not exactly closed. Instead, each particle moves forward a very small distance (exaggerated in this figure) during each orbit. Therefore, there is a very small net mass transport of water in the direction of the wave.

Water immediately below the surface moves in an orbit whose diameter is equal to the wave height, whereas water farther below the surface moves in orbits whose diameter decreases as depth increases. The decrease in orbital diameter with increasing depth depends on the wavelength. The diameter is reduced to one-half of the wave height when the depth is equal to one-ninth of the wavelength, and it is almost zero at a depth of one-half of the wavelength (**Fig. 9-10**).

The speed of a wave (the speed at which the wave form of the wave moves forward) in deep water also depends on wavelength and is approximated by the following equation:

$$C = \sqrt{\frac{gl}{2\pi}} \tag{9.7.1}$$

where g is the acceleration due to gravity. Because *g* (9.8 m•s^{-2}) and π (3.142) are constants, we can simplify this equation to:

$$C = 1.25\sqrt{L} \tag{9.7.2}$$

C is measured in meters per second (m•s^{-1}), and *L* in meters (m).

Thus, the speed of deep-water waves (depth >*L*/2) increases with increasing wavelength. Because wave speed is also equal to *L*/*T*, we can calculate the wavelength of a deep-water wave if we know its period:

$$C = \frac{L}{T} = 1.25 \tag{9.7.3}$$

$$L = 1.25^2 \times T^2 = 1.56 \times T^2 \tag{9.7.4}$$

Table 9-1 lists wavelengths and celerities for deepwater waves of different periods. The table can be useful for scuba divers in planning a dive. If the divers count the number of waves that pass their boat per minute, they can calculate the average wave period. From **Table 9-1**, they can then determine the *L*/2 and *L*/9 depths for this wave period. If the *L*/9 depth exceeds the depth of the planned dive, the divers can expect to feel a strong wave surge, though it will be less than that at the surface, throughout the dive. At depths greater than *L*/2, there will be no wave surge.

Table 9-1a. *Characteristics of Deep-Water Waves (metric)*

Wavelength (m)	Period (s)	Frequency (waves per minute)	Depth of No Motion (*L*/2) (m)	Depth at Which Motion Reduced by 50% (*L*/9) (m)	Deep Water Celerity ($m \cdot s^{-1}$)
1.6	1	60	0.8	0.2	1.6
6.2	2	30	3.1	0.7	3.1
14	3	20	7.1	1.6	4.7
25	4	15	13	2.8	6.2
39	5	12	20	4.3	7.8
56	6	10	28	6.2	9.4
77	7	8.6	38	8.5	11
100	8	7.5	50	11	12
126	9	6.7	63	14	14
156	10	6.0	78	17	16
225	12	5.0	113	25	19

Wavelength (m)	Period (s)	Frequency (waves per minute)	Depth of No Motion (*L*/2) (m)	Depth at Which Motion Reduced by 50% (*L*/9) (m)	Deep Water Celerity ($m \cdot s^{-1}$)
306	14	4.3	153	34	22
400	16	3.7	200	44	25
506	18	3.3	253	56	28
624	20	3.0	312	69	31
975	25	2.4	488	108	39
1405	30	2.0	703	156	47

Table 9-1b. *Characteristics of Deep-Water Waves (imperial)*

Wavelength (ft.)	Period (s)	Frequency (waves per min)	Depth of No Motion (*L*/2) (ft.)	Depth at Which Motion Reduced by 50% (*L*/9) (ft.)	Deep Water Celerity (mph)
5	1	60	0.5	0.6	3.5
20	2	30	10	2.2	7.0
45	3	20	23	5.0	10
80	4	15	40	8.9	14
126	5	12	63	14	17
181	6	10	90	20	21
246	7	8.6	123	27	24

Wavelength (ft.)	Period (s)	Frequency (waves per min)	Depth of No Motion ($L/2$) (ft.)	Depth at Which Motion Reduced by 50% ($L/9$) (ft.)	Deep Water Celerity (mph)
322	8	7.5	161	36	28
407	9	6.7	203	45	31
502	10	6.0	251	56	35
724	12	5.0	362	80	42
985	14	4.3	493	109	49
1287	16	3.7	644	143	56
1628	18	3.3	814	181	63
2008	20	3.0	1004	223	70
3138	25	2.4	1569	349	87
4521	30	2.0	2261	502	105

Wave Trains

Waves usually travel in groups called “wave trains.” As the leading wave in a wave train travels, some of its energy is transferred to the molecules of undisturbed water in front of it. This energy is used to initiate orbital motion of the previously undisturbed water molecules. However, only half of the energy of the individual wave is transferred forward; the remaining energy is transferred back to the second wave in the train. By examining the pressure gradient under the wave, we can see how this happens (**Fig. 9-4**). The second wave in the train also transfers energy forward to the first and back to the third wave. As the wave train progresses, energy is lost from the front wave of the wave train. The front wave progressively loses height and eventually disappears. The energy transferred backward from the first wave is transferred progressively rearward from one wave to another, until it passes the last wave and builds a new wave at the rear of the wave train.

In deep water, a wave train and the energy associated with it move at one-half the speed of individual waves within the wave train. The time it takes for waves of a particular wavelength to cross the ocean must be calculated from this group speed, which is one-half of the individual wave speed specified in **Table 9-1**.

As a result of the rearward transfer of energy in the wave train, the front wave in the train continuously decays, and a new wave is continuously created at the rear of the train (**Fig. 9-11**). Individual waves form at the back of the train progress forward as

their predecessors decay and, eventually, reach the front of the train, where they decay themselves. We can see this process by carefully watching ripples caused by a stone thrown into a lake or the waves of a ship's wake (**Fig. 9-12**).

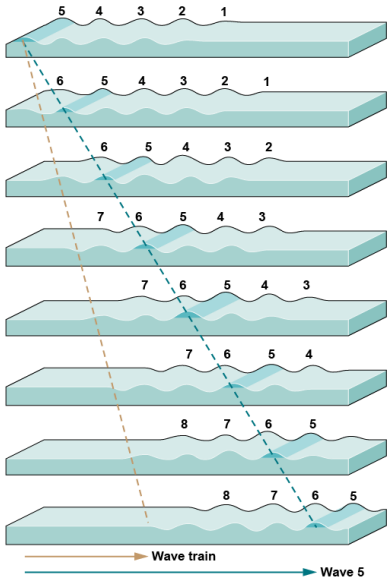


Figure 9-11. Waves travel in groups called wave trains. As the leading wave of the group travels forward, it transfers half of its energy forward to initiate motion in the undisturbed surface ahead and transfers the other half to the wave behind it to maintain the wave motion. The leading wave is constantly decaying (losing energy) as it transfers energy forward to displace the undisturbed water surface ahead of it. As a result, the leading wave in the wave train is constantly disappearing, while a new wave is constantly being formed at the back of the train. To visualize this process, follow wave number 5 in this diagram. Notice that the train travels at half the speed of an individual wave.

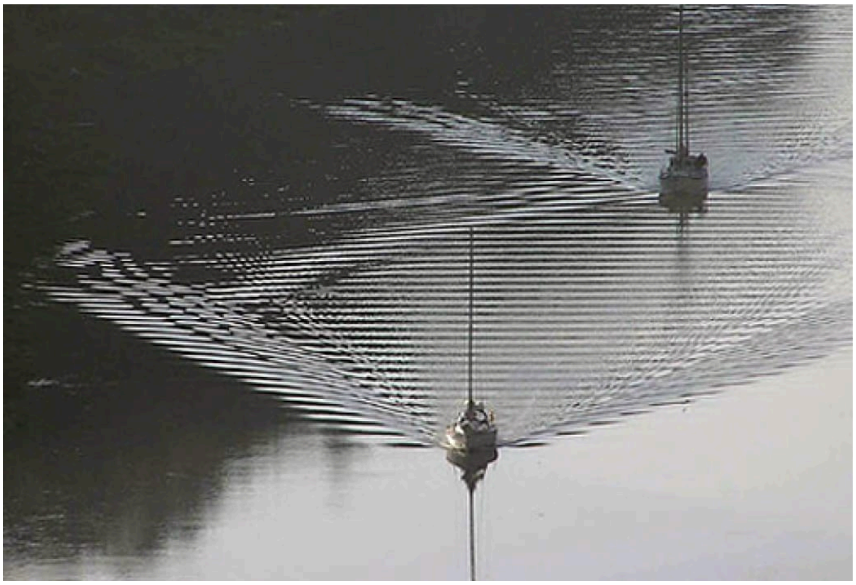


Figure 9-12. A boat's bow and stern waves travel out from the ship as wave trains. If you look carefully at the bow wave of the leading boat, you can see that the leading wave disappears as it moves away from where the boat has passed and is replaced by another wave behind it. The second wave in turn disappears to be replaced by another one behind it, and so on. The wave energy is also slowly spread as the wave crest lengthens by dispersion, and the wave train eventually disappears. Note how the bow and stern waves of the two boats combine to form complex wave interference patterns.

Wave Interference

Only very rarely are waves in a specific part of the ocean regularly spaced and all of the same height. The reason for the irregularity in height and spacing is that waves of several different periods and moving in different directions pass a given location simultaneously. If we plot the sea surface height as a function of time, we get the type of complicated wave pattern shown in **Figure 9-13a**. Surprisingly, the pattern in **Figure 9-13a** is simply the sum of the five simple waves shown in **Figure 9-13b**.

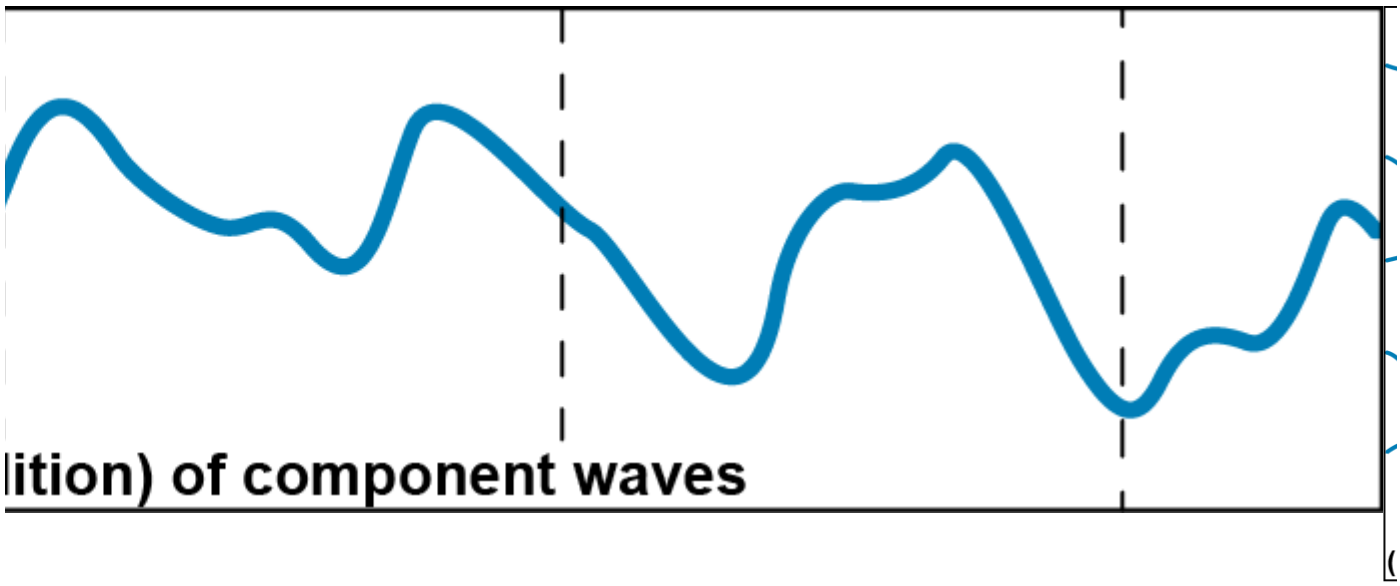


Figure 9-13. The waveforms observed at any point in the ocean are often complex. The complex waveform in part (a) is simply the result of interference (addition) of the five simple sine waves in part (b).

When wave trains of similar wave heights but slightly different periods arrive at the same time, the sea surface will appear to alternate between periods of low and high waves. During relatively calm periods (the segments where the waveform depicted in **Figure 9-14b** is almost flat), the crest of one wave arrives at almost the same time as the trough of another wave, and the elevations of the two waves cancel each other out (**Fig. 9-14**). During the intervals when the waves are high, the crests of the two waves arrive at the same time and combine to form a larger wave (**Fig. 9-14**). When waves of several different, but similar, periods arrive simultaneously, the situation is more complex. However, the process of wave addition and subtraction, called "wave interference," is what causes waves to vary in height in irregular patterns.

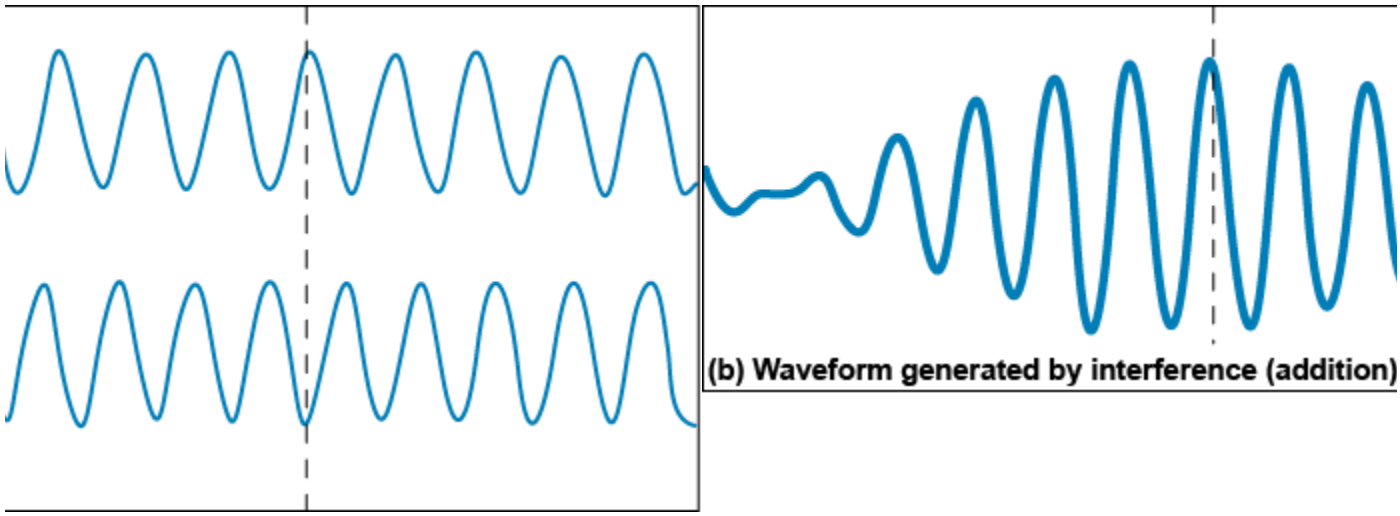


Figure 9-14. When two sets of waves of very similar wavelengths arrive at the same location at the same time, their addition can produce large, long-period variations of the wave height in the resulting waveform. (a) These two simple waves have slightly different wavelengths. (b) When they are combined, they produce a very different waveform. This is why we may see periodic sets of very much higher (and lower) waves than normal arriving at the shore.

The irregular pattern of waves caused by wave addition is important to surfers, who await sets of big waves. Contrary to some popular beliefs, the periodicity in wave heights does not always follow a consistent pattern. Sets of waves that are especially high may appear every few minutes or only every few hours at any given location on any day. People who visit coastlines when the sea is rough should be aware of this phenomenon. It is not safe to assume that a dry position on the rocks or beach is beyond the reach of the waves. More than a few unfortunate individuals are washed off the rocks or swept off beaches and drown every year because they have made that assumption. It is wise always to anticipate that waves much larger than those that have arrived in

the preceding hours might suddenly appear.

Wave Dispersion

Storms create waves with a wide range of periods. The speed of deep-water waves increases as period (and wavelength) increase (**Table 9-1**). Hence, waves with longer periods and longer wavelengths travel faster than shorter-period waves, and longer-period waves move away from the area where they were generated more rapidly than shorter-period waves do. Wave trains of different periods become separated as they move away from the storm center where they were created (**Fig. 9-15**). This phenomenon is called "wave dispersion" and causes waves to be sorted by period (and wavelength). Because waves are sorted by period as they move away from their origin, the longer-period waves from a storm arrive at locations far from the storm before the shorter-period waves do. If the storm that created the waves occurred about 100 km away, waves with a 5-s period will arrive about 3 h after waves with a 10-s period. If the storm occurred about 1000 km away, the 5-s waves will arrive about 36 h after the 10-s waves. The waves from the more distant storm will have traveled farther and been dispersed by wavelength to a greater extent (**Fig. 9-15**).

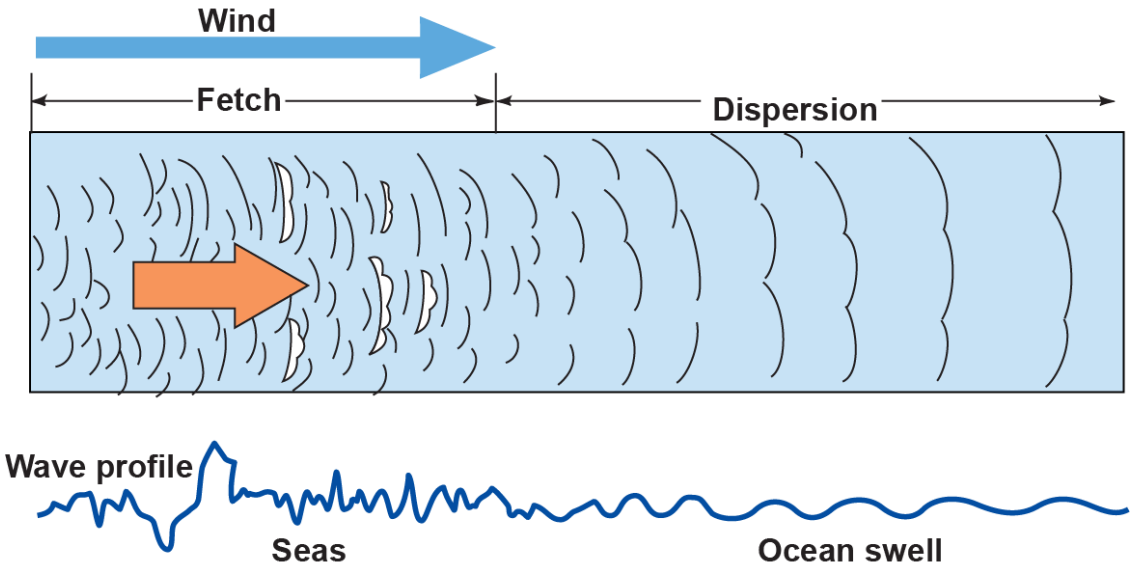


Figure 9-15. Within the fetch of a storm, the waves are irregular because they are the sum of many separate waves of different wavelengths, as shown in Figure 9-12. Long-wavelength waves move faster than shorter-wavelength waves, such that, as waves move away from the storm, the long-wavelength waves move ahead of the shorter-wavelength waves. This sorting of the waves by wavelengths is called wave dispersion. Because they are separated by wavelength, the waves appear to become smoother as they travel away from the storm fetch. The greater the distance from the fetch, the greater the difference in the arrival times of waves of two different wavelengths.Normally, waves of more than one storm arrive simultaneously, so the shorter-period waves from any given storm are usually obscured by longer-period waves from another storm. The shorter-period waves can be observed if detailed wave records are analyzed mathematically. Oceanographers are able to calculate the locations of storms many thousands of kilometers away by analyzing waves arriving at two or more locations. The wave dispersion at each location reveals the distance of the storm from that location, and the distances of the storm from two widely separated locations will reveal the storm location. For example, storms near Antarctica can be located by wave measurements made in locations such as the United Kingdom and Brazil. Although this method is reliable, satellite observations of cloud patterns are now a much easier way to locate storms.

As waves disperse by wavelength in their direction of travel, wave steepness decreases because the waves also spread laterally, and their energy is distributed over a wider area. The reduction in steepness is accompanied by a change in the wave shape, making it closer to the ideal smooth sine wave shape (**Fig. 9-2a**). Thus, at some distance from the area where they are generated, the waves are sorted by wavelength and form a swell—smooth undulations without sharp or breaking crests (**Fig. 9-15**). A swell originates from a sea and is caused by the combination of wave (wavelength) dispersion and lateral dispersion of wave energy (spreading of the wave front like the spreading of a ripple as it moves away from the point of impact of a stone) as the waves travel from their origin.

This page titled 9.7: Deep-Water Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.8: Waves in Shallow Water

On a calm-wind day at the shore, when there is a swell from a distant storm, we observe that offshore in deep water, the waves roll smoothly toward the shore without breaking. Once they reach shallow water, the same waves form breakers. Clearly, something happens in shallow water that alters the behavior of waves.

Interaction with the Seafloor

When waves enter water shallower than half their wavelength (*L*/2), they begin to interact with the seafloor. The wave motion becomes inhibited because the solid seafloor prevents water molecules from moving in circular orbits. Molecules immediately above the seafloor can move only back and forth. As a result, the orbits of water molecules within the wave are distorted into elongated ellipses (**Fig. 9-16**). Compression of the orbits and friction between water and seafloor slow the forward motion of the wave.

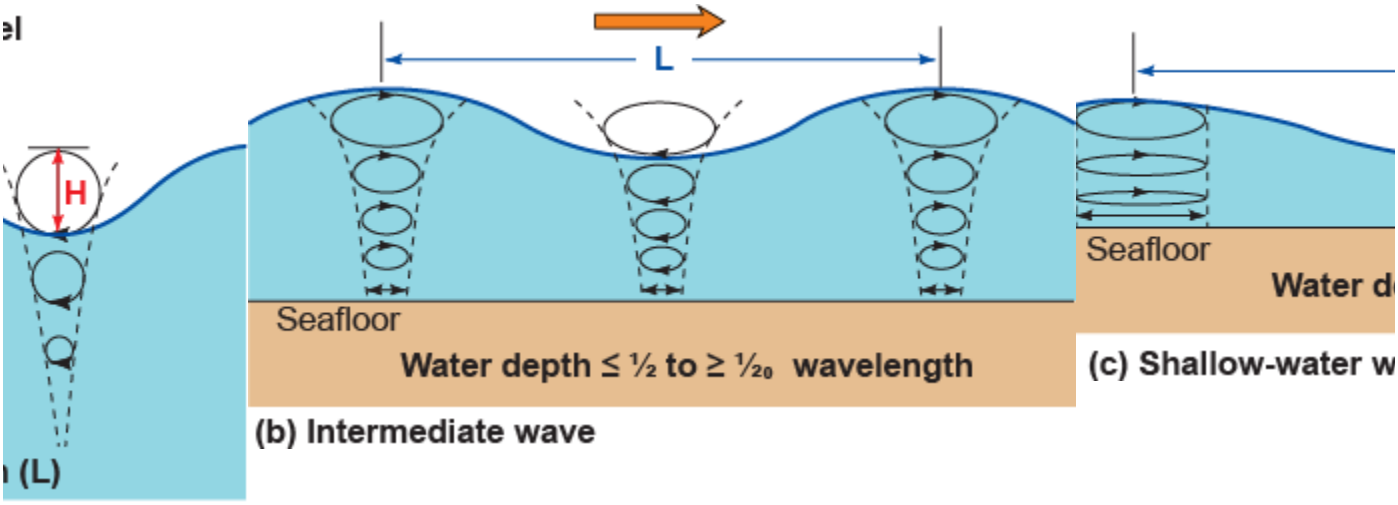


Figure 9-16. Effects of water depth on waves. (a) In waves traveling across the ocean where the water depth is greater than one-half of the wavelength, the orbits of water particles in the wave are circular and wave speed is determined only by the wavelength. (b) Where the water depth is less than one-half but more than one-twentieth of the wavelength, wave motion is slowed and distorted by friction with the seafloor, and the particle orbits are flattened into ovals. (c) Where the water depth is less than one-twentieth of the wavelength, the particle orbits are flattened so much that they become nearly horizontal, back-and-forth motions. In this case, the wave speed is determined only by the depth. Note that each part of the figure does not depict a single wave moving into shallower water. As explained later in the chapter, wavelength is progressively shortened once a wave enters shallow water and becomes an intermediate or shallow water wave. Also, the wave height decreases somewhat when it is an intermediate wave and then increases rapidly once it becomes a shallow water wave and moves into progressively shallower water.

As water depth decreases between *L*/2 and *L*/20, the forward speed of the wave decreases progressively (**Fig. 9-17**). The equation used to calculate the speed of waves when they are in water depths between *L*/2 and L/20 is somewhat complex. In these depths, waves are called **intermediate waves**. In contrast, at water depths of *L*/20 or less, the wave speed is controlled only by water depth. Waves in water depths less than *L*/20 are called **shallow-water waves**. The speed of shallow-water waves is given by the equation:

$$C = \sqrt{gD} = 3.13\sqrt{D} \tag{9.8.1}$$

where *C* is the celerity ($m \bullet s^{-1}$), *D* is the depth (m), and *g* is the acceleration due to gravity ($m \bullet s^{-2}$). Thus, all waves, regardless of their period, travel at the same speed when they are in water of the same depth, provided that the depth is less than *L*/20. Because wavelength and period are related, longer-period waves will become shallow-water waves in deeper water than short-period waves do (because *L*/20 is greater when *L* and *T* are larger).

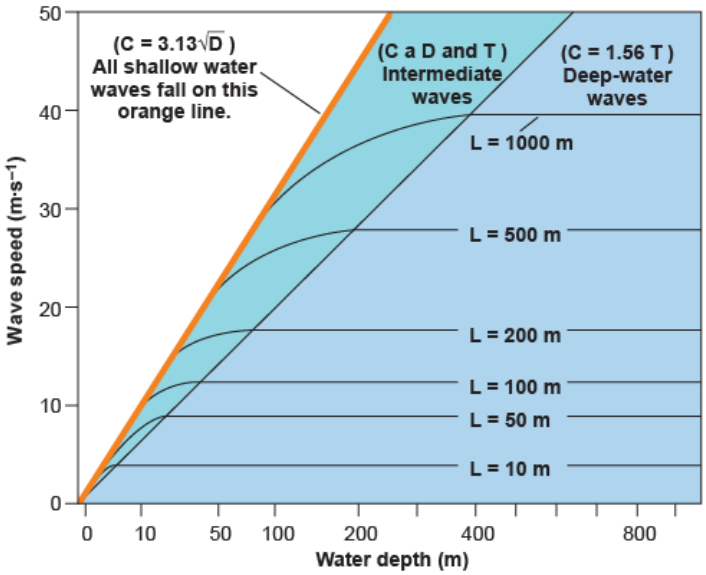


Figure 9-17. The speed (or celerity, *C*) of a deep-water wave is constant and depends on the wavelength (or wave period). For example, a deep-water wave of wavelength 1000 m has a speed of almost exactly 40 m•s^{-1}. The speed of shallow-water waves is determined only by depth and decreases as water depth decreases. To determine the speed of a shallow-water wave in this figure, simply find the speed on the orange line that corresponds to the water depth. The speed of intermediate waves is determined by both water depth and wavelength. For example, the speed of an intermediate water wave of wavelength 1000 m decreases from almost exactly 40 m•s^{-1} when the water is 400 m deep, to a little more than 30 m•s^{-1} when the depth equals one-twentieth of the wavelength (50 m) and the wave becomes a shallow-water wave.

As waves enter shallow water and are slowed, their period does not change. Because wave speed is the ratio of wavelength to period ($C = L/T$) and the period does not change, the wavelength must decrease as the speed decreases (Fig. 9-17). We can easily see why the wavelength decreases if we consider what happens to two successive waves following each other into shallow water. The first wave enters shallow water and is slowed before the second wave. The second wave comes closer to the first wave because it does not slow until it also reaches the shallow water. As they move into shallower water, both waves continue to be slowed, but the first wave continues to be slowed sooner than the second wave, which is always in deeper water. Therefore, the wavelength decreases progressively as waves move into shallower water and wave speed continues to decrease (Fig. 9-17). As wavelength decreases, wave steepness increases (Fig. 9-8).

Wave Refraction

Waves usually approach a shoreline at an angle. Consequently, because one end of the wave-crest line enters shallow water and slows while the rest of the wave is still in deeper water and traveling at its original speed, the wave is **refracted**, or bent (Fig. 9-18a). Lines perpendicular to the crest lines of successive waves in Figure 9-18a show the path that a specific point on the wave crest follows as it moves inshore. These lines are called **wave rays** or "orthogonals." As the wave continues to move inshore, more of the wave enters shallow water and slows. However, the section of the wave that first entered shallow water is still moving into progressively shallower water and slowing further, so wave refraction continues.

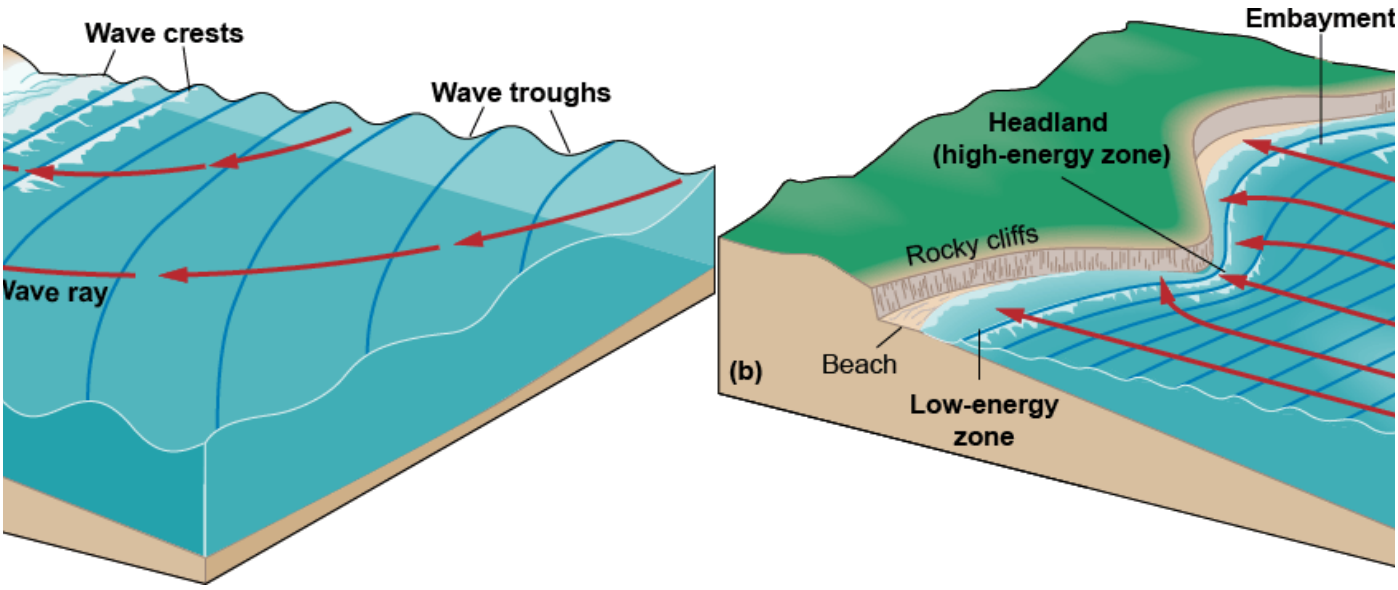


Figure 9-18. Wave refraction. (a) Waves generally approach the coastline at an angle. Consequently, part of the wave front reaches shallow water and is slowed before the rest of the wave. It continues to slow as the other parts of the wave move more rapidly toward shore until they, too, are progressively slowed. As a result, the wave front is bent and becomes aligned more closely parallel to the beach. The path followed by any point on the wave crest is called a wave ray. (b) As a wave approaches a bay flanked by headlands whose topography extends onto the offshore seafloor, the first parts of the wave to slow are those over the underwater extensions of the headlands, and the last part to be slowed is that over the deeper water at the center of the bay. Thus, the wave is refracted, and wave energy is concentrated on the headlands and spread out inside the bay. Breaking waves are the strongest at the headlands.

Note that the wave rays bend toward shallower water. As the refraction process continues to bend the wave, the wave crests tend to become aligned parallel to the shore. The refraction is seldom complete, so a wave crest rarely reaches the shore or breaks at precisely the same time along its entire length. However, even if waves approach the shore at a large angle, they are refracted and always reach the shore almost parallel to it (**Fig. 9-18a**).

If waves move into shallow water where seafloor ridges or depressions are present, the parts of each wave where the water is shallower slow down, while other parts in deeper water do not. Thus, the refraction pattern differs from the simple pattern depicted in **Figure 9-18a**. In many areas, seafloor **topography** immediately offshore is a continuation of land topography. For example, many coastlines have horseshoe-shaped bays where valleys reach the shore (**Fig. 9-18b**). The valley often continues offshore from the center of the bay, and the seafloor falls away in a valley-shaped depression between submerged ridges that extend out from the bay's headlands. Most bays of this type have a horseshoe-shaped sandy beach at the center and rocky headlands at each end (**Fig. 9-18b**). As we shall see, the headlands are rocky, and the bay's interior is sandy because of the wave refraction.

Figure 9-18b shows the locations of successive wave crests as waves move toward a bay. As a wave approaches shore, the first part of the wave to encounter shallow water will be that over the undersea ridge extending out from the headland closest to the direction from which the wave arrives. While this part of the wave slows, the rest of the wave continues at its original speed. Then another part of the wave slows where it encounters the submerged ridge extending from the second headland. Thus, while parts of the wave are slowed at each headland, the rest of it continues at its full speed into the center of the bay. As the wave front enters the bay, the center section travels farther in deep water than the parts that enter the bay at each side. The wave is refracted so that it breaks at almost the same time along the entire length of the beach. Note again that the wave rays always bend toward shallower water.

Refraction redistributes the wave energy. In deep water, the wave has the same amount of energy per unit length along the entire length of its wave crest. As the wave is refracted in a bay, the total length of the wave crest is increased (**Fig. 9-18b**), and the same wave energy is distributed over this greater length. Lengthening of the wave crest also results in a lowering of the wave height. In contrast, at a headland, refraction reduces the length of the wave crest and, consequently, increases the wave energy per unit area and the wave height.

Refraction focuses wave energy on headlands while spreading wave energy along the beach within the bay. This is why we generally swim on the beach near the center of the bay and not at the headlands. The same wave that breaks gently at the middle of the beach smashes violently at the headland. This is also the reason we should exercise caution while walking on rocks of a headland. The headland is where waves are highest and crash most violently ashore. In addition, on narrow headlands where the

bottom topography is appropriate, a single wave with the right period that approaches from the right direction may be refracted to hit the headland simultaneously from two directions (**Fig. 9-18b**).

The wave refraction that concentrates wave energy at headlands and spreads energy within a bay determines the character of the shore at these locations. At the headlands, where wave energy is focused, sand does not accumulate because it is carried away by the wave action, and the shore is steadily eroded. In contrast, within the bay, gentler wave action transports sand toward the shore, where it builds and maintains a beach (**Chap. 11**).

In areas where the seafloor has complex topography with offshore rocks, ridges, and depressions, wave refraction can be far more complicated than the simple patterns depicted in **Figure 9-18**. Sometimes we can tell where such underwater features are by carefully watching wave refraction patterns from a beach or headland.

Breaking Waves

When waves enter shallow water and interact with the seafloor, their height is altered by the interaction. At first, the wave height is reduced slowly as the water depth decreases below *L*/2. This loss in wave height is caused by flattening of the orbital paths of water molecules in the wave (**Fig. 9-16b**). However, in water depths of about *L*/10 and less, the trend reverses and wave height increases rapidly as water depth decreases (**Fig. 9-19**). The reason is that wavelength decreases as water depth decreases, each wave is "squeezed" by its neighbors, and kinetic energy is converted to potential energy.

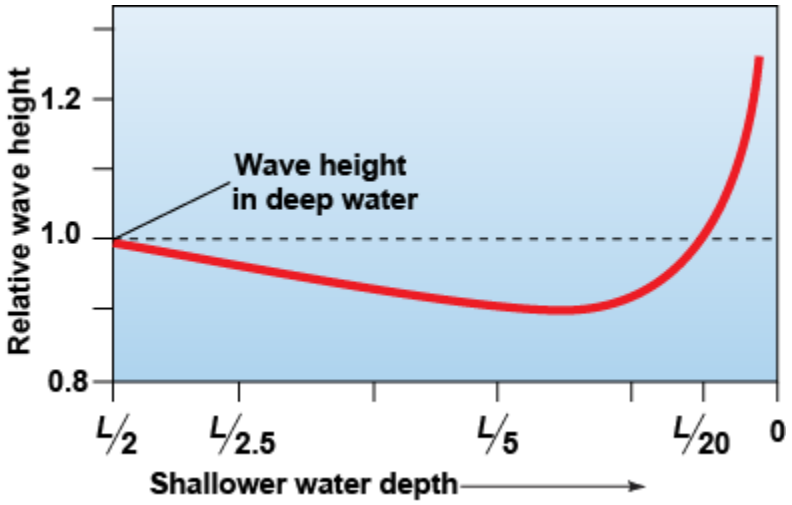


Figure 9-19. Wave height changes as a wave moves into progressively shallower water. The wave height declines slowly until the water depth is about one-tenth of the wavelength, then increases rapidly as depth decreases. Most waves become unstable and break before their height is much greater than it was in deep water (usually before the wave height exceeds 1.1 times its height in deep water). However, very long-wavelength waves, such as tsunamis, may be dramatically increased in height before they break.

Wavelength decreases faster than wave height as water depth decreases below *L*/2. Hence, wave steepness increases (*H* decreases but *L* decreases more rapidly, and *H*/L therefore increases). The speed and wavelength continue to decrease in water shallower than *L*/20 as the wave becomes a shallow-water wave (**Fig. 9-17**), but wave height increases (**Fig. 9-19**), wave steepness increases more rapidly as depth decreases (*H* increases, *L* decreases, and *H*/*L* increases rapidly), and wave shape is much modified.

Wave steepness increases until the wave becomes unstable and breaks. Wave heights and wavelengths vary among waves reaching the shore at any one time. Waves that follow each other, therefore, become unstable and break at different depths and thus different distances from shore. The area offshore within which waves are breaking is called the **surf zone**.

Although waves are refracted as discussed above, waves rarely reach the shore from a direction exactly perpendicular to the shoreline, and the seafloor rarely has exactly the same slope along the entire shoreline, so almost always one part of a wave breaks before another. A wave often breaks progressively along the crest of the wave as each part of the crest moves into water shallow enough to cause the wave to break.

Waves break in different ways that depend on several factors, including wave period, wave height, and the slope of the ocean floor. These factors determine how quickly the wave becomes oversteepened and unstable.

Spilling breakers are formed when the seafloor over which the wave is traveling is almost flat. When a wave reaches a water depth of about 1.2 times its wave height, it becomes unstable and the crest begins to tumble down the forward face of the wave (**Fig. 9-20a**). The forward face fills with churning, turbulent water and air bubbles that we see as a white foam. As the wave continues inshore, wave steepness increases slowly because of the almost flat seafloor, but as water spills off the wave crest, it reduces wave height and steepness. The spillage occurs fast enough to maintain the wave at its critical steepness against the tendency for the shallowing water to increase the steepness. Thus, spilling breakers break progressively as they travel inshore. Wave height is reduced progressively by the spilling action as the water depth decreases, until finally the turbulent wave crest encounters the seafloor and wave motion ends.

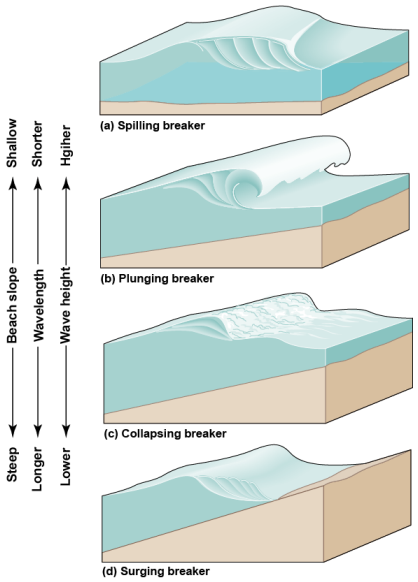


Figure 9-20. Breaking waves. (a) The peak of a spilling breaker becomes oversteepened, and it breaks by spilling water from the crest down the front face of the wave. (b) The peak of a plunging breaker travels fast enough to outrun the rest of the wave, and the crest curls over and smashes down on the front of the wave. (c) The base of the front side of a collapsing breaker becomes unstable before the crest, and the front face of the wave collapses. (d) A surging breaker never really breaks. The front face of the wave simply surges up and down on the beach, and the wave energy is partially reflected back out to sea.

Plunging breakers are the spectacular curling waves that many surfers covet (**Fig. 9-21**). They form when the seafloor slope is moderately steep. When the wave reaches the depth at which it becomes unstable, the bottom part of the wave is slowed more quickly than the upper part can slow or spill. Thus, the bottom of the wave lags behind as water in the wave crest outruns it while still traveling in its wave orbit. As a result, the wave crest curls over in front of the wave and plunges downward until it crashes into the trough preceding the wave (**Fig. 9-20b**).

Figure 9-21. This surfer is riding in the tube formed by a plunging breaker in the north end of the Banzai Pipeline, the famed North Shore surfing area of Oahu, Hawaii.

Collapsing breakers, which are relatively rare, occur where the seafloor has a steep slope, and the lower part of the wave is slowed so rapidly that the leading face of the wave collapses before the crest arrives. As the wave breaks, foam and bubbles are concentrated at the base of the forward face of the wave, and the crest collapses behind, usually with little splash (**Fig. 9-20c**).

On very steep shores, waves may appear not to break at all. The waves simply surge up and down a very steep beach with little bubble production (**Fig. 9-20d**). Surging breakers are rare, but the very small, gentle waves that lap onto some beaches in very calm conditions behave somewhat like surging breakers because even a flat beach has significant slope in relation to the tiny wave height. Surging breakers do not break up in foam and bubbles because most of the wave energy is reflected by the steep shore face.

Waves are reflected off vertical or nearly vertical solid objects such as cliffs or **seawalls** with little loss of energy, just as light is reflected by a mirror. When waves are reflected, they pass through and interfere with the incoming waves. Wave patterns created by such reflection can be very complex and are of great concern to engineers who build harbors, marinas, and other coastal structures. Waves are also diffracted by solid objects. Diffraction occurs when part of a wave is blocked by a solid object and the edge of the remaining wave spreads out after passing the object (**Fig. 9-22**).

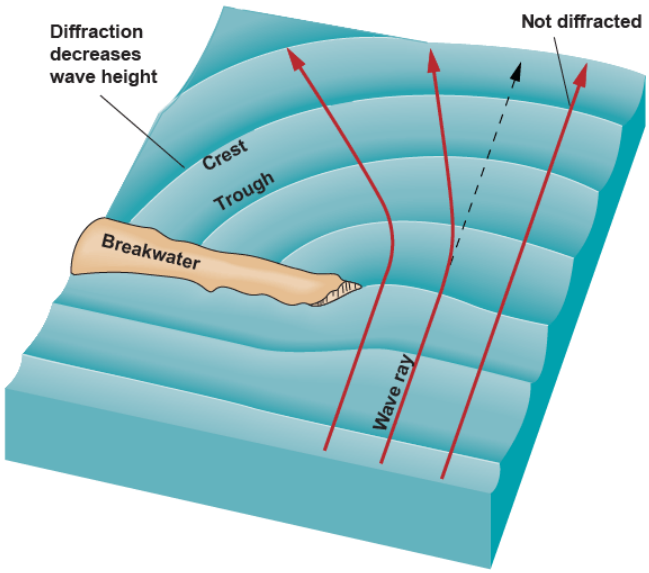


Figure 9-22. A wave is diffracted when the wave crest encounters a solid object, such as a breakwater. The part of the wave that passes by the end of the breakwater is bent such that the wave spreads out behind the breakwater.

Often, we see waves breaking at some distance from shore. Reduced in height, the same waves continue toward shore until they break again near the water's edge. This pattern indicates that the seafloor is shallower in the offshore surf zone than it is just inshore of that area. In tropical locations, the shallow area is commonly a fringing **coral reef** (Chap. 13). In other locations, it is a **longshore bar** (Chap. 11). **Fringing reefs** and long-shore bars are aligned parallel to many coastlines. They help to protect coasts from erosion by waves because they dissipate some wave energy, particularly from the highest and most energetic waves, before they reach the shore.

Surfing

To surf a wave, surfers must propel themselves forward on the board to join the wave motion just at the point where the wave becomes unstable and begins to break. They must position themselves on the forward face of the chosen wave and point the board "downhill." If the surfers "catch" the wave correctly, they will move forward with the wave. In the correct location, surfers are balanced such that their tendency to fall down the wave because of gravity is just offset by the pressure on the board from the water pushing upward in its orbital motion. Therefore, they must place the board where water is moving upward into the breaking crest. Plunging breakers (**Figs. 9-20b, 9-21**) provide unique conditions for surfing because most of the front face of the wave is traveling upward and then forward to join the curling crest.

Many surfers believe the best surfing is found where waves approach the shore from a large angle. In such places, refraction does not align the waves exactly parallel to the shore before they break. A wave breaks progressively along the crest as each section reaches shallow water. Surfers simply ride laterally along the wave crest. They remain on the section of the wave crest where the forward face of the wave is moving upward in its last orbital motion before breaking. Where plunging breakers break in this way, surfers can ride under the wave crest in the tube formed where the crest plunges forward ahead of the lower part of the wave (**Fig. 9-21**).

Humans are not the only animals that enjoy riding the orbital motion of waves. Porpoises ride ships' bow waves by positioning themselves where the water in the bow wave is moving forward in its orbital motion. They often ride this way for hours, "pushed" along by the ship.

Surf Drownings and Rip Currents

Unfortunately, many people drown while playing or swimming in the surf. Drownings are often blamed on "undertow" that supposedly sucks unwary individuals underwater away from the beach. However, "undertow" does not exist. People drown because of two simple phenomena, each of which can be easily avoided or escaped. When large waves break very close to the beach, the force of a wave can easily knock over a person who walks out into the surf. Once knocked over by a wave, an unwary individual may be washed seaward as the water flows back down the beach from the collapsed wave. The person is then at a place where the next wave crashes down. This wave washes the person first toward and then away from the beach as it breaks and the water runs back. In these circumstances, many people simply panic and drown because they cannot recover their balance and stand up.

If you find yourself in this situation, you will not drown if you follow two simple rules. First, grab a breath each time a wave has passed, before you are hit by the next one. Second, use the waves to your advantage. Let a wave carry you inshore. When your inshore motion ceases, do not try to stand up. Instead, either dig your hands and feet into the sand or start crawling toward the beach as you grab that next important breath. As each new wave arrives, let go and allow it to carry you farther inshore. If you take this approach, you will soon be sitting safely on the beach telling others how much fun you had playing in the surf. If you are a good swimmer, an alternative is to swim out beyond the breaking waves, where you can float comfortably and rest before swimming back to shore or calling for help.

The second major cause of beach drownings is **rip currents** (often incorrectly called "rip tides"). Waves create a small but significant net movement of water in the direction of wave travel (**Fig. 9-10b**). This net forward movement is increased in the surf zone because water in the wave crest outruns water in the wave trough as the wave breaks. Accordingly, in the surf zone, water is continuously transported toward the beach. However, it cannot simply accumulate on the beach, but must flow back through the surf zone. When the waves approach the beach at an angle, as they usually do, water that is transported onto the beach is also transported along the beach in the direction of the waves. Eventually the water encounters an area where it can flow back out as a rip current through the surf zone against the waves more easily than it can continue to accumulate and flow along the beach (**Fig. 9-23a**). In such areas, wave heights are generally somewhat smaller because a depression or shallow channel runs offshore from the beach.

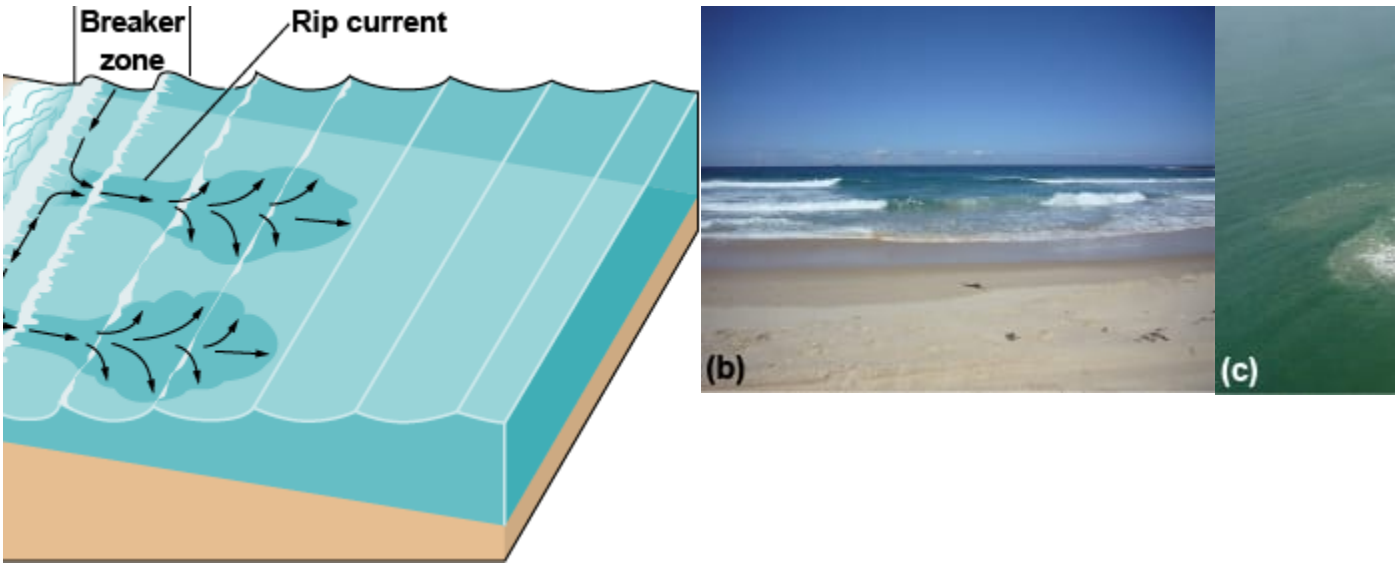


Figure 9-23. (a) Rip currents are fast, narrow currents that return water offshore after it has been transported toward the beach that form at intervals along a beach. (b) Rip currents are sometimes visible as foam or discoloration of the water caused by sediments suspended by turbulence in the swiftly moving current. This type of rip current is not in a fixed location on the beach and may appear quickly and dissipate just as quickly. (c) A more permanent type of rip current may form where water runs through a channel on the seafloor. This type can often be seen as a gap in the breaking crests of the waves.

The corridors in which water flows away from the beach are usually narrow and may be spaced well apart. The rip currents that flow offshore through these corridors consist of all the water that was transported onshore over a broader width of beach (**Fig. 9-23**). Because large amounts of water are moving through the narrow corridors across the surf zone, the flow rate or current through the return corridors can be fast.

Swimmers who enter a rip current will find themselves carried rapidly out into deeper water by the current. A rip current can be so fast that even the most accomplished swimmer cannot swim against it back to the beach. Tragedies occur when a panicked swimmer tries to fight the rip current, becomes exhausted, and drowns. However, even poor swimmers can easily avert such tragedies if they understand that rip currents are narrow, often only a few meters wide. A swimmer who meets a rip current should turn and swim parallel to the beach. A few strokes will bring the swimmer out of the rip current to an area where it is easy to swim safely ashore. A less desirable alternative is simply to ride with the rip current until it stops, usually only 200 or 300 m offshore, and then signal for help.

The locations of rip currents cannot always be seen easily, particularly by a swimmer in the water, but they are most likely to occur where depressions run down the beach into the surf. Rip currents may reveal themselves in plumes of increased turbidity (**Fig. 9-23b**), reduced wave heights in the surf zone, or, occasionally, as lines of floating debris or foam moving offshore.

This page titled 9.8: Waves in Shallow Water is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.9: Tsunamis

Although tsunamis are relatively rare and most tsunamis are small enough to cause no harm, a large tsunami can be devastating. For example, the December 2004 tsunami caused the deaths of many thousands of people around the Indian Ocean, and the March 2011 tsunami also killed thousands in Japan. Tsunamis as large as these are not frequent events but most people do not know that far bigger tsunamis have occurred in the past and will occur again.

Tsunamis can be generated when an earthquake or volcanic eruption moves a section of seafloor or coast, when a meteorite or coastal land collapse impacts the sea surface, or when an undersea landslide takes place. Each of these events can produce an abrupt displacement of the ocean water that causes the water column to oscillate, generating waves that generally have much longer periods than wind waves. Tsunamis are not related to tides, but they are often incorrectly called “tidal waves.” Tsunami is a Japanese word that is translated as “harbor wave,” but tsunamis are definitely not restricted to harbors.

Not all undersea earthquakes, volcanic eruptions, or landslides cause tsunamis. Tsunamis are most likely to be created when such seismic events cause a section of seafloor to move vertically or to slump. The sudden vertical movement either pushes up the seafloor and overlying water to form a wave crest or lowers the seafloor and overlying water to form a wave trough. The event may cause only one, almost instantaneous, vertical movement of the seafloor. Nevertheless, a series of several waves is created as the water oscillates before returning to a level configuration. This series resembles the waves created by a stone thrown into a pond, radiating outward from the stone’s impact point. Hurricanes and **extratropical cyclones** can cause a different phenomenon, called a **storm surge**, that elevates the sea surface and may cause flooding and damage similar to that caused by tsunamis. These storm surges are not considered to be wind-driven waves, although they behave in a similar manner. They are caused by elevation of the sea surface created by the lower atmospheric pressure at the center of the storm (**Chap. 7**).

Tsunamis consist of a series of waves with extremely long wavelengths (typically 100 to 200 km) and periods (typically 10 to 30 min). Only a very small fraction of the ocean is deeper than 6 km (**Fig. 4-4**), and half of the ocean floor is less than 4 km deep. Hence, the water depth is almost always less than one-twentieth of the tsunami’s wavelength, which causes tsunamis to behave as shallow-water waves. The speed of a shallow-water wave is determined by the water depth. In water 4 km deep, tsunamis travel at approximately 200 $m \bullet s^{-1}$ (720 $km \bullet h^{-1}$), or nearly the speed of a jet airliner. Because tsunamis are shallow-water waves, their speed changes with depth (**Fig. 9-17**), and they are refracted as they pass over seafloor topography (**Fig. 9-24**). Tsunamis can be spread out or focused by undersea ridges or depressions, just as wind waves are as they approach a coastline.

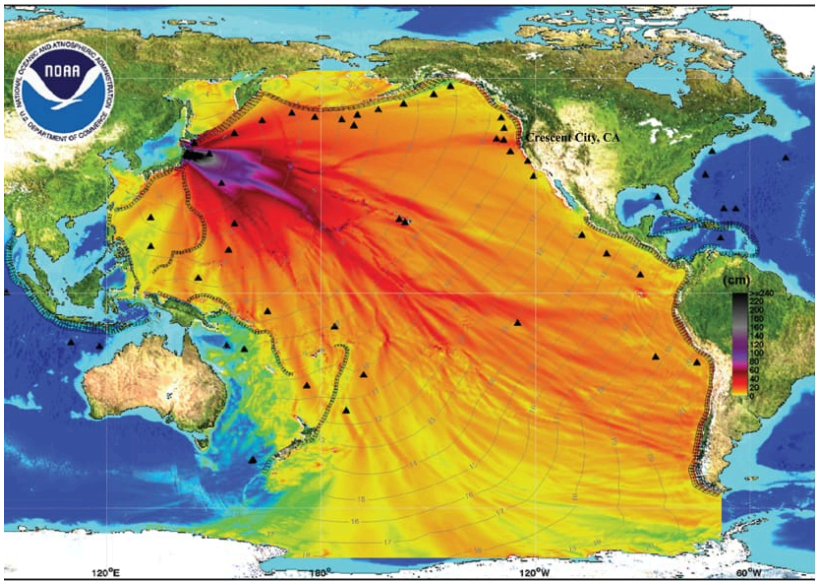


Figure 9-24. Computer model of the paths of the tsunami wave caused by the earthquake that occurred in 2011 offshore from Sendai, Japan. The wave energy of a tsunami is refracted by seafloor topography because tsunamis are shallow water waves. Refraction focuses or spreads the tsunami wave energy in the same way that wind wave energy is focused and spread at headlands and in bays. Note that one of the darker red streaks, which indicate paths along which the tsunami wave energy is focused, travels to Crescent City, California. Crescent City is extremely vulnerable to tsunamis, as are some Pacific Islands and locations along the South American Pacific Coast, due to this focusing effect.

When a tsunami travels over deep-ocean waters, its wave height rarely exceeds 1 or 2 m. Therefore, ships at sea are not affected by tsunamis. Indeed, it is almost impossible to detect a tsunami at sea because of its very long wavelength and limited wave height. A tsunami raises and lowers a ship only a meter or two, and each rise and fall takes several minutes. Tsunamis become dangerous only when they enter shallow water.

As a tsunami enters shallower water, it slows, and its wavelength is reduced, but its period is unchanged, as is true for any other wave (**Fig. 9-17**). As a wave slows, the wave height increases. Because water depth is very small in comparison to a tsunami's wavelength, wave height builds very rapidly (**Fig. 9-19**). The tsunami does not break because its wavelength is so long that even a large increase in wave height does not produce steep, unstable waves. Nevertheless, the leading edge of the tsunami wave can produce tremendous surf as it flows turbulently across the shore and coast.

The tsunami reaches the shore as a wave that can be tens of meters high and can take 5 to 10 min to pass from trough to crest. As the tsunami moves inshore, sea level rises several meters above normal, and enormous quantities of water are transported onshore and into any **estuaries** or rivers. The water keeps pouring onshore for several minutes as the wave crest approaches. The enormous energy stored in the wave is released as the water in the wave flows turbulently onto land and past any structures it encounters. Very strong currents and the equivalent of large breaking waves are generated as the water is concentrated in flows through harbors and channels and between structures. Buildings and trees can be destroyed, and boats and debris can be carried far inshore to be left stranded when the wave recedes (**Fig. 9-1**). If the ocean is not calm when the tsunami arrives, wind waves will add to the tsunami wave and may contribute to the destruction as they break far inland from the normal surf zone.

The impact that creates a tsunami may initially form either a trough or a crest. Accordingly, the first indication of the tsunami's arrival at the coast may be a rise or recession in the normal sea surface level that lasts several minutes and exposes large areas of seafloor that are not normally exposed. Regardless of whether a trough or crest arrives at the coast first, the tsunami will consist of several waves following each other. After the first wave, successive waves may be larger, but they will eventually decrease in height. Waves can continue to arrive for 12 h or more. Many drownings occur when curious sightseers or beachcombers walk out onto a beach exposed by a tsunami trough and are caught by the following crest. Despite its 10- to 20-min period, the crest of a tsunami moves onshore much too fast for someone on the beach to escape after the water begins to rise.

In 1883, the Indonesian island volcano Krakatau erupted with an extremely violent explosion that almost instantly blew a large fraction of the island's mass into the air. The explosion and the subsequent collapse of the remaining sides of the volcano into the underwater **caldera** (crater) created by the explosion caused a tsunami with waves of unusually long periods (estimated to be as much as 1 to 2 h). On the island of Java, about 60 km away from the eruption, the tsunami hit with waves about 30 m high. The waves destroyed many structures and carried a ship more than 3 km inland, where it was stranded almost 10 m above sea level. Krakatau's tsunami killed an estimated 35,000 people and was observed by water-level recorders as far as 18,000 km away in Panama.

In 1946, an earthquake in the Aleutian Trench off Alaska caused a tsunami at Scotch Cap, Alaska, that destroyed a concrete lighthouse 10 m above sea level, killed the lighthouse operators, and tore down a nearby radio mast mounted 33 m above sea level. About 5 hours later, the tsunami heavily damaged the Hawaiian Islands and swept away 150 people to their deaths. In response to this type of disaster, a tsunami warning system was developed for the Pacific Ocean. As a result, when a tsunami hit Hawaii in 1957 with waves larger than those in 1946, no lives were lost.

On 26 December 2004, a great earthquake struck in the **subduction zone** just offshore from the Indian Ocean coastline of the northern part of the island of Sumatra in Indonesia. The earthquake generated a tsunami that was estimated to be 30 m high when it hit the coast of Sumatra. The tsunami spread across the Indian Ocean, smashing into Thailand to the east and India and Sri Lanka to the west. Continuing across the Indian Ocean to the west, the tsunami then impacted the western coastline of Africa. More than 220,000 people were killed or disappeared, with the majority of deaths occurring in Indonesia. Many deaths also occurred in Thailand, Sri Lanka, and India; and several deaths were even reported in Somalia, more than 5000 km from where the earthquake occurred. In addition to the deaths, hundreds of thousands were rendered homeless, and the damage to property is estimated to have been many billions of dollars. In March 2011, a tsunami approximately the same size as the 2004 Indonesian tsunami hit Japan. This tsunami caused massive destruction and took many lives, but not nearly as many died in 2011 as in 2004. Some of the credit for the smaller death toll is due to the tsunami detection and warning system built after the 2004 event, and some also undoubtedly belongs to the Japanese preparedness. Earthquake and tsunami awareness are a high priority in Japan, and frequent emergency drills to respond to such events are a routine part of Japanese schoolchildren's lives.

No doubt most of you who are reading this will have seen videos of the December 2004 or the March 2011 tsunami. Although some of the videos are horrific, they are also instructive. Many of the video records show that a tsunami does not arrive as a single short-lived wave. They also show that the often turbulent front edge of the advancing wave is followed by the mass of water in the wave continuing to pour ashore in a relentless current for several minutes before finally receding. Eyewitness accounts in

both tsunamis also attest to the fact that several waves came ashore, 10 to 20 min apart, and that, in some places, the crest of the first wave arrived without warning, whereas in others, the trough arrived first, providing some warning as the ocean receded rapidly and much farther out than normal. The videos also show that it is not possible to outrun a tsunami wave unless one is lucky enough to be very close to the highest point the wave will reach. However, there were many survivors in each of these tsunamis. Some survivors saved themselves by running to higher ground at the first sign that a tsunami might hit, which was usually the shaking caused by the earthquake that generated the waves.

We should all learn from these events. It is imperative to seek higher ground immediately if we are on or near the shoreline and we either feel an earthquake or see the ocean rapidly recede in an abnormal fashion. We must also leave immediately if we see a massive wave moving toward shore, but unfortunately, fleeing might not guarantee safety. Finally, if a tsunami has occurred, even a small one, we must stay well away from the shore for at least 12 h.

Although the 2004 Indian Ocean and 2011 Japan tsunamis were events of epic human scale, they are both dwarfed by giant tsunamis that occurred before recorded history, and will probably occur again. For example, **Chapter 6** describes the monstrous tsunami that is thought to have been created when a meteorite hit the Earth at Chicxulub, Mexico. In addition, studies of the Hawaiian Islands have shown that huge tsunamis may be created when large segments of volcanic islands break loose and slump to the ocean floor (**Chap. 11**). A slump of 350 km^3 of the island of Hawaii's coastline about 120,000 years ago led to the deposit of marine **fossils** in a location on the island that was at that time at least 5 km inland and at least 400 m above sea level. Blocks of **coral** have been found at a height of 325 m on the island of Lanai, perhaps swept upward by this same tsunami. The earliest historical record of a tsunami appears to be almost 4000 years old, when the Minoan city of Ugarit was documented on clay tablets as having been destroyed by giant waves, possibly created by a massive volcanic eruption far more powerful even than the eruption of Krakatau.

In 2022, the Hunga Tonga volcano eruption caused a tsunami 20 m high. Only 6 people died, a testament to the success of tsunami awareness education programs around the Pacific Ocean. However, this event was notable because it caused two different tsunamis, a traditional one caused by displacement of the water by the explosion, plus a type never observed before, a meteotsunami. This meteotsunami was caused by the intense atmospheric pressure (shock) wave from the volcanic explosion, which apparently pushed a tsunami ahead of it as it passed over the oceans. This meteotsunami travelled even faster than the ocean wave tsunami and arrived at Japan's coastline about two hours before the ocean wave tsunami.

Tsunamis are most common in the Pacific Ocean because it has many subduction zones, where earthquakes are likely to cause vertical seafloor movements. However, tsunamis may occur wherever vertical movements of the land or seafloor occur. Tsunamis are most likely to be damaging on island or continental coasts with narrow continental shelves because much of the wave energy can be dissipated as a tsunami moves over the shallow waters of a wide continental shelf. A tsunami may cause no damage at all on **atolls** or other islands that have no continental shelf, because the tsunami has no opportunity to build in height before it reaches the coast. **Table 9-2** lists a number of destructive tsunamis that have occurred around the Pacific Ocean since 1990.

Table 9-2. *Selected Destructive Tsunamis since 1990*

Date	Location	Estimated Maximum Wave Height (m)	Number of Fatalities
2 September 1992	Nicaragua	10	170
12 December 1992	Flores Island, Indonesia	26	>1000
12 July1993	Okushiri, Japan	31	239
2 June 1994	East Java, Indonesia	14	239

Date	Location	Estimated Maximum Wave Height (m)	Number of Fatalities
14 November 1994	Mindoro Island, Philippines	7	49
9 October 1995	Jalisco, Mexico	11	1
1 January 1996	Sulawesi Island, Indonesia	3.4	9
17 February 1996	West Papua, Indonesia	7.7	161
21 February 1996	Northern coast of Peru	5	12
17 July 1998	Papua New Guinea	15	>2200
26 December 2004	Sumatra, Indonesia	30	>220,000
27 February 2010	Maule, Chile	8	ca. 500
11 March 2011	Sendai, Honshu, Japan	47	>25,000
15 January 2022	Hunga Tonga Ha'apai, Tonga	20	6

The tsunami warning system, headquartered in Hawaii, provides warnings of possible tsunamis whenever an earthquake occurs in the Pacific that might cause such waves. Often, tsunami warnings are not followed by a dangerous series of tsunami waves because the wave refraction patterns are complex and different for each tsunami. Therefore, even when a tsunami is actually observed near its source, predicting whether the tsunami will significantly affect any section of coast at a more remote location is very difficult. Even though some warnings may prove false, heeding such warnings is always wise.

In Hawaii and Oregon, tsunami warnings are broadcast through sirens at points along the coast and through the emergency radio and television broadcasting system. In the U.S. and some other nations, tsunami warnings are now broadcast on an emergency warning network. The US system is the Wireless Emergency System (WEA) that is active on all cell phones (but may be turned off, so make sure WEA is activated in your phone setting). If the tsunami is generated by a nearby earthquake the warning time will be very short so you should always leave the beach for higher ground immediately after you feel an earthquake. What feels like a "mild" earthquake may actually be a major earthquake centered some kilometers offshore, and a tsunami warning may be issued too late.

This page titled 9.9: Tsunamis is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.10: Internal Waves

Wind waves are oscillations that occur at the ocean surface where there is a sharp density discontinuity between water and the considerably less dense overlying atmosphere. Waves can be created at any density interface with a sharp density gradient where a fluid of low density overlies a fluid of higher density. Such sharp vertical density gradients, called **pycnoclines**, occur in the water column in many areas of the oceans (**Chap. 8**).

If a pycnocline is displaced up or down, the displacement will create **internal waves** that travel on the pycnocline. Relatively little is known about internal waves, but their principal causes appear to be tidal motions, the effects of topography on currents moving in ocean water mass layers, and shear stress created between water layers as they flow across one another. Occasionally, where the pycnocline is shallow, internal waves can be generated by propellers or bow waves of ships.

The wave height of internal waves can be great (up to 100 m) in comparison with that of surface waves because a given amount of energy causes waves of larger amplitude where the density difference across the wave interface is smaller. The density difference across the pycnocline is much smaller than that across the sea surface. Internal waves travel approximately one-eighth as fast as surface waves with similar period, and they typically have periods of 5 to 8 min and wavelengths of 0.6 to 1 km. They have much less energy than surface waves, but they can be very important to sea life, submarines, offshore oil platforms, and vertical mixing of ocean waters.

Internal waves behave just like surface waves as they enter shallow water and interact with the seafloor. They slow, their wavelength is reduced, and eventually their wave height increases until they break. Because of their long wavelength, internal waves generally break on the outer part of the continental shelf. In some locations, including the continental shelf offshore of New York and New Jersey, breaking internal waves may mix **nutrient**-rich, cold water from below the pycnocline into warmer, nutrient-poor surface waters. This mechanism may supply nitrogen, phosphorus, and other nutrients to **phytoplankton** and thereby support larger populations of fishes and other animals than would otherwise be present (**Chaps. 12, 13**). Internal waves may also move **sediments** up and down canyons.

Submarines must be wary of internal waves because such waves can substantially change the submarine's depth in an uncontrollable and unpredictable way. Internal waves are speculated to have caused the sinking and loss of the nuclear-powered submarine USS *Thresher*, with its crew of 129, in 1963. Some types of drilling and oil production platforms can also be affected by internal waves. These are designed to withstand surface waves and are constructed on a tower resting on the seabed or on pontoons that float well below the depth of significant surface wave motion. In 1980, a series of internal waves rotated a production platform to face almost 90° from its original direction, although the platform did not topple.

Internal waves are generated not only on the main pycnocline but also at the interfaces between water mass layers and by interactions of the tide wave (**Chapter 10**) with seafloor topography. Thus, internal wave action occurs throughout the depths of the oceans and is a major contributor to vertical mixing that forms part of the MOC by returning deep-ocean water progressively to the surface layers.

This page titled 9.10: Internal Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.11: Rossby and Kelvin Waves

Rossby waves, often called “planetary waves,” are caused by the interaction of the gradient in the magnitude of the Coriolis effect with latitude (**CC12**), and the **geostrophic** flow of water around a sea surface depression or with the flow of air around a depression or elevation of atmospheric pressure (**Chap. 10**, **CC13**). Thus, these waves occur both in the ocean and in the atmosphere. Their generation is a somewhat complex process, but in the ocean, they have wavelengths of hundreds to thousands of kilometers and wave heights of just a few centimeters. Because they are so long and low, their presence in the oceans was not confirmed until satellites were able to measure sea surface height very precisely across wide areas of ocean within short periods of time.

Rossby waves move only from east to west. However, they can appear to move from west to east if they are carried by a much faster-moving air or water-mass movement. For example, satellite images show atmospheric Rossby waves as clouds following the large-scale **meanders** of the **jet stream**. Though weather systems generally move from west to east with the main flow of air masses in the atmosphere, Rossby waves embedded in this flow travels from east to west. Rossby waves move at speeds that vary with latitude, but the speeds of oceanic Rossby waves are on the order of only a few kilometers per day. Thus, for example, a single wave can take months or even years to cross the Pacific Ocean at mid latitudes.

There are two types of Kelvin waves: coastal and equatorial. Coastal Kelvin waves have sufficiently long periods and slow speeds that they are deflected into and constrained to move along a boundary by the Coriolis effect. Coastal Kelvin waves thus flow along the coastal margin in a counterclockwise direction in the Northern Hemisphere and a clockwise direction in the Southern Hemisphere. At the equator, where the Coriolis effect is zero, there is a special case where Kelvin waves flow directly along the equator from west to east. Equatorial Kelvin waves travel about three times as fast as Rossby waves, so they can cross the Pacific Ocean in about 70 days.

Rossby and Kelvin waves and their effects are poorly understood, but they are believed to affect phytoplankton distribution in the open oceans. They are also believed to affect weather—dramatically in some instances—by creating periodic but irregular variations in the location of ocean currents, such as the Gulf Stream, and in surface water temperatures (and thus atmospheric energy) in various ocean regions. Kelvin and Rossby waves are also associated with the development and relaxation of the El Niño/Southern Oscillation (**ENSO,** **Chap. 7**).

This page titled 9.11: Rossby and Kelvin Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.12: Standing Waves

Standing waves are completely different from the various types of progressive waves described previously in this chapter. Standing waves are sometimes called “stationary waves” or “seiches.” Their crests and troughs alternate at fixed locations, and they do not progress across the water surface.

A simple standing wave can be produced in a rectangular cake pan partially filled with water. If you slowly lift one end of the pan a few centimeters and then set it down quickly, the water will slosh back and forth from end to end of the pan. As you tilt the pan, the water surface remains level, but as you set it down, the water must move quickly so that the surface returns toward the level position. However, the surface overruns the level position, and water continues to flow until the surface is tilted in the direction opposite the tilt that occurred when you first set the pan down. An oscillating motion of the water surface continues back and forth (**Fig. 9-25a**) as the wave energy is reflected off either end of the pan until friction slows and finally stops the motion. The wavelength of the standing wave equals twice the length of the pan, and a standing wave is formed rather than progressive waves because the wave is blocked by (cannot progress past) the ends of the pan.

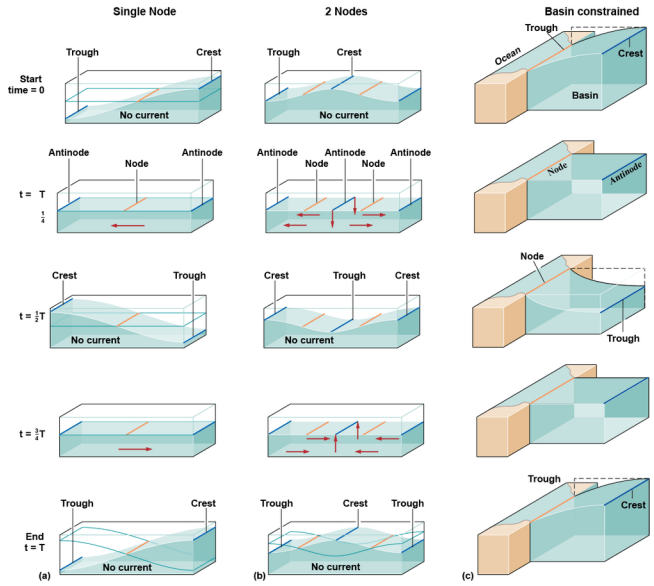


Figure 9-25. The motion of water in standing waves. The standing-wave motion is a seesaw oscillation of the surface. Standing waves have one or more nodes (where there is no change in the water surface height but there are oscillating horizontal currents) and one or more antinodes (where there are no currents but the surface oscillates up and down). (a) A standing wave with a single node. (b) A standing wave with two nodes. (c) A standing wave that is essentially a half wave with a node at the basin entrance and only one antinode, at the closed end of the basin.

The water surface at the center of a standing wave does not move up and down. This position is called a **node** (**Fig. 9-25**). At the **antinodes** (**Fig. 9-25**), which are at the ends of the standing wave, water within the waves can move only up and down; hence, there are no horizontal currents at these antinodes. Because water must be moved from one end of the wave to the other as it oscillates, water at the node moves back and forth horizontally. The current flows in one direction at the node as the wave moves in that direction, and then reverses when the wave moves back. Horizontal or nearly horizontal reversing currents occur at all points within the wave, except at the antinodes. The maximum speed of these currents is highest at the node and decreases with distance from the node until it is zero at the antinodes.

Standing waves are important within lakes and restricted ocean basins. They form especially easily in basins with steep sides because the sides reflect wave energy and little energy is lost by friction with the seafloor. Standing waves are refracted like progressive waves. Because water moves horizontally except at the antinodes, standing waves in very large basins, such as the Great Lakes of North America, are deflected by the Coriolis effect. The influence of the Coriolis effect on a standing tide wave is examined in **Chapter 10**.

The standing wave just described is essentially one-half of a wave because the crest and trough of the wave are at opposite ends of the basin. Standing waves with more than one node can be established in certain basins (**Fig. 9-25b**).

When a progressive wave of exactly the right period arrives at the entrance to a basin of exactly the correct length, a standing wave is established in the basin. Each successive crest of the standing wave must arrive at the basin entrance at the same time that a crest of a progressive wave arrives from the other direction. This type of **tuned** oscillation is often important in tidal motions within enclosed bays or estuaries (**Chap. 10**). The oscillation is tuned in much the same way that organ pipes are tuned to specific wavelengths of air oscillations that create different musical notes. The pure note in an organ pipe is a single-node standing wave, whereas harmonics are standing waves that have two or more nodes within the pipe.

This page titled 9.12: Standing Waves is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.13: CHAPTER SUMMARY

Complexity of Ocean Waves.

Waves usually seem to come in haphazard sequences of heights and periods. The reason is wave interference, the combination of waves of several different periods and from different directions.

Wave Motion.

Water within a progressive wave moves in almost circular orbits whose diameters decrease with depth and are almost zero at a depth of half the wavelength. There is a very small net forward motion of water with the wave. Waves are created when the water surface is displaced. Once started, the wave motion is sustained by a restoring force that tends to return the sea surface to a flat, level state. Gravity is the principal restoring force for most ocean waves, and surface tension is important for very short-period capillary waves. The Coriolis effect affects very long wavelength waves, such as tides and tsunamis.

Tides are created by gravitational attractions of the sun and moon. Some ocean waves are caused by earthquakes or the passage of vessels. Most ocean waves are created by winds. Winds blowing over a calm water surface create tiny capillary waves that increase sea surface roughness. The wind increases the wave height and causes capillary waves to be combined into waves with longer wavelengths.

Capillary waves are dissipated quickly as energy is lost through water viscosity. Most ocean waves are gravity waves and lose little energy until they reach the shore, break, and dissipate much of their energy as heat.

Wave Breaking and Wave Height.

Winds are usually variable, causing seas with waves of many different wavelengths and heights. If winds are strong and blow long enough, waves begin to break when their height is one-seventh of their wavelength. Maximum wave heights are determined by the wind strength and duration and by the fetch. Wave heights up to 34 m or more have been reported. Wave period and wave steepness increase when waves travel directly into a current.

Wave Trains and Wave Dispersion.

Progressive waves travel in groups called “trains” that move at half the speed of the individual waves. The leading wave in the train is continuously destroyed and a new one created at the rear. Wave trains having waves of different wavelengths combine to produce the confused sea state that is characteristic of storms. The speed of deep-water waves increases with wavelength, so waves of different wavelengths move away from a storm at different speeds. Thus, waves are sorted by wavelength dispersion as they move away from a storm. As wavelength dispersion occurs, wave steepness declines and a swell is formed.

Waves in Shallow Water.

When waves enter water shallower than half their wavelength, the water's orbital motion encounters the seafloor. As a result, wave orbits are flattened and both wave speed and wavelength are reduced, while period remains unchanged. Waves continue to slow as they enter progressively shallower water until the water depth is less than one-twentieth of the wavelength, after which the speed is determined only by depth. Waves are refracted in shallow water because any part of the wave in shallower water is slowed more than any part in deeper water. Refraction tends to turn waves to align parallel to the shore, focuses wave energy on headlands, and spreads energy within bays. Waves are reflected off barriers such as seawalls, and they are diffracted when part of the wave passes a barrier that blocks the rest of the wave.

As waves enter shallow water, wave height slowly decreases until water depth is about one-tenth of the wavelength and then increases rapidly as depth decreases further. Waves break when they become oversteep. There are four types of breakers: spilling breakers, in which the crest tumbles down the front of the wave; plunging breakers, in which the crest outruns the bottom of the wave and curls over before crashing down; collapsing breakers, in which the front face of the wave collapses; and surging breakers,

in which the water simply surges up and down the shoreface.

Waves transport water forward onto the beach. Water returns offshore through narrow rip currents at intervals along the beach. Rip currents are a major cause of drownings, but even weak swimmers can survive them if they are aware that rip currents are narrow.

Tsunamis and Storm Surges.

Tsunamis are trains of very long-wavelength (100 to 200 km) waves with periods of 10 to 30 min created by earthquakes or other such events that abruptly displace a section of seafloor. Tsunamis travel at speeds in excess of 700 km•h–1 but behave as shallow-water waves because their wavelength far exceeds the ocean depths. Tsunamis are rarely more than a meter or two in height until they enter shallow coastal waters, where their height builds rapidly. Tsunamis can do tremendous damage because they pour onshore for 10 min or more before the wave crest passes.

Hurricanes and other major storms can cause storm surges in which the water level ahead of the storm is elevated. Storm surge can reach far inland of the normal high water line.

Rossby and Kelvin Waves.

Rossby waves are very low-amplitude waves that move slowly from east to west, have wavelengths of tens or hundreds of kilometers, and are responsible for periodic but irregular variations in weather along their track. Kelvin waves flow along the coastlines of the ocean basins—counterclockwise in the Northern Hemisphere, clockwise in the Southern Hemisphere, and directly west to east at the equator.

Internal Waves.

Internal waves form on pycnoclines. They have long periods and wavelengths, have large heights in comparison with surface waves, and tend to break near the edges of continental shelves, where they enhance the vertical mixing of water.

Standing Waves.

Standing waves are formed in basins where the ends of the basin prevent progressive waves from passing. At the node of a standing wave, the water surface does not move vertically but there are horizontal reversing currents. At the antinodes, water motion is vertical and there are no horizontal currents.

9.13.1: Key Terms

9.13.2: Chapter Questions

9.13.3: References

This page titled 9.13: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.13.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- amplitude
- atolls
- beach
- caldera
- capillary waves
- climate
- coastlines
- continental shelf
- coral reef
- coral reef
- Coriolis effect
- crest
- current
- deep-water wave
- density
- ENSO
- erode
- estuaries
- extratropical cyclones
- fetch
- fossils
- frequency
- fringing reefs
- geostrophic
- gravity
- gravity waves
- greenhouse effect
- hurricane
- intermediate waves
- internal waves
- jet stream
- joules
- kinetic energy
- leeward
- longshore bar
- longshore currents
- meanders
- nutrient
- phytoplankton
- potential energy
- pressure gradient
- progressive waves
- pycnoclines
- reef
- refracted
- rip currents
- scuba
- sea
- seawalls

- sediments
- shallow-water waves
- shear stress (friction)
- shore
- shoreline
- slumps
- standing wave
- storm surge
- subduction zone
- surf
- surf zone
- surface tension
- swell
- tide
- topography
- trade winds
- trough
- tsunamis
- turbidity currents
- turbulent
- wave dispersion
- wave heights
- wave interference
- wave period
- wave rays
- wave speed
- wave steepness
- wavelength
- westerly
- windward

This page titled 9.13.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.13.2: Chapter Questions

Study Questions

1. Why do successive waves that arrive at the beach have different heights?
2. You are standing at the end of a pier with a stopwatch. How would you measure the average wavelength of the waves passing by?
3. What is a restoring force, and why is it necessary if waves are to develop?
4. List the factors that determine the maximum wave height in a sea. How do they differ in the Atlantic and Pacific Oceans?
5. Why do deep-water waves sometimes break instead of just getting bigger?
6. Why is it not a good idea to enter a narrow harbor mouth when the tide is flowing out of the entrance and large waves are entering the harbor mouth from offshore?
7. Why do waves disperse as they move out from a storm? What would you see if you stood at the beach and observed the waves for many hours after the first waves from a distant storm arrived on an otherwise calm sea?
8. What are wave trains? Why do they move more slowly than individual waves?
9. What are the principal differences between internal waves and surface waves? Why do these differences exist?
10. What are rip currents, and why do they occur?
11. Why do tsunamis travel almost as fast as jet aircraft? Why would you not even notice a passing tsunami if you were on a boat far out to sea in deep water?
12. Describe standing waves. How do they differ from progressive waves? How do they resemble progressive waves?

Critical Thinking Questions

1. If the surface tension of water were much smaller than it is, how would you expect this to affect the formation of waves on the ocean? How would the waves behave differently when they approached a beach?
2. You are sailing across the water into an almost regular swell that comes from directly ahead of you. Suddenly you notice that the same swell now appears to be coming at you from two slightly different directions at a small angle from each side of the boat's bow and the wave pattern ahead looks confused compared to the smooth swell you were sailing on a few minutes before. What are you seeing, and why? Should you be concerned?
3. Sailing ships exploring the oceans before there were maps were able to safely enter lagoons of atolls or those behind fringing coral reefs, even though in many such reefs there are only a few narrow entrances where the water is deep enough for safe passage. (a) How did they do this? (b) How might they have done this differently on days when the ocean surface was extremely calm, lacking even a gentle swell?
4. Describe how waves would be refracted if they entered a very long basin whose underwater sides described a perfect V shape, with the depth of the V slowly decreasing with distance into the basin. Draw your answer showing the shapes of several successive waves as they travel up the basin. Include wave rays in your diagram.
5. In some parts of the oceans, there are often strong seas with breaking waves. In other areas, the waves are usually smooth swells. In yet other areas, the sea surface is often calm, and finally, in some areas, ships may encounter large swell waves with steepened fronts. Using what you have learned in this chapter, identify on a map at least one area of the Pacific Ocean where you would expect to find each of these situations. Explain the reasons for your choices.
6. In strong storms at sea, ships often alter course, for safety, so that they travel directly into the direction from which the wind and waves are coming, even though this may be far from their intended course of travel. Why do you think it is safer to travel into the wind and seas than at an angle across this direction?
7. In extreme storms, ships may turn and travel in the same direction as the wind and waves are traveling, even if doing so takes them in almost the opposite direction from their intended course. Why do you think this might be safer than traveling into the wind and waves?
8. Are there internal waves in the atmosphere? If so, where would you be most likely to find them? If there were internal waves in the atmosphere, would you expect their wavelengths to be longer or shorter than those of internal waves in the oceans? Why?

This page titled 9.13.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

9.13.3: References

Figure 9-1 – Welch, G., 2011: A large ferry boat rests inland amidst destroyed houses, ID 110320-M-0145H-063. United States Navy, accessed 26 June 2025, Source.

Figure 9-5 – Adapted from Kinsman, B., 1965: *Wind Waves, their generation and propagation on the ocean surface*. Prentice Hall, 676 pp.

Figure 9-7d – Adapted from Sverdrup, H. U., M. W. Johnson, and R. H. Fleming, 1942: *The Oceans: their physics, chemistry, and general biology*. Prentice Hall, 1087.

Figure 9-8c – Kharif, C., E. Pelinovsky, and A. Slunyaev, 2009: Observation of Rogue Waves, *Rogue Waves in the Ocean*. Springer, Berlin, Heidelberg, 11–31, Source.

Figure 9-12 – Arpingstone, 2004: Two boat wake, Avon Gorge. Accessed 9 December 2025, Source.

Figure 9-17 – Thurman, H. V., 1988: *Introductory oceanography 5th Edition*. Charles E. Merrill, 515 pp.

Figure 9-21 – Segar, D., and E. Stamman Segar, 1997: Surfer riding a tube, Oahu, Hawaii.

Figure 9-23b,c – NWS, 2012: Rip current science. Accessed 6 January 2026, Source.

Figure 9-24 – NTCR PMEL, 2011: Energy of tsunami of The Honshu, Japan tsunami of 11 March 2011. NOAA, accessed 27 June 2025, Source.

This page titled 9.13.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10: TIDES

Critical Concepts

- **CC12 The Coriolis Effect**: Water masses move freely over the Earth's surface while the solid Earth itself is constrained to move with the Earth's rotation. This causes moving water masses, including long wavelength waves that comprise the global tide wave motion, to appear to follow curving paths across the Earth's surface. The apparent deflection is called the Coriolis effect. Coriolis deflection can create a rotary motion of the tide wave, called amphidromic systems, within certain ocean basins. Amphidromic systems rotate counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere.

Figure 10-1. These two sets of photographs were taken from the same location 6 h apart at the port of Anchorage, Alaska, at approximately (a,b) high tide and (c,d) low tide. The tidal range this day was almost 11 m. For scale, note the cargo containers on the stern of the ship. They are the type hauled on 18-wheel trucks or flatbed railcars.

Most of us are familiar with the water movements that slowly expose and then cover the seaward part of the **shore** during a day. This rise and fall of sea level is called the **tide**. If we observe tidal motions long enough at one location, we will see that they are periodic. At some locations, the tide rises and falls twice during a day; at others, it rises and falls only once. The times of high and low tide vary predictably, and the height of the tide also changes somewhat from day to day. Indeed, the **tidal range** is often very different for the two tides on a single day. In addition, we may observe that the sea-level change between high and low tide is large in some places, while in other places there appears to be no tide at all.

Tides are important to mariners because many harbors and channels are not deep enough for vessels to navigate at low tide. Tides are an energy source that has been harnessed for electricity generation in some parts of the world. They are also important to many marine creatures, especially **species** that live in the **intertidal zone** between the **high-tide line** and **low-tide line** and must cope with alternate periods of immersion in water and exposure to air.

Tides cause tidal **currents** that can be very swift in coastal waters and within harbors and **estuaries**. In estuaries, tidal currents reverse direction as the tide rises and falls. When the current flows in from the sea, it is a **flood** current. When it flows out, it is an **ebb** current. The reversals between flood and ebb currents do not necessarily occur at high or low tide. However, where only one tide occurs each day, there is one flood and one ebb; and where two tides occur each day, there are two floods and two ebbs.

In the deep ocean, tides produce generally weaker currents than in coastal waters. However, throughout the ocean depths, tidal motions are responsible for generating currents that interact with density differences between water layers and with seafloor topography to cause internal waves, eddies, and turbulence that is a major contributor to the vertical mixing of deep layer water upwards as part of the Meridional Overturning Circulation (MOC) discussed in **Chapter 8**.

It has been known for more than 2000 years that tides are related to the movements of the sun and moon. However, the relationship was not fully explained until 1686, when Sir Isaac Newton published his theory of **gravity**. Tides are a phenomenon that are controlled entirely by a few simple physical principles, yet they can be extraordinarily complex.

10.1: Tide-Generating Forces

10.2: Characteristics of Tides

10.3: Tides on an Ocean Covered Earth

10.4: Tides in the Earth's Oceans

10.5: Tides in the Open Ocean

10.6: Tidal Currents

10.7: Tide Prediction

10.8: Marine Species and the Rhythm of Tides

10.9: Energy from Tides

10.10: Chapter Summary

10.10.1: Key Terms

10.10.2: Chapter Questions

10.10.3: References

This page titled 10: Tides is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.1: Tide-Generating Forces

Tides continuously raise and lower the sea surface. The vertical movement of the water surface is as much as several meters, twice a day in some areas. Clearly large amounts of energy are necessary to move the water involved in this process. The energy is supplied by the gravitational attraction between the water and the Earth, moon, and sun—but how? Tides are cyclical, but they do not come and go at the same times each day, and the height to which the water rises or recedes varies from day to day. Why does this happen? To answer these questions, we must first learn some basic information about gravity and the motions of the moon around the Earth and the Earth around the sun.

Gravity

Newton's law of gravitation states that every particle of mass in the universe attracts every other particle of mass with a force that is proportional to the product of the two masses and inversely proportional to the square of the distance between the two masses. The gravitation force (*F*) between two masses can be expressed as

$$F = G \times \frac{(M_1 \times M_2)}{D_1} \tag{10.1.1}$$

where *G* is the gravitational constant (a constant defined by Newton's law and equal to 6.673 ×10^{-11} m^3•kg^{-1}•s^{-2}), M_1 and M_2 are the two masses, and *D* is the distance between them.

For approximately spherical objects, such as planets, the entire mass can be considered to be at the object's center of gravity, and the distance *D* is measured between the centers of the two masses. The gravitational force increases as either mass increases or the distance between the two objects decreases.

We are familiar with gravity because it is the force that attracts all objects on the Earth's surface, including ourselves, to the much greater mass of the Earth itself. In addition, the Earth and every object on it are subject to gravitational attractions toward the sun, the moon, other planets, and every other celestial body. In the following discussion in this chapter, we refer to Earth's gravitational force that we experience on its surface as gravity, while we refer to the gravitational force between celestial objects such as the Earth and moon as gravitational attraction. This is done simply to minimize possible confusion as the phenomena are the same and the laws of gravity apply to both.

Gravitational forces are extremely small in relation to the gravity we experience on Earth's surface. Although the sun, stars, and some planets each have a mass much greater than the Earth's mass, they are much farther away from us than we are from the Earth's center. For example, gravitational attractions that the sun and moon exert on any object at the Earth's surface are approximately 0.06% and 0.0003% of the Earth's gravity, respectively. Gravitational attractions of other celestial bodies are much smaller.

Although they are almost negligible in comparison with the Earth's gravity, gravitational forces exerted by the moon and sun on the Earth are the cause of the tides. To understand how these forces cause the tides, we must first consider the characteristics of orbits in which celestial bodies move.

Orbital Motions and Centripetal Force

Any two bodies that orbit together in space, orbit around their common center of mass. This works like a seesaw. If the two riders are of equal weight, the seesaw balances when the riders are at equal distances from the center. Similarly, two planetary bodies of equal mass would orbit around a point exactly midway between them. However, if one rider is heavier than the other, the heavier rider must sit closer to the center of the seesaw to make it balanced. Similarly, two planetary bodies of unequal mass orbit around a point closer to the more massive body. The Earth's mass is 81 times that of the moon. Therefore, the common point of rotation is about 4680 km from the Earth's center, or approximately 1700 km below the Earth's surface (**Fig. 10-2**). Similarly, because the sun has far greater mass than the Earth, the common point of rotation of the Earth and the sun is deep within the sun.

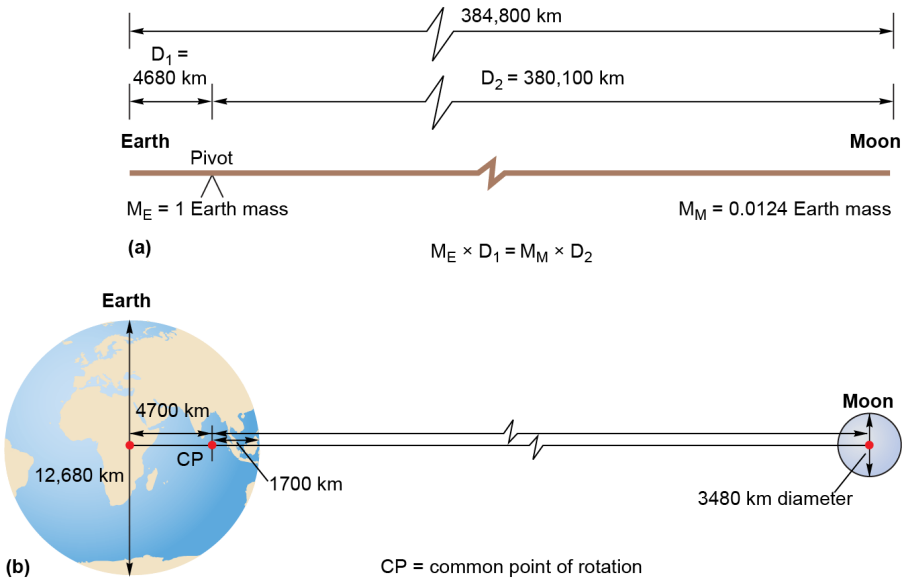


Figure 10-2. Any two bodies in space orbit around a common point of rotation. Just as, on a seesaw, the heavier person must sit nearer the pivot point to provide balance, the common point of rotation for two bodies orbiting in space must be closer to the more massive body. (a) Thus, the Earth's center must be much closer to the common center of rotation than the moon's center is. (b) Because the Earth is much more massive than the moon, their common point of rotation is approximately 1700 km below the Earth's surface.

As two planetary bodies orbit each other, they must be constrained in their orbit by a **centripetal force** that prevents each from flying off in a straight line (**CC12**). The centripetal force is supplied by the gravitational attraction between them. The centripetal force required is the same for all parts of each of the two bodies. Because centripetal force varies with distance from the center of rotation (**CC12**), this may be difficult to understand. However, within either of the orbiting bodies, all particles of mass move at the same speed in circular paths of the same diameter. We can understand this motion if we examine the motions of the two objects as they orbit each other without considering other motions, such as the Earth's spin on its axis or orbits that involve a third body.

Figure 10-3 shows a non-spinning planet (a hypothetical "Earth") orbiting with another planetary object (a hypothetical "moon"). In this diagram, the common center of rotation, which is also the center of mass of the two bodies, is located beneath the surface of the larger planetary body (the hypothetical Earth). Careful examination of **Figure 10-3** reveals that, as the two bodies rotate around each other, each particle of mass on the planet moves in a circular orbit, and all of the orbits have the same diameter. However, as **Figure 10-3** shows, the centers of the circular orbits are all displaced from the common center of rotation, except the orbit of the planet's center of mass. Because the particles of mass in the planet move in circles of the same diameter and complete the circles in the same amount of time, all particles require the same centripetal force to maintain their orbits. This is true no matter where the common center of rotation is. It is true for the Earth-moon system, where the center of rotation is inside the Earth, and for the Earth-sun system, where the center of rotation is inside the sun.

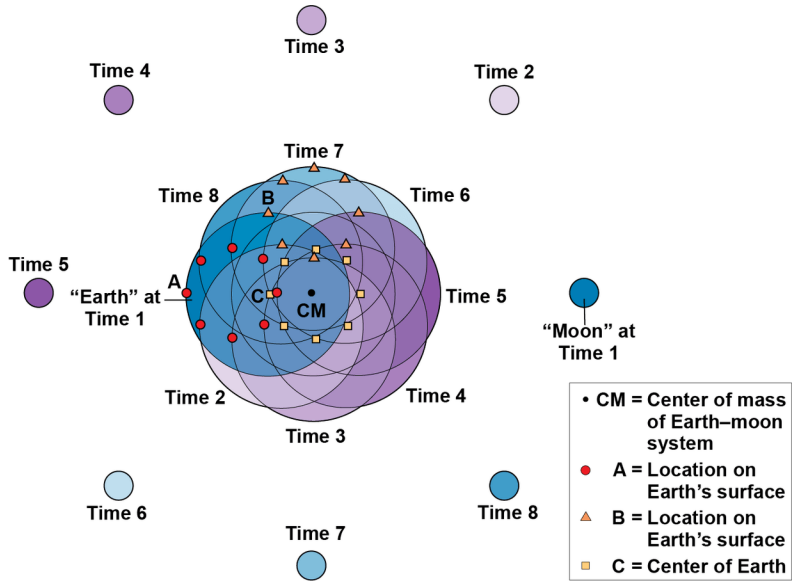


Figure 10-3. This figure is a representation of the location of the Earth in relation to the moon as they orbit around their common center of mass, which is inside the Earth. In this representation the Earth is assumed to be nonrotating (not spinning on its axis). As the two bodies rotate around their common center of rotation, points A and B, which are at the Earth's surface, and point C, which is the center of the Earth, would move through the color-coded locations shown in the figure. Notice that all points on or within the Earth move in circular paths that have the same diameter. This is also true for all points on or within the moon, or on or within any other orbiting bodies. Because the orbital diameter and rotation rate are always the same, the centripetal force needed to keep each body in orbit is equal at all points within each of the two bodies. Note that a centripetal force is also needed to maintain parts of the Earth in the circular motion of the Earth's spin, but this is a separate motion with a different distribution of its centripetal force and does not affect the centripetal force between the Earth and moon.

The Balance between Centripetal Force and Gravitational Force

The total gravitational force between the moon and the Earth (or the sun and the Earth) must be equal to the total centripetal force needed to maintain these bodies in their orbits, or else the orbits would change. Although all particles of mass within the Earth are subject to the same centripetal force, the gravitational force on each particle of mass within the Earth varies with its distance from the moon or sun. For example, the distance from the moon's center to the Earth's center is approximately 384,800 km, and the Earth's diameter is approximately 12,680 km. Therefore, a point on the Earth's surface nearest to the moon is only 378,460 km from the moon's center, and a point farthest from the moon is 391,140 km from the moon's center. Because the gravitational force is inversely proportional to the square of the distance, the change in the moon's gravitational force between the point on the Earth nearest to the moon and the Earth's center is 378,4602 divided by 384,8002, or about 1 to 1.034. Hence, the moon's gravitational attraction is about 3% greater at the Earth's surface nearest the moon than it is at the Earth's center (the average gravitational attraction between moon and Earth). Similarly, the moon's gravitational attraction at the Earth's surface farthest from the moon is about 3% less than the average gravitational attraction.

The average centripetal force for each orbiting body must equal the average gravitational force between the two if the orbit is to remain stable. Because the centripetal force is the same at all points on the Earth and gravitational force is not, there is a net excess gravitational force on the side of the Earth nearest the moon, and a net deficit of gravitational force on the side farthest from the moon (**Fig. 10-4**). These imbalances are tiny in relation to the Earth's gravity and are easily compensated in the solid Earth by small changes in **pressure gradient** (the gradient of pressure within the atmosphere, ocean, or solid Earth that increases toward the Earth's center). Pressure within the Earth (and other objects) represents the force needed to resist the gravity tending to pull all material toward the Earth's center). For people, the compensation in the pressure gradient causes a change in weight. Each of us weighs about 0.000,000,1 kg (0.1 mg) less when the moon is over the opposite side of the Earth than when it is directly overhead, much too small a difference to notice.

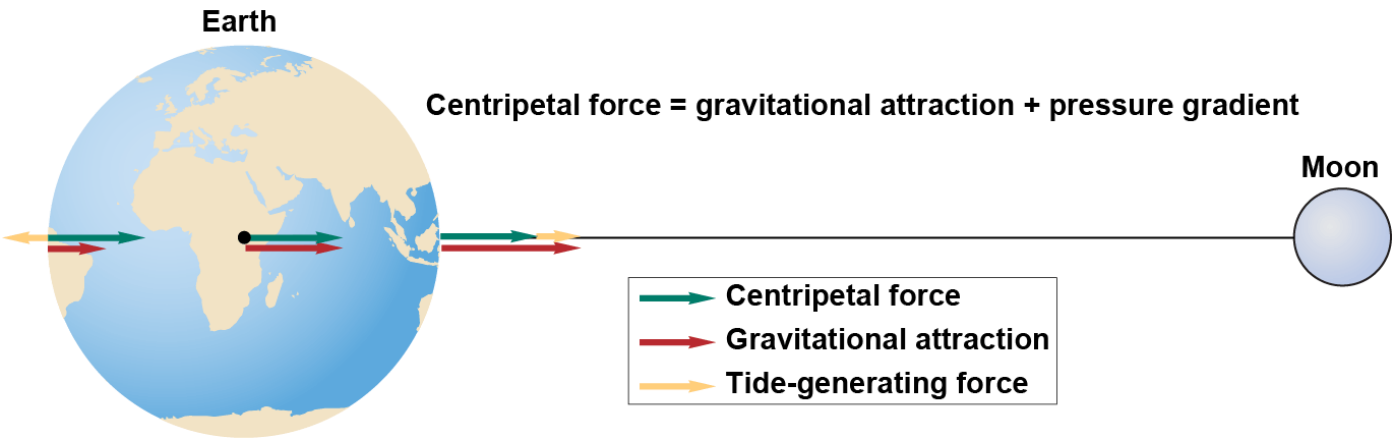


Figure 10-4. The total gravitational force between the Earth and moon (or sun) must equal the total centripetal force needed to maintain the two bodies in their common orbit. However, because the gravitational force varies slightly at different points on the Earth, whereas the centripetal force is the same at all points, there is a slight imbalance between them everywhere except at the exact center of mass of the Earth. On the side of the Earth nearest the moon, the gravitational force due to the moon is slightly higher than it is at the Earth's center. The slight excess of gravitational force over centripetal force at this point is easily compensated by the pressure gradient. Similarly, there is a slight excess of centripetal force over gravitational force on the Earth at the point directly away from the moon. Thus, objects at points directly under the moon weigh very slightly less than they do at points where the moon is directly overhead.

Unlike solid objects, ocean water can flow in response to the imbalances between gravitational force and centripetal force. These flows are the tides. For convenience, we refer to the net excess of gravitational force over centripetal force as the "tidal pull" or "tide-generating force."

Distribution of Tide-Generating Forces

At the point on the Earth's surface nearest the moon, there is a slight tidal pull. Water is pulled upward toward the moon at this point. However, the Earth's gravity, which is comparatively very strong, acts directly opposite the tidal pull at this location. Therefore, the tidal force at this point (and at the opposite side of the Earth, which is farthest from the moon) has little effect, just as the tidal force has only an extremely small effect on our weight.

At other points on the Earth's surface, the tidal pull is exerted in a direction different from that of the Earth's gravity because the direction between that point on the Earth's surface and the moon, and the direction between that point and the Earth's center of mass, are aligned at an angle. Thus, at every point except directly under the moon and exactly on the other side of the Earth, the tidal pull has a component that acts parallel to the Earth's surface (**Fig. 10-5**). This component of tidal pull, which is referred to as the "tidal force" cannot be compensated by the Earth's gravity and therefore causes water to flow in the direction of the force.

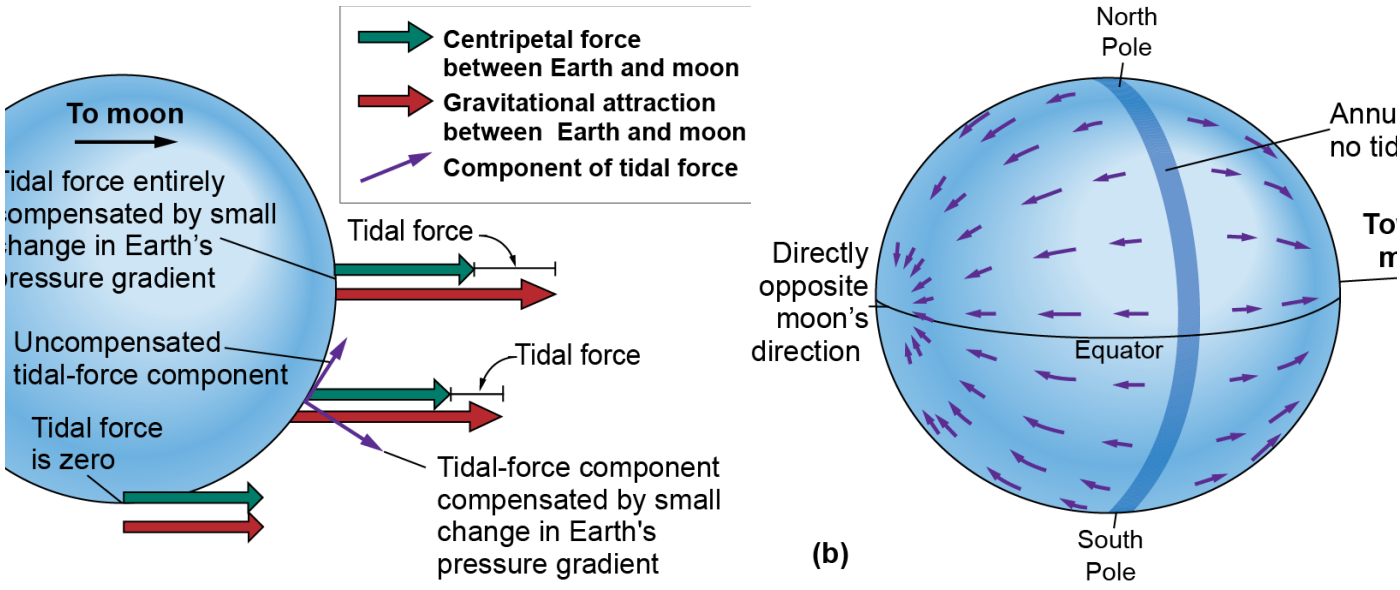


Figure 10-5. Horizontal tidal forces. (a) Components of gravitational attraction, centripetal force, and the Earth's pressure gradient. The vertical component of the tidal force is easily compensated by a minute change in the pressure gradient. Magnitudes of the tidal force and tidal force components are highly exaggerated in the figure. (b) The horizontal tidal force is zero at a point directly under the moon (or sun) and a point located exactly on the opposite side of the moon. It is also zero in a ring (annulus) around the Earth that is equidistant from these two points. The horizontal tidal force increases to a maximum and then decreases between the point directly under the moon and the annulus where the force is zero. The same is true between the point directly away from the moon and the annulus, but the tidal force is in the opposite direction.

The net result of the tidal forces acting on the Earth's oceans is to move water toward the points nearest to the moon and farthest from the moon (**Fig. 10-5b**). This movement creates bulges of elevated water surface at these points, and a depression of the water surface in a ring around the Earth approximately halfway between these points. The diagrams in this book necessarily greatly exaggerate the bulges. Even on an Earth totally covered by oceans, the bulges would be less than a meter high.

The tide bulges are oriented directly toward and away from the moon. Because the Earth is spinning, the point on the Earth facing the moon is continuously changing. Therefore, as the Earth spins on its axis the bulge remains aligned with the moon, but the locations on the Earth that are under the tide bulge change.

Relative Magnitude of the Lunar and Solar Tide-Generating Forces

Although all planets and stars exert tidal forces on the Earth, only tidal forces of the moon and the sun are significant. Tidal forces are caused by the small difference in gravitational force from one side of the Earth to the other. This difference depends on the distance between the Earth's center and the other body's center, and on the other body's mass. Although we need not perform the calculation here, we can show mathematically that the magnitude of the tidal force is proportional to the mass of the attracting body (sun or moon) and inversely proportional to the cube of the distance between the centers of the two bodies (r^3). This means that the tidal force is much more dependent on the distance between bodies than it is on mass.

For the Earth–moon system, the tide-generating force (F_M) is given by:

$$F_M = K \times \frac{Mass of Moon}{(Earth - to - moon Distance)^3} \tag{10.1.2}$$

and for the Earth–sun system, the tide-generating force (F_S) is given by:

$$F_S = K \times \frac{Mass of Sun}{(Earth - to - moon Distance)^3} \tag{10.1.3}$$

where K is a constant that is always the same for tidal forces between the Earth and any other planetary body.

The sun's mass is about 27 million times greater than the moon's. However, the sun is about 390 times farther from the Earth.

The ratio of solar to lunar tidal force can be calculated with the following equations:

$$\frac{F_S}{F_M} = \frac{Mass of Sun}{Mass of Moon} \times \frac{(Earth - to - moon Distance)^3}{(Earth - to - sun Distance)^3} \tag{10.1.4}$$

$$\frac{F_S}{F_M} = \frac{27,000,000}{1} \times \frac{(1 \times 1 \times 1)}{(390 \times 390 \times 390)} \tag{10.1.5}$$

$$\frac{F_S}{F_M} = 0.46 \tag{10.1.6}$$

Despite the sun's much greater mass, its tide-generating force is only about 46% of the moon's tide-generating force because the sun is much farther away from the Earth.

This page titled 10.1: Tide-Generating Forces is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.2: Characteristics of Tides

Tides are measured by a variety of gauges that continuously monitor the sea surface height. The tide is described by a plot of the water surface height as a function of time, called a "tidal curve" (**Fig. 10-6**). The tidal curves in **Figure 10-6** show that tidal motions are similar to the **progressive waves** described in **Chapter 9**. In some locations, the tidal curve resembles a simple progressive wave. In other locations, it resembles the complex waveforms produced when waves of different **wave periods** and **wave heights** interfere. Tidal curves are, in fact, the net result of several tide waves of different periods, as explained in the next section.

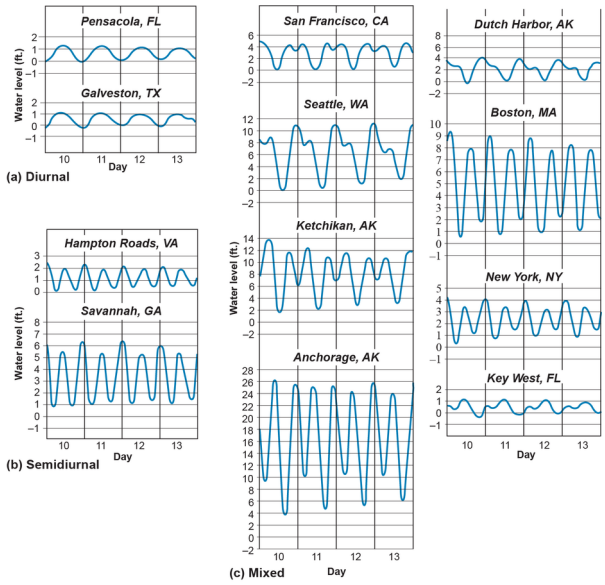


Figure 10-6. Selected tide records. (a) Diurnal tides occur at Pensacola, Florida, and Galveston, Texas. (b) Semidiurnal tides occur at Hampton Roads, Virginia, and the Savannah River entrance, Georgia. (c) Mixed tides occur at most locations, including San Francisco, California; Seattle, Washington; Ketchikan, Anchorage, and Dutch Harbor, Alaska; Boston, Massachusetts; New York, New York; and Key West, Florida. Note that even the Hampton Roads and Savannah River tides are not pure semidiurnal tides, but they are nearly so. Tidal heights are in feet because tide tables issued by the U.S. government still use these units.

The most important characteristics of a tidal curve are the times and relative elevations of high tide and low tide, and the tidal range, which is the difference between the height of the high tide and that of the low tide (**Fig. 10-6**). Because the tide is a wave, the tidal range is an expression of the tide wave height.

Diurnal, Semidiurnal, and Mixed Tides

Tides are classified into three general types based on the number and relative heights of the tides each day at a given location along a **coast**. **Diurnal tides** (or daily tides) have one high tide and one low tide in each tidal day, which equals approximately 24 h 49 min (**Fig. 10-6a**). The next section explains why the tidal day is 49 min longer than the solar day. **Semidiurnal tides** (or semidaily tides) have two high and two low tides each tidal day, and thus a period of 12 h 24½ min. For pure semidiurnal tides, the two high tides (and the two low tides) each day are equal in height (**Fig. 10-6b**).

Mixed tides also have two high tides and two low tides each tidal day, but the heights of the two high tides (and/or of the two low tides) in each tidal day are different (**Fig. 10-6c**). Mixed tides have a higher high water (HHW) and a lower high water (LHW), as well as a higher low water (HLW) and a lower low water (LLW), each day. **Figure 10-6c** shows that the relationship of these four daily extreme levels varies greatly among different locations.

Mixed tides are the most common. On the North American continent, tides are mixed along the entire Pacific coast, on the Atlantic coast of Canada north of Nova Scotia, in the Caribbean Sea, and in parts of the Gulf of America (Golfo de México) (**Fig. 10-7**). Tides are primarily semidiurnal (or mixed) along the Atlantic coast of the U.S. and southern Canada, and are diurnal only in certain parts of the Gulf (**Fig. 10-7**). However, pure semidiurnal and diurnal tides are rare. All tides have some mixed characteristics,

but the classifications are useful to mariners and to illustrate the complex tidal motions. In some locations, tides may change in character during the tidal month. For example, they may be semidiurnal for one part of the month and mixed for another part.

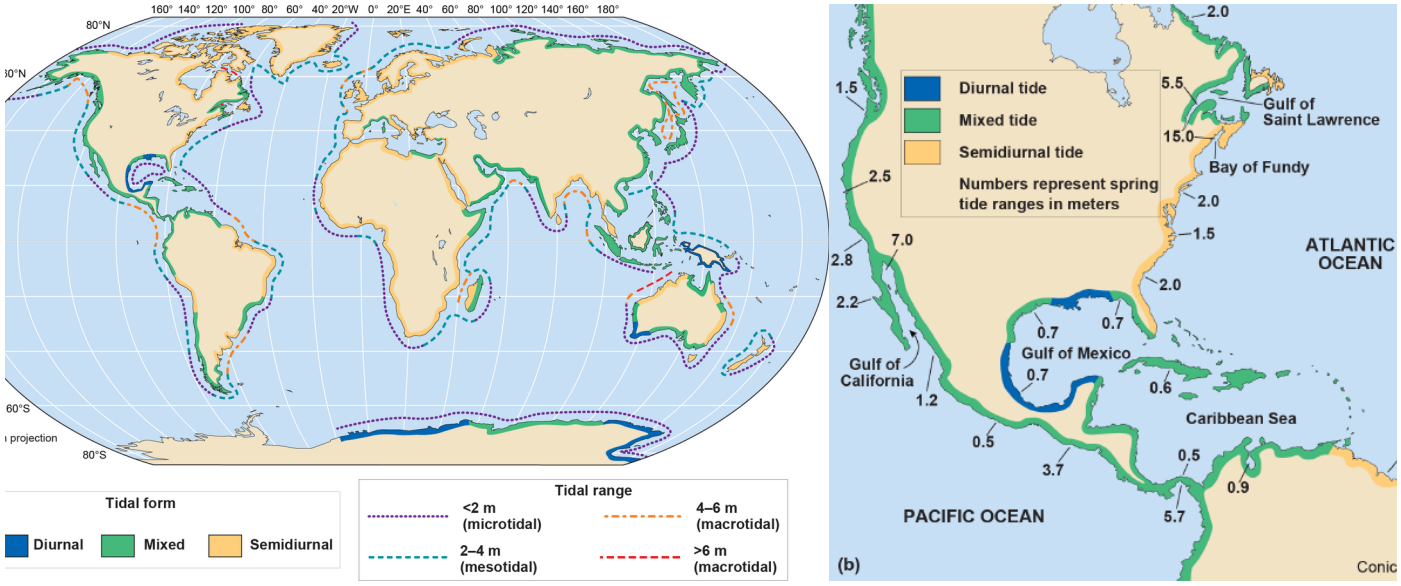


Figure 10-7. Although the characteristics of tides may change during the month at a given location, all locations have tides that can be characterized as primarily diurnal, semidiurnal, or mixed. (a) Global distribution of tides. (b) North American tides. Most of the coasts on the Atlantic Ocean have semidiurnal tides, and most coasts on the Pacific Ocean have mixed tides. Diurnal tides are relatively rare. This pattern is generally true for the North American continent, except that much of the Gulf of America (Golfo de México) has diurnal tides, and parts of the Atlantic coast of Canada have mixed tides.

Spring and Neap Tides

If we look at tide records for various locations over a period of a month, we see that the daily tidal range varies during the month (**Fig. 10-8**). This variation occurs at all locations, regardless of whether tides are diurnal, semidiurnal, or mixed. Although the variations are often complex, the daily tidal range generally reaches a maximum twice during each lunar month (29½ days). The tidal range thus oscillates back and forth twice each lunar month. Tides with the largest tidal range during the month are spring tides. Tides with the smallest tidal range are neap tides. Two sets of spring tides and two sets of neap tides occur each lunar month.

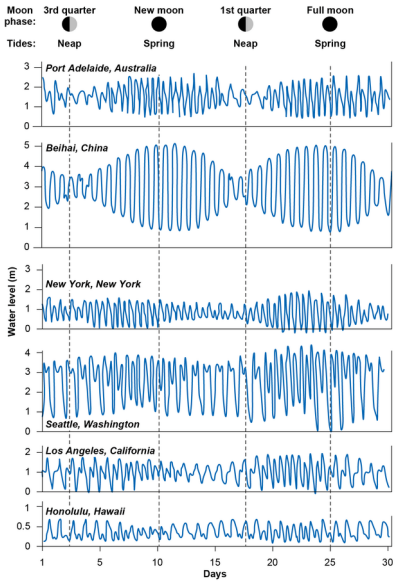


Figure 10-8. Monthly tide records show the twice-monthly occurrence of spring tides associated with new moon and full moon, and the twice-monthly occurrence of neap tides associated with the quarter moons. This pattern occurs at all of the locations shown, but it is more pronounced in some areas, such as Port Adelaide, Beihai, and New York, New York.

If we look at tide records for an entire year or for several years, we find that the tidal range varies during the year and between years. However, these variations are much less than the variation between spring and neap tidal ranges during the lunar month.

This page titled 10.2: Characteristics of Tides is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.3: Tides on an Ocean Covered Earth

We can understand many features of tides and their variability by examining the theoretical effect of the sun and moon on a hypothetical planet Earth entirely covered by deep oceans with no **friction** between water and the seafloor. This approach is the basis of equilibrium tide theory.

The Fundamental Equilibrium Tides

Earth spins on its axis once every 24 h. At the same time, the moon orbits the Earth in the same direction as the Earth's spin, but much more slowly (**Fig. 10-9**). By the time the Earth completes one rotation (24 h), the moon has moved forward a little in its orbit. The Earth must turn for an additional 49 min before it "catches up" with the moon. Hence, 24 h and 49 min elapse between successive times at which the moon is directly overhead at a specific location on the Earth's surface. This is why the moon rises and sets almost an hour later each night.

On a hypothetical planet covered by oceans and without friction, we can consider what would be the effect of the moon alone during a solar day (**Fig. 10-9**). The moon would create upward bulges on the oceans at points aligned directly toward and directly away from the moon. As the Earth spun, the tidal forces would continue to pull water toward these points. The tide bulges would remain aligned with the moon as the Earth rotated, so the bulges would appear to migrate around the Earth each day. However, the moon does not always orbit directly over the equator. Instead, the angle between the moon's orbital plane and the equator, called the **declination**, varies with time during the lunar month and on longer timescales (**Fig. 10-9a**). For now, we need only consider the much simplified case of the moon at its maximum declination to see how diurnal, semidiurnal, and mixed tides might result (**Fig. 10-9b**).

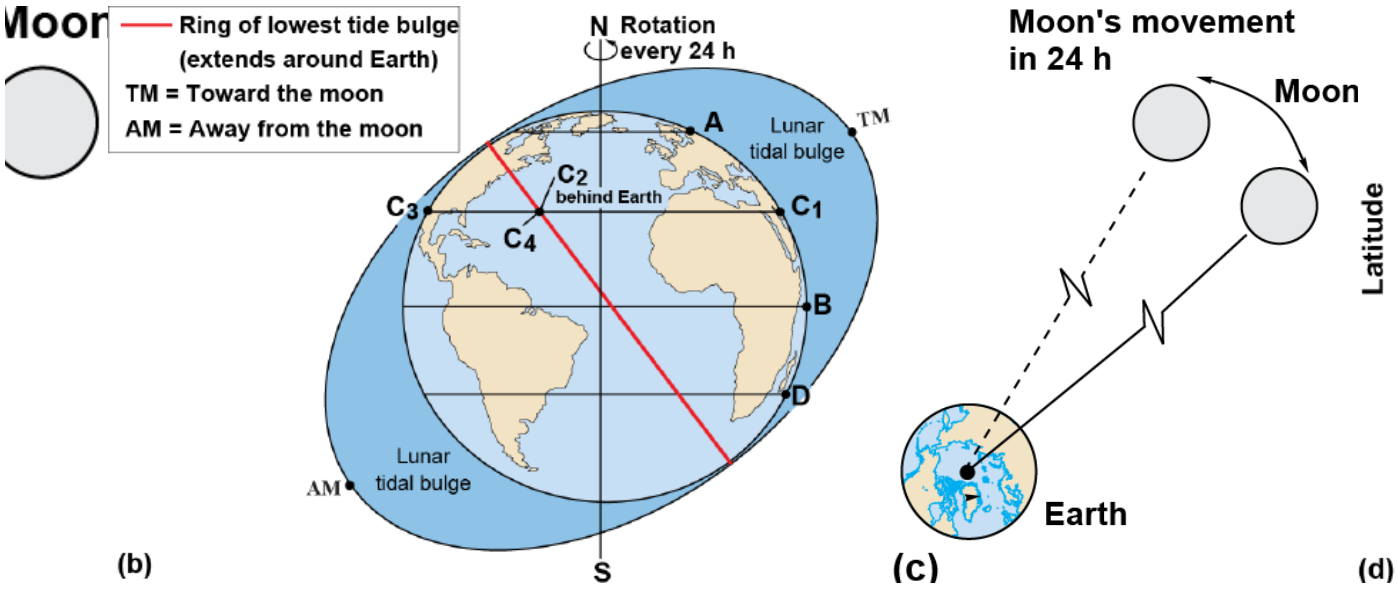


Figure 10-9. A diagram with the moon and an ocean-covered Earth illustrates the principal features of the tides. In this figure, the continents are shown only to make it easier to identify Earth's orientation. (a) The angle of the moon's orbit with respect to the equator (the declination) oscillates regularly. The maximum declination is about 25°. (b) The tide bulges caused by the moon's tidal forces are shown here with the declination and size of the bulges greatly exaggerated, and the effects of the continents ignored. (c) Looking down from above the North Pole at the Earth and moon, we would see that the moon moves through a small segment of its orbit each time the Earth rotates. Hence, the moon is "overhead" at any point on Earth a little later each day, so the tidal day is 24h 49min long. (d) In the simplified system depicted in (b), an observer at latitude A would see a diurnal tide, an observer at C would see a semidiurnal tide, and observers anywhere else on Earth, such as B and D, would see mixed tides. As discussed later in the chapter, the tides are more complex than shown in this figure.

Imagine an observer standing at a point along latitude B in **Figure 10-9b** and rotating with the Earth. When the observer is at point B1 there is a high tide. As the Earth rotates, the observer passes through the low tide at the back of the Earth (B_2) after the Earth has rotated a little more than 90°, or a little more than 6 h later. After 12 h and 24½ min, the observer passes a second high tide (B_3), but it is not as high as the original high tide. B_3 is further from AM than B_1 is from TM. Thus, as the bulge passes the

longitude of this location (B_3), the highest point of the bulge is farther away from the observer than it was when the bulge passed the longitude of the observer's original position (B_1). The observer rotates farther with the Earth, passing a second low tide after about 18 hr (B_4), and another high tide after 24 h and 49 min (B_1). The extra 49 min are due to the moon's progression in its orbit during the day (**Fig. 10-9c**).

We can follow the tidal patterns observed at points A, C, and D in a similar way. A diurnal tide is observed at A, a semidiurnal tide at C, and mixed tides at B and D (**Fig. 10-9d**). Note that, in this simplified model of tides, pure semidiurnal tides occur only at the equator. Note also that, in mixed tides, low tides do not occur exactly midway between high tides.

In the simple model shown in **Figure 10-9**, low tide would always be the same height at any specific location. However, the tide records in **Figure 10-6** and **Figure 10-8** show that this is not true. The reason is that the simplified system in **Figure 10-9** does not include solar tides that are added to lunar tides or the effects of continents, both of which create much greater complexity in the actual tides on the Earth.

The Origin of Spring and Neap Tides

The sun exerts tidal forces on the Earth that are about half as strong as the lunar tidal forces. We can see how the solar and lunar tides interact by again considering a simplified ocean-covered Earth. Tide bulges are created by both the sun and the moon (**Fig. 10-10**). The tidal height at any point on our model Earth is the sum of the tidal heights of the lunar and solar tides. The moon's orbit around the Earth and the Earth's orbit around the sun are not quite in the same plane, but they are nearly so. We can ignore this small angle for now.

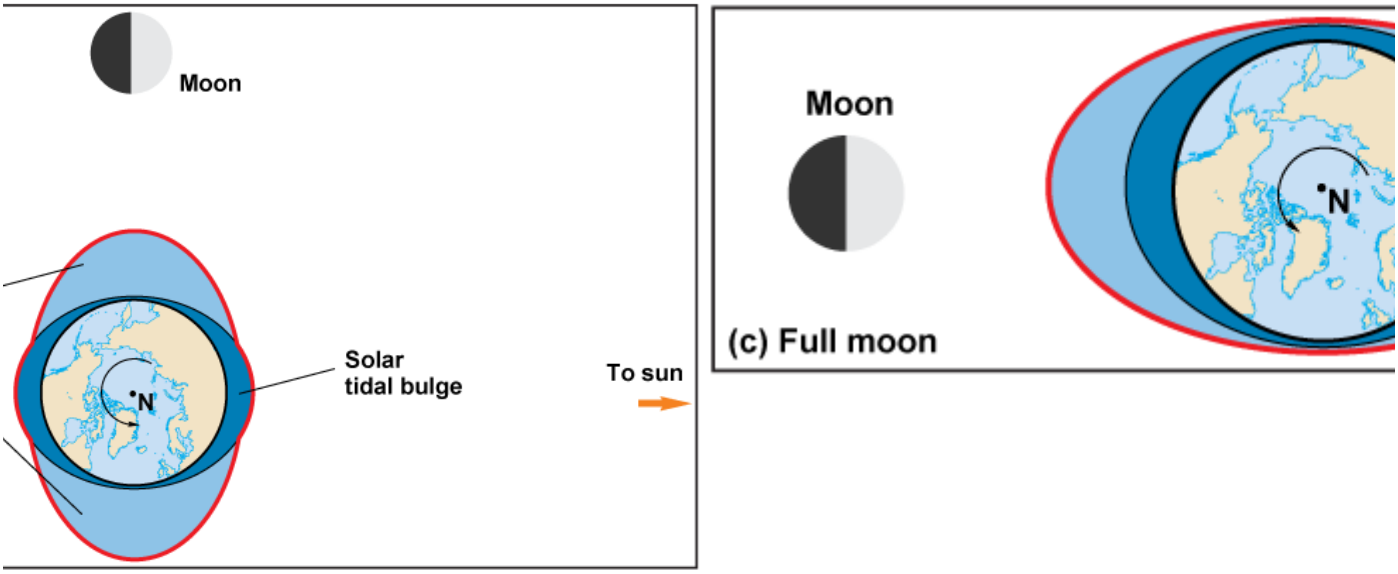


Figure 10-10. A simplified depiction of the Earth-moon-sun system shows how the solar and lunar tides interact to create spring and neap tides. (a) At the new moon, spring tides with a greater tidal range (higher high and lower low tides) occur because the tide bulges caused by the moon and the sun (both greatly exaggerated here) are aligned so that their maxima are at the same locations. (b) At the first-quarter moon, neap tides occur because the moon's bulge and the sun's bulge are aligned at 90° to each other and the maxima of the moon's tide coincide with the minima of the sun's tide. (c) At the full moon, spring tides occur again as the moon's and the sun's bulges are again aligned with their maxima at the same locations. (d) At the third-quarter moon, neap tides occur again as the moon's and the sun's bulges are again aligned at 90° to each other and the maxima of the moon's tide coincide with the minima of the sun's tide. The observed spring and neap sequences on the Earth are more complicated for various reasons, including that the orbits of the moon and sun are not exactly in the same plane as this simplified figure depicts. These orbits are inclined to each other, and the inclination changes with time.

Figure 10-10a shows the relative positions of the sun, moon, and Earth at a new moon. Because the entire disk of the moon is in shade as seen from the Earth, the moon must be between the sun and the Earth. Lunar eclipses occur when the three bodies are precisely aligned. When the alignment is not exact, we see the moon as a thin sliver of light. This is a new moon. At a new moon, we can see that the high-tide bulges of the solar and lunar tides are located at the same point on the Earth. The lunar high-tide bulge directly under the moon and the solar high-tide bulge directly under the sun combine in the same location, and the lunar and solar high tide bulges on the side of Earth directly away from the sun and moon also coincide. Finally, the locations on Earth where the low tide areas of the solar and lunar tide bulge also coincide. Thus, at a new moon the solar and lunar high tides are

added together and the solar and lunar low tides are added together, producing the maximum tidal range with the highest high tides and also the lowest low tides during the lunar month. These are spring tides.

Each day, the moon's movement in its orbit is the equivalent of the angle through which the Earth rotates in about 49 minutes. Approximately 7½ (7.38) days after full moon, the moon, Earth, and sun are aligned at right angles (**Fig. 10-10b**). The right half of the moon (as seen from the Northern Hemisphere) is now lit by the sun. The other part of the moon that is lit is hidden from an observer on the Earth. This is the moon's first quarter. (The lunar month is counted from a new moon.) At the first-quarter moon, the lunar tide bulge coincides with the low-tide region of the solar tide, and vice versa. Because the tides are additive, the low tide of the solar tide partially offsets and reduces the height of the lunar high tide, and the high tide of the solar tide partially offsets the lunar low tide. The tidal range is therefore smaller than at full moon. These tides are the neap tides.

Fifteen days after a new moon, the moon is directly between the Earth and the sun (**Fig. 10-10c**). Now the moon is entirely lit by the sun as seen from the Earth and there is a full moon. At full moon, the solar and lunar tide bulges again coincide and there is a second set of spring tides for the month. These tides are equal in height and range to the spring tides that occurred 15 days previously at new moon. Similarly, a little more than 22 days (22.14) after new moon, there is a third-quarter moon. The left side of the moon is lit for Northern Hemisphere observers (**Fig. 10-10d**). The solar and lunar tides again offset each other, and there is a second set of neap tides.

After about 29½ (29.53) days, the moon is new again, we have a new set of spring tides and begin a new lunar month. From this simple model, we can see why we have two sets of spring tides and two sets of neap tides in each 29½-day lunar month (**Fig. 10-9**).

Other Tidal Variations

From the simple model we have been discussing, we have seen that the height of the tide due to the moon varies on a cycle that is 24 h 49 min long. High tides actually occur every 12 h 24½ min because there are two bulges. Similarly, the solar tide varies on a 24-h cycle, and high tides occur every 12 h. The motions of the moon and the Earth in their orbits are much more complex than the simple model suggests, and tidal variations occur on many other timescales. These additional variations are generally smaller than the daily or monthly variations, but they must be taken into account when precise tidal calculations are made. They are due to periodic changes in the distances between the Earth and the sun and between the Earth and the moon, and the various declinations. Declinations are the angles between the Earth's plane of orbit around the sun, the Earth's axis of spin, and the moon's orbital plane. **Table 10-1** lists some of the more important variations of Earth–moon–sun orbits, along with the periodicity of the partial tides that they cause.

Table 10-1. *Selected Tidal Components*

Tidal Component	Frequency	Period (h)	Relative Amplitude	Description
Principal lunar	Twice per day	12-42	100.0	Rotation of Earth relative to moon
Principal solar	Twice per day	12-00	46.6	Rotation of Earth relative to sun
Larger lunar ecliptic	Twice per day	12-66	19.2	Variation in moon-Earth distance
Lunisolar semidiurnal	Twice per day	11.97	10.7	Changes in declination of sun and moon
Lunisolar diurnal	Daily	23.93	58.4	Changes in declination of sun and moon

Tidal Component	Frequency	Period (h)	Relative Amplitude	Description
Principal lunar diurnal	Daily	25.82	41.5	Rotation of Earth relative to moon
Principal solar diurnal	Daily	24.07	19.4	Rotation of Earth relative to sun
Lunar fortnightly	Biweekly	327.9	17.2	Moon's orbit declination variation from zero to maximum and back to zero
Lunar monthly	Monthly	661.3	9.1	Time for moon-Earth distance to change from minimum to maximum and back
Solar semiannual	Semiannually	4382.4 (182.6 days)	8.0	Time for sun's declination to change from zero to maximum and back to zero
Anomalistic year	Annually	8766.2 (365.2 days)	1.3	Time for Earth-sun distance to change from minimum to maximum and back

This page titled 10.3: Tides on an Ocean Covered Earth is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.4: Tides in the Earth's Oceans

Tides in the Earth's oceans behave somewhat differently from those on the hypothetical ocean-covered Earth. For example, on the model Earth, pure semidiurnal tides would occur only at the equator (**Fig. 10-9**), except when the moon's declination was zero, and then all points on the Earth would have pure semidiurnal tides. However, on the real Earth semidiurnal tides occur at places other than the equator (**Figs. 10-6**, **10-7**, **10-8**).

Four major interrelated factors alter the Earth's tides from the equilibrium model: the Earth's landmasses, the shallow depth of the oceans relative to the wavelength of tides, the latitudinal variation of orbital velocity due to the Earth's spin around its axis, and the Coriolis effect (**CC12**). When these factors are included in tide calculations, the resulting tides are called "dynamic tides."

Effects of Continents and Ocean Depth

In discussing the factors that affect tidal motions, it is convenient to envision tide waves moving across the Earth's surface from east to west, even though, in fact, it is the Earth that is spinning. The peaks of the tide waves remain fixed in their orientation toward and away from the moon or sun, moving only slowly as the moon, Earth, and sun move in their orbits.

The presence of landmasses prevents the tide wave from traveling around the world. The continents are generally oriented north–south and bisect the oceans. The equilibrium tide wave moves from east to west. When the tide wave encounters a continent, its energy is dissipated or reflected, and the wave must be "restarted" on the other side of the continent. Because continents, landmasses, and the ocean basins are complicated in shape, the tide wave is dispersed, **refracted**, and reflected in a complex and variable way within each ocean basin.

The average depth of the oceans is about 4 km, and the maximum depth is only 11 km. In contrast, the wavelength of the tide wave is one-half of the Earth's circumference, or about 20,000 km at the equator. Hence, the water depth is always considerably less than 0.05 times the wavelength of the tide wave (that is, the water depth is considerably less than $L/20$), and the tide wave acts as a **shallow-water wave** (**Chap. 9**). The speed of a shallow-water wave is controlled only by the water depth. In the average depth of the oceans (4 km), the tide **wave speed** is approximately 700 km•h^{-1}. The equilibrium model tide wave always travels across the Earth's surface aligned with the movements of the moon and sun. However, to travel across the Earth's surface and remain always exactly lined up with the moon (or sun), the tide wave would have to travel at the same speed as the Earth's spin. At the equator, the wave would have to travel at about 1620 km•h^{-1}. Because it is a shallow-water wave that can travel at only 700 km•h^{-1}, the tide wave at the equator must lag behind the moon as the moon moves across the Earth's surface.

The interaction between the moving tide wave that lags behind the moon's orbital movement and the tendency for a new tide wave to form ahead of the lagging wave is complex. However, the result is that the tide wave lags behind the moon's orbital movement, but not by as much as it would if it were not continuously recreated. The tide wave is said to be a forced wave because it is forced to move faster than an ideal shallow-water wave.

Because the tide wave is a shallow-water wave and its speed depends on the water depth, the tide wave is refracted in the same way that shallow-water wind waves are refracted as they enter shallow water. Dynamic-tide theory therefore must include the refraction patterns created by the passage of tide waves over **oceanic ridges**, **trenches**, shallow **continental shelves**, and other large features of the ocean basins.

Latitudinal Variation of the Earth's Spin Velocity

The tidal time lag changes with latitude because the orbital velocity (due to the Earth's spin) of points on the Earth's surface decreases with latitude (**Fig. 10-11**). At latitudes above about 65°, the shallow-water wave speed is equal to or greater than the orbital velocity. Hence, there is no tidal time lag at these latitudes, and the tide wave even tends to run ahead of the moon's orbital movement. Another complication in calculating the tide wave time lag is the fact that the Earth's axis is usually tilted relative to the moon's orbit (and to the Earth's orbit with the sun). Therefore, even on a planet with no continents, the tide waves would not normally travel exactly east–west.

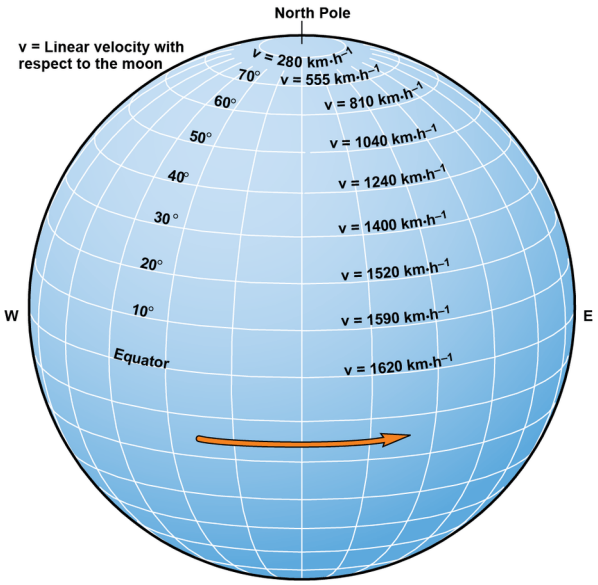


Figure 10-11. The tide wave moves across the Earth's surface as the Earth spins in relation to the moon and sun. If it were to do so exactly in time with the Earth's spin, it would have to move at different speeds across the Earth's surface at different latitudes. This is because the orbital velocity of points on the Earth's surface due to the Earth's spin is at a maximum at the equator and decreases with increasing latitude until it is zero at the poles.

Coriolis Effect and Amphidromic Systems

Tide waves are shallow-water waves in which water particles move in extremely flattened elliptical orbits. In fact, the orbits are so flattened that we can consider the water movements to be horizontal. Water moving within a tide wave is subject to the Coriolis effect (**CC12**). One important consequence of the Coriolis deflection of tide waves is that a unique form of **standing wave** can be created in an ocean basin of the correct dimensions. This type of standing wave is called an **amphidromic system**, in which the high- and low-tide points (the wave **crest** and **trough**, respectively) move around the basin in a rotary path—counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere.

Figure 10-12 shows how an amphidromic system is established in a wide Northern Hemisphere basin. In the Northern Hemisphere, a standing-wave crest (**antinode**) enters the basin at its east side. As water flows westward, it is deflected ***cum sole*** (meaning "with the sun") to the north, causing a sea surface elevation on the north side of the basin as the wave height at the east boundary decreases behind the crest. Water now flows toward the south with the north-to-south pressure gradient created by the sea surface elevation on the north side of the basin. As water flows south, it is deflected *cum sole* to the west. The sea surface elevation continues to move around the basin counterclockwise in this way until it returns to the east side of the basin. If the standing-wave crest returns to the east side of the basin after exactly 12 h and 24½ min (or any multiple thereof), it will meet the crest of the next lunar tide wave and the oscillation is then said to be **tuned**.

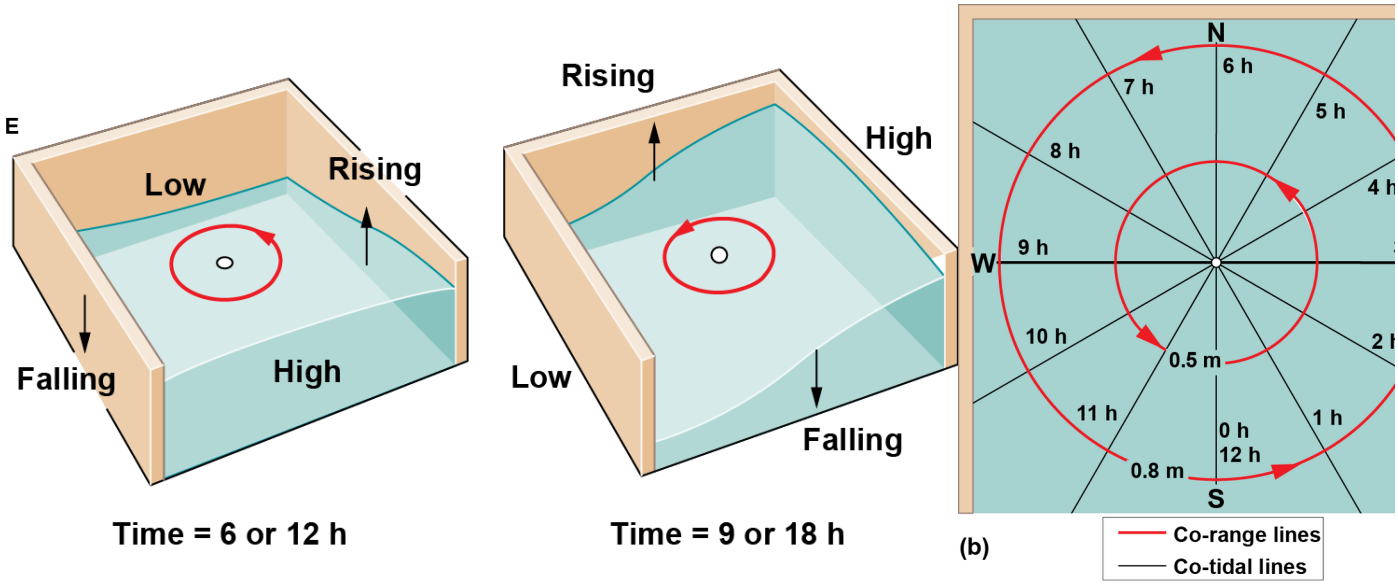


Figure 10-12. Development of an amphidromic standing-wave oscillation in response to the Coriolis effect in the Northern Hemisphere. (a) An amphidromic system is created in the principal solar semidiurnal or diurnal component of the tide when the tide wave enters a basin and is deflected by the Coriolis effect so that it rotates around the basin and returns to the entrance exactly 12 or 24 h later. (b) In an idealized Northern Hemisphere amphidromic system, the tidal range is the same for all places on any circle centered on the amphidromic point. The line along which the low tide or high tide is located progresses around the basin counterclockwise (clockwise in the Southern Hemisphere). In this plan view, high tide or low tide occurs along each of the radial lines at the indicated number of hours it occurs along the line labeled "0 h." (c) A cross-sectional view of the basin shows the up-and-down motion about the node upon which the rotation around the basin is superimposed.

One important characteristic of amphidromic systems is the amphidromic point (**node**) near the center of the basin, at which the tidal range is zero (Fig. 10-12). Amphidromic systems are established in all the ocean basins (Fig. 10-13). However, different amphidromic systems are set up for each tidal component (Table 10-1) because the components have different wavelengths. Hence, a location in an ocean basin that is an amphidromic point for one component of the tide will have a zero tidal range for that component, but will have tides generated by tidal components with other periods. In addition, some basins tune more easily with a particular tidal component. In parts of such basins, the tidal range due to that tidal component can be enhanced in relation to the ranges due to other components. This effect explains the presence of dominant diurnal and semidiurnal tides at many locations where they would not be predicted by equilibrium tide theory.

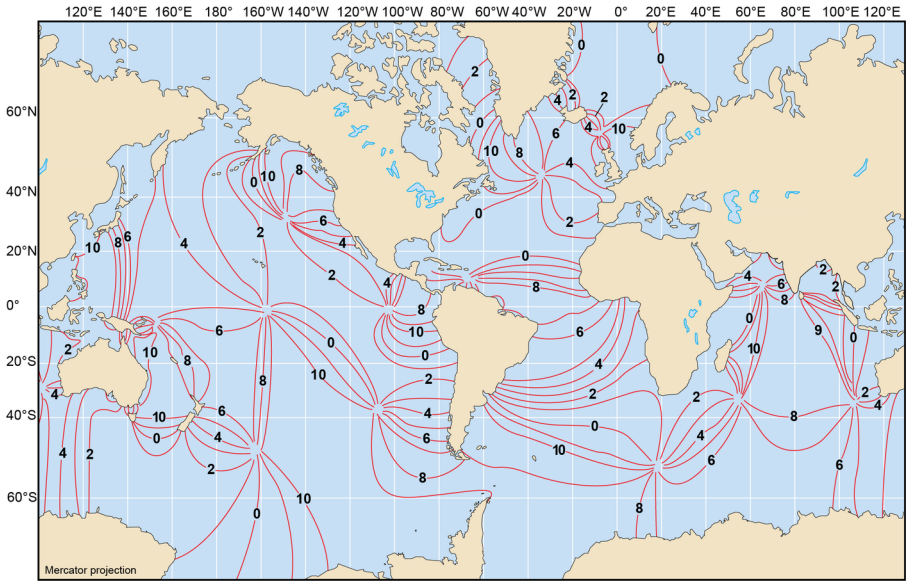


Figure 10-13. Amphidromic systems of the principal lunar semidiurnal tidal component in the Earth's oceans. The lines show the location of the tidal maximum. The numbers indicate the number of hours that elapse as the maximum tide (crest) travels from the line labeled 0 to the indicated line. Notice that the crest of the tide wave moves in rotary amphidromic systems in most of the ocean basins, and that the rotation is counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. Because it is a shallow-water wave, the tide wave is also refracted as water depth changes, resulting in some of the complex bending of the wave in certain parts of the oceans.

This page titled 10.4: Tides in the Earth's Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.5: Tides in the Open Ocean

Having examined the many factors that control tides, we can now look at some characteristics of real tides in the ocean basins. Much of what we know about the behavior of the tides is based on tidal-height measurements made along the coasts and on islands. The details of tidal movements in the deep oceans generally are derived from these measurements and dynamic-tide theory, although a number of deep-sea tide gauges have been deployed to make direct observations during the past several decades. Satellites, moored current meter arrays, and AUVs now also provide data that is helping to steadily refine our understanding of deep-ocean tidal motions.

Figure 10-14 shows the progression and height of the principal lunar semidiurnal component of Atlantic Ocean tides. This component is only one of the many different partial tides that must be added to determine tidal height at any time and location.

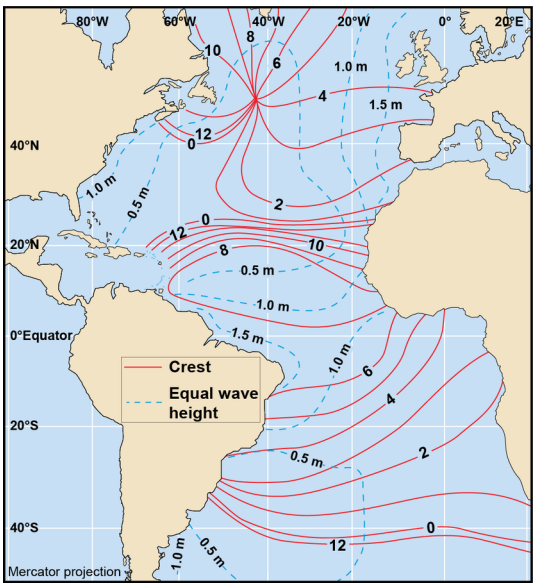


Figure 10-14. The principal lunar tidal component wave enters the Atlantic Ocean from the Indian Ocean around Antarctica and travels north into the North Atlantic, where it forms an amphidromic wave. Red lines show the location of the tidal maximum. Dashed blue lines are contours of tidal range. Note the higher tidal range on the west side of the South Atlantic compared to the east side. Also note that the tidal range increases in all directions from the center of the amphidromic system.

The most striking feature of **Figure 10-14** is that the tide wave does not move from east to west with the moon, as equilibrium tide theory predicts. This is because the Atlantic Ocean is a relatively narrow basin in which only a small tide wave can be generated during a single pass of the moon. The small east–west tide wave soon encounters the American continents, where it is partially reflected and much of its energy is dissipated.

The only segment of the Earth where the tide wave can travel east to west around the world without encountering a landmass is near Antarctica. At this high latitude, the orbital velocity of the Earth's surface is low enough (**Fig. 10-11**) that the shallow-water tide wave can travel fast enough to keep up with the moon's orbital movement. The tide wave around Antarctica is therefore well developed (**Fig. 10-13**). It enters the Atlantic Ocean between the tip of South Africa and Antarctica and is partially deflected and dispersed into the South Atlantic Ocean. The tide wave moves northward through the South Atlantic Ocean as a progressive wave. As it travels north, it interacts with the weaker east–west wave formed in the Atlantic and is reflected and refracted in complex ways. It is also deflected by the Coriolis effect. Although the deflection is obscured by other factors in the southern part of the South Atlantic, it causes tides to be slightly higher on the South American coast north of Rio de Janeiro than on the opposite African coast (deflection to the left in the Southern Hemisphere). The Coriolis deflection is also partially responsible for tides being slightly higher on the coasts of Europe and North Africa than on the North American coast (deflection to the right in the Northern Hemisphere).

In the North Atlantic, the progressive wave traveling north from Antarctica is converted into the standing wave of an amphidromic system (**Fig. 10-14**). The high tide moves around the North Atlantic basin counterclockwise and arrives back where the next crest of the Antarctic progressive wave arrives almost exactly 12 hr and 24½ min later. The North Atlantic Ocean basin is therefore well

tuned to the semidiurnal tidal component, and this component dominates and produces semidiurnal tides in this region (**Fig. 10-7**).

Amphidromic systems similar to the system in the North Atlantic are also present in the North Sea and the English Channel. However, the Gulf of America (Golfo de México) has a natural period of about 24 h. Therefore, the semidiurnal tidal component in the Gulf is poorly developed, and the diurnal component is stronger, so diurnal tides dominate in some parts of the Gulf (**Fig. 10-7**). The Pacific Ocean basin is wider than the Atlantic Ocean basin. Therefore, the Pacific has a more developed east–west tide wave and greater complexity than the Atlantic Ocean. In both the Pacific Ocean and the Caribbean Sea, diurnal and semidiurnal tide waves are relatively well tuned, and tides are generally mixed tides.

This page titled 10.5: Tides in the Open Ocean is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.6: Tidal Currents

Orbits of water particles in tide waves are so flattened that water movements in the tide wave are essentially oscillating currents. The magnitude and periodicity of these currents are as important to mariners as the tidal range is, particularly in coastal and estuarine regions. Although tidal currents are relatively weak in the open oceans, their speed increases as the tide wave moves inshore and its energy is compressed into a shallower depth of water. Tidal-current speed also increases where the tide wave must move large quantities of water through a narrow opening into a large bay or estuary, such as the entrances to San Francisco Bay and New York Harbor.

Open-Ocean Tidal Currents

Tidal currents are generally weak and rotary in character in deep-water areas away from the coast in the large ocean basins (**Fig. 10-15**). In areas of the Northern Hemisphere with semidiurnal tides, the tidal-current direction rotates progressively, usually clockwise, and completes a 360° rotation in 12 hr 24½ min (**Fig. 10-15a**). The current speed varies but is never zero.

In areas with mixed tides, the progression of tidal-current speed and direction is more complicated (**Fig. 10-15b**) because both variables represent a combination of two different tidal components. Note that the times of maximum and minimum tidal-current speed are apparently not related to the times of high and low tide.

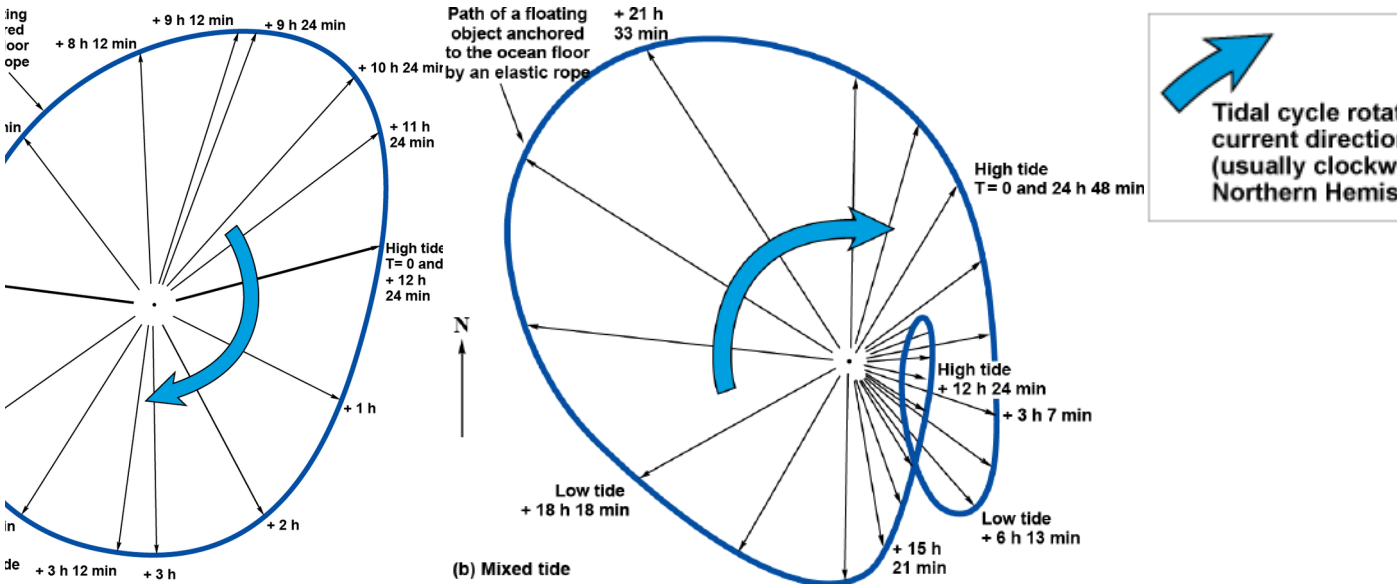


Figure 10-15. Changes in the direction and magnitude of open-ocean tidal currents in the Northern Hemisphere. The arrows indicate the direction of the current at approximately hourly intervals, and the length of the arrows is proportional to the current speed. The outer blue line describes how an object would move in the water if there were no currents or water movements other than tidal currents. (a) Where there is a simple semidiurnal tide, the current speed varies little, but its direction rotates clockwise through 360° every 12 hr 24½ min. (b) Where there is a mixed tide, the variations of current speed and direction are more complex. In this case, the current direction rotates clockwise twice within 24 h 49 min, but the rate of rotation and the current speeds follow very different patterns within the two successive 12-h periods.

Tidal currents flow at all depths in the oceans. While current speeds due to tides in the deeper parts of the oceans are slow, these currents interact with seafloor topography in complex ways. Although little is known about these interactions, it is known that increased turbulence as tidal currents flow over seafloor topography is a significant factor in vertical mixing of water masses.

Temporal Variation of Tidal-Current Speeds

Many people mistakenly believe that tidal currents stop when the tide reaches its highest and lowest points. This is rarely true. In fact, if the tide were a pure progressive wave, the tidal current would be at its fastest at high or low water. We can see why by considering the motions of water particles in the tide wave (**Fig. 10-16a**). However, the maximum tidal currents rarely coincide

with high or low tide for several reasons. The most important is that all tides are complex combinations of many components. Some components may be progressive waves, some may be standing waves, and some may have both progressive-wave and standing-wave characteristics. In addition, tide waves are reflected and refracted, so the observed tide may be the result of several different waves moving in different directions.

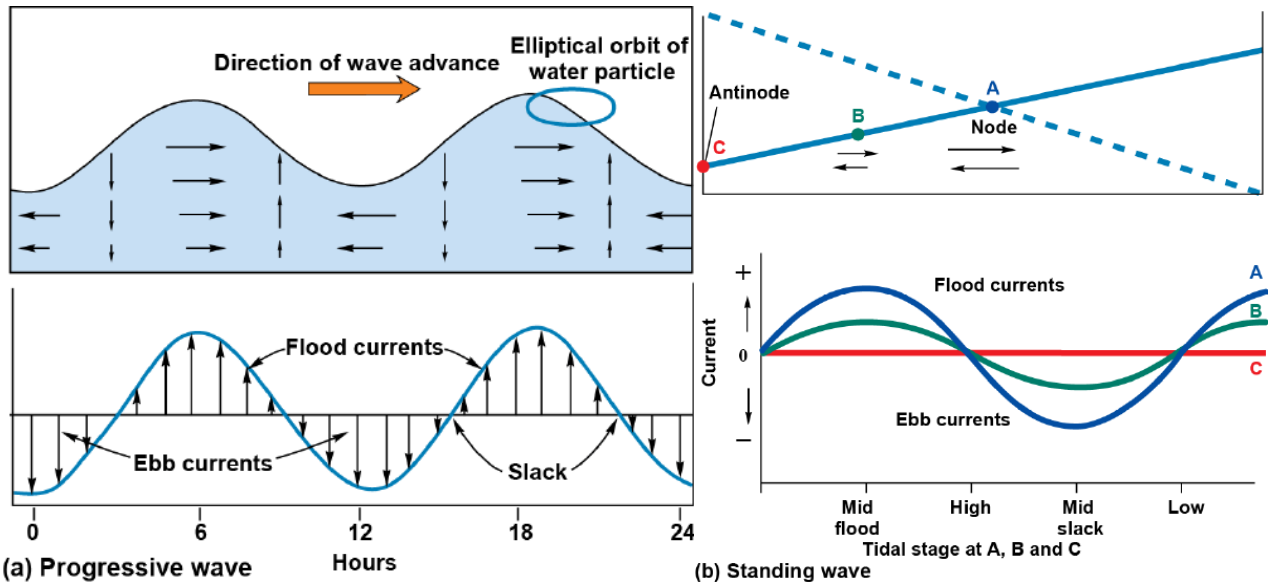


Figure 10-16. Tidal currents. (a) In a pure progressive wave, the forward motion in the wave is at a maximum as the crest passes, and in the opposite direction when the trough passes. Thus, if the tide wave were a pure progressive wave, the flood currents would reach a maximum at high tide and the ebb currents would reach a maximum at low tide. (b) In a standing wave, the horizontal currents are always zero at the antinodes, and they are reversing currents whose maximum speed increases from zero at the antinode to a maximum at the node. The maximum current speeds are reached at the mid-flood and mid-ebb stages. Because most tides have both progressive-wave and standing-wave components, there is no fixed relationship between the times of high and low tide and the times of maximum tidal current and slack water at different locations. Tidal currents are also complicated by seafloor friction.

In a standing wave, currents are at their maximum midway between high and low tide, when the sea surface is exactly level (**Fig. 10-16b**). Current speed varies not only with time, but also with location in the wave. It is highest at the node and is always zero at the antinodes, where the vertical water surface displacement, or tidal range, is greatest.

In most places, the tide is a complex mixture of progressive-wave and standing-wave components that vary from location to location. Hence, the relationship between the timing of **slack water** (minimum current speed) and high and low tide is different for each location. Curiously, at certain times, some locations have tidal currents but no tide (zero tidal range), while others have a large tidal range but no tidal currents. We can envision such locations as the node and antinode, respectively, of a standing wave that is the dominant component of the tides in that location, although the situation is generally more complicated because the tide is the sum of many different components of both solar and lunar tides.

Tide tables that list only times of high and low tide are useless for determining tidal-current speeds because there is no consistent relationship between the times of high and low water and the times of highest current speed and slack water. The only generalization about the tidal currents is that their maximum speed will increase or decrease as tidal range increases or decreases from day to day. Therefore, tidal currents must be measured in each location for which a forecast of the current speed is important, just as tidal ranges must be measured. Tidal-current information from such measurements is subject to harmonic analysis that is similar to the analysis of tidal-height data described later in this chapter. From this analysis, tidal-current tables are produced.

Tidal Currents in Estuaries and Rivers

Tides extend far into many bays, estuaries, and rivers. Tides in such locations are affected by the extremely shallow water depths, freshwater flow, and friction with the seafloor. In very shallow water, the crest of the tide wave moves in significantly deeper water than the trough. Therefore, the high tide tends to catch up to the low tide in estuaries or rivers where the tide travels long distances through shallow water. As a result, river tides can be modified so that there is a long period between high and low tide, but a very short period between low and high (**Fig. 10-17**).

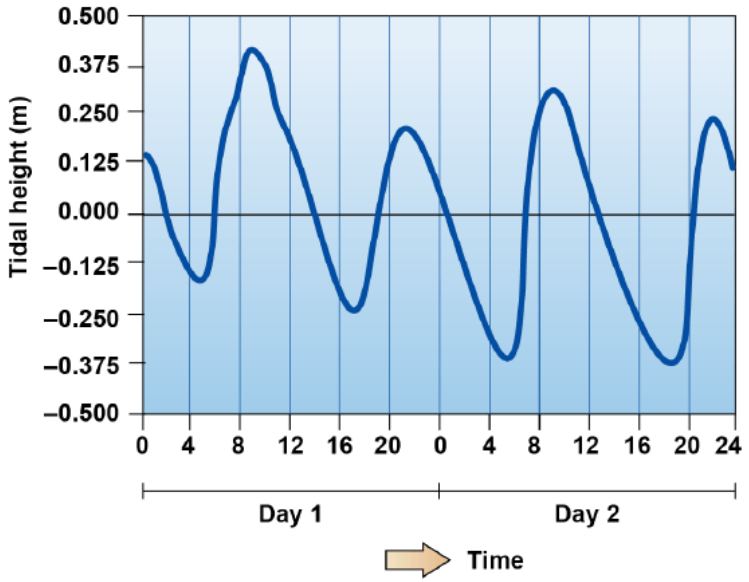


Figure 10-17. This tidal-height plot for the Hudson River near Albany, New York (about 220 km from New York Harbor), shows that the tide rises much faster (that is, the plot line is steeper) than it falls.

In some areas where tidal ranges are large and the tide enters a channel or bay that narrows markedly or has a steeply sloping seafloor, tidal bores may occur. A tidal bore is created when the currents in the flooding tide are faster than the speed of a shallow-water wave in that depth. The leading edge of the tide wave must force its way upstream faster than the wave motion can accommodate. The wave therefore moves up the bay or estuary as a wall of water, much like a continuously breaking wave (**Fig. 10-18**). Well-known tidal bores occur in the Bay of Fundy in Nova Scotia and in Turnagain Arm off Cook Inlet in Alaska. Most tidal bores are less than 1 m high, but they can reach as much as 10 times that height. Some bores, notably one in the Qiantang River in China, are high enough that they attract surfers who are looking for the longest ride of their lives.

Figure 10-18. Tidal bore in a channel of the Kent River near Arnside, Westmoreland, England.

Tidal currents in estuaries and rivers are affected by freshwater flowing toward the ocean. Freshwater flowing from a river into an estuary prolongs the ebb tide in relation to the flood tide because more water must be transported out of the river during the ebb than enters during the flood to accommodate the freshwater discharge. River flow rates can substantially change the nature of tides and tidal currents in estuaries. Accordingly, tide tables and tidal-current tables for bays, estuaries, and rivers must be used only as a general guide, particularly when river flow rates are abnormally high or low.

Tidal currents within bays, rivers, and estuaries can behave as progressive waves, standing waves, or, in some cases, a combination of the two. In long bays with progressive tides, tidal currents can flow in opposite directions at the same time in

different sections of the bay. Chesapeake Bay has this type of circulation (**Fig. 10-19**). Slack water from each tide migrates up the bay about 60 km•h^{-1} and takes about 10 h to travel the length of the bay. Successive slack tides enter the bay approximately every 6 h 12 min because each semidiurnal tide has slack water associated with both flood and ebb. This progressive wave is not modified significantly by an east–west wave, because the bay is too narrow for a significant wave to be generated directly by the tide-generating forces.

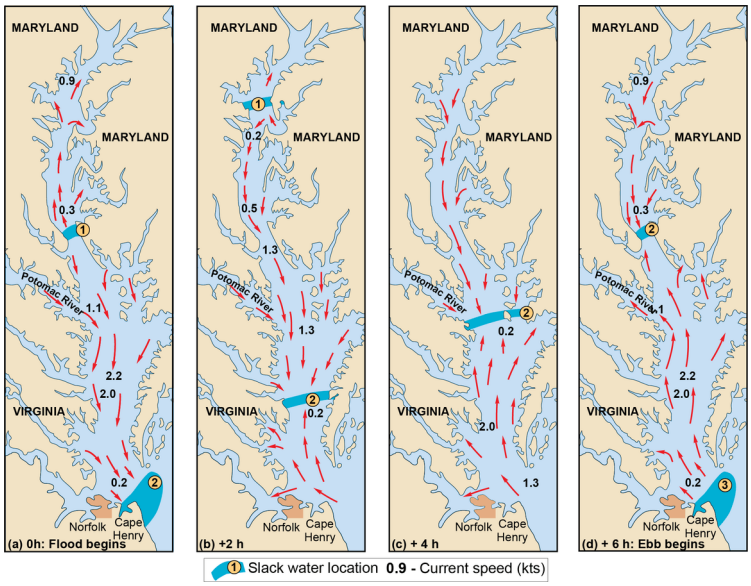


Figure 10-19. Chesapeake Bay has a primarily progressive-wave tide. The tide wave enters the estuary and moves progressively up the bay, reaching the north end about 10 h after entering the bay. Follow the successive slack-water occurrences numbered 1, 2, and 3 in these diagrams as they move north. (a) Slack water occurs at the mouth of the bay just as the flood begins and also at a location just above mid bay. (b) Two hours after the start of the flood, the two areas of slack water have migrated northward. (c) Four hours after the start of the flood, the northernmost slack has reached the north end of the bay and the second slack is almost in mid bay. (d) Six hours after the start of the flood, the slack that entered the bay 6 h earlier is now where its preceding slack had been at that time, and a new slack-water area occurs at the mouth of the bay as the ebb begins.

Tidal currents in Chesapeake Bay are somewhat more complex than those shown by the red arrows in **Figure 10-19** that show the relative current speeds, because the bay does not have uniform depth. The tide wave slows more rapidly at the sides of the bay, where the water is shallower and friction with the channel sides is enhanced. In this and other bays with a deep central channel and shallower sides, the tidal current reverses sooner at the sides than in the deep channel. This effect can be useful to boaters who want to avoid the fastest opposing currents. It is also a mechanism for mixing water from the deep main channel with water from the shallow sides, and therefore is of interest in studies of the dilution of discharges.

Long Island Sound has a predominantly standing-wave tide (**Fig. 10-20**). Tidal currents first flow westward throughout the length of the sound, then slack water occurs throughout the sound, and subsequently the current reverses and flows eastward throughout the sound. The area of nearly permanent slack water at the west end of the sound is the standing wave antinode and has a large tidal range, but little or no tidal current. The east end of the sound is the node and has a small tidal range.

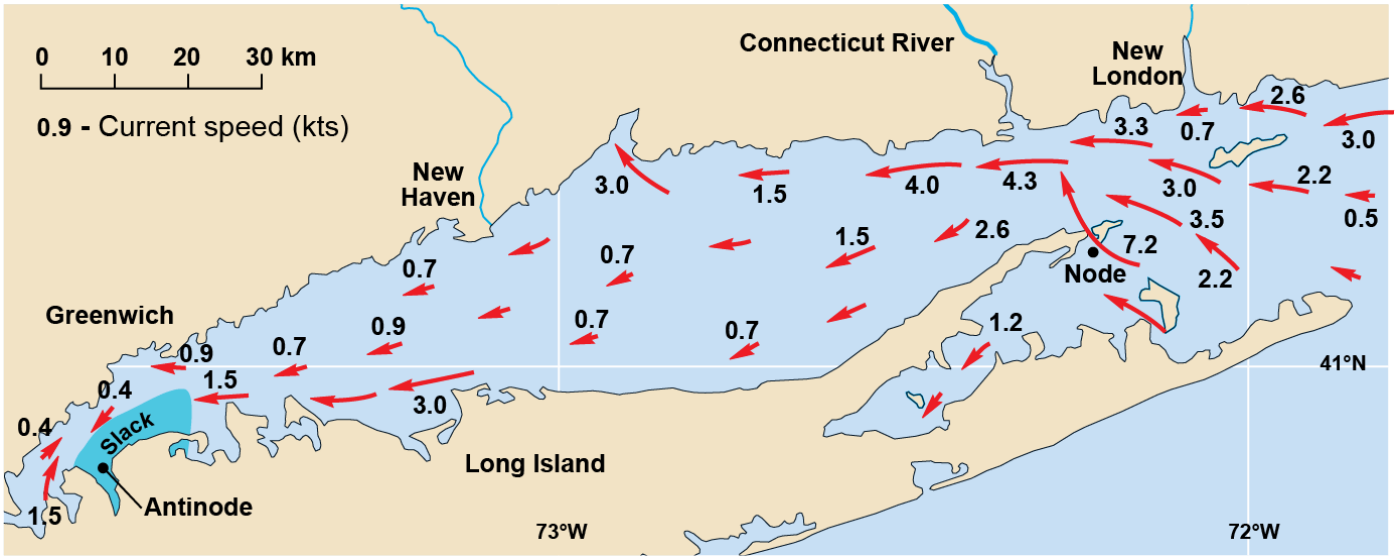


Figure 10-20. Long Island Sound has a predominantly standing-wave tide, with a node at the entrance to the sound and an antinode at the western end. There is little or no tidal current at any time during the tidal cycle at the western end where the antinode is located. In the remainder of the sound there are reversing tidal currents, with the maximum current speed increasing from the western end to the eastern opening to the ocean, where a node occurs.

This page titled 10.6: Tidal Currents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.7: Tide Prediction

As we have seen, the dynamic-tide theory must take into account many variables to predict tidal height at any given location at a specific time. The best predictions made by applying dynamic-tide theory closely approximate the actual tides. However, the predictions are not accurate enough for use by mariners, particularly in complex estuaries and bays. Therefore, tide predictions are made from tide observations. Tidal variations are continuously observed and recorded at many locations worldwide. Reasonably accurate tide predictions can be made from a tide record covering only a few months, but much longer records produce more accurate predictions.

Once the tide record is available, computers can perform harmonic analysis of the data to identify components of the complex wave. For example, harmonic analyses would distinguish the five component waves in **Figure 9-13b** from the complex wave record in **Figure 9-13a**. Harmonic analysis determines the wave heights and time lags (in relation to the moon's orbital movement or other motions) of each of the many different tidal components (**Table 10-1**). More than 390 different tidal components have been identified. Once the timing, periodicity, and **amplitude** of each component have been determined for a particular location, computing the magnitude of each component for any future time is relatively easy. Simple addition of the components gives the tidal height at the future time and location.

Tide predictions made in this way are used extensively. They are the basis for the tide predictions published daily on many news and marine websites and the printed tide tables offered in bait-and-tackle stores, dive shops, and marinas. The NOAA Tides & Currents site also provides constantly updated, detailed tide information.

Satellite-based radar sensors accurate enough to measure tidal heights have allowed detailed mapping of the tides in the open-ocean areas where few tidal-height measurements had previously been made. Satellite-collected tidal-height data sets are steadily accumulating and will eventually enable extremely accurate computer models of the tides for most locations worldwide.

Tidal currents vary locally, often in locations just a few tens or hundreds of meters apart, and predictions must be based on at least several months of current measurements at each specific location. Unfortunately, data for one location are inadequate to make predictions for other locations, especially within complex estuaries.

This page titled 10.7: Tide Prediction is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.8: Marine Species and the Rhythm of Tides

Tides play an important role in the life cycles of many marine species. The most obvious connection is with species that live in the intertidal zone, for which the tides expose them and their **habitats** to the atmosphere for variable periods during the day. These species have developed strategies to avoid dehydration and/or being eaten by birds and other predators during periods when their habitat is exposed to the air. For example, most species that live in sandy or muddy intertidal **environments** bury themselves in the sand when the tide is out. Many species that live on rocky shores have shells to retreat into or leathery outer skins that, in addition to deterring predators, resist loss of water through evaporation

Although we are only just beginning to learn of the scope of the many subtle influences of tidal motions on marine species, numerous well-documented connections exist between the tides and the reproductive behavior of many species, especially regarding the timing of the reproductive cycle. For example, many species that **spawn** eggs and **larvae** into the water column exhibit spawning behavior timed to benefit the species' survival. One example is that many **coral** species spawn en masse at a specific phase of the lunar tide, most often 1 or 2 days after a full moon in the spring months, when the tidal range is at its greatest. Spawning at this time ensures the greatest possible dispersal of the eggs and larvae to potential new colonization sites. Spawning strategies are discussed in more detail in **Chapter 14**.

There is one connection between tides and species' reproductive strategies that you may be able to see for yourself. This is the use of high spring tides by some species at a certain time of year to place fertilized eggs in the sand high up on a **beach** where they cannot be reached by most water-dwelling predators. Perhaps the best-known examples are sea turtles that haul themselves high up the beach, usually at night when there is a high spring tide, dig a hole in the sand, lay and bury their eggs, and then leave the eggs to hatch some weeks later. Many other species use a very similar strategy. Some of these species can be seen on beaches or mudflats around North America. For example, on the Atlantic coast, horseshoe crabs (*Limulus polyphemus*) spawn on the beaches between the high- and low-tide lines on nights when the spring high tide occurs. Female horseshoe crabs crawl out of the water as the tide advances, dig a cavity about 15 cm deep in the sand, and deposit several thousand large, greenish eggs. The male, which is attached to the female's back, fertilizes the eggs, and the eggs are then buried by sand moved by the advancing tide. The eggs hatch 14 days later when the next high spring tide covers the sand.

On the Pacific coast, several fishes use a similar strategy. The most famous of these fish species is the grunion (*Leuresthes tenuis*). This small, silvery fish comes completely out of the water onto the southern California and Baja beaches to spawn just after each maximum nighttime spring tide during March through September. The eggs are protected from wave action for the next 14 days because they are laid above the lower high-tide lines that occur at high tide between one spring tide and the next. Although they mature in about 9 days, the eggs do not hatch until disturbed by water and wave **turbulence**, which occur as the next spring high tide arrives 14 days after they were laid. The eggs hatch within minutes of being exposed to the water.

Although the grunion is the most famous fish that spawns on the beach at high tide, other species also use this strategy, but they do not come completely out of the water to do so. One example of such a species is the surf smelt (*Hypomesus pretiosus*), a small fish that spawns at high tide on the beaches of Puget Sound.

This page titled 10.8: Marine Species and the Rhythm of Tides is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.9: Energy from Tides

The tides are a potential source of inexhaustible and clean energy. Tidal energy dissipates as heat by friction between water masses and the seafloor. Some of the energy can be captured by electricity-generating turbines installed in areas with very fast tidal currents. However, very few locations have maximum tidal currents fast enough to drive turbines efficiently. Even where tidal currents are swift, they generally need to be enhanced by channeling the tidal flow into the turbines through narrow openings. The best locations for capturing tidal energy are bays or estuaries with a large tidal range and swift tidal currents, and where a dam can be built across the entrance. The dam is constructed with a number of narrow openings that house the turbines. Water entering the bay is channeled through the turbines as the tide enters the bay, and again as the tide flows back out of the bay.

For a tidal power-generating plant of this type to be feasible, the tidal range needs to exceed about 5 m, and it must be possible to build a dam that isolates an area into which such a tide flows from the open ocean. The most promising locations are few and widely distributed (**Fig. 10-21**). A large tidal power-generating plant has operated successfully at La Rance in France since the 1960s (**Fig. 10-22**). There has been continuing interest in constructing other plants, notably in the Bay of Fundy, Nova Scotia. Such projects face several problems, two of which are especially important.

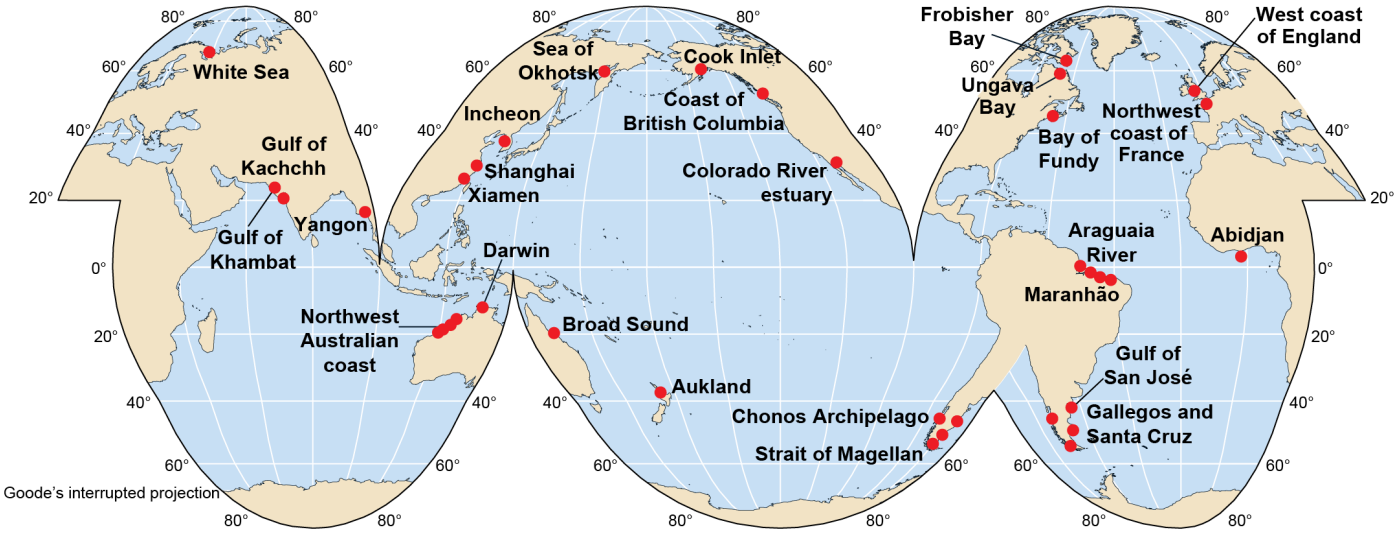


Figure 10-21. There are relatively few locations with a tidal range sufficiently large (more than about 5 m) that large scale tidal power-generating plants are considered to be feasible.

Figure 10-22. La Rance tidal power-generating plant in France.

First, a tidal power-generating plant cannot generate energy throughout the 24 h in a day. Power cannot be generated during the period of slack water when tidal currents are low. Judicious operation can retain water outside or behind a dam to some extent and thus lengthen the period during which power can be generated, but generating power uniformly over a 24-h period is apparently not possible.

The second problem is that building a dam across a bay substantially alters the **ecology** and other aspects of the bay. For example, the Bay of Fundy sustains rich and **diverse** marine **communities** and fisheries and has a famous tidal bore. Building a dam across the mouth of the bay would drastically alter or destroy both the fishery and the tidal bore. Fishers who live around the bay would lose both their local catch and access to the open ocean for their boats. Tourism would decline with the loss of the tidal bore. The benefits and losses associated with tidal power-generating stations that use dams to close off arms of the sea will be difficult to balance. A decision to build or not build such a plant at the Bay of Fundy (or anywhere else) will be controversial and will generate opposing views from reasonable people.

As an alternative to tidal power plants that rely on dams, submerged turbines can be deployed, preferably in a location where the tide wave is restricted and becomes a reversing current rather than rotary one. Some such tidal power plants have been tested and deployed, but they generate limited amounts of power and do not generate power consistently at times of day when it is needed. Therefore, as with several renewable energy options, efforts are being made to produce power storage capabilities to use with such installations.

While tidal power potentially available from the oceans is very large, its real potential for the foreseeable future is, like wave energy, believed to be small, and the technology is most suited to use in specific local situations.

This page titled 10.9: Energy from Tides is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.10: CHAPTER SUMMARY

Tide-Generating Force.

Bodies orbiting together do so around their common center of mass. A centripetal force acting toward the other body is needed to keep each body in orbit. The gravitational attraction between the bodies supplies this force. The centripetal force needed to maintain the Earth in its common orbit with the moon (or sun) is the same at all points within the Earth. However, the gravitational force of the moon (or sun) is inversely proportional to the distance from the moon's (or sun's) center of mass. The gravitational force exerted by the moon (or sun) is therefore slightly higher on the side of the Earth facing the moon (or sun) than on the other side. Hence, there is a slight excess of gravitational force over centripetal force at the Earth's surface facing the moon (or sun), and a slight excess of centripetal force over gravitational force at the opposite side of the Earth. The slight imbalances are the tide-generating forces. Tide-generating forces exerted by the sun are only 46% as strong as those exerted by the moon because even though the sun is more massive than the moon, it is much farther away from the Earth than the moon is.

Characteristics of Tides.

Tides resemble patterns of oscillating sea surface elevation caused by simple progressive waves or by the addition of two or more progressive waves. Because the moon's orbit is inclined to the equator, tides are usually predominantly one of three types. Diurnal tides have one high and one low tide per tidal day (24 h 49 min). Semidiurnal tides have two high and two low tides each tidal day, and the two lows (and two highs) are of equal height. Mixed tides have two high and two low tides each tidal day, but the two lows (and two highs) are not of equal height. Tidal range varies with location and from day to day within a lunar month. Spring tides (highest tidal range) and neap tides (lowest tidal range) occur twice each lunar month. Spring tides occur when the moon and sun are both on the same side of the Earth (new moon) or on opposite sides (full moon), so their tidal pulls are additive. Neap tides occur when the moon and sun are 90° apart (first and third quarters). Tidal range also varies from month to month. Tides are composed of numerous components with different periods, each caused by a regular periodic change in the Earth–moon and Earth–sun orbits, such as distance or declination.

Tide Waves.

Tide waves appear to progress from east to west as the Earth spins under the tide bulges. Continents interrupt the tides everywhere except around Antarctica and cause tide waves to be dispersed, refracted, and reflected in complex ways that affect tidal characteristics differently at different locations. Because its wavelength is so long, a tide wave is a shallow-water wave that is refracted by seafloor topography. The tide wave is a forced wave because it is too slow, even in deep oceans, to match the orbital velocity of the Earth's surface as it spins, except near the poles. The Coriolis effect deflects tide waves. In ocean basins of suitable dimensions, this deflection can create tuned-oscillation standing waves called amphidromic systems. The crest of one tidal component enters the basin and passes counterclockwise (Northern Hemisphere) or clockwise (Southern Hemisphere) around the basin to arrive back at the entrance exactly when another crest arrives.

Tidal Currents.

Orbits within tide waves are so compressed, particularly in coastal and estuarine regions, that vertical tidal motion is very small in relation to the horizontal motions. The horizontal motions are tidal currents. Open-ocean tidal currents are weak, and their direction progresses in a rotary pattern during the tidal cycle. In coastal and estuarine areas, tidal currents generally reverse direction 180° as flood and ebb currents during the tidal cycle. Tides in such areas are usually a combination of progressive and standing waves, and the relationship between high or low tide and times of slack water or maximum currents is different at each location. Tidal currents can be very strong, tidal range particularly high, and the progression of the tide wave complicated in estuaries where standing waves occur for some tidal components.

Tide Predictions.

Tides are too complex to be predicted without measurement at each location for which predictions are needed. Tidal-height measurements may now be made with satellite sensors. Tidal-height measurements recorded for several months or longer can be

subjected to harmonic analysis. This analysis can accurately predict the tidal range and times of high and low tides at the studied location. Tidal currents are more difficult to predict.

Energy from Tides.

Electricity is generated from tidal energy in a few locations where turbines and dams have been placed across a bay or estuary. However, very few locations have a sufficiently large tidal range to make such projects feasible. Suitable locations have a unique ecology that depends on those tides.

10.10.1: Key Terms

10.10.2: Chapter Questions

10.10.3: References

This page titled 10.10: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.10.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- amphidromic system
- amplitude
- antinode
- beach
- centripetal force
- coast
- communities
- continental shelves
- coral
- crest
- cum sole
- currents
- declination
- diurnal tides
- diverse
- ebb
- ecology
- environments
- estuaries
- flood
- friction
- gravity
- habitats
- high-tide line
- intertidal zone

- larvae
- low-tide line
- mixed tides
- node
- oceanic ridges
- pressure gradient
- progressive waves
- refracted
- semidiurnal tides
- shallow-water wave
- shore
- slack water
- spawn
- species
- standing wave
- tidal range
- tide
- trenches
- trough
- tuned
- turbulence
- wave heights
- wave periods
- wave speed

This page titled 10.10.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.10.2: Chapter Questions

Study Questions

1. If the mass of the moon were four times what it is, the distance between the Earth and the moon would be twice what it is now. Is this a correct statement? Explain why or why not.
2. Describe the imbalance between the centripetal force and the gravitational attraction force at different locations on the Earth's surface. How is it created?
3. Where on the Earth's surface, in relation to the moon's direction, is the tide-generating force due to the moon equal to zero?
4. Why does the moon exert a greater tide-generating force than the sun, even though the sun's gravitational attraction at the Earth's surface is almost 200 times stronger than the moon's?
5. How are diurnal, semidiurnal, and mixed tides different? Why is it important to some sailors to know which type of tide is present in a port that they visit?
6. Spring tides and neap tides occur twice each month but are not the same height each month. What variations of the Earth's orbit around the sun might explain seasonal variations in the tidal range of spring or neap tides?
7. Why is the tide wave a forced wave in some places and not in others? Where is it a forced wave, and where is it not a forced wave? Why?
8. Why are amphidromic systems not established in all ocean basins? Why do some tidal components in a particular basin form amphidromic systems, whereas other components do not?
9. Why are tidal currents in the deep oceans slower than those in coastal waters? Why are deep-ocean tidal currents generally rotary, whereas coastal tidal currents are reversing?
10. How are tide tables able to predict the future times and heights of high and low tides? These predictions are generally accurate, but weather conditions can cause the heights and times of high and low tide to be shifted somewhat from those predicted. How can this happen? List and explain more than one possible reason.

Critical Thinking Questions

1. What might happen if the mass of the moon doubled? How would the moon's orbit be affected? How would ocean tides on the Earth be affected?
2. Why is the tide wave a shallow-water wave? How would it be refracted as it approached a large island with a wide continental shelf?
3. Why is it necessary to build a dam across an estuary to get the greatest possible amount of energy from tidal power? Why couldn't we just put a number of turbines side by side across the entrance?
4. There was once a proposal to build a sea-level canal with no locks to connect the Atlantic and Pacific Oceans across the Isthmus of Panama near the existing Panama Canal although the proposal has not been acted upon. Imagine that this canal was built instead across Mexico from the Gulf of America (Golfo de México) and that it was several hundred meters wide and 30 m deep.
   (a) Would it affect the nature of the tides in the Gulf of America? If so, how?
   (b) In the canal itself, would there be tidal-height variations of water level?
   (c) Would there be tidal currents? If so, hypothesize what characteristics the tides and tidal currents would have.
5. Tidal bores are formed when a flood tide moves up an estuary or inlet faster than the speed that a shallow-water wave of tidal wavelength can sustain. When the tide ebbs in such inlets, is there a tidal bore that flows seaward? Explain why such an ebb tide bore does or does not exist.

This page titled 10.10.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

10.10.3: References

Figure 10-1 – Segar, D., and E. Stamman Segar, 1997: Port of Anchorage, Alaska

Figure 10-6a – NOAA Tides & Currents, 2025: Tides at Pensacola, Florida. Accessed 1 July 2025, Source.

Figure 10-6a – NOAA Tides & Currents, 2025: Tides at Galveston, Texas. Accessed 1 July 2025, Source.

Figure 10-6b – NOAA Tides & Currents, 2025: Tides at Sewells Point (Hampton Roads), Virginia. Accessed 1 July 2025, Source.

Figure 10-6b – NOAA Tides & Currents, 2025: Tides at San Francisco, CA. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at Seattle, WA. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at Ketchikan, AK. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at Anchorage, AK. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at Dutch Harbor, Amaknak Island, AK. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at Boston, MA. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at LONG BEACH, NEW YORK, NY. Accessed 1 July 2025, Source.

Figure 10-6c – NOAA Tides & Currents, 2025: Tides at Key West, FL. Accessed 1 July 2025, Source.

Figure 10-7 – Adapted from Cooper, A., Ed., 1983: *The Times Atlas of the Oceans*. Van Nostrand Reinhold Company, 272 pp.

Figure 10-8 – Adapted from Naval Oceanographic Office, 1968: Section 1: Tides and Currents. *The Oceanographic Atlas of the North Atlantic Ocean*. H.M. Publication No 700, 75 pp.

Figure 10-9 – Bowditch, N., 1965: *American Practical Navigator*. H.M Publication No. 9, U.S. Navel Oceanographic Office, 1524 pp.

Figure 10-13 – Adapted from Cartwright, D., 1969: Deep-sea tides, *Sci. J.*, **5**, 60–67.

Figure 10-14 – Adapted from Defant, A., 1969: *Ebb and Flow: The tides of earth, air, and water*. University of Michigan Press, 132 pp.

Figure 10-17 – NOAA Tides & Currents, 2025: Tides at ALBANY, HUDSON RIVER, NY. Accessed 1 July 2025, Source.

Figure 10-18 – Price, A., 2007: Tidal bore. Accessed 1 July 2025, Source, CC BY-SA 2.0.

Figure 10-19 – Data from NOAA Tides & Currents, 2025: Tides at Cape Henry, VA. Accessed 1 July 2025, Source.

Figure 10-20 – Data from NOAA Tides & Currents, 2025: Tides at Cove Point MD. Accessed 1 July 2025, Source.

Figure 10-22 – Clipper, 2006: Rance tidal power station. Accessed 1 July 2025, Source, CC BY-SA 3.0.

This page titled 10.10.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11: COASTS

Critical Concepts

- **CC2 Isostasy, Eustasy, and Sea Level**: Earth's crust floats on the plastic asthenosphere. Sections of crust rise and fall isostatically as temperature changes alter their density or as their mass loading changes. This, in turn, causes isostatic changes of sea level. Eustatic changes of sea level occur globally when the volume of water in the oceans changes or when the volume of the ocean basins themselves change. Sea level changes create new coasts. Oceanic crust cools progressively after it is formed and sinks because its density rises. Thus, the hot spot volcanic islands slowly sink after they move away from the hot spot.
- **CC4 Particle Size, Sinking, Deposition, and Resuspension**: Suspended particles in the ocean sink at rates primarily determined by particle size: large particles sink faster than small particles. Once deposited, particles can be resuspended if current speeds are high enough. Generally large particles are more difficult to resuspend, although some very fine particles may be cohesive and therefore, also difficult to resuspend. Sinking and resuspension rates are primary factors in determining the grain size characteristics of beach sands and sediments at any given location.
- **CC8 Residence Time**: The residence time of seawater in a given segment of the oceans is the average length of time the water spends in that segment. In restricted arms of the sea, such as lagoons behind barrier islands and fringing reefs, residence time can be long, in which case the nutrients may be depleted, limiting the growth of corals and other marine species.

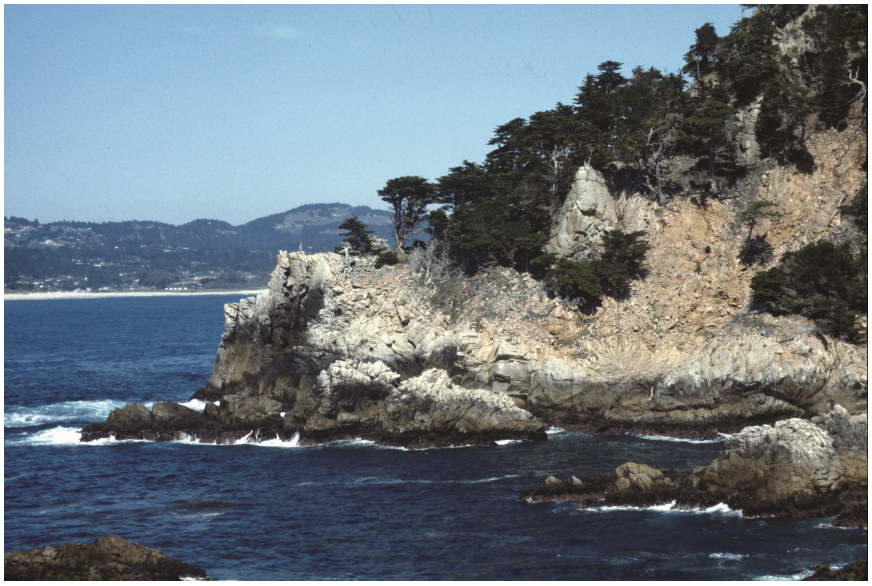


Figure 11-1. This beautiful rocky coastline is in the Point Lobos State Park just south of Monterey, California. In the distance, at the far side of the bay shown in the image, is the wide, white sand beach of Carmel, California.

A large proportion of the world's population lives within a few tens of kilometers of a **coast**. In 2020, 40% of the total U.S. population, or more than 129 million people, lived in counties directly on the shoreline, which account for less than 10% of the nation's land mass. Tens of millions more Americans live in counties that, while not shoreline counties, are close to a river or estuary. The concentration of human populations on or near the coast reflects the importance of the **coastal zone** to civilization.

The term coast has various definitions, but it is generally considered to be the strip of land between the **coastline** (where water meets land) and the inland location where there is no longer any environmental influence of the ocean. We build houses, factories, piers, docks, and marinas in this zone, primarily because the ocean is valuable for transport, recreation, and disposal of wastes, particularly treated sewage. In a human lifetime, most coastlines seem fixed and permanent. In reality, they are in a continuous state of change and movement. This chapter examines different types of coasts and the processes that form and reshape them. It also examines human attempts to control and mold the coasts, and the futility of many such efforts.

11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures
11.5: Coral Reefs and Atolls
11.6: Wetlands
11.7: Deltas
11.8: Chapter Summary
11.8.1: Key Terms
11.8.2: Chapter Questions
11.8.3: References

This page titled 11: Coasts is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.1: Processes that Form and Modify the Coast

All coasts change in character along their length, but their general character tends to be similar for thousands of kilometers. **Figure 11-2** and **Figure 11-3** show just some of the various coast types of the U.S. and Canada. Worldwide, every coast is continuously changing in response to various processes that form and modify it over decades, centuries, and millennia.

Figure 11-2. Pacific and Arctic coastlines of the U.S. (clockwise from top left). (a) Permafrost beach on the Beaufort Sea in the Arctic Wildlife Refuge, Alaska. (b) Cliff and beach terrace in Olympic National Park, Washington. (c) Storm-eroded coast with tree debris in Willapa Bay, Washington. (d) Rocky coast near Monterey, California. (e) Beach and cliffs near Point Magu in southern California). (f) Lava flows on the south shore of Hawaii. (g) Island beach and cliffs near Polihale, Kauai, Hawaii. (h) Fringing coral reef in Guam. (i) Fjord in Turnagain Arm, Alaska.

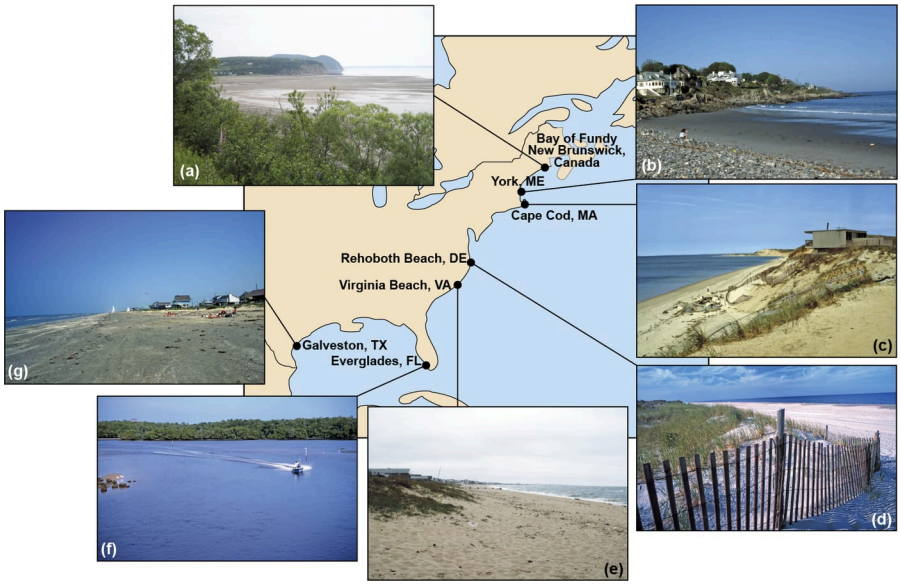


Figure 11-3. Atlantic coastlines of the U.S. and Canada (clockwise from the top left). (a) High tidal range in the Bay of Fundy off of New Brunswick, Canada. (b) Rocky beach with high wave energy in York, Maine. (c) Sand dunes in Cape Cod, Massachusetts. (d) Beach and dunes near Rehoboth Beach, Delaware. (e) Barrier island beach in Virginia Beach, Virginia. (f) Mangrove forest in the Florida Everglades. (g) Barrier island beach in Galveston, Texas.

Most coasts can be classified as either **erosional** or **depositional**, depending on whether their primary features were created by erosion of land or deposition of sediments. Erosional coasts develop where the shore is actively eroded (worn away) by wave action or where rivers or glaciers caused erosion when sea level was lower than it is now. Depositional coasts develop where

sediments accumulate either from a local source or after being transported to the area by rivers, glaciers or by ocean **currents** and waves (**Fig. 11-4**). Erosional coasts are often dominated by sea cliffs and rocky shores, whereas depositional coasts include **deltas**, **mangrove** swamps, **salt marshes**, **barrier islands**, and beach **sand dunes**.

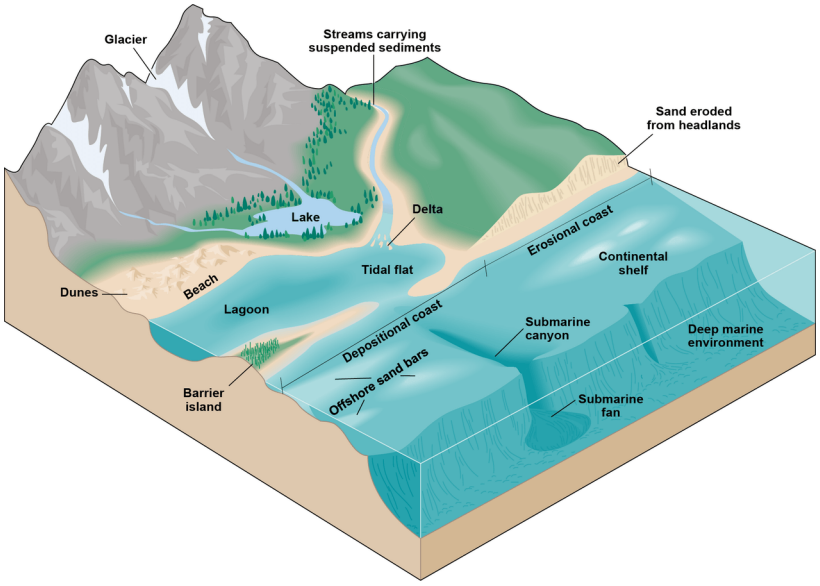


Figure 11-4. The major sources of sediments in the coastal ocean are erosion of the shoreline and transport of particulate matter by streams, rivers, and glaciers. Many river-borne particles are deposited in estuaries or lagoons along the coastal plain to form sediment deposits. Once in the coastal marine environment, sediment particles can be transported along the coastline and deposited on beaches, barrier islands, and offshore bars. Small sediment particles can be transported offshore by currents and deposited on the deep-ocean floor. Larger particles can be transported offshore by turbidity currents that sometimes travel down submarine canyons.

Coasts can also be classified as either primary or secondary. Primary coasts are shaped predominantly by terrestrial processes, including erosion or deposition by rivers, streams, glaciers, volcanism, and tectonic movements. Secondary coasts are shaped by marine erosion or deposition caused by wave action, sediment transport by currents, and marine organisms (e.g., those that form reefs). Many coasts have characteristics of both marine and terrestrial processes.

Formation of Coasts

New coasts are formed either when the relative levels of the ocean surface and coastal landmass change or when the edge of the landmass is added to or removed. Some processes that form coasts, such as volcanic eruptions and earthquakes, can occur instantly or over a very short period, but most other processes, such as sea-level change and **coral reef** growth, continue slowly over centuries. Many of these processes can occur on the same coast at the same time, but at different rates or frequencies.

Tectonic Processes

Chapter 4 describes movements of **lithospheric plates** that create the major **topographic** features of the ocean floor. It also explains how these movements build mountain chains and **magmatic arc** and **sedimentary arc** islands at **convergent plate boundaries**, create new oceans at **continental divergent plate boundaries**, and create volcanic islands at **hot spots** and particularly-active locations on **oceanic ridges**. Tectonic processes take millions of years to re-form the planet. For example, the plate movements that broke apart Pangaea to form the present configuration of continents required about 150 to 200 million years.

Motions of the lithospheric plates are slow but continuous, modifying coasts at plate boundaries and hot spots. At hot spots, new coast is formed when volcanoes erupt. For example, several hundred hectares were added to the island of Hawaii by an eruption of the Kilauea Volcano's East Rift Zone that started in 1983 and continued until 2018. Lava from East Rift Zone eruptions often flows into the sea, hardens, and extends the coastline (**Fig. 11-2f**). Tens of thousands of years from now, a new island, Loihi, will emerge south of Hawaii (**Chap. 4**).

New coast is also formed by volcanic eruptions in magmatic arcs at convergent plate boundaries. Indonesia and the Aleutian Islands are good examples (**Chap. 4**). At convergent plate boundaries or at **transform faults**, new coast is formed when

earthquakes uplift a continent edge as the oceanic plate is **subducted** beneath it, or when earthquakes uplift ocean sediments at a sedimentary arc. Although earthquakes along these boundaries that uplift land to create new coast are infrequent, coastal erosion processes are also slow. Therefore, uplifted coast is formed at some convergent plate boundaries periodically but almost instantaneously, much faster than the coast is modified by ocean processes. For example, the Loma Prieta earthquake in October 1989 raised the coastal mountains and coast near Santa Cruz, California, by as much as 1.5 to 2 m. A strong earthquake in Alaska in 1964 raised parts of the seafloor of Prince William Sound by as much as 8 m, creating a new strip of land several hundred meters wide from the former seafloor.

Coasts can also be destroyed by tectonic processes. For example, coasts of what is now northern India were destroyed as India collided with Asia (**Chap. 4**). Earthquakes at convergent plate boundaries or at transform faults can also cause sections of coastal land to move vertically downward, although such changes at these boundaries are often temporary because subsequent earthquakes may uplift this same section as the often complex subduction process continues.

Landslides

The simultaneous destruction of old coast and formation of new is particularly dramatic where volcanoes form islands, such as Hawaii, whose underwater flanks are much steeper than most other terrestrial margins. The steep flanks can become unstable as lava accumulates from continuing eruptions or when the island cools and sinks isostatically after moving off the hot spot. When this happens, a section of the island can break loose and slide down to the deep-ocean floor like a giant avalanche, destroying the old coast and creating a new coastline where the break occurs. Huge sections of the Hawaiian Islands have apparently broken off in this way in the past. As much as 10% to 20% of Oahu has instantly broken loose at one time. There is evidence that about 70 more such landslides have occurred around the Hawaiian Islands during the past 20 million years. The remains of these giant landslides are littered over vast areas of seafloor extending more than 200 km around the islands (**Fig. 11-5**).

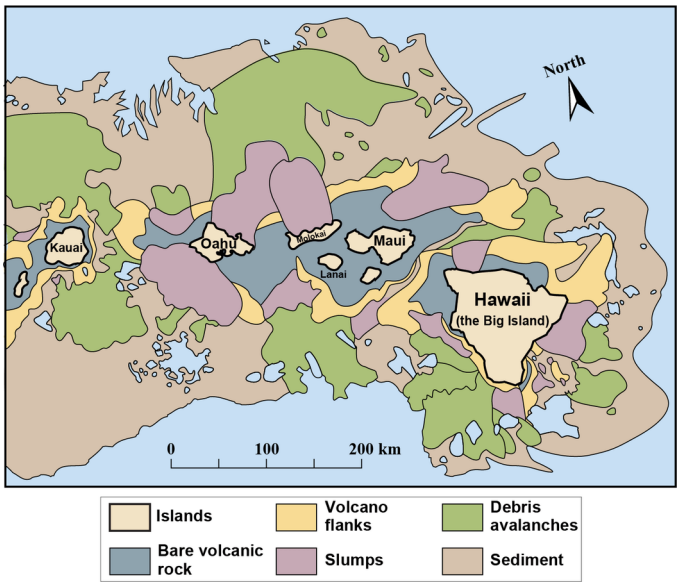


Figure 11-5. Large areas of the ocean floor surrounding the Hawaiian Islands are covered by slumps, fields of large debris, and sediments that originated from the islands. The data depicted on this map were obtained by the extensive use of precision sonar and three-dimensional sonar and the collection of many sediment samples. The slumps are areas where large blocks of the side of the volcano slipped downward. Slippage may occur both by a slow creeping motion and by periodic surges of several meters that cause large earthquakes. The debris fields and sediments are the result of catastrophic landslides in which huge blocks of the islands, up to 10 km or more in size, break loose instantaneously in massive avalanches that slide and flow down the steep volcano sides to the deep-sea floor. Although not recorded in recent history, these landslides must cause massive turbidity currents and, probably, massive tsunamis that may be hundreds of meters high when they reach the adjacent islands.

Little is known about these monster landslides or the probability of another one occurring on Hawaii or other volcanic islands. However, in November of 2000, a 20-km-long by 10-km-wide section of the southeast slope of the Kilauea Volcano on Hawaii slipped about 10 cm in only 36 h, millions of times faster than most tectonic plate motions. This occurrence may have been a forewarning of an imminent (in geological time) collapse of this section of the island. Such a slide not only could destroy a large section of the islands and their inhabitants, but also could cause a huge **tsunami** (**Chap. 9**), which might be several tens of meters high when it impacted the west coast of North America. Fortunately, such large slides apparently occur only at intervals of about 100,000 years or more in Hawaii and about once in every 10,000 years on average worldwide.

Although not as dramatic in size or impact, landslides smaller than those observed around Hawaii are important processes of coastline modification on most uplifted coasts. On eroding coastlines such as those found in southern California, many homes built years ago and tens of meters from the then existing cliff edge have been destroyed as the cliffs have progressively collapsed.

Isostatic and Eustatic Sea-Level Changes

If sea level rises, coasts are drowned and a new coastline is formed inland from the previous location. Similarly, if sea level falls, the ocean floor is exposed and becomes the new coastline. Sea level can change on a particular section of coast because the continent edge rises or sinks **isostatically**, while global sea level remains the same (**Chap. 4**, **CC2**). Sea level can also change **eustatically** if the volume of water in the oceans changes or the volume of the ocean basins themselves changes (**CC2**). Eustatic changes take place at the same time throughout the world's oceans, whereas isostatic leveling occurs locally or regionally. At present, worldwide sea level is rising slowly because of eustatic processes (melting of glaciers and the warming and expansion of ocean water), but sea level is not observed to be rising on all coasts. Some coasts are rising isostatically as fast as, or faster than, the rate of eustatic sea-level rise. The net result is that the observed sea level is stable or falling on these coasts. Other coasts are sinking isostatically, and the observed sea level on these coasts is rising faster than the rate of eustatic sea-level rise.

Eustatic sea-level changes have led to major shifts in the locations of continental coastlines. During approximately the past 19,000 years, sea level has risen about 120 m (**Fig. 6-18**). The history of sea-level change during this 19,000-year period can be determined by, for example, studies of the ages of **relict sediments** or buried sediments on the **continental shelf** (**Chap. 6**). Determining the history of sea level before 19,000 years ago is more difficult, but sea level seems to have oscillated many times during the present **spreading cycle**, from about 130 m or more below its present level, to about 40 to 50 m above the present level. Consider where the coastline would be if no isostatic changes had occurred during this oscillation of sea level. At the highest sea level, the Gulf of Mexico would extend across the central plains states of the U.S. as far north as southern Canada. At the lowest sea level, the Texas coastline would be about 150 km farther south in the present-day Gulf of Mexico and the Florida Peninsula would be about twice as wide.

During the past 4000 years, the eustatic rise in sea level has been slower than in the immediately preceding period. Therefore, most present-day coasts were formed several thousand years ago as sea level rose rapidly over what is now the continental shelf. Sea level is expected to rise more rapidly in the future as a result of global **climate** changes caused by enhancement of the **greenhouse effect**. In any event, the relatively slow sea-level change of the past 4000 years cannot continue indefinitely. If sea level does rise more quickly, the types of coasts at various locations will change because the rate of sea-level rise greatly affects the formation and migration of coastal features such as barrier islands and estuaries. Coastal changes may disrupt the Earth's ecosystems, thereby possibly causing more damage than even that caused by the **flooding** of coastal cities.

Because sea level is expected to rise more rapidly in response to climate change induced by the enhanced greenhouse effect, oceanographers are currently mounting intensive studies of coasts. Critical questions that remain to be answered include how fast sea level is rising eustatically, whether it will continue to rise, whether the rate of rise will continue to accelerate, how isostatic changes will enhance or mitigate eustatic changes in sea level on specific coasts, and how coasts will change if the rate of eustatic sea-level rise increases.

Glaciers

As glaciers flow, they scour out steep-sided valleys (**Fig. 11-6a**). Rocks and smaller particles that have been eroded from the valley walls and floor are carried by the glacier and deposited where the ice melts at the glacier's end. During the last 19,000 years, the Earth's climate warmed and the glaciers retreated, each one leaving one or more sedimentary deposits called moraines at the former location of the glacier's end. At the same time, sea level was rising as ocean waters warmed and expanded, and as more water entered the oceans from melting glaciers. The rising sea inundated many steep-sided valleys cut by glaciers and created deep, narrow **fjords**, many of which are partially closed off from the ocean by a submerged **sill**, which is usually a moraine (**Fig. 11-6b,c**).

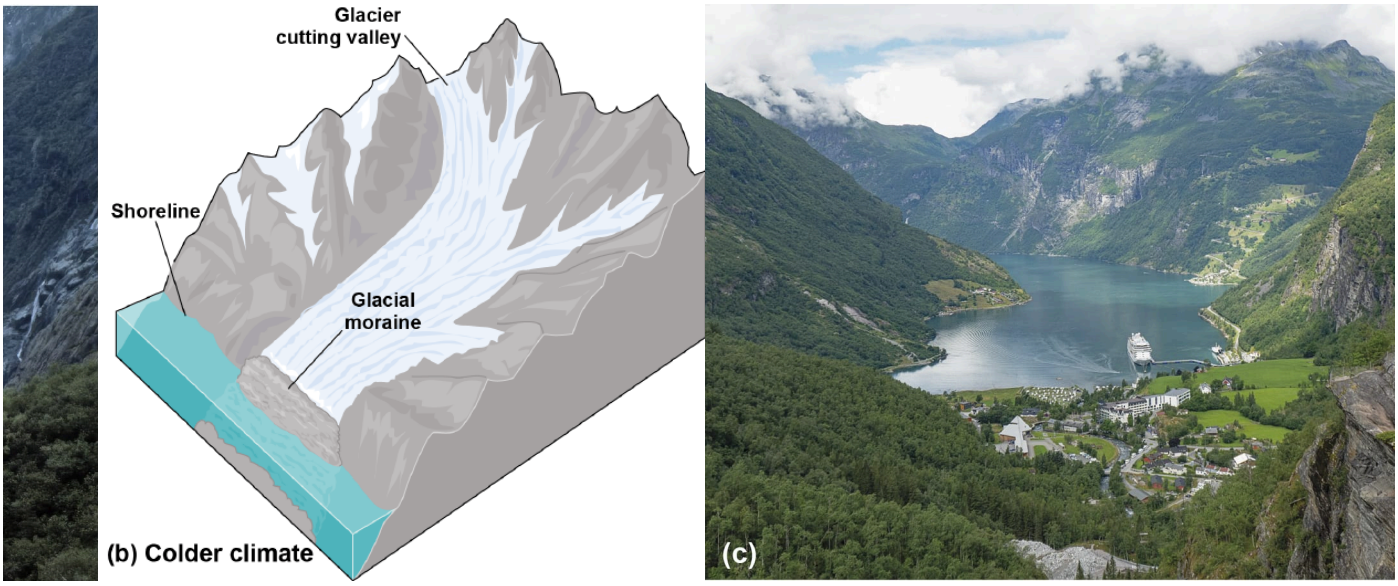


Figure 11-6. Glaciers and fjord formation. (a) The Franz Josef Glacier on the South Island of New Zealand has retreated substantially in recent decades. Note the very steep-sided valley that the glacier has left behind as it has receded. (b) When sea level was lower and glaciers more extensive, the glaciers cut deep, steep-sided valleys down to the ancient sea level. Rock fragments carried by the glaciers were deposited to form one or more glacial moraines at the foot of each glacier. (c) Geirangerfjord, Norway where rising sea level has filled a glacial valley that now allows ocean cruise ships to navigate to the head of the fjord, which is 60 km inland. (d) As the Earth's climate warmed and sea level rose, the glaciers retreated, and their steep-sided valleys were filled by seawater to form fjords. The glacial moraines left behind by the glaciers formed shallow submerged sills at the mouths of many fjords.

Because fjords are long, narrow inlets, they are generally well protected from erosion by ocean waves, and their shores are little altered from the original sides of the glacial valley. Many high-**latitude** areas where glaciers cut through coastal mountain ranges have extensive fjord systems. Excellent examples are found on the South Island of New Zealand, in Scandinavia, on the Pacific coast of Canada and Alaska, and in Patagonia, Chile.

River-Borne Sediments

New coasts are formed where large amounts of river-borne sediments are deposited. The extended delta of the Mississippi River (**Fig. 11-7**) and similar deltas elsewhere are examples of coasts formed by river-borne sediments. Deltas, discussed in more detail later in this chapter, are present at the mouths of relatively few rivers. Most of the world's rivers flow across a gradually sloping **coastal plain** before reaching the sea. The sediment load of the river is deposited in the river valley as the flow slows in this flatter area.

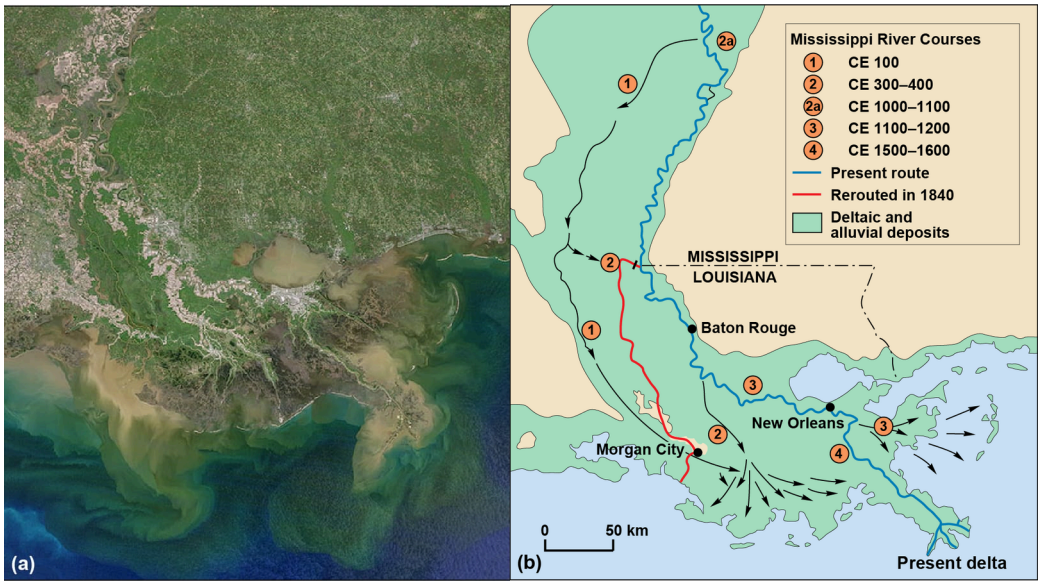


Figure 11-7. The Mississippi River delta. (a) This satellite image shows the extensive delta with its many distributary channels. (b) The Mississippi River periodically changes its course through the coastal plain and delta, distributing its sediments across the entire area. Flood-control levees now prevent this natural process.

Only a few rivers other than the Mississippi carry such large sediment loads that their river valleys have filled enough for large quantities of sediment to be transported to the sea. Most rivers that flow across coastal plains, such as those on the Atlantic coast of North America, carry considerably less sediment than the Mississippi. In addition, rivers emptying to the Atlantic Ocean have only recently (in geological time) begun to flow toward that ocean. In the region now drained by the rivers emptying to the Atlantic Ocean, rivers flowed away from the Atlantic Ocean until about 100 million years ago, when the newly formed **passive margin** of the Atlantic coast sank isostatically sufficiently far to reverse the slope (**Chap. 4**).

Rivers that flow across tectonically active coastal margins generally flow through steep coastal mountain ranges. Because they drain relatively small land areas and flow through steep valleys to the sea, many carry relatively little sediment, but most of it is transported to the oceans. The continental shelf is steeper and narrower on these active coastal margins, so sediment is transported to the deep-sea floor more easily than at passive margins.

Biological Processes

Reef-building **corals** cannot grow and build reefs unless they are underwater. However, reefs grow fastest in shallow waters where light intensity is high (**Chap. 15**), and some reef tops emerge above water at low **tides**. Although they are not truly land, these reefs constitute an important feature of the coast because ocean waves break on them and lose much of their energy. Fringing reefs and barrier reefs are present on many coasts in tropical and subtropical regions. A small drop in sea level or a small coastal margin uplift can raise coral reefs above the sea surface, where the corals cannot survive.

When corals die, they leave behind their hard “skeletons.” Many tropical and subtropical islands and coasts are characterized by rocks composed of old coral reefs. These coral-rock shores are eroded to form a jagged surface, often called “ironshore,” that makes walking difficult (**Fig. 11-8**). The Cayman Islands in the Caribbean Sea have excellent examples of reef-dominated shore. In fact, these islands are predominantly uplifted coral reefs. The ironshore on part of one island is so jagged that the local community has been named “Hell.”

Figure 11-8. An uplifted and eroded coral reef forms a jagged shore often called "ironshore." (a) Coastlines of coral islands, like this island in Palau, often have interspersed sandy beaches and eroding ironshore. (b) The jagged rocks of this ironshore segment in Fiji are razor-sharp.

Many coral reefs are located on islands or submerged pinnacles that are sinking isostatically (**Fig. 4-29**). Isostatic sinking and sea-level rise both tend to increase the depth of the water column over a reef. If the combined rate of these deepening processes exceeds the rate at which a coral reef can build upward, the top of the reef becomes progressively deeper and may eventually become too deep to sustain the **photosynthesis** on which the corals depend. Sea level has been rising for about the past 19,000 years, and many coral reefs appear to have been drowned in this way. If global climate change leads to an increase in the rate of sea-level rise, many more coral reefs may die, primarily those with lower maximum growth rates, even if they can survive ocean **acidification**.

The maximum upward growth rate of coral reefs varies with latitude, depth, and water clarity. Upward growth is slower at higher latitudes, where water temperatures are cooler, and they are lower at deeper depths or in less clear waters, where the light levels are reduced. However, the maximum upward growth rate is about 1 to 10 mm per year at 10 m depth, and the current rate of sea-level rise is estimated to be about 3 to 4 mm per year.

Modification of the Coast

All shores and coasts are continuously, but slowly, modified by waves, tides, winds, and biological processes. The present form of coasts represents a balance between modification processes and formation processes. Older coasts and coasts with higher wind, wave, and tide energies are generally more extensively modified. The extent of modification also depends on the type of rock constituting the coast and, in some cases, on the types of vegetation on the coastal land and in the nearshore zone.

Waves

The breaking of waves on the shore is the principal coast-modifying process. On rocky coasts, breaking waves progressively erode the rock away. Soft sedimentary rocks are eroded much faster than harder volcanic rocks. In addition, erosion is faster on coasts that are exposed to greater wave action. Wave action is greater in areas of frequent storms or where the coastline is impacted by waves that travel far across the ocean.

As discussed in **Chapter 9**, wave energy is focused on headlands and spreads out along the interior shores of bays. Consequently, on an indented coastline, erosion occurs fastest at headlands, and the products of erosion (sand) accumulate within the intervening bays. Preferential erosion of headlands along a coastline tends to straighten the coast progressively until no headlands remain. This process is complete on many coasts, particularly where rocks are easily eroded. The New Jersey coast is an example (**Fig. 11-9**).

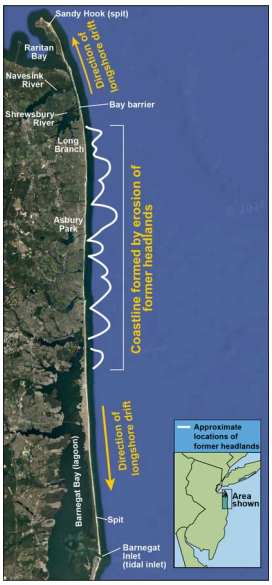


Figure 11-9. The New Jersey coastline has long, straight beaches and barrier islands. Headlands that used to exist along the northern shore have been eroded away, providing sand for the formation and maintenance of the spits to the north and south.

On rocky coasts, waves cut away rock between the **high-tide lines** and **low-tide lines**. As rock is cut away, the land becomes unstable and breaks away or slumps, leaving behind a cliff that may be nearly vertical (**Figs. 11-2g, 11-10**). The debris from the cliff temporarily alters the shape and nature of the beach where the cliff face has collapsed. However, these rocks fall into the wave-breaking zone, where they are eroded away relatively quickly. The beach is thus restored, and the waves renew their attack on the base of the cliff. In many locations, there is no beach at the base of the cliffs because wave energy is too high or wave erosion has not continued long enough for sand to accumulate (**Fig. 11-11**).

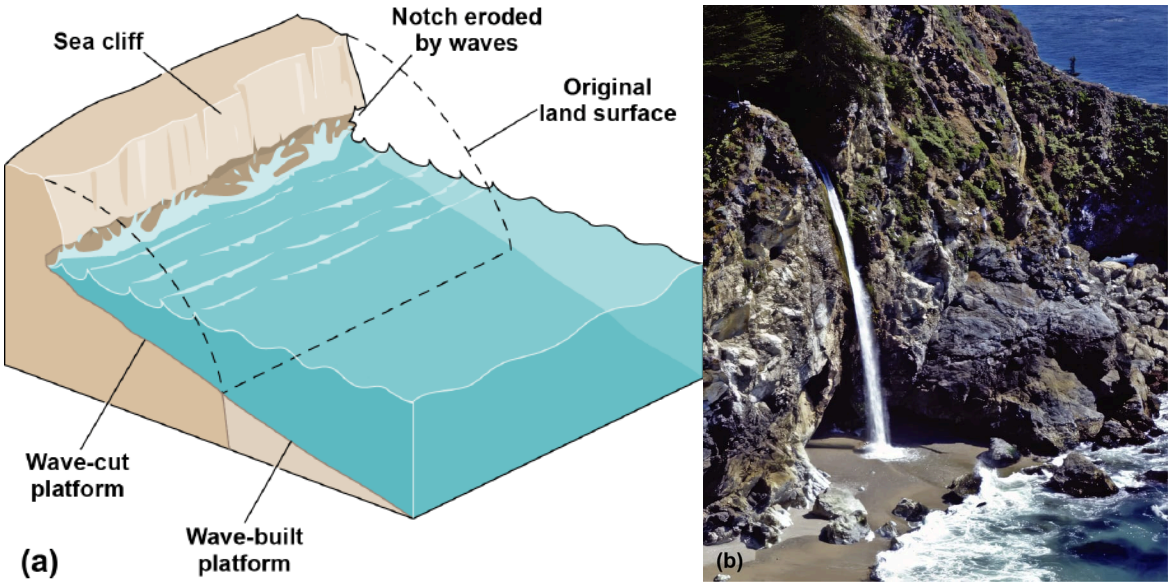


Figure 11-10. Wave erosion on steep coastlines can create cliffs. (a) As rock is eroded by waves, the land is undercut, becomes unstable, and slumps. The slumped material is eroded and transported away by the wave action. Thus, a gently sloping wave-cut platform is maintained at the base of the cliff. (b) Cliff and beach near Monterey, California.

Figure 11-11. Wave erosion, combined with boring and dissolution by marine organisms, undercut this rocky coastline on the island of Palau in the Pacific Ocean. Here, erosion is confined to a very narrow height range because the tidal range is nearly zero.

As waves cut into coastal cliffs and headlands, they encounter rocks of variable resistance to erosion. Rapid erosion of the less resistant rock often leads to the formation of sea caves at the base of a cliff (**Fig. 11-12a,b**). Sea caves that are cut into either side of a headland can continue to be eroded until they meet under what remains of the headland, resulting in the formation of a sea arch (**Fig. 11-12b,c**). As the headland erodes further, the arches collapse, and the remaining pinnacles of rock, called "stacks" (**Fig. 11-12b,d**), are eventually eroded away.

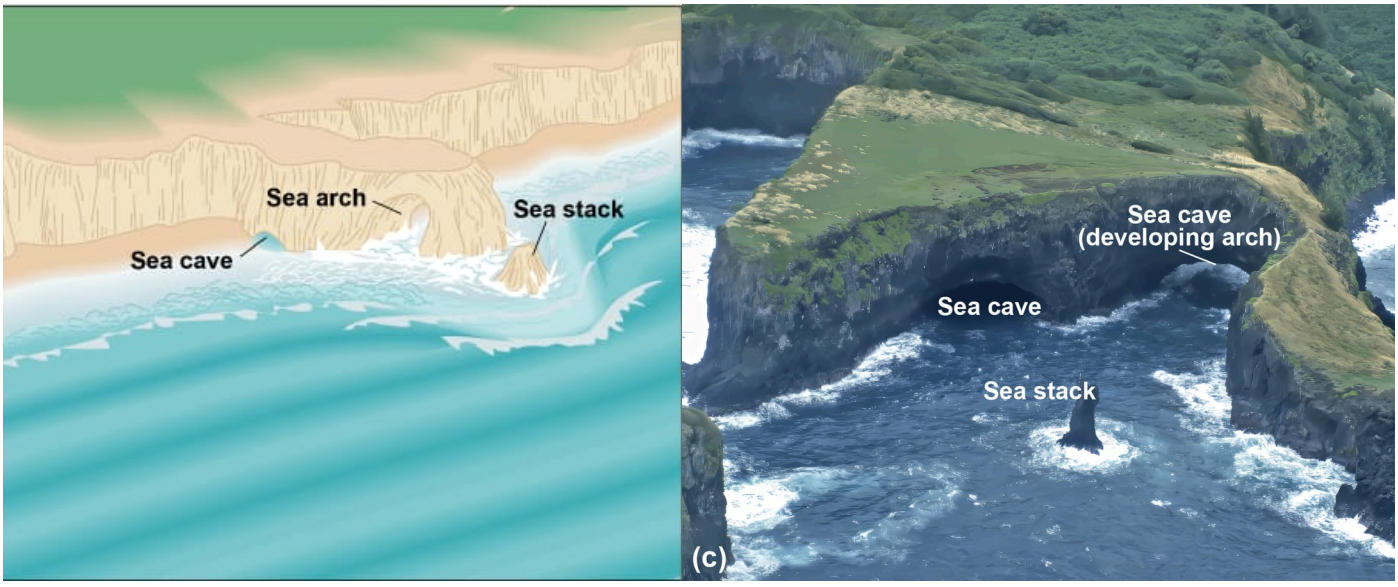


Figure 11-12. (a) Wave energy is concentrated on headlands by wave refraction. Because the seafloor topography is rough and varies as slumps occur from different parts of the headland, wave energy is distributed unevenly. In addition, the rocks of the headland differ in composition and susceptibility to erosion. As a result, the wave action causes the headlands to be eroded unevenly. (b) The uneven erosion of the headland can cause sea caves and sea arches to be formed. (c) This headland on the island of Molokai in Hawaii has been eroded by the intense wave action to form sea caves and sea stacks. (d) There are many sea stacks on the Pacific coast of North America, including these on the Big Sur coast south of Monterey, California.

Coasts that have been substantially eroded usually have beaches. Off these coasts, waves transport and distribute particles (e.g., sand, silt, or pebbles) that make up the beach. These processes are discussed later in the chapter. Beaches help protect the coast from wave erosion.

Tides

The shore is defined as the area along the coast that lies between the lowest point exposed at the lowest tide and the highest point reached by storm waves. The shore consists of the **foreshore**, which is the area between the low-tide line and the high-tide line, and the **backshore**, which is the area above the high-tide line that is affected by storm waves (Fig. 11-13). **Tidal range** determines the height range over which wave erosion occurs and, thus, the width of the shore. Where the tidal range is large, wave-erosion energy is spread over a large vertical range. Generally, coasts with small tidal ranges are eroded faster because wave energy is concentrated in a narrow zone. The swift currents associated with large tidal ranges have little erosional effect because they are slow in relation to the orbital speed of water in waves (**CC4**).

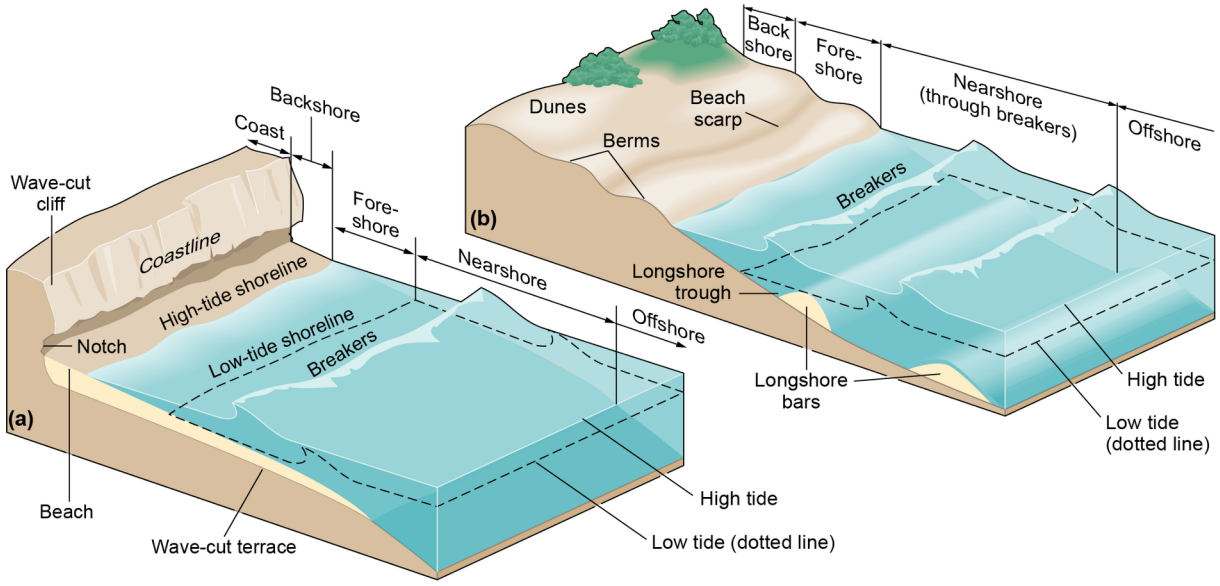


Figure 11-13. The coastline is separated into several zones that reflect different exposures to erosion and sediment movement due to the variation in waves and tides. The backshore is covered by water only during storms, but it is subject to seawater wind spray constantly. The foreshore is the region between the high- and low-tide lines, and the nearshore is the region between the low-tide line and the depth at which wave action no longer affects the seafloor. (a) When the shoreline is more resistant to weathering, wave action may cut steep cliffs with a notch underneath and a terrace offshore. (b) Longshore bars and troughs may be present on the seafloor within the nearshore zone. The foreshore is often separated from the backshore by a berm. Also, a scarp may be present in the foreshore at the height reached during a recent high tide by waves that are not strong storm waves but somewhat higher and steeper than waves that reached the beach during previous days or weeks and on the backshore at the base of sand dunes (see also Figs. 11-14, 11-15c,d).

Tides are particularly important in shaping and maintaining wetlands. Tidal motions in shallow bays and estuaries transport and redistribute sediments. In some bays and estuaries, extensive mudflats form just below the high-tide line. These mud flats are usually dissected by drainage channels. During the tidal cycle, the mudflats of tidal wetlands are alternately exposed and covered by shallow water.

Winds and Weather

Onshore winds can carry sand from beaches and deposit it on the backshore above the highest point reached by waves. This process leads to the formation of the sand dunes (Fig. 11-14) that characterize many coastlines. Conversely, sand dunes and soils can be eroded by offshore winds and deposited in the oceans. Just as they are elsewhere, coastal rocks, sand dunes, and soils are eroded by water in streams and rainfall, and by various processes, including alternate cycles of freezing and thawing and dissolution in acidic rainwater. Wind-driven particles and ocean spray may also erode rocks or soils.

Figure 11-14. Sand dunes are characteristic of many coasts. (a) Dunes usually separate the beach from a low marshy area, as they do on this Aruba coast. (b) Dunes can be 10 m or more high, like these in Monterey Bay, California. Note the patches of vegetation on the dunes that help reduce erosion.

Vegetation

The type and extent of vegetation on the coast affect the rate at which winds, streams, and storm waves erode the land. Grasses are particularly important in protecting sand dunes from erosion. Similarly, rooted plants that grow in the water, including **sea grasses** and mangroves (**Chap. 13**), help reduce erosion of mudflats by waves and currents. In contrast, tree roots and animal activities, such as burrowing, contribute to the continuous erosion of rocks and soils of land near the coast just as they do elsewhere. In addition, especially on rocky coastlines, many animals that live in the zone between high and low tides erode the rocks as they bore or chemically dissolve their way into the rock or probe into cracks in the rocks, either to find food or shelter or to be able to “hold on” to the rock against the power of the waves.

This page titled 11.1: Processes that Form and Modify the Coast is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.2: Beaches

Beaches are endlessly fascinating. We love to sunbathe on them, play in the **surf**, search for shells and other things washed up by the tide, and simply walk along them, enjoying their natural splendor. To many people, beaches are the best place to enjoy what is usually an unspoiled natural **environment**. Civilizations have dramatically altered much of the dry land adjacent to the oceans, but beaches have been substantially changed in only a few locations.

Although beaches appear to be resistant to change, in reality, they are places of continuous movement and transformation. Beaches are formed by processes that occur in the zone between the seaward boundary of land vegetation (or the base of a cliff, if present) and the point where the seafloor reaches a depth at which sediment is no longer disturbed by waves, commonly 10 to 20 m depth (**CC4**). This zone includes the coast and shore and is known to geological oceanographers as the **littoral zone** and this meaning is used in this chapter. Unfortunately, the term littoral zone means something different to biological oceanographers: the zone between the high- and low-tide lines.

The Littoral Zone

The landward side of the beach is delineated by either a cliff or sand dunes. Cliffs vary in height and steepness. Sand dunes vary from a few centimeters to more than tens of meters high and can be hundreds of meters wide. The beach that stretches out to sea from the cliffs or dunes consists of three zones. First, the backshore zone is the low-slope area of the upper beach that is normally dry, even at high tide (**Fig. 11-13**). Although it is generally flat and slopes only gradually downward toward the sea, the backshore may have well-defined areas where the downward slope steepens abruptly (**Fig. 11-13**). Such areas are called berms. The top of a berm, built by sand deposited on the beach during periods of quiet wave action, is usually flat, and the seaward side slopes downward relatively steeply (**Fig. 11-15a**). Sometimes the berm appears as a slight mound or ridge (**Fig. 11-15b**). Berms generally continue along the length of the beach at a fixed height above the water. In many places, we see several berms on the same beach at different heights above the water.

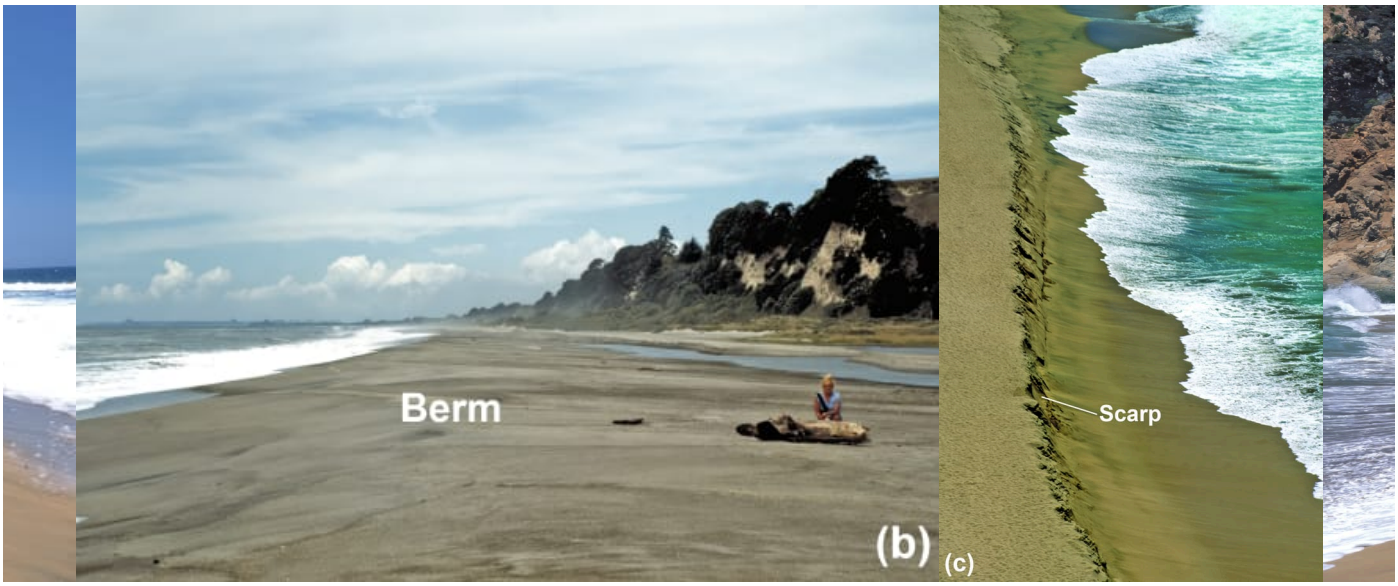


Figure 11-15. Berms and scarps. (a) There are two berms on this beach just south of San Francisco, California: One in the center of the photograph, where the waves of a recent storm have reached, and another much higher on the left. This upper berm must have been formed during a strong storm. (b) The berm on this beach on the east coast of the North Island of New Zealand forms a distinct ridge with a depression on its landward side. During major storms, the waves will pass far over this berm to erode the cliffs seen in the background. (c) This beach scarp on the California coast just south of San Francisco is about 1 m high. (d) The same scarp as in part (c), seen from a different angle.

Berms are created by storm waves that cut away sand to form the steeply sloping side of the berm. The highest berms are created by storm waves that reach farthest up the beach. Because the strongest storms occur in winter in most locations, the uppermost berm is often called the “winter berm,” even though it may actually have been created by an infrequent great storm that occurred

one or more years earlier. Berms are not permanent features. They are continuously destroyed, and new ones are formed as **wave heights** and tidal ranges vary. Where there is more than one berm on a beach, the uppermost berm must have been formed first, followed in succession by each lower berm, as storms obliterate any berm that is lower than the height on the beach to which its waves reach.

The second zone of the beach is the foreshore (**Fig. 11-13**). It is the area between the highest point reached by waves at high tide and the lowest point exposed at low tide. This area is commonly an almost flat and featureless slope and is often called the **low-tide terrace**. Foreshore slopes can vary greatly from beach to beach and from month to month at the same beach, but they are generally greatest on beaches composed of coarse-grained material. At the landward edge of the low-tide terrace (the high-tide line), there is often an abrupt vertical face of sand, usually a few centimeters high, before the beach slopes normally upward again as dry backshore. This feature is called a **scarp** (**Fig. 11-15c,d**). Like a berm, a scarp stretches along the beach, and there may be more than one on the foreshore. Scarps are caused by waves that are somewhat stronger than those that have reached the beach in previous days or weeks, cutting away sand from the beach, and the location of the scarp represents the height that these waves reach at high tide. Waves flatten the low-tide terrace when it is covered with water. A scarp often develops on the backshore at the base of any sand dunes and where cliffs develop the cliff itself may be considered to be also a scarp (**Fig. 11-15b**).

If two scarps are present on the foreshore, the one higher on the beach was formed before the lower one at a time when either the tide (**spring tide**), waves, or both were higher. The lower scarp would have been flattened and destroyed if waves had passed over it to cut the higher scarp. The boundary between backshore and foreshore is defined as the location reached by waves at high tide. This location is not fixed, because the maximum tidal height changes from day to day (**Chap. 10**) and wave action varies. When waves are small, they do not reach far up the beach. However, storm waves can crash far beyond the highest point reached by smaller waves. In addition, when strong coastal winds blow, **Ekman transport** can raise or lower local sea level on a coast by a meter or more (**Chap. 8**).

Seaward of the low-tide line, the beach continues to slope downward. However, if we were able to remove the water, we would see, at some distance offshore, a trough parallel to the beach. Seaward of the trough, there is a rounded, elongated mound of sand, called a **longshore bar**, before the seafloor resumes its downward slope (**Fig. 11-13**). Sometimes there is more than one longshore bar, each located at a different depth and distance from shore. Longshore bars are created and maintained by wave-driven sand movement. Although we cannot see submerged longshore bars, we may see evidence of them. When high waves with a long **wave period** approach the shore, a line of surf generally forms tens or hundreds of meters offshore. The offshore surf line is caused by a longshore bar or reef where the seafloor is elevated and water is shallower than it is immediately inshore or offshore. Waves may break over this bar (**Chap. 9**), then re-form over the deeper water inshore and break again near the beach.

In some areas, longshore bars have been built up high enough to emerge above the normal water line and become barrier beaches or even barrier islands. Longshore bars and barrier islands are dynamic features that are continuously created, destroyed, and moved. Some consequences that often ensue when we ignore this basic truth are discussed later in this chapter.

Sources of Beach Materials

For most of us, the word beach conjures up visions of a long strip of land next to the water that is composed of sand grains. The grains and beach usually have a color so familiar that it is called “sandy.” However, not all beaches are composed of this familiar sandy material. Some are composed of rounded pebbles or cobbles (**Fig. 11-16a**), and the color of the sand on a beach can vary dramatically.

Figure 11-16. Many beaches are not soft and sand-colored like those that are so popular for recreation. (a) This beach in New England is covered by large cobble-sized deposits, a sure sign that intense storm waves often reach this beach and transport all of the fine-grained material elsewhere. (b) The sand on this beach on the south shore of the island of Hawaii is black. The sand particles that make up the beach are eroded from nearby black volcanic rocks. Unfortunately, the beach shown here has now been destroyed by lava flowing from the Kilauea volcano.

For example, volcanic island shores where the coast consists of recently solidified black **basaltic** rocks have black sand beaches (**Fig. 11-16b**). The island of Hawaii has several black sand beaches. Other beaches on Hawaii and Maui consist of green or red sand, where the coasts consist of volcanic ash with high concentrations of iron or other elements that form colored minerals. In some cases, these odd-colored Hawaiian beaches are only a few hundred meters around a headland from beaches that have a more typical sandy color. The sand grains on such beaches originate from erosion of local rocks. Erosion is continuous on all shores and is due primarily to the action of waves.

When rocks are **weathered**, some minerals slowly dissolve and others change in composition. After prolonged exposure and erosion, only the most resistant minerals remain. The most resistant rocks include granite, which is composed mostly of the solution-resistant silicate minerals, primarily quartz and feldspars. Beaches on many shores consist mainly of grains of these minerals that, with their impurities, have the common sandy color. Many of these sand grains may have been eroded upland and brought to the beach by storm flows of rivers and then extensively weathered by water exposure.

Many low-lying coasts have no rivers that can carry large amounts of quartz and other mineral sand grains to the ocean and have little coastal erosion. Beaches in such areas are absent or poorly developed and may be composed of other materials. In many tropical or subtropical areas, particularly on low-lying coasts, beaches are composed primarily of calcium carbonate in the form of shell fragments from **foraminifera**, **mollusks**, **echinoderms**, and of platelets from **calcareous algae** such as *Penicillus sp.* and *Halimeda sp.*

If we look carefully at a handful of sand, we see that its grains consist of several materials. Sand is a mixture of particles from different sources. Sand composition may differ on two beaches that are close to each other, indicating local river input, local erosion, or shell fragment washup contributes most of the material on at least one of the beaches. However, along most coasts, sand has much the same composition on all the beaches between any two river mouths, aside from some variations in **grain size**. On these coasts, river transport may be the dominant source of beach sands, and coastal erosion may be less important. However, for many locations the relative contribution of each source is not well known. If rivers supply most of the sand for beaches stretched out along many kilometers between river mouths, sand must move along the coast from the river mouths to form the beaches. This movement is achieved by **longshore drift**.

Longshore Drift

As waves travel through shallow water, some of their energy is transferred to sand grains, which sets them in motion (**CC4**). This interaction continuously forms and re-forms beaches and transports sand along the coast.

When a wave breaks, it rarely does so exactly parallel to the **shoreline**. Usually the wave approaches from a slight angle, and the water that crashes on a beach does not move directly up the beach slope (**Fig. 11-17**). As the breaking wave moves up the beach (the **swash**), it **resuspends** sand grains from the beach and carries them with it. Thus, sand grains move up the beach, and a small distance downcoast, in the direction of the waves. When the water from the broken wave flows back down the beach (the **backwash**), sand grains flow with **gravity** directly down the beach slope (**Fig. 11-17**). Each swash and backwash cycle may move the grain only a centimeter or two along the beach. The net result is that sand grains are carried along the beach in the direction of the waves by a series of saw-toothed swash and backwash movements. Water from the wave is also transported along the beach, creating a **longshore current** in the zone landward of the breakers. Movement of the sand with this current is called **longshore drift** or **littoral drift**.

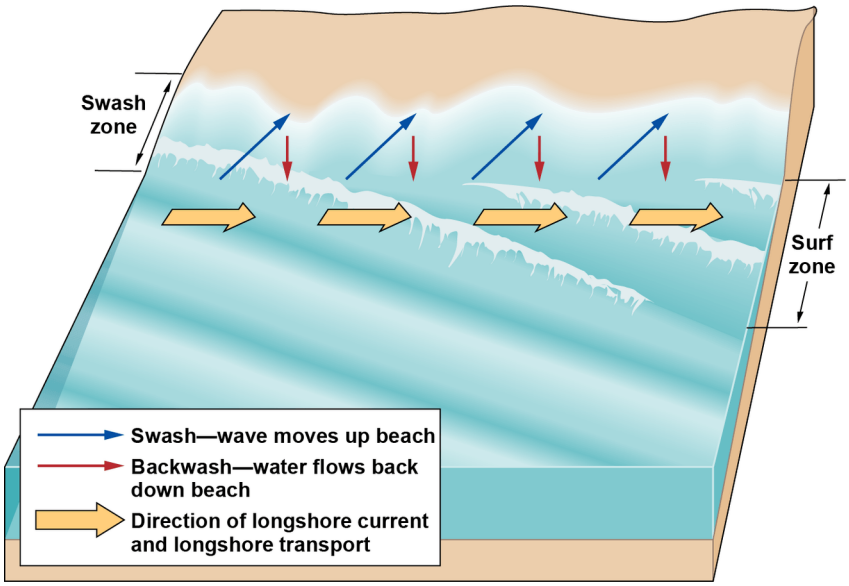


Figure 11-17. Water moves up a beach face at an angle with the wave direction but returns directly down the beach slope. Sand picked up by the incoming wave is washed up the beach and is returned seaward a small distance further down the beach in the direction of the waves. Successive waves move sand progressively along the beach—a process known as longshore drift or littoral drift.

Although each wave moves sand grains only a very small distance along the beach, waves follow each other every few seconds in a continuous series. As a result, sand can be moved by longshore drift at speeds that range from the typical rate of a few meters per day to as fast as 1 km per day. Large quantities of sand are moved along the coast by longshore drift. Transport along the east and west coasts of the U.S. is estimated to range from several hundred to more than 5000 m^3 of sand per day, depending on factors such as the height and **frequency** of waves.

Longshore drift is directed downcoast, in the direction away from which the waves approach, and it can reverse if that direction changes. Waves on both the east and west coasts of the continental U.S., particularly the larger and more energetic storm waves, come predominantly from the north. Therefore, longshore drift is generally to the south along both coasts.

If so much sand moves along the coast every day, where does it go and why is there any sand left on the beaches? As sand moves along the coast, it may eventually meet the head of a **submarine canyon**. Instead of continuing to move along the coast, it is funneled into the canyon and flows down onto the deep-sea floor. As this sand is lost, new sand is brought to the beach by river flow and erosion of the shore. The beach is maintained by the balance between sand supplied from these sources and sand lost down the canyon. Thus, along the coast there are a series of separate cells within which beach sand is supplied and transported until it meets the head of a submarine canyon. For example, southern California has four well-defined coastal cells of this type (**Fig. 11-18**). Little or no sand passes south from one cell to the next. If you drive down the California coast, you can see the southern ends of these longshore drift cells where the sand beach ends abruptly. You see cliffs or cobble beaches as you continue south from these points until you reach an area sufficiently supplied with new sand to form the first beach of the next cell.

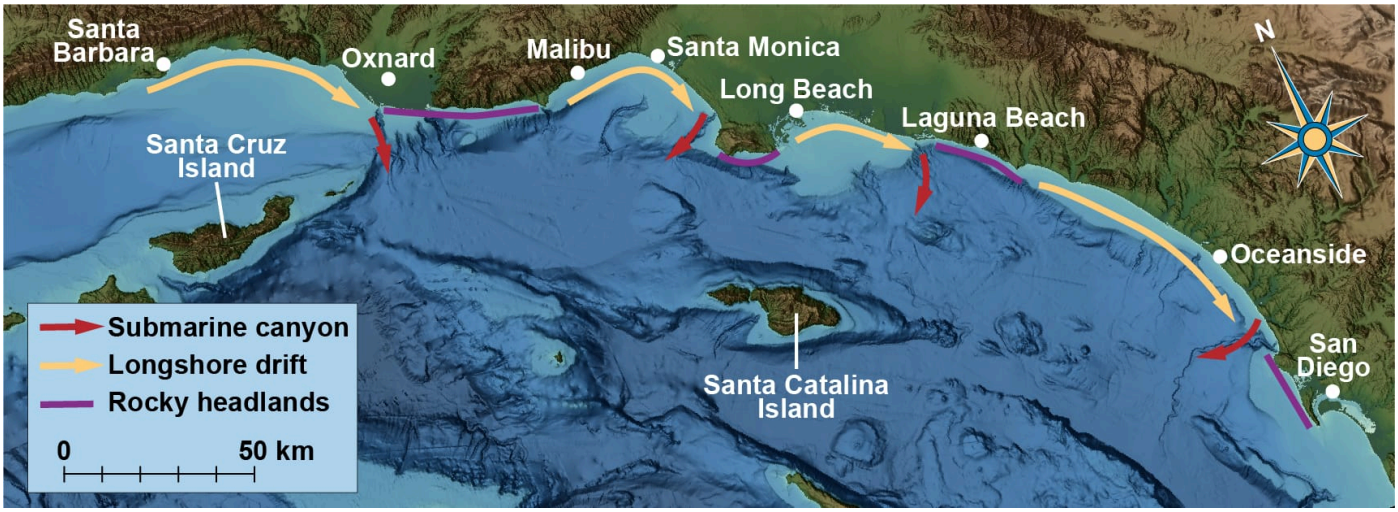


Figure 11-18. There are several north-to-south littoral drift cells in southern California. Each of these cells ends at the head of a submarine canyon where the sand is lost as it is transported down the canyon.

The future of California's beaches is in some doubt. Most rivers emptying into the Pacific Ocean have been dammed. Each dam acts as a trap for sand moving down the river. In addition, the dams and water withdrawals reduce the rivers' flow rate and speed, thereby reducing their ability to erode rocks and to carry sand (**CC4**). If rivers, rather than coastal erosion, are the major historical sources of most sand on California's beaches, this reduction in river flows will lead, or may already have led, to the progressive depletion of sand on the beaches. Beaches will, or may already have, become narrower, and waves will, or may have been able to, more easily reach the backshore, where communities such as Malibu are already susceptible to wave damage. Loss of river borne sand may be compensated by an increase of sand derived from coastal erosion, which is increasing due to climate change driven increase in wave energy and rising sea level. Sand supply by rivers and by coastal erosion are both highly variable and, as discussed later in this section, sand is also continuously moved between beaches and longshore bars. Because of this variability, studies of the future fate of, and the existing effects of dams on, these beaches have been inconclusive. However, many California beaches have required regular beach nourishment (placement of new sand by humans) since the 1930s.

Longshore drift cells on many other coasts are not as well defined as those on the California coast. For example, on the North Atlantic coast of the U.S., the cells are irregular and complex because of the many local barriers to longshore drift, such as headlands, deep river mouths, and rocky shores. In addition, this coast has a broad, flat continental shelf and few steep canyons through which sand can be transported offshore.

Wave Sorting of Beach Sands

Waves move sand if the speed of orbital motion is fast enough to resuspend the sand grains (**CC4**). Water in the wave moves toward the beach with the wave **crest** and then away from the beach when the **trough** passes over. Therefore, we might expect resuspended sand grains to move forward and back and return to their original location with each wave pass. However, the **orbital velocity** of the wave is distorted by its interaction with the seafloor. Water particles are farther from the seafloor as they flow toward the beach in the wave crest than they are as they flow back away from the beach in the trough. Because bottom **friction** increases nearer the seafloor, orbital velocity is lower in the wave as water moves offshore (trough) than it is when the wave moves onshore (crest). Although offshore (trough) flow is slower, it lasts longer than onshore (crest) flow. If this were not so, the wave orbit would not be complete (Fig. 11-19).

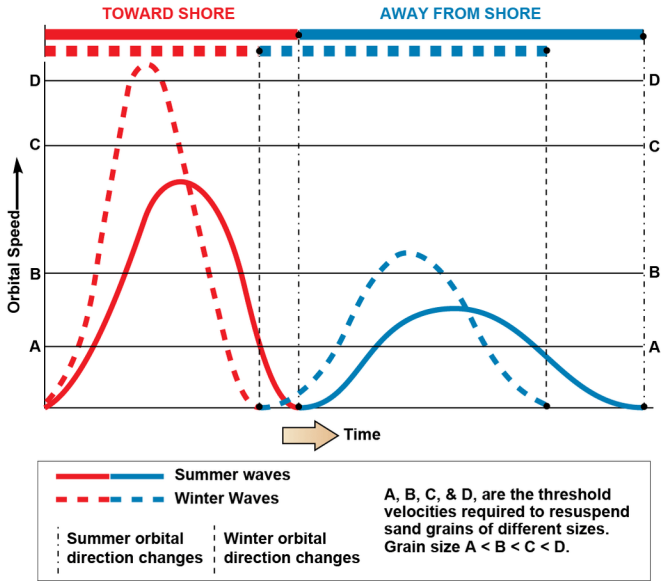


Figure 11-19. Water in a wave moves onshore with the crest at higher speeds and for a shorter period than it moves offshore with the trough. Winter waves generally have shorter wavelengths, greater wave heights, and thus, faster orbital speeds than summer waves. The text describes how variations in wave orbital velocities affect the transport of different sizes of sand grains (A, B, C, D) on a beach and how these processes sort sand grains by size.

The responses of sand grains to the difference in orbital velocity between crest and trough and to the varying height and **wavelength** of waves reaching the beach are the factors that control the movement of sand grains across the littoral zone. Winter storm waves are generally higher than summer waves, and they have shorter wavelengths and periods. The orbits are larger (higher waves) and must be completed in less time (shorter period) in winter waves, so they have higher orbital velocity.

Consider the movements of sand grains of different sizes in response to winter and summer waves. In **Figure 11-19**, sand grains of size A are small enough to be resuspended easily by passage of both the crest and the trough of winter and summer waves. Even if there is time for such grains to be deposited temporarily between crest and trough, they are immediately resuspended and will not be deposited on the beach. These particles eventually are carried offshore by rip currents. Once in deeper water, they settle to the seafloor, where waves will no longer resuspend them, because the orbital velocity of waves in deeper water is reduced with depth below the surface (**Chap. 9**). Thus, fine-grained sand is picked up from the beach, transported seaward, and deposited in deeper water below the influence of waves. The finest-grained sand is resuspended and transported offshore, where it is carried by ocean currents until deposited elsewhere (**Chap. 6**).

Sand grains of size B (**Fig. 11-19**) are resuspended under the crest of summer waves, but they cannot be resuspended by the slower orbital velocity under the trough. These particles are resuspended, moved shoreward with the crest, and redeposited as the velocity drops. They remain in place through the return flow (trough), which has lower velocity, and are resuspended and moved farther shoreward by the next crest. Thus, summer waves move sand grains of this size range from offshore onto the beach. When larger winter waves arrive, size-B grains are resuspended under both crest and trough and therefore are moved offshore. Because size-B grains are relatively large, they are not transported by ocean currents (whose velocities rarely approach the orbital velocities of waves), so they are deposited offshore just beyond the depth of wave influence. Hence, during winter, size-B particles are removed from the foreshore, transported offshore, and deposited to form a longshore bar. The longshore bar builds until it is just deep enough that the orbital velocity of passing waves is not quite fast enough to resuspend the sand grains.

Larger sand grains, of size C (**Fig. 11-19**), are moved up the beach by strong winter waves, but they are too large to be moved back offshore, even by the most intense storms. The largest grains (size D) cannot be moved at all and will remain until they are physically broken down into smaller grains that can be moved.

If wave action were the only process occurring, we would expect beaches to be composed of a mixture of sand grains of different sizes, lacking only those small enough to be continuously resuspended and removed to the deeper ocean. Beaches protected from strong wave action would be an unsorted mixture ranging from fine-grained sand to pebbles and boulders. Beaches with greater wave action would have no sand grains below a certain size. This grain size minimum would be larger for beaches subjected to strong wave action, but all grain sizes above the minimum would be present. However, we rarely find such a mixture. Each beach tends to be characterized by a narrow range of grain sizes, although the range varies from beach to beach (e.g., **Fig. 11-16**).

Why are virtually no large grains found on a fine-sand beach and no pebbles found on a sandy beach? The answer is that the range of grain sizes present depends on the sediment supply as well as on wave energy. The sand that makes up a beach is supplied mainly by river inflow or erosion of rocky headlands and is carried to the beach by longshore drift. Large particles, such as those of size D in **Figure 11-19**, cannot be transported, because they cannot be resuspended and moved by longshore drift, even by the most intense waves. Therefore, any boulders and other large particles present on a beach must have been eroded from the adjacent cliffs.

The size range of beach sand particles is thus limited because particles must be small enough to be resuspended and moved by longshore drift, but large enough not to be resuspended easily by the troughs of the waves. This explains why we normally see only a narrow range of grain sizes among particles on most beaches. Sedimentary deposits are said to be well **sorted** when they have only a narrow range of grain sizes (**Chap. 6**, **CC4**). Beaches with extremely fine sand are those best protected from wave action (and any coarser material present is rapidly buried). Pebble or cobble beaches, such as that shown in **Figure 11-16a**, are regularly exposed to intense storm waves that winnow away smaller particles.

Seasonal Changes in Beach Profiles

If we look carefully, we can see that the sand of beaches with calm summers and stormy winters is finer-grained in summer than in winter. In winter, both large and small sand grains are transported along the beach by longshore drift, but the finer grains are also moved offshore and deposited as longshore bars. In summer, these finer grains are moved back onto the foreshore, where they cover or mix with the coarser grains of the winter beach.

Seasonal movement of sand changes the beach profile. In winter, the foreshore moves back toward the cliffs or dunes as sand is winnowed by winter waves and deposited offshore. The foreshore is more gently sloped in winter than in summer, but the backshore is steeper (**Fig. 11-20**). When "quiet" conditions of summer return, fine-grained sand is returned to the beach from offshore. However, the summer waves do not reach as far up the beach as the winter waves, so a rounded ridge of sand may be formed at the highest point that these summer waves reach (**Fig. 11-20a**). This beach ridge, or crest, is a summer berm. When a summer storm temporarily increases wave energy on a steep summer beach, the beach below the berm can be cut away temporarily to form a scarp (**Fig. 11-15c,d**). In most cases, the scarp dries and is flattened by **slumping** and winds, or is destroyed by waves of subsequent storms.

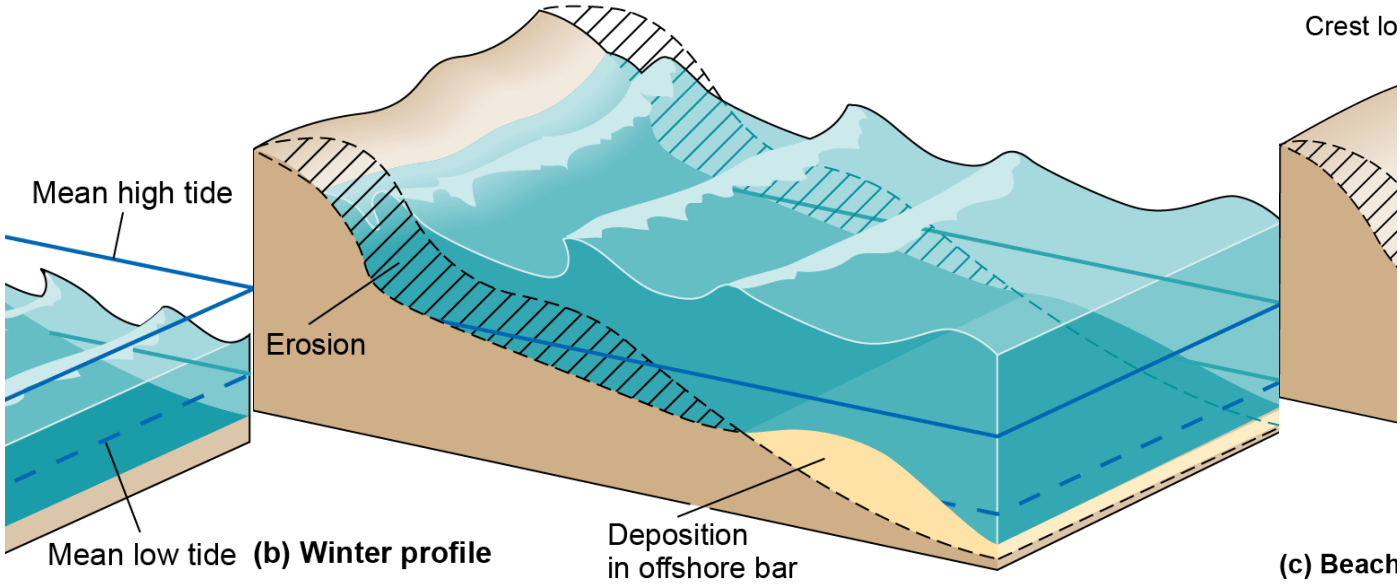


Figure 11-20. Seasonal variation in beach profile. (a) During summer, when waves are relatively gentle, the beach profile is steep, and one or more summer berms form immediately above the high-tide line. (b) Winter or storm waves, especially when combined with storm surges, reach farther up the beach and erode sand from the berm, moving the sand seaward to form one or more offshore bars. The result is a narrower flat-beach foreshore that is cut back to dunes or a winter berm left from periodic extreme winter storms. (c) When storm waves cease, the gentler waves return sand from the offshore bar to the beach, rebuilding its summer profile. If storms are strong in a particular winter, the beach may recede inland, and dunes may be somewhat lowered.

Reconfiguration of the beach and winter removal of its sand to longshore bars is an important process that protects the coast from erosion by the huge waves generated by exceptionally large storms. Longshore bars are built by shorter-wavelength waves

from the less intense storms that characterize most of the winter. Therefore, they are shallow enough to cause the exceptionally huge waves to break. Wave energy is partially dissipated as the wave breaks over the longshore bar, and the weakened wave causes less damage when it hits the shore. Some of the consequences of ignoring the protective nature of longshore bars are discussed later in this chapter.

Beach Slope and Grain Size

Beach slope between high- and low-tide lines depends not only on wave size, but also on the grain size of beach materials. Pebble and cobble beaches can have slopes of 10° to 20°, whereas the finest sand beaches have slopes of 1° or less. The slope reflects the conditions needed for the equilibrium of particle transport up the beach in the swash and down the beach in the backwash.

Beach slope is greater for large grain sizes because water can percolate downward more easily through large grains than it can through smaller, more closely packed grains. This process reduces the amount of water in the backwash and thus its ability to carry particles seaward. Consequently, sediment moves onshore and builds the beach slope. The slope increases until the backwash is strong enough to move particles down the beach as fast as they are moved up.

This page titled 11.2: Beaches is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.3: Barrier Islands and Lagoons

Many U.S. coasts, including large stretches of coast along the Atlantic Ocean, Gulf of Mexico, and Beaufort Sea (Alaskan Arctic) (**Figs. 11-2, 11-3**), have low, elongated sandy islands offshore from the mainland coast (**Fig 11-21**). These islands are aligned approximately parallel to the coast and are called barrier islands. Landward of the barrier island, there is usually a calm shallow **lagoon** in which sea grasses, mangroves, or marsh grasses may grow. Barrier islands and their sheltered lagoons are important because they protect the coast from erosion caused by storm waves. In addition, the sheltered lagoons provide habitat for the juveniles of many marine **species**.

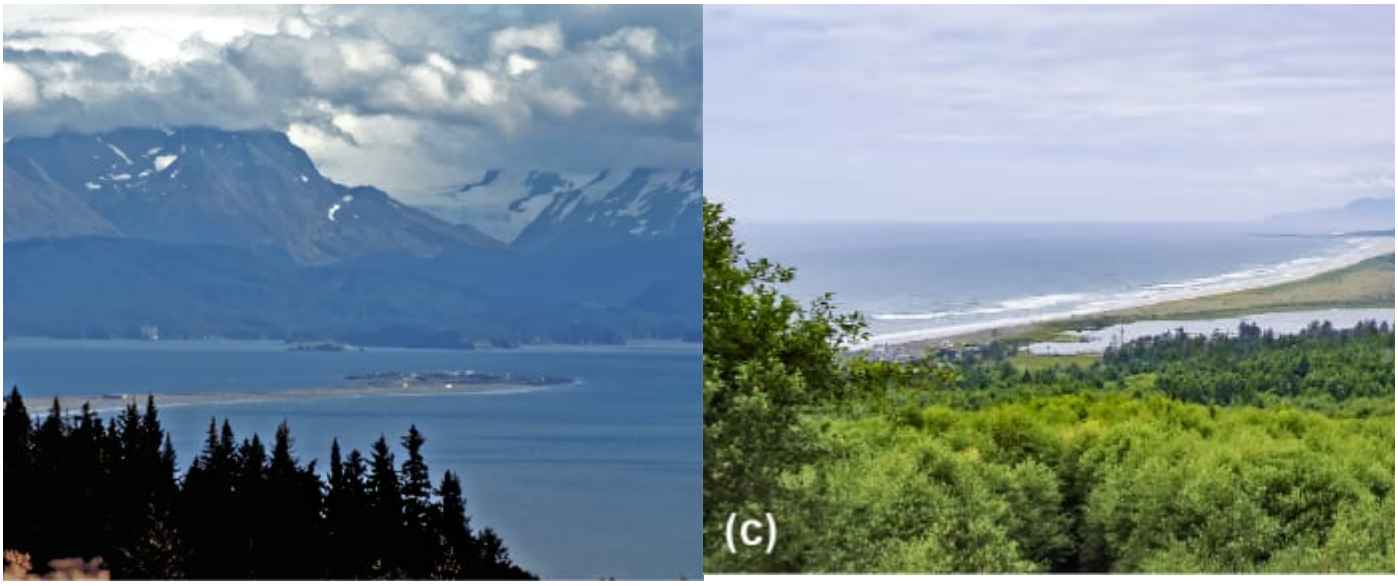


Figure 11-21. (a) Spits, baymouth bars, barrier islands, and lagoons are all formed on coastlines where erosion supplies large amounts of sand to the longshore drift. (b) Homer Spit, Alaska. (c) A baymouth bar at the mouth of Tillamook Bay, Oregon. (d) A barrier island, Assateague Island, just south of Ocean City, Maryland looking north.

Formation of Barrier Islands

The mechanisms that lead to the formation and migration of barrier islands are not fully understood. Some barrier islands may be formed from **spits** created by longshore drift, or they may be coastal sand dunes behind which rising sea level has flooded. Others may originate as a longshore bar when particularly strong storms temporarily raise the sea level and transport sand from the beach to the bar. The enlarged bar emerges when the storm has abated and sea level has dropped. Almost all possible mechanisms depend on or are aided by rising sea level.

Many or most barrier islands are thought to have been formed by erosion of the flat, sediment-filled coastal plains of passive-margin coasts as sea level rose during the past 19,000 years. As sea level rose, the soft sediments and rock of the newly flooded coastal plain were rapidly eroded, providing an abundant supply of beach sands that entered the longshore drift. As it migrated along the coast past headlands, the abundant sand accumulated downcurrent of the headlands to form spits (**Figs. 11-9, 11-21**). Spits often grow completely across the mouths of bays and inlets between two headlands to become a **baymouth bar** (**Fig. 11-21c**).

As sea level rose further, the spits and bars were breached in places, allowing the sea to inundate more of the coastal plain behind the accumulated beach sands. Where sections of beach were separated from the headlands on both sides, barrier islands were formed (**Fig. 11-21**). Once formed, barrier beaches (including islands, spits, and baymouth bars) continued to be fed with large amounts of sand from the easily eroded coast and began to retreat landward in concert with the rising sea level (**Fig. 11-22**). As sea level rose, storms piled sand onto and across the barrier islands, eroding their ocean shores but building the lagoon side by the accumulation of over-washed sand (**Fig. 11-22**).

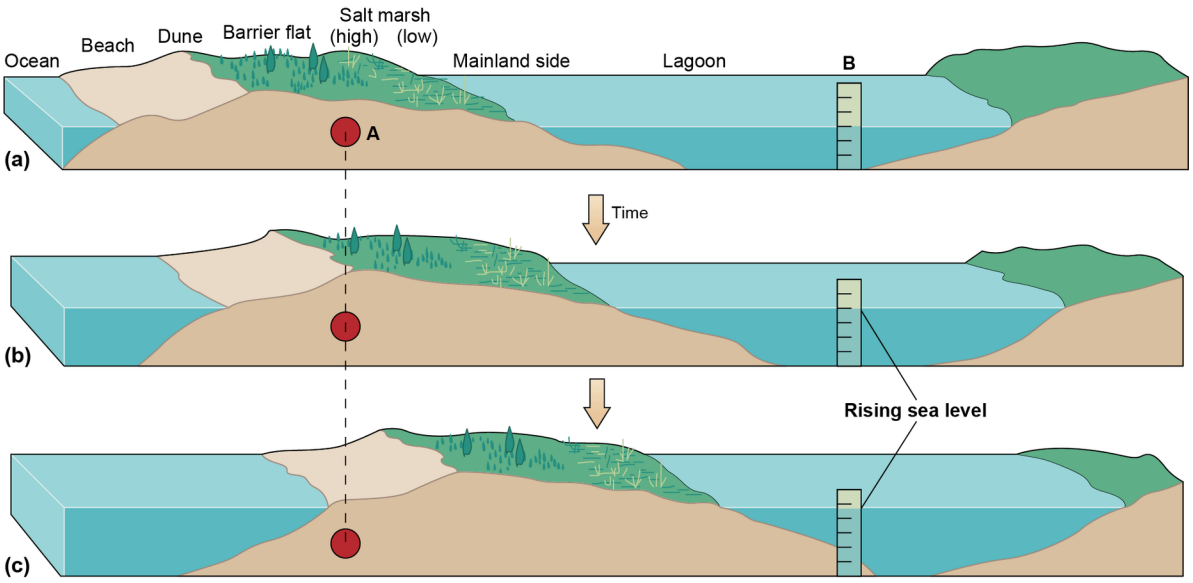


Figure 11-22. As sea level rises, barrier islands retreat toward the mainland as storm waves erode sand from the beach and transport it over the barrier island to accumulate on the landward side of the island. Note the sea-level rise shown at B, the inundation of the coast as sea level rises, and the landward progression of the island as time progresses from (a) to (c). Also note that the sand at point A remains in the same place during this process as the island migrates. This sand would eventually be eroded if sea level continued to rise and the barrier island continued to retreat. Vegetation stabilizes the barrier island dunes, but it is continuously eroded on the ocean side of the island and replaced by new growth on the advancing landward side.

As sea level rose rapidly between about 19,000 and 4000 years ago, the barrier islands retreated with the advancing sea (**Fig. 11-22**). However, barrier beaches do not retreat continuously. The abundant sand in the longshore drift system is accumulated on the seaward side of a barrier island until an exceptionally strong storm and elevated sea surface carry large amounts of the sand over the island. Consequently, as the barrier beaches retreated, they did not catch up to the retreating shoreline in most places. The rising sea continuously flooded more coastal plain through breaches in the barrier beach system, while the shoreline, lagoon, and barrier beach complex retreated onto the continent as sea level rose. The shorelines, lagoons, and barrier beach breaches underwent many changes in their configuration during this process. These changes were related to variations in the coastal-plain topography and in the distribution of storms along the coast.

About 4000 years ago, the rapid rise in sea level diminished, and sea level has risen only slowly since that time. The barrier beaches of the world have not yet responded completely to the virtual halt in sea-level rise, and many of them are still retreating. The retreat can be quite rapid. In some cases, barrier islands have retreated by a distance equal to their entire width within the past several decades.

Future of Barrier Islands

If sea level remains the same as it is today, barrier islands should retreat more slowly, but the retreat may be necessary for their continued existence. Abundant sand eroded from newly inundated coast is needed to sustain a barrier beach, and rising sea level is needed to prevent shallow lagoons behind the barrier island from slowly filling with sediment. If sea level remains stable, barrier islands may slowly disappear. If sea level falls significantly, barrier islands may no longer form. In contrast, more rapidly rising sea level, predicted as a consequence of the enhanced greenhouse effect, will probably result in a renewed cycle of active barrier island formation and retreat.

Development on Barrier Islands

Barrier islands are particularly inviting places to build houses and resorts. In addition, the lagoons behind them are well suited to be harbors for small boats if the channel to the sea between islands is maintained at an adequate depth for navigation. Unfortunately, barrier beaches are dynamic features that undergo continuous change. Each time a major storm or **hurricane** hits one of the many heavily developed barrier islands, massive damage occurs and “undesirable” sand accumulates in the lagoon and often in channels. Large sums of money must be spent to rebuild and to move sand out of the lagoon and back to the depleted beach.

Although periodic destruction by storms will continue indefinitely, it may be too technologically difficult, too costly, or too environmentally undesirable to indefinitely “restore” the barrier islands as natural processes continue to cause their landward retreat. In addition, we do not know the long-term consequences of continuous beach restoration. Will the lagoon slowly fill with sediment, and will the protected shoreline marshes continue to sustain their vital **ecological** role (**Chap. 13**)? Clearly, if sea level rises more rapidly, the task of maintaining civilization’s temporary presence on barrier islands will become increasingly difficult and expensive.

At many places on U.S. coasts, development has led to an endless cycle of efforts to arrest the natural landward retreat of barrier islands. Among the best known of the barrier islands are Miami Beach and Palm Beach in Florida, and Galveston and Corpus Christi in Texas. In Galveston, a seawall was constructed to protect houses that were rebuilt after a strong hurricane in 1900 (**Fig. 11-23**). However, during hurricane Ike in 2008, considerable areas of Galveston were flooded as ocean water flowed into the city mostly around the ends of the wall where the adjacent sections of the barrier island had already moved a considerable distance landward (**Fig. 11-23b**).

Figure 11-23. Seawalls are designed to protect beachfront property from erosion. (a) The Galveston seawall was constructed after the destructive hurricane of 1900. Since the wall was completed in 1962, the beach has eroded along much of its length, leaving the shoreline more exposed to attack by the highest waves of future hurricanes. (b) The barrier island has continued to retreat since the Galveston seawall was constructed. The beach just beyond the end of the wall has now retreated substantially farther shoreward than the wall’s location as a result of the natural process of barrier island retreat and the consequent reduction in the amount of sand in the littoral drift system that occurred once the beach had been eroded away in front of the wall. (c) Seawalls are undermined from the sides and eventually destroyed as sea level rises.

Corpus Christi has similar problems. Instead of building a seawall, the approach in Corpus Christi was to try to replenish the vanishing beach. Unfortunately, the sand used for replenishment had a larger grain size than the natural beach sand. As a result, the beach stabilized with a much steeper slope than it previously had, and thus it became more dangerous for people entering the water. In addition, beach replenishment must be repeated periodically and continued indefinitely if the beach is to be maintained. Meanwhile, the barrier island adjacent to the area where the beach is replenished continues to retreat, further increasing the exposure of Corpus Christi to ocean waves.

Both Galveston and Corpus Christi suffered flooding and considerable damage during hurricane Harvey in 2017.

This page titled 11.3: Barrier Islands and Lagoons is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.4: Beaches and Human Structures

The beachfront is a preferred location for constructing houses, condominiums, hotels, marinas, and other commercial establishments. Structures are built next to the beach despite the knowledge that beaches are constantly changing and subject to erosion and, hence, that the structures are vulnerable to damage by storm waves.

Most beaches migrate inland as coastal storm erosion transports sand offshore. The migration occurs more quickly in some places, particularly some barrier beaches, than in others. Nevertheless, coastal erosion continues inexorably on most coasts, carving away the base of coastal cliffs, removing sand by longshore drift, and pushing barrier islands back toward the mainland. In a few locations, beaches are built up, sometimes very rapidly, by coastal sand transport. Usually, however, beaches are built and extended seaward only where the land edge is uplifted by tectonic processes or where isostatic processes raise the land edge more quickly than erosion can occur. Accordingly, in most locations, maintenance of beachfront property requires a defense against the continual loss of beach due to erosion and the consequent increasing vulnerability of the property to storm waves.

When the beach in front of beachfront property loses a significant amount of sand, the traditional engineering response is to build seawalls or **groins** to restore or retain beach sand and protect the property from wave damage. These structures interfere with the normal movement of sand in the beach system and often have unintended consequences.

Seawalls

When the beach in front of buildings or highways has eroded or retreated so far that winter storm waves can reach and damage the structures, a seawall is often built to break up or reflect storm waves. The seawall is usually constructed parallel to the beach and behind the area of beach used for summer recreation (**Fig. 11-23**). It is made of large boulders, concrete, or steel, and either replaces or covers beach sand that normally would be mobile and eroded by the strongest winter storm waves. The wall usually works well for some years, but beach erosion continues on either end of the walled section. As the adjacent coastline retreats, the wall protrudes progressively farther seaward on the beach (**Fig. 11-23**) and is increasingly exposed to waves. Eventually the beach in front of the wall becomes very narrow, and the wall itself is undermined and eroded by waves. At this point, a continual cycle of rebuilding and destruction of the wall begins, and the beach essentially disappears.

A seawall has another unintended effect. Because it is built within the area of wave action, it prevents waves from eroding sand from what was previously the upper part of the beach. Consequently, the amount of sand carried offshore to form longshore bars is reduced. The seawall also prevents the beach from migrating landward and the waves from progressively eroding the coast to provide new sand. Thus, it reduces the amount of sand available to replace the sand lost through longshore drift. Because less sand is available, the beach becomes narrower and longshore bars are further depleted. The narrower beach and smaller longshore bars absorb less wave energy, increasing the wave energy reaching and eroding the seawall, even though storms are no more intense.

Groins and Jetties

Beaches progressively lose sand and become narrower when structures prevent the normal erosional retreat of the coastline or when the sand supply to the longshore drift system is reduced by other means, such as the damming of rivers. When the beach erodes and narrows in front of a valuable piece of real estate, groins are often built to try to restore it. A groin is a wall built perpendicular to the beach from the backshore out to beyond the **surf zone** (**Fig. 11-24a**). The groin's purpose is to block the longshore drift so that sand accumulates on the upcurrent side of the groin, widening the beach at that point. The problem with groins is that they further deplete sand supply to the beach on their downcurrent side, where severe erosion may ensue. A common solution to this problem is to build another groin downcurrent from the first. Eventually a long series of groins may be built along the entire length of the beach (**Fig. 11-24b**). The net effect is to alter the beach so that wave erosion is enhanced on the downcurrent sides of the groins while segments of "normal"-width beach are maintained on the upcurrent sides (**Fig. 11-24a**). In many areas, groins are a temporary solution at best because they do not stop shoreline retreat, but merely redistribute a declining sand supply.

Figure 11-24. Groins designed to stop beach erosion modify the longshore sand transport and, consequently, the beach. (a) Sand accumulates on the upcurrent (longshore drift current) side of the groin and is eroded from the downcurrent side. (b) Often a line of groins is constructed along the length of a beach, which creates a saw-toothed beach shoreline like this one at Cape May, New Jersey. The groins do slow the recession of the barrier island but will not provide effective protection as sea level rises. (c) You can see that the barrier island continues to recede beyond the location where the line of groins ends in this image of Beach Haven, New Jersey where the barrier island has already retreated in the undeveloped and unprotected area just south of the line of groins. Within perhaps fifty years, all the structures that you see on this barrier island will be gone or will require massive seawalls entirely around the perimeter of the island to protect them.

Other structures built on the coast also cause problems by interfering with the longshore drift system. For example, rock **jetties** are built to protect inlets between barrier beaches (Fig. 11-25). Such inlets are dynamic features that normally appear, disappear, and move up and down the coast as sand is moved in the longshore drift system. Jetties stabilize an inlet by blocking waves that otherwise would erode its banks and by blocking longshore drift into the inlet that would cause **shoaling**. Beaches often become wider on the upcurrent sides of inlet jetties. In contrast, downcurrent beaches can be eroded and severely depleted of sand, and the inlet behind the island can be silted up as the island retreats inland.

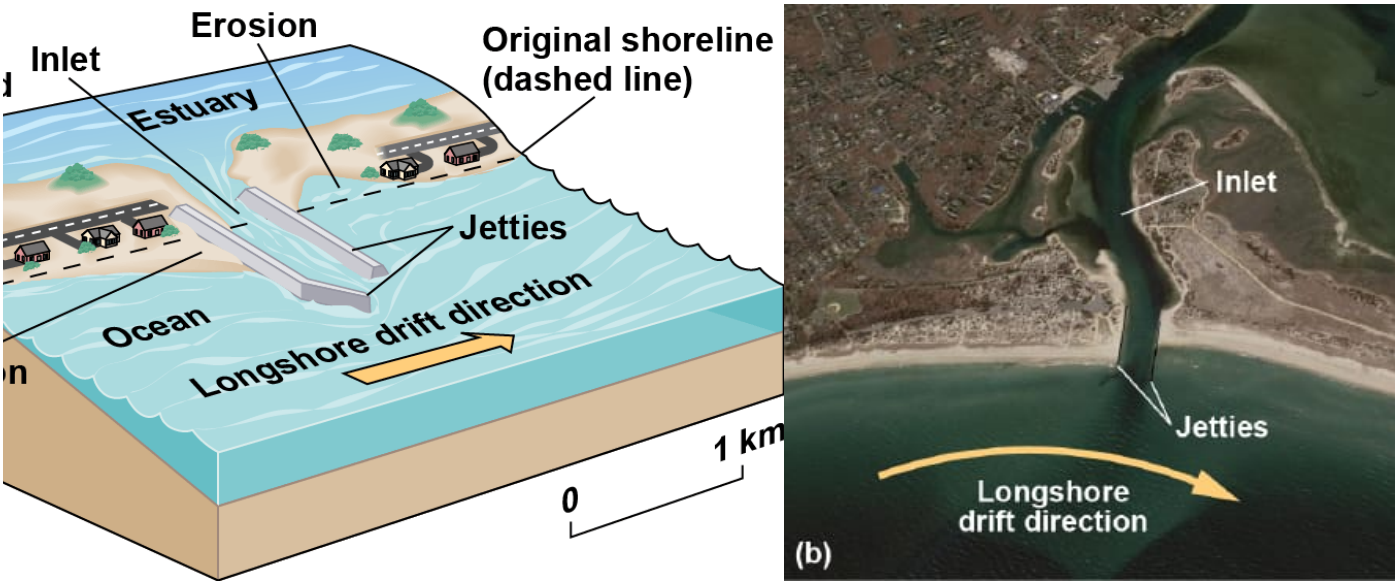


Figure 11-25. Jetties. (a) Rock jetties are often constructed on either side of an inlet between barrier islands to prevent silting of navigation channels and erosion of the sides of the inlet by waves from boats' wakes. These jetties obstruct the longshore sand transport, leading to the accumulation of sand and widening of the beach on the upcurrent side, and erosion and loss of the beach on the downstream side. (b) Three Mile Harbor, East Hampton, Long Island, New York.

Harbors

On coasts with few natural inlets, jetties or breakwaters are commonly built to provide a safe harbor for small boats. Two such harbor designs are shown in **Figures 11-26** and **Figure 11-27**. The dogleg jetty on the upcurrent side of the Santa Barbara harbor (**Fig. 11-26**) interrupts the longshore drift so that it carries sand around the jetty but not across the deep harbor entrance, which is too deep for waves to sustain the sand movement. Sand flows around the jetty and accumulates in the harbor, making parts of the harbor too shallow for navigation. The beach that is downcurrent of the harbor is depleted of sand and threatened by severe erosion. These problems are addressed by a very expensive (millions of dollars per year), energy-intensive annual dredging project. Sand is normally dredged from the harbor, pumped through a pipeline, and discharged downcurrent, where it re-joins the longshore drift.

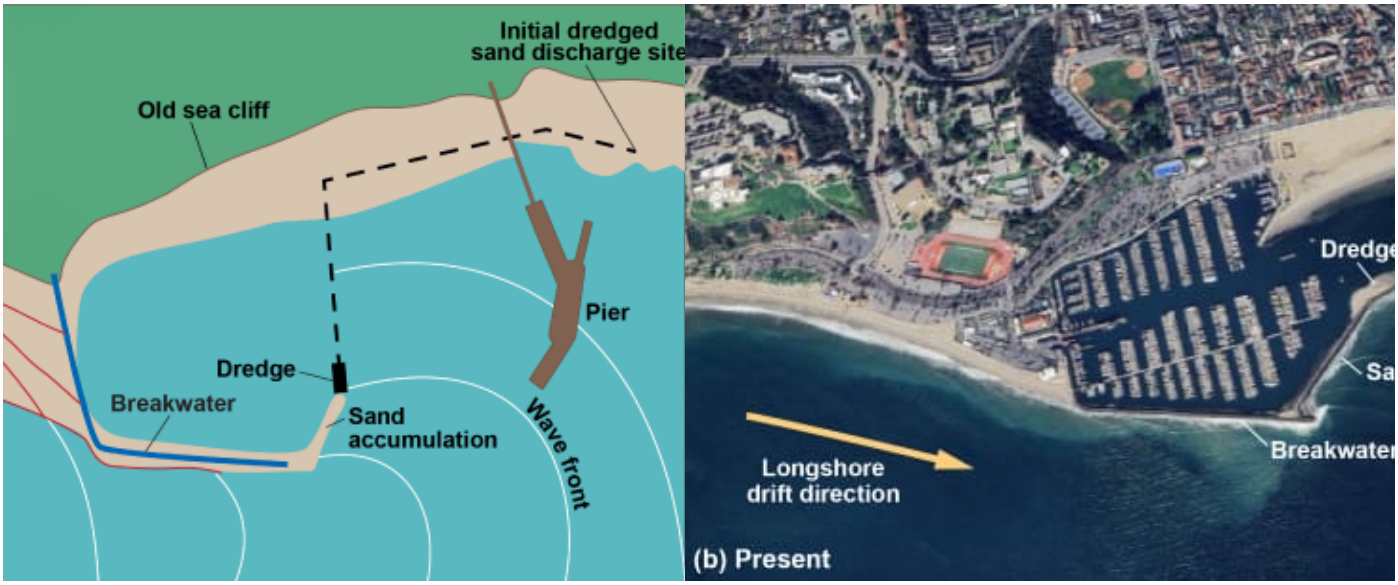


Figure 11-26. The jetties of the Santa Barbara harbor breakwater have interfered with the longshore drift transport system. (a) When the breakwater was first built, sand accumulated and progressively widened the beach on the north side of the main jetty (the direction from which waves most often come, the left in this photo) until the sand began to be transported south again. However, sand transported around this breakwater cannot be transported across the deep harbor mouth and accumulates in a spit inside the harbor mouth. The beach on the south side of the harbor has had its supply of sand from longshore drift cut off. Periodic but continuous dredging is needed to keep the harbor from silting up. The dredged sand is deposited on the south-side beach to prevent this beach from eroding further. (b) Since the breakwater was first constructed this satellite image shows how the deeper channel in the harbor mouth provides a break in the wave pattern that interrupts the longshore drift. The sand spit that developed has now been armoured with rocks, and dredging continues from the area at the end of the spit where sand still accumulates. This sand is still dredged annually, pumped through a pipeline, and discharged on the beach farther down the coast than the original discharge location. Note the sports stadium, roads, parking lots and other buildings now built on the area of sand accumulation on the left of the breakwater that was previously ocean where longshore drift carried sand accumulated to extend the beach within the first few years after the breakwater was built.

Figure 11-27. (a) Some boat harbors consist of a breakwater or jetty built parallel to the shore. The breakwater protects boats but also interferes with the longshore drift by blocking the waves from reaching the beach behind the breakwater. (b) In this image the submerged remains of the Santa Monica harbor wall (the dark line almost parallel to the shore) are parallel to the beach line. When it was first built, before it sank, the wall blocked enough wave energy to allow sand to build up on the beach under the pier so the harbor became too shallow and unusable. The wall has now sunk deep enough that its effect on the longshore drift is much reduced The pier is elevated on pilings, which allows the wave energy to pass through. As a result, the pier does not block the longshore drift. (c) Offshore breakwaters are used not only to create harbors but also to control coastal erosion as in this satellite image of an area near Norfolk, Virginia. The built up spits of sand behind each of the segments of the offshore breakwater are caused by the interruption of longshore drift. This type of segmented breakwater temporarily slows overall erosion but many would consider that it also makes the beach unnatural and unsightly.

In the 1930s, in Santa Monica, California a harbor was created by constructing a detached breakwater offshore from the end of the Santa Monica Pier. The breakwater was a simple structure built parallel to the beach beyond the surf zone, and was designed to shelter boats. Because this breakwater did not encroach on the surf zone, it seemingly should not have interfered with the long-shore drift. Unfortunately, this was not the case, because this type of offshore breakwater prevents waves from reaching a segment of the beach behind it. Since only low-energy waves break on this segment, the longshore drift is reduced just as effectively as it would be by a groin passing through the surf zone. Soon after the Santa Barbara harbor was constructed, the beach behind the breakwater became greatly enlarged, reducing the area of safe anchorage for boats. The only solution would have been an expensive dredging program. However, the breakwater was constructed of rock piles on a sandy seafloor and quickly sank into the seabed. You can see the submerged structure in (**Fig. 11-27b**). Multiple segmented detached breakwaters are now used along many coastlines like this one in Norfolk, Virginia and especially in Europe to deliberately slow down the longshore drift and attempt to maintain sand on the beach (**Fig. 11-27c**). As with other structures that attempt to slow down loss of beach sand these segmented breakwaters are, at best, a temporary solution.

This page titled 11.4: Beaches and Human Structures is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.5: Coral Reefs and Atolls

Reef-building corals are **communities** of microscopic animals that have photosynthetic **dinoflagellates**, called **zooxanthellae**, living symbiotically within their tissues. These reef-building corals, unlike many other corals, cannot grow successfully without the zooxanthellae, or where there is not enough light for the zooxanthellae to photosynthesize. Reef-building corals also require warm water (≥18°C) and are intolerant of low **salinity** and high concentrations of suspended sediments that block sunlight. Hence, coral reefs grow only in shallow tropical and subtropical waters that are not subject to low salinity and high **turbidity** caused by stream **runoff** or abundant **plankton**.

The three basic types of coral reefs are fringing reefs, barrier reefs, and **atolls** (**Fig. 4-28**). Fringing reefs grow in shallow waters along the shore and are best developed off coasts with arid climates and where river runoff is limited. Fringing reefs are distributed in a patchy fashion along most coasts where they occur because they are easily damaged by the low salinity of excess freshwater runoff and by high concentrations of suspended sediment in the runoff or resuspended by storm waves. Human activities such as flood-control projects, dredging, sewage discharge, and coastal modification often result in elevated concentrations of suspended sediments and enhanced runoff of low-salinity water during storms. Consequently, fringing coral reefs are damaged by these activities in many locations, but it is often difficult to distinguish damage due to human influence from natural variations.

Barrier reefs and atolls are formed only where there has been a change of sea level on the adjacent coast. If sea level rises or the coast subsides in an area that has a fringing reef, the reef grows upward. The upward growth is faster at some distance offshore where conditions are optimal. In the shallowest water close to the shoreline, wave resuspension of sediments and the effects of runoff inhibit coral growth. In deeper offshore water, reduced light levels inhibit photosynthesis by zooxanthellae and thus coral growth. Therefore, the optimal zone for coral growth is a strip of ocean some distance offshore, but not extending into deep water. The width of this zone varies with factors such as the seafloor slope and the amount of runoff. As the land subsides, the reef grows upward in this zone faster than coral can grow in either shallower or deeper water, and a barrier reef is formed (**Fig. 4-29b**). Once formed, the barrier reef continues to grow upward. The rate of upward growth is usually just sufficient to match the changing sea surface level. The top of the reef remains just below the water surface, because corals are damaged if exposed to air for prolonged periods. If the maximum possible rate of upward growth is less than the rate at which the land subsides, the reef is submerged and eventually dies.

Once a barrier reef is formed, it closes or partially closes a lagoon between the reef and the coastline. Corals grow in this lagoon, but they are limited by higher turbidity due to land runoff, by salinity variations, or by lack of **nutrients** as a result of the long **residence time** of the lagoon water (**Chap. 12**, **CC8**).

Barrier reefs are especially abundant on the Earth at present because sea level has been rising eustatically for approximately the past 19,000 years. The Great Barrier Reef of Australia, which is about 150 km wide in places and more than 2000 km long, is the best-known barrier reef.

Islands formed by volcanoes on oceanic **crust** are particularly good locations for the development of barrier reefs. Once a volcanic island has moved away from the oceanic ridge or hot spot on which it was formed, it cools and sinks isostatically (**Chap. 4**, **CC2**). Hence, even in times of falling sea level, barrier reefs can form around such islands, as long as the island sinks faster than the sea level falls. Sinking of volcanic islands leads to the formation of the third type of coral reef, the atoll (**Figs. 4-29c, 4-30**). As the island sinks, a barrier reef develops from the fringing reef and continues to grow upward, remaining at or close to the sea surface as the island continues to sink. Eventually the island sinks entirely below the ocean surface, and its reef is left in the form of an atoll. The atoll continues to grow and maintain itself as the island sinks if the coral growth rate exceeds the combined sinking rate of the island plus the rate of any sea level rise.

Low islands may form on an atoll, particularly on the side of the atoll from which wind and waves most often approach. Such islands are composed of debris from the reef and of calcareous algae that grow in very shallow water and can survive exposure at low tide. Islands can also be formed on an atoll by earthquakes that uplift or tilt the sinking volcano.

This page titled 11.5: Coral Reefs and Atolls is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.6: Wetlands

On many shores protected from wave action, tidal wetlands develop. Tidal wetlands are flat, muddy areas covered by water during only part of the tidal cycle. Most tidal wetlands are characterized by an abundance of emergent plants: marsh grasses in salt marshes, and mangroves in mangrove swamps. Both plant types grow well in muds rich in organic matter that are inundated periodically by salt water. Grasses grow in salt marshes at all latitudes, but mangroves grow only in tropical and subtropical latitudes between about 30°N and 30°S. Where both are present, mangroves usually rapidly outgrow and eliminate marsh grasses.

Wetlands are created by the accumulation of sediments from a variety of sources. The most important are usually river-borne muds laden with organic **detritus**. Sediments accumulate in shallow protected waters, particularly at the edges of estuaries and on lagoon shores, until they extend above the low-tide line and a wetland forms.

A tidal wetland consists of a muddy, flat terrace between the low- and high-tide lines that is dissected by shallow channels. As the tide rises, water moves through the channels and spreads over the mudflats. The current speed within the channels is faster than the water's speed as it spreads over the flats. As the speed decreases over the mudflats, suspended particles carried up the channels are deposited (**CC4**) as wetland mud. Particles of organic detritus have low **density** and can be fragmented to very small sizes. Consequently, wetlands tend to accumulate muds that are rich in organic matter, with much of the detritus contributed by the local marsh grasses or mangroves.

The abundant supply of detritus makes wetlands very attractive places for marine animals to feed and for the juvenile stages of many marine species to spend part of their life cycle before migrating to the sea. Wetlands provide not only abundant food, but also safety from predators. Large marine predators are not able to enter these shallow areas, and the grasses, mangrove roots, and abundant leaf litter offer many excellent places to hide from other predators. Wetlands have such an abundance of vegetation and juvenile marine animals that many birds use wetlands as their principal, or only, feeding areas.

Wetlands are extremely valuable ecologically, but generally they are not considered to have great scenic beauty. Therefore, because their proximity to the water makes them very attractive for development, vast areas of wetlands in the U.S. and elsewhere have been filled or drained. The result has been the destruction of habitat that is important for many commercially valuable fishes and **shellfish** species and a number of species of ducks and other birds. Filled wetlands are not ideal places to build or farm because they are low-lying and susceptible to flooding and to storm surges or storm waves. Structures built on former wetlands also are susceptible to subsidence as the former wetland sediments upon which they are built are compacted.

Even more importantly, structures built on filled wetlands are very vulnerable to earthquake damage. The sediments underlying structures in filled wetlands, if not properly compacted during the construction process, can be "liquefied" by the energy of earthquake waves, allowing the structures to collapse and sink. This is one of the reasons that the Loma Prieta earthquake of 1989 caused far more damage to buildings and freeways in Oakland and San Francisco, about 100 km north of the earthquake epicenter, than it did in urban areas closer to the epicenter. Most of the badly damaged or destroyed structures in Oakland and San Francisco were constructed in areas that were formerly wetlands.

This page titled 11.6: Wetlands is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.7: Deltas

Rivers flowing toward the sea carry particles as suspended sediment. When they reach the flat coastal plain, river current speed decreases because of diminishing slope and widening of the channel. Consequently, the larger size fractions of suspended sediment are deposited (**CC4**) before reaching the ocean, slowly filling the river valley. Many rivers flow through valleys cut by glaciers or rivers when sea level was lower during the most recent **ice age**, and the lower ends of these valleys have not yet filled with sediment. Because most rivers on the East Coast of the U.S. are in this category, they deliver little sediment to the Atlantic Ocean. Some rivers, such as the Mississippi, create a delta as they flow into the ocean because they carry a large amount of sediment and slow down abruptly when they flow into the ocean. These sediments build up over time creating a roughly triangular delta at the river's mouth as the river flows in distributary channels through the deposited sediment. In some places, such as the Mississippi Delta, the channels pass through long extensions or lobes of these deposits (**Fig. 11-5**).

Deltas are built and supplied with sediment by the same basic processes that occur in wetlands. During normal river flow, relatively small quantities of suspended sediment are transported down the river channels and deposited in the channel or carried out to sea. When the river floods, it picks up large quantities of suspended sediments and can overflow its banks and inundate the flat land (flood plain) around it where it is slowed and deposits some of this suspended sediment including fine-grained, organic-rich particles. Deltas form when the river valley is filled by sediments at its lower end so that the river's flood plain extends to the ocean. As the flood plain is created, the river cuts a narrower channel, or channels, and its speed increases accordingly. The faster-moving river then transports larger particles to the oceans where, since the current speed decreases, larger grained suspended sediments are deposited to form the base of the delta. The delta is built seaward as larger grained sediment is deposited along the edges of the narrow river mouth where river speed decreases as it enters the ocean. Once delta channels are formed, flooding deposits finer-grained sediment to form wetlands across the delta. Sediment distribution across the delta is aided by the river occasionally abandoning of channels as they fill up with sediment and creating new ones (**Fig. 11-5**). In this way, surface soils of the entire delta are periodically enriched by the addition of organic matter and nutrients from the river.

In some deltas, the deposition rate has been high enough that large quantities of organic-rich muds have been continuously buried by newly deposited sediments over long periods of geological time. These buried sediments are compacted by the overlying layers, water is squeezed out, and in many such areas, oil and gas deposits have been formed as the organic matter has decomposed. Many oil and gas deposits are located in deltas or offshore in areas where deltas existed when sea level was lower. Major deltas of the world include the Mississippi Delta, the San Francisco Bay Delta, the Copper River Delta in Alaska, the Mackenzie Delta in northwest Canada, the Nile Delta in Egypt, the Niger Delta in Nigeria, and the Ganges–Brahmaputra Delta in Bangladesh and India.

Muddy delta soils are slowly compacted as water is squeezed out of them by the weight of overlying sediment. This process can be accelerated by the weight of structures built on the delta. In addition, the organic matter in the delta soils decomposes if it is not buried by additional river-borne sediments, and the soils themselves are further eroded by winds and rainfall. Therefore, periodic influxes of new sediment are needed to maintain the delta land surface slightly above sea level.

Because they are particularly rich in nutrients and organic matter, delta soils are among the best for agriculture. Consequently, many deltas, such as the Mississippi Delta and Nile Delta, are populated and intensively farmed. Unfortunately, deltas are also very easily flooded. Almost the entire population of Bangladesh lives on a delta that is regularly inundated by swollen rivers and hurricane storm surges (**Chap. 9**), often with great loss of life.

To control the frequent flooding of two major deltas in the U.S.—the Mississippi Delta and the San Francisco Bay Delta—extensive engineering projects have been carried out in which banks called **levees** have been built along each side of the river channels. Levees have worked extremely well by some measures because they have dramatically reduced flooding, but they also have unintended consequences. Without the periodic influx of nutrient-rich suspended sediment to the delta during floods, the fertility of the soils steadily declines. In addition, without the periodic inputs of new soil, the land slowly sinks as soil compacts and erodes. Sinking of the land surface is actually accelerated by agriculture through increased soil erosion and decomposition of soil organic matter. Worse still, withdrawal of freshwater from aquifers beneath the delta further accelerates sinking, particularly in the Mississippi Delta.

As the land sinks in the Mississippi and San Francisco Bay Deltas, more and higher levees are needed to prevent flooding. In the San Francisco Bay Delta, most of the land is now meters below the normal river water level. Any breach of the levee floods the entire area, even when the river is not swollen by flood rainwater. For example, a 100-m-wide breach occurred in a levee in the San Francisco Bay Delta in the summer of 2004. The breach flooded nearly 50 km^2 of farmland, destroying crops, homes,

and other structures. It took about 200,000 tonnes of rock to seal the levee breach. The total cost of the crops and structures destroyed, repairing the levee, and pumping water out of the flooded land was nearly $100 million. Significant as this event was, it was just a hint of the devastation, chaos, and loss of life and property that was caused by the breaks in the protective levees of New Orleans that flooded most of that city to a depth of 10 m or more in 2005 following the passage of Hurricane Katrina.

Higher and stronger levees could be built, but they would be very expensive and difficult to construct because they would need to be capable of restraining increasing **hydrostatic pressure** as the delta floor sinks and sea level rises. In addition, levees constrain the river to flow very swiftly through a narrow channel, so erosion of existing levees is a major problem and would remain so even if levees were enhanced.

What is the future of the Mississippi and San Francisco Bay Deltas, particularly if global climate change causes sea level to rise more quickly? If we continue on our present course, these delta lands will continue to sink, and levees will continue to be built higher until holding back the water becomes impossible. Meanwhile, soils will become impoverished and the land's agricultural value will decline. One alternative is to tear down the levees. Within a few years or decades, the deltas would return to their normal functioning, and valuable agricultural lands would again be available for the future. If we are to retain the highly productive deltas, the price we must pay is periodic flooding.

In the Mississippi Delta and the San Francisco Bay Delta, restoration programs have begun to establish breaches in the levees to replenish delta sediments in key areas, converting these areas back to wetlands. So far, on the Mississippi Delta, the program has focused mainly on areas that had become shallow coastal lagoons as a result of the levee diversions of sediments. Strong opposition to the program from fishers who used the lagoons to collect oysters had to be overcome. Even greater opposition is, and will continue to be, encountered when the proposed replenishments involve land that has been drained for agriculture. It remains to be seen whether these replenishment programs will ultimately succeed and can be applied to other deltas.

This page titled 11.7: Deltas is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.8: CHAPTER SUMMARY

Processes That Form and Modify the Coastline.

Coasts can be erosional or depositional. Primary coasts are formed by volcanoes, earthquake movements, landslides, deposition of river-borne sediment, and sea-level change. Sea level has risen about 130 m in the past 19,000 years and continues to rise. Secondary coasts are formed by erosion or deposition by waves, tidal and ocean currents, and the formation of coral reefs.

Beaches.

The littoral zone extends between the seaward boundary of land vegetation and the depth where wave action does not disturb the sediments (about 10 to 20 m). Beaches are part of the littoral zone and consist of a back-shore, foreshore, and offshore. The backshore is normally dry at high tide and may be characterized by berms created by storm waves. The foreshore is the gently sloping area between the high- and low-tide lines, where scarps may form. The offshore extends seaward of the low-tide line and may have a longshore bar near its seaward boundary.

Beaches are a mixture of grains of varied composition supplied by erosion or by longshore drift. Longshore drift, caused by the incomplete refraction of waves, transports sand downcoast of the direction from which the waves approach. Sand is moved along the coast until it is transported down a submarine canyon. Some beaches may be in danger of losing sand because the supply of river-borne sand to the longshore drift system has been reduced by dams and coastal structures designed to reduce wave erosion.

Particles that make up a beach are sorted by waves. Small particles are removed to deep water. Large particles are not moved by longshore drift and are present only if they are formed locally by the erosion of cliffs next to the beach. Grain size is larger on beaches with greater wave action. Quiet beaches are composed of fine-grained sand, whereas those with greater wave action are composed of coarse-grained sand or gravel.

Large winter waves move sand from the beach offshore to build a longshore bar. In summer, gentler waves return sand to the beach. In winter, a beach is narrower with a flat foreshore and steep backshore. In summer, the beach is broader, the foreshore is steeper and a summer berm or berms are formed. Beach slope increases as grain size of the beach materials increases.

Barrier Islands and Lagoons.

Barrier islands protect the coast from wave erosion and enclose lagoons that are nursery areas for many marine species. They are formed and maintained by sand that is moved by longshore drift. Barrier island–lagoon systems are retreating landward as sea level rises. Development on barrier islands presents problems because of this landward migration. Dredging of the lagoons and beach replenishment are necessary to reverse migration that would destroy developed structures on the islands, but these measures cannot be successful indefinitely. Developments on barrier islands repeatedly sustain expensive damage when major storms occur.

Beaches and Human Structures.

As beaches migrate inland or lose sand because river sources are reduced, seawalls and groins are built to retain the sand or protect coastal structures. Seawalls temporarily protect coastal structures from waves, but eventually they are undermined and destroyed as the shoreline recedes. Seawalls also reduce erosion that would otherwise supply sand to the longshore drift system, and consequently, they reduce beaches and longshore bars that would absorb wave energy. Groins are built across the littoral zone to retain sand on beaches by blocking longshore drift. Sand builds up on the upcurrent side of the groin and is lost from the downcurrent side. Groins are often only temporarily effective and can cause a loss of sand on down-current beaches that leads to the building of more groins. Jetties or breakwaters protect harbors from wave action but interfere with longshore drift and often cause sediment to accumulate in the harbor, making continuous dredging necessary.

Coral Reefs and Atolls.

Reef-building corals grow only in warm water shallow enough that their zooxanthellae can photosynthesize. They grow best in waters with lower turbidity and little salinity variation. Fringing coral reefs grow in shallow water offshore where erosion and freshwater inputs are low. Barrier reefs and atolls form where coral growth is faster some distance offshore than immediately along the shore, and where sea level rises or the coast subsides. Most atolls are formed from barrier reefs as volcanic islands sink isostatically below the sea surface.

Wetlands.

Tidal wetlands are flat, muddy areas covered by water part of the time. They retain detritus from river flow and vegetation that grows in them. The detritus provides food, and the sea grasses or mangroves and shallow water provide shelter for juveniles of many marine species. Large areas of coastal wetlands have been filled for development and agriculture.

Deltas.

Deltas are formed as sediments accumulate at the mouth of a river. They are flat areas maintained at an elevation slightly above sea level by periodic river floods that bring sediment, detritus, and nutrient-containing particles to be deposited on the flooded delta. They are good agricultural land because of these periodic inputs, but they are often flooded. Although levees have been built to protect many deltas from flooding, these structures cut off the supply of new sediment, so the land slowly subsides, is eroded away, and loses its nutrients and detritus.

11.8.1: Key Terms

11.8.2: Chapter Questions

11.8.3: References

This page titled 11.8: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.8.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- acidification
- atolls
- backshore
- backwash
- barrier islands
- basaltic
- baymouth bar
- calcareous algae
- climate
- coast
- coastal plain
- coastal zone
- coastline
- communities
- continental divergent plate boundaries
- continental shelf
- convergent plate boundaries
- coral reef
- crest
- crust
- currents
- deltas
- density
- depositional
- detritus
- dinofalgellates
- echinoderms
- ecological
- Ekman transport
- environment
- erosional
- eustatically
- fjords
- flooding
- foraminifera
- foreshore
- frequency
- friction
- grain size
- gravity
- greenhouse effect
- groins
- high-tide lines
- hot spots
- hurricane
- hydrostatic pressure
- ice age
- isostatically
- jetties
- lagoon

- latitude
- levees
- lithospheric plates
- littoral drift

- littoral zone
- longshore bar
- longshore current
- longshore drift
- longshore drift
- low-tide lines
- low-tide terrace
- magmatic arc
- mangrove
- mollusks
- nutrients
- oceanic ridges
- orbital velocity
- passive margincorals
- photosynthesis
- plankton
- relict sediments
- residence time
- resuspends
- runoff
- salinity
- salt marshes
- sand dunes
- scarp
- sea grasses
- sedimentary arc
- shellfish
- shoaling
- shoreline
- sill
- slumping
- sorted
- species
- spits
- spreading cycle
- spring tide
- subducted
- submarine canyon
- surf
- surf zone
- swash
- tidal range
- tides
- topographic
- transform faults
- trough
- tsunami
- turbidity
- wave heights
- wave period
- wavelength
- weathered
- zooxanthellae

This page titled 11.8.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.8.2: Chapter Questions

Study Questions

1. Describe the processes that modify a coastline once it has been formed. How and why are the coastlines of passive and active margins different?
2. Describe the features of a beach at the end of a summer in which no large storm waves arrived at the beach. How would these features change if a storm with very large waves hit the shoreline?
3. Why are the particles of sand on most beaches limited to a narrow range of sizes? Why can the size ranges be different on two beaches located near each other?
4. Where does beach sand come from, and how is sand moved along a coastline?
5. It has been proposed that the energy of waves can be harnessed by various devices placed seaward of the surf line. If many such devices extended along a coastline and they collected a substantial fraction of the incoming wave energy, how would the beaches be changed?
6. Why is replacing sand lost from a barrier island beach or building a seawall to protect beachfront property only temporary solutions for beach erosion?
7. In many indented bays there is a wide, flat, sandy beach at the bay's center, but the beach disappears near each headland. Another sandy beach is usually found in the next bay. Describe the processes responsible for this distribution of beach sand.
8. Why don't fringing coral reefs grow on all tropical coastlines? What is the relationship between rising sea level and the number and distribution of atolls and barrier reefs in the oceans?
9. Describe the differences between fringing coral reefs, barrier reefs, and atolls. The island of Hawaii has mostly fringing reefs. Why?
10. What are the important physical, biological, and geological features of a wetland? Many wetlands have had roads built across them, either on gravel or dirt roadbeds or on elevated concrete pilings. Which type of road is environmentally preferred and why? What could be done to lessen the impact of the lesser-preferred option, short of rebuilding it?

Critical Thinking Questions

1. In the Hawaiian Islands the youngest island, Hawaii, is the biggest, has sandy beaches along the smallest fraction of its coastline, and has more black sand beaches than the other islands. Oahu, which is older than Hawaii, is smaller, has a greater fraction of its coastline occupied by beaches, and does not have black sand beaches. Kauai, which is older than both Oahu and Hawaii, is smaller than either of these other islands, has the largest fraction of its coastline occupied by beaches, and also does not have black sand beaches. Explain why this progression of island characteristics with age occurs. (These processes are described individually in several different chapters of this text.)
2. Sea level may soon begin to rise faster than has ever occurred in the history of the Earth. Hypothesize the possible effects of such a rapid sea-level rise on the number and health of coral reefs in the world's oceans and on the geographic extent and distribution of wetlands.
3. Describe and explain what happens to deltas when humans build levees to prevent frequent flooding. Many lives and huge expenditures needed to fix property damage have been saved by the construction of levees on the Mississippi and San Francisco Bay Deltas. Deltas in other parts of the world, such as the delta on which most of the population of Bangladesh lives and farms, are still regularly flooded by hurricane storm surges or intense monsoon rains, often with great loss of life and damage to property.
   (a) Discuss what you think should be done to manage the Mississippi River, San Francisco Bay, and Bangladesh Deltas in the future.
   (b) What information should be considered in making such decisions?
   (c) How do you think these decisions should be made and by whom?
4. Describe barrier islands and how they form and move. Explain why rising sea level is important for barrier island formation, and discuss what would happen to barrier islands if sea level were to fall.
   (a) Should these be rebuilt?
   (b) Many of the owners of such homes have no flood insurance. Should federal disaster funds be used to help them rebuild their homes?
   (c) If it is decided that these homes should not be rebuilt, should the owners be compensated? If so, how and by whom?
5. Imagine that you are a member of the U.S. Congress representing a district that includes a barrier island community, such as Miami Beach or Galveston, that has just been hit by a hurricane causing such extensive property damage that many homes

and hotels near the beach must be torn down. The existing seawall is clearly too low to fully protect the community from future hurricanes. You are aware that emissions of greenhouse gases during the past century will almost certainly result in climate warming, more rapidly rising sea level, and increased intensity and frequency of hurricanes. You are also aware that any reduction in the current rate of greenhouse gas emissions will be very expensive and probably detrimental to the economy. Furthermore, it will only slow, not stop, the warming trend, and it will have no effect unless other nations also reduce their use of fossil fuels, including those underdeveloped nations that need to increase such use if they are to improve their living standard. The national budget deficit is getting worse, and money for social programs is being reduced.

(a) On the basis of what you have learned about barrier islands, the effects of sea-level rise, and the current state of knowledge of the enhanced greenhouse effect, what would you do?

(b) Discuss your reasons for choosing the action or course of actions that you have chosen.

(c) Would you feel differently if you owned a home on the barrier island in question? If so, how and why?

(d) Would you feel differently if you represented a congressional district in Colorado? If so, how and why?

6. Explain what happens to a beach if a solid jetty is built out through the surf zone. Describe how the situation might be different if the jetty were built on widely spaced pilings.
7. What would happen to the beach if a solid jetty were built at an angle of 45° instead of at right angles to the beachfront? Explain your answer.
8. Why are there few deltas in the United States? In the next few million years, where would you expect to see new deltas form in the United States? Explain why.

This page titled 11.8.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

11.8.3: References

Figure 11-1 – Segar, D., and E. Stamman Segar, 1997: Point Lobos State Park just south of Monterey, California.

Figure 11-2a – Stanek, A., 2021: Scenic view of the Beaufort Sea, Alaska. USGS, Source.

Figure 11-2b – Segar, D., and E. Stamman Segar, 1997: Cliff and beach terrace, Olympic National Park, Washington.

Figure 11-2c – Segar, D., and E. Stamman Segar, 1997: Storm eroded coast with tree debris, Willapa Bay, Washington.

Figure 11-2d – Segar, D., and E. Stamman Segar, 1997: Rocky coast, San Diego, California.

Figure 11-2e – Segar, D., and E. Stamman Segar, 1997: Beach and cliffs, Point Magu, southern California.

Figure 11-2f – Segar, D., and E. Stamman Segar, 1997: Lava flows, South shore, Hawaii.

Figure 11-2g – Segar, D., and E. Stamman Segar, 1997: Island beach and cliffs, Polihale, Kauai, Hawaii.

Figure 11-2h – Segar, D., and E. Stamman Segar, 1997: Fringing coral reef, Guam.

Figure 11-2i – Segar, D., and E. Stamman Segar, 1997: Fjord, Turnagain Arm, Alaska.

Figure 11-3a – ShareBear, 2005: Fundy National Park, NB, Canada. Accessed 18 February 2026, Source.

Figure 11-3b – Segar, D., and E. Stamman Segar, 1997: Rocky beach with high wave energy York, Maine.

Figure 11-3c – Segar, D., and E. Stamman Segar, 1997: Sand dunes, Cape Cod, Massachusetts.

Figure 11-3d – Folsom, W., 2013: Looking from the dune line to the north at Delaware Seashore State Park. NOAA Digital collections, accessed 18 February 2026, Source.

Figure 11-3e – Segar, D., and E. Stamman Segar, 1997: Barrier island beach, Virginia Beach, Virginia.

Figure 11-3f – Folsom, W., 2013: A recreational fisherman cutting a swath through the waterways. NOAA Digital collections, accessed 18 February 2026, Source.

Figure 11-3g – Segar, D., and E. Stamman Segar, 1997: Barrier island beach, Galveston Island, Texas.

Figure 11-5 – Adapted from Moore, J. G., W. R. Normack, and R. T. Holcomb, 1994: Giant Hawaiian underwater landslides. *Science*, **264**, 46-47, Source.

Figure 11-6a – Segar, D., and E. Stamman Segar, 1997: The Franz Josef Glacier on the South Island of New Zealand.

Figure 11-6c – Segar, D., and E. Stamman Segar, 1997: Geirangerfjord, Norway where rising sea level has filled a glacial valley.

Figure 11-7a – Kurig, N., 2022: Southern Mississippi River watershed and the northern Gulf of Mexico. NASA PACE, accessed 18 February 2026, Source.

Figure 11-7b – Adapted from Scott, R. C., 1996: *Introduction to physical geography*. West Publishing Company, 560 pp.

Figure 11-8a – Segar, D., and E. Stamman Segar, 1997: Coastlines of Palau.

Figure 11-8b – Segar, D., and E. Stamman Segar, 1997: Jagged rocks of an ironshore segment in Fiji.

Figure 11-9 – Base map from Google Earth 10.100.0.1, 2026: New Jersey coastline, 40°01'10.16"N 74°07'09.72"W. Accessed 18 February 2026, Source.

Figure 11-10b – Segar, D., and E. Stamman Segar, 1997: Cliff and beach near Monterey, California.

Figure 11-11 – Segar, D., and E. Stamman Segar, 1997: Rocky coastline on the island of Palau.

Figure 11-12c – Segar, D., and E. Stamman Segar, 1997: Headland on the island of Molokai in Hawaii.

Figure 11-12d – Segar, D., and E. Stamman Segar, 1997: Sea stacks on the Big Sur coast south of Monterey, California.

Figure 11-14a – Segar, D., and E. Stamman Segar, 1997: Dunes separate the beach from a low marshy area, Aruba coast.

Figure 11-14b – Segar, D., and E. Stamman Segar, 1997: Dunes 10 m, Monterey Bay, California.

Figure 11-15a – Segar, D., and E. Stamman Segar, 1997: Two berms on a beach just south of San Francisco, California.

Figure 11-15b – Segar, D., and E. Stamman Segar, 1997: Berm on the east coast of the North Island of New Zealand.

Figure 11-15c – Segar, D., and E. Stamman Segar, 1997: Beach scarp on the California coast just south of San Francisco.

Figure 11-15d – Segar, D., and E. Stamman Segar, 1997: Another view of a beach scarp on the California coast just south of San Francisco..

Figure 11-16a – Segar, D., and E. Stamman Segar, 1997: Beach in New England is covered by large cobble-sized deposits.

Figure 11-16b – Segar, D., and E. Stamman Segar, 1997: Black sand on a beach on the south shore of the island of Hawaii.

Figure 11-18 – Base map from NCEI Bathymetric Data Viewer, 2026: Southern California coastline DEM color shaded relief, 32°52'21.96"N 117°16'49.70"W. NOAA, accessed 18 February 2026, Source.

Figure 11-21b – Segar, D., and E. Stamman Segar, 1997: Homer Spit, Alaska.

Figure 11-21c – Segar, D., and E. Stamman Segar, 1997: A baymouth bar at the mouth of Tillamook Bay, Oregon.

Figure 11-21d – Bledsoe, S. 1998: Aerial view of Assateague Island on the Atlantic Ocean in Maryland, USA. Army Corps of Engineers Digital Visual Library, accessed 8 July 2025, Source.

Figure 11-23a – Segar, D., and E. Stamman Segar, 1997: The Galveston seawall with the beach eroded along much of its length and the shoreline more exposed.

Figure 11-23b – Segar, D., and E. Stamman Segar, 1997: The beach just beyond the Galveston seawall has retreated substantially shoreward.

Figure 11-24b – Base map from Google Earth 10.100.0.1, 2026: Groins near Cape May, New Jersey, 38°55'40.00"N 74°55'18.79"W. Accessed 18 February 2026, Source.

Figure 11-24c – Base map from Google Earth 10.100.0.1, 2026: Groins near Beach Haven, New Jersey, 39°33'09.41"N 74°14'27.84"W. Accessed 18 February 2026, Source.

Figure 11-25b – Base map from Google Earth 10.100.0.1, 2026: Jetties in Three Mile Harbor, Long Island, NY, 41°02'06.43"N 72°11'18.96"W. Accessed 18 February 2026, Source.

Figure 11-26b – Base map from Google Earth 10.100.0.1, 2026: Santa Barbara Harbor, CA, 34°24'16.64"N 119°41'18.61"W. Accessed 18 February 2026, Source.

Figure 11-27b – Base map from Google Earth 10.100.0.1, 2026: Breakwater off Santa Monica pier, CA, 34°00'28.11"N 118°30'07.77"W. Accessed 18 February 2026, Source.

Figure 11-27c – Base map from Google Earth 10.100.0.1, 2026: Series of breakwaters on the coast of Norfolk, VA, 36°56'57.30"N 76°14'24.76"W. Accessed 18 February 2026, Source.

This page titled 11.8.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

12: FOUNDATIONS OF LIFE IN THE OCEANS

Technical Editor: Emily Brownlee, Ph.D. (St. Mary's College of Maryland)

Additional contributions and edits by Elaine Stamman, M.S. (Texas A&M University, retired) and Michelle Valkanas, Ph.D. (Pennsylvania Western University)

Critical Concepts

- **CC4 Particle Size, Sinking, Deposition, and Resuspension:** Suspended particles in ocean water, including plankton, sink at rates primarily determined by particle size: large particles or plankton sink faster than small particles. Larger plankton species have various adaptations to reduce their sinking rate. Small organic particles can be aggregated into fecal pellets that are larger than the individual particles and, thus, sink faster.
- **CC8 Residence Time:** The residence time of seawater in a given segment of the oceans is the average length of time the water spends in the segment. In the deep ocean trenches water column water mass residence time can be long and this affects the character of the biota that inhabits this zone.
- **CC10 Modeling:** Complex environmental systems including the cycling of nutrients among ocean water, living matter, suspended particles and sediments can best be studied by using conceptual and mathematical models. Many oceanographic and climate models are extremely complex and require the use of the fastest supercomputers.
- **CC14 Phototrophy, Light, and Nutrients:** Phototrophy (which includes photosynthesis) and chemosynthesis are the two processes by which simple chemical compounds are made into the organic compounds of living organisms. Photosynthesis depends on the availability of carbon dioxide, light, and certain dissolved nutrient elements including nitrogen, phosphorus, and iron. Non-photosynthetic phototrophy is the use of rhodopsins in place of chlorophyll and traditional photosynthesis allowing organisms to grow in low nutrient environments. Chemosynthesis does not use light energy and instead depends on the availability of chemical energy from reduced compounds which occur only in limited environments where oxygen is depleted.
- **CC15 Food Chain Efficiency:** All organisms use some of their food as an energy source in respiration and for reproduction, and also lose some in excretions including wastes. On average, at each level in a food chain, only about 10% of food consumed is converted to growth and biomass of the consumer species.
- **CC19 Metagenomics:** Metagenomics is the study of all of the genetic material in a sampled environment. These studies provide insight on the organisms present and what their metabolic potential is. Metagenomics has been used to understand the microbial biomes that exist within the ocean. Studies have shown that the majority of microbes living within the oceans have yet to be discovered and that viruses play a much larger role than originally described

Figure 12-1. Although it takes many wonderful forms, life in the oceans is all made possible by a few simple physical and chemical processes. When we found it, this midring blue-ringed octopus (*Hapalochlaena sp.*, Indonesia) was a drab, dark color and blended in almost totally with the background. After we approached, it changed color rapidly and displayed the blue rings as a warning. A very closely related species of blue-ringed octopus is venomous with a toxin in its saliva for which there is no antidote. This species is thought to be less toxic, but little is known about it. This individual octopus entertained us by swimming and stopping several times, all the while displaying the blue rings and rapidly varying its body color before it settled and assumed the shape and color to blend perfectly with the rock on which it had landed.

In many ways, the oceans are an ideal **environment** for life. Temperatures are much more uniform than on land, so most marine organisms do not have to contend with the temperature variations and extremes that terrestrial organisms experience. In addition, all elements essential to life are present as dissolved **ions** in seawater, albeit sometimes at very low concentrations, and they are readily available to marine **archaea**, **bacteria**, and **algae**, which perform the task of converting these dissolved substances into living matter. Living in the ocean also presents special problems, and the abundance of ocean life is limited by a variety of physical and chemical factors.

This chapter examines the basic processes of life in the oceans, and the physical and chemical parameters that control the quantities and types of organisms in different regions of open oceans. The parameters that determine the distribution of life include the physical and chemical nature of seawater, the distribution of winds and **currents**, and even **plate tectonics**. **Chapter 13** discusses how life in coastal and **estuarine** zones differs from open-ocean marine life as a result of the different physical and chemical environments.

12.1: How do we describe life?
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
12.4: Studying Microbes in the Ocean
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.7: Food Webs
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
12.10: Organic Carbon
12.11: Biological Provinces and Zones
12.12: Plankton

12.13: Nekton
12.14: Benthos
12.15: Chapter Summary
12.15.1: Key Terms
12.15.2: Chapter Questions
12.15.3: References

This page titled 12: Foundations of Life in the Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.1: How do we describe life?

Before we examine the many complex physical, chemical, and biological interactions that are the foundations of life in the oceans, it will help to look briefly at how biologists group and classify living organisms. All living things are arranged into formal groups according to their anatomy, physiology, and, more recently, genetic differences, through a system known as "taxonomy." Taxonomy is discussed more fully in Appendix 3. For now, we need to know that, at the highest level of organization, called the "tree of life," all **species** are classified into one of three domains (**Fig. 12-2**, **Table 12-1**): Bacteria, Archaea, and Eukarya.

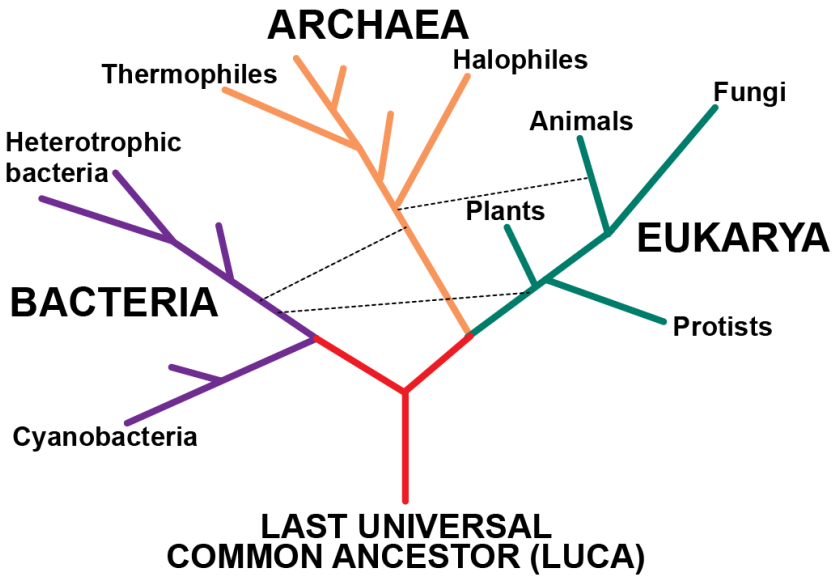


Figure 12-2. The tree of life. The prokaryotes—Bacteria and Archaea—are single-celled organisms with no membrane-bound (enclosed by a membrane) organelles, including a nucleus. The eukaryotes (Eukarya)—protists, fungi, plants, and animals— are single- or multicellular organisms whose cells contain membrane-bound organelles (e.g., nucleus, mitochondria, endoplasmic reticulum, golgi apparatus). All life is assumed to derive from a common ancestor, a progenote, referred to as LUCA (Last Universal Common Ancestor of all life). Bacteria and Archaea are known to have developed before Eukarya (to which humans belong). DNA indicates that the Eukarya developed from Archaea but Eukarya also inherited one or more genes from Bacteria through endosymbiotic events and horizontal gene transfer (represented conceptually by the dashed lines in the figure). Horizontal gene transfer (transfer of one or more genes between species) is now known to be common so the original Darwinian theory of evolution of one species into another in a linear fashion that was the basis for the tree of life concept must now be revised. Protists include diatoms, dinoflagellates, foraminifera, radiolaria, marine algae and others (see **Table 12-1**).

As technology continues to advance, our ability to look at organisms on a molecular level continues to shed light on relatedness and unrelatedness of organisms. For example, Archaea were once considered part of the Bacteria domain but are now considered separately. While the three-domain tree is the widely accepted model, there is now some evidence that suggests that the tree of life might only contain two branches with the Eukarya being considered simply a subbranch of Archaea. As advances in **genome sequencing** progress, there may be a move toward a wider acceptance of the two-domain model.

Together, bacteria and archaea make up a group of species known as **prokaryotes**. All other living species belong to a group called **eukaryotes**. Scientists traditionally divided eukaryotes into four groups, three kingdoms Fungi (**fungi**), Plantae (plants), and Animalia (animals) and a fourth group, the **protists**. However, the protists, traditionally assigned as kingdom Protista, is not a single, unified group, but rather a designation of diverse organisms, most of which are single-celled, that do not fit into the other eukaryotic kingdoms. Viruses have traditionally been considered to be non-living because they cannot reproduce without a host, are acellular, not made of cells, and cannot generate energy. However, this view has more recently been debated and there is now evidence that viruses may have preceded all of the other kingdoms and that they stand at the base of the tree of life and may have evolved into cellular organisms and LUCA, the last universal common ancestor of all species of cellular life. However, the origin of viruses remains unknown and the issue of whether they are a branch of "life" remains controversial.

As discussed in Appendix 3, species within each kingdom are classified into groups that are arranged in a hierarchy containing a number of levels. All species eventually are given a formal name that consists of **genus** and species, the two lowest levels in the hierarchy. For example, humans are genus *Homo*, species *sapiens*; we are all members of the species *Homo sapiens* (note the italics always used for genus and species names).

Although the formal taxonomic classification of species addresses the need to classify organisms according to their physiological and genetic differences, it does not always serve well to classify species with regard to their functions in the **ecosystem**. Therefore, many other functional grouping schemes are used for specific situations. For example, marine biologists may group organisms according to whether they live in the water column or on the seafloor, or whether they are able to swim against currents or primarily drift with the water, or whether they are capable of synthesizing living organic matter from inorganic matter or not.

Table 12-1. The Major Taxonomic Groupings of Living Organisms

Group	Description
Prokaryotes	Generally microscopic, single-celled organisms that have no membrane-bound (enclosed by a membrane) organelles (e.g., nucleus, mitochondria, endoplasmic reticulum, golgi apparatus, etc.). May be photosynthetic, chemosynthetic, or heterotrophic.
Domain Bacteria	Generally, microscopic and relatively simple single-celled organisms. Bacteria have no organelles (membrane-bound nucleus or internal structural features). However, most possess a carbohydrate-based cell wall composed of peptidoglycan.
Domain Archaea	Archaea are similar to bacteria but have different cell wall structures, different constituent organic compounds, and many different genes. Many archaea live in extreme environments, such as those found at hydrothermal vents, those where the temperature exceeds the boiling point of water (e.g., geysers), those where salinities are extremely high (e.g., salt lakes), or those that are strongly acidic or alkaline. Some archaea are single-celled, whereas others form filaments or aggregates.
Eukaryotes: Domain Eukarya	Single-celled or multicellular organisms whose cells have a nucleus and other internal structure. Generally larger than prokaryotes.
Protists *Informal Group*	A diverse group of species, including all the eukaryotes except for the plants, fungi, and animals. The vast majority of protists are single-celled organisms, typically only 0.01 to 0.5 mm in size, yet generally larger than the prokaryotes. Examples are diatoms and dinoflagellates, foraminifera, radiolaria, marine algae, and seaweeds. Protists commonly can survive dry periods in the form of cysts; some are important parasites; a few forms are multicelled—for example, the brown and red algae.
Kingdom Fungi	Mostly multicellular organisms but single-celled organisms exist (yeast), including many decomposers and parasites that infect animals and plants. Fungi release enzymes that break down organic matter that they absorb for food and energy. Examples are molds and mushrooms.
Kingdom Plantae	Multicellular photosynthetic autotrophs. Flowering plants, including sea grasses, mangroves, ferns, and mosses. Note that marine algae and seaweeds are *not* generally considered to be plants as they are non-vascular lacking roots, stems, and leaves.
Kingdom Animalia	Multicellular heterotrophs, including all invertebrates and vertebrates.

This page titled 12.1: How do we describe life? is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.2: Production, Consumption, and Decomposition

Life is based on the production of an enormous variety of organic compounds, each of which serves a different function in the cells of living organisms. Organic compounds are created when carbon atoms are combined in chains or rings. Chemical properties of a compound are determined by the number of carbon atoms and how they are arranged, and by the number, position, and elements in other groups of atoms (containing, for example, oxygen, nitrogen, sulfur, or phosphorus) attached to the carbon chains or rings. Organic compounds made by living organisms can contain hundreds of carbon atoms and attached groups arranged in an enormous number of ways. Even the simplest living organisms consist of a bewildering array of organic compounds.

The most important use of organic molecules by living organisms is to provide energy for the **biochemical** processes that control their growth, movement, feeding, and reproduction. Energy is needed to combine carbon atoms into organic compounds. This energy can be released from organic compounds by decomposition (breakdown) of their molecules through catabolic pathways like **respiration** and **fermentation**. All living organisms use catabolic pathways to provide their needed energy.

There are two fundamental types of organisms. Those of the first type, **autotrophs**, create their own food from inorganic compounds by using an external source of energy. The process of converting inorganic compounds to organic matter is called **primary production**. Autotrophs use a portion of the food they synthesize to fuel their own life processes through respiration. Plants, algae, and some bacteria and archaea are autotrophs. Organisms of the second fundamental type, **heterotrophs**, cannot make organic compounds from inorganic compounds and must obtain organic matter as food. Animals and most bacteria and archaea are heterotrophs. Recent research has shown that the distinction between autotrophs and heterotrophs is not as simple as once thought. There is now strong evidence that certain microbial species in the oceans, collectively called **mixotrophs,** possess the genetic machinery for both autotrophy and heterotrophy and it is likely that these species can behave either as an autotroph or a heterotroph depending on the availability of light, nutrients and dissolved food (carbon in organic compounds that can be “burned” to produce fuel for metabolic processes).

The distribution of life depends on the distribution and growth rate of autotrophs. Heterotrophs can live only where and when autotrophs supply enough food and **nutrients**. Autotrophs synthesize organic matter by one of two fundamentally different mechanisms: **phototrophy** (predominately **photosynthesis**) and **chemosynthesis.**

Phototrophy

The dominant process of primary production in the surface mixed layer of the oceans is photosynthesis (**CC14**). The basic raw materials needed for photosynthesis are carbon dioxide, water, and nutrients. An ample supply of carbon dioxide is dissolved in seawater in the form of carbonate and bicarbonate ions (**Chap. 5**). Of course, water is readily available. The source of energy used to combine carbon atoms is solar radiation. Therefore, light availability is an important determinant of where photosynthesis can occur and at what rate. Photosynthesis in the oceans is dominated by phytoplankton that include species of algae and cyanobacteria. The general term for organisms using light energy to produce organic matter for growth of living organisms is **phototroph** (**CC14**). Organisms that perform traditional photosynthesis are **photoautotrophs**.

Non-Photosynthetic Phototrophy

Certain archaea, and bacteria are phototrophs that do not use chlorophyll or photosynthesis and do not release oxygen. Instead they use a simpler metabolic pathway to capture light energy using compounds called **rhodopsins**. Rhodospins do not rely on the electron transport chain to generate energy, but instead use light-activated proton pumps. When these pumps are activated by light, they pump hydrogen ions across the membrane generating ATP.

Microbes utilizing rhodopsin-dependent phototrophy are heterotrophs, unable to fix carbon dioxide on their own (i.e., **photoheterotrophs**). Organisms capable of photoheterotrophy tend to thrive in waters where nutrient concentrations are very low. This is especially true in iron-depleted environments where chlorophyll synthesis, which requires iron, is limited. Additionally, rhodopsins are not dependent on temperature and can absorb a wide range of light waves, allowing these microbes to thrive in marine environments where other organisms may not be able to. Recent research also shows that certain marine bacteria and phytoplankton use the rhodopsin based pathway to supplement photosynthesis when conditions are unfavorable for photosynthesis (e.g., cold, low iron concentrations, and/or high light intensity). This versatility speaks to why the use of rhodopsins is more widespread than originally thought. In fact, it is estimated that over 60% of microbial cells found in ocean surface

environments carry rhodopsin genes. While it is not exactly clear what impact rhodopsins have on biogeochemical cycling in marine environments, it is clear that the use of rhodopsins is advantageous, increasing energy efficiency and survival in microbes living in an ever changing ocean.

Nutrient elements, such as nitrogen, phosphorus, magnesium, and sulfur, are also necessary for phototrophy. For example, nitrogen atoms are part of all protein and **chlorophyll** molecules, which are essential to photosynthesis (**CC14**). The following sections of this chapter explain how the distribution of light and nutrients and the processes that affect their distribution control the distribution of life in most of the oceans.

Chemotrophy

Some types of bacteria and archaea can convert carbon dioxide and water to organic matter by using energy from chemical reactions rather than from light. This process, called "chemosynthesis," uses energy obtained from chemical reactions including the conversion of methane gas to carbon dioxide and water; and the oxidation of inorganic compounds a process known as lithotrophy) including:

- hydrogen sulfide to sulfate
- metals from a reduced to an oxidized form, including iron, arsenic, manganese and uranium
- hydrogen to water
- ammonia to nitrite
- nitrite to nitrate

Note that "oxidation" does not necessarily require oxygen atoms to be present. Oxidation is the general term for a chemical reaction in which an electron is removed from a molecule. This means that molecules without oxygen atoms can act as the electron acceptor for example in the conversion (oxidation) of compounds, for example ferrous iron (Fe^{2+}) compounds to compounds of ferric iron (Fe^{3+}). These reactions release energy that can be used by organisms. Some organisms can perform chemosynthesis in the complete absence of oxygen so chemosynthesis is possible almost anywhere on Earth and other planets even if they have no free oxygen. Indeed, there was no (or extremely low concentrations of) oxygen in Earth's atmosphere or free dissolved oxygen in Earth's oceans when life first evolved. The first life forms were almost certainly chemosynthetic.

Hydrogen sulfide, metal sulfides, hydrogen, and methane are all oxidized readily by chemical processes in environments where oxygen is present. Therefore, most of the reduced compounds that fuel chemosynthesis are not currently present in the atmosphere or in most ocean waters, because these environments contain appreciable concentrations of oxygen. In reducing environments where chemosynthetic fuels are abundant, free oxygen is absent because it has been consumed by respiration or chemical reactions. In the oceans chemosynthesis currently occurs in geographically limited environments where both reducing and oxygenated waters meet and mix, including **hydrothermal vents** and some seeps. Chemosynthesis also occurs in seafloor or near seafloor anaerobic environments such as **sediment** pore waters and oxygen free dead zones (**Chap. 15**). It also occurs in transitional environments, such as oxygen minimum zones discussed later in this chapter, where dissolved oxygen is not totally depleted but where complex microbial communities include chemosynthetic species.

Chemosynthesis is common near hydrothermal vents on **oceanic ridges** and volcanoes, but it also occurs in water released to the oceans by vents or in seeps from ocean sediment. Some vents or seeps are in **subduction zones**, where water with dissolved methane and hydrogen sulfide is squeezed out of **subducted** ocean sediment. Others are in certain areas of the **continental shelf** where **groundwater** from the continents migrates through the sediment and accumulates methane or hydrogen sulfide. Chemosynthesis also occurs in the surface sediments of salt marshes and swamps, at the interface between oxygenated surface waters and oxygen-deficient bottom waters in **fjords**, and in other ocean areas where oxygen is depleted. Living microorganisms have been found in many extreme environments, including deep within ocean sediments and even within oceanic **crust** that is millions of years old. Many of these organisms are probably chemosynthetic, obtaining their energy needs by chemical reactions such as the combination of hydrogen and carbon dioxide to form methane.

Secondary Production and Decomposers

Nonautotrophic marine organisms are heterotrophic and must obtain food to supply all or part of their needs for organic matter and energy. Heterotrophs convert organic compounds to carbon dioxide and water during respiration, thereby releasing energy that was originally derived from the sun (phototrophy) or chemical sources (chemotrophy). The released energy fuels the metabolic processes of the heterotrophs.

Heterotrophs include all animals and many species of archaea, bacteria and fungi. Animals eat plants, algae, bacteria, archaea, other animals, and organic **detritus** formed by the partial decomposition of dead organisms. Animals that are **herbivores** eat photosynthetic organisms, **carnivores** eat only animals, **omnivores** eat herbivores, carnivores or other omnivores, and **detritivores** eat detritus. The production of animal **biomass** by animals that consume primary producers or other heterotrophs is called “secondary production.”

Heterotrophs use food inefficiently (**CC15**). They **excrete** part of their food as solid waste organic particles, called **fecal pellets**, or as dissolved organic matter in liquid excretions. Humans are no exception to this rule.

Many species of microorganisms, including some bacteria, archaea, and fungi, are heterotrophs but are also called **decomposers**. Aerobic decomposers obtain energy from organic particles or dissolved organic compounds, which they convert to carbon dioxide and water. Decomposition can also occur in oxygen-poor environments through anaerobic processes that result in the production of methane gas. Some species of microorganisms are capable of growing either autotrophically or heterotrophically, depending on their environment.

Phototrophic and chemosynthetic organisms, as well as food-eating and decomposing organisms, function together as a **community** within an ecosystem (**Fig. 12-3**). In an ecosystem, carbon dioxide, water, and nutrients are synthesized to organic matter, which is then processed through herbivores, carnivores, omnivores, detritivores, and decomposers. Each of the organisms in this **food chain** breaks down some organic matter and releases carbon dioxide, water, and nutrient elements back to solution.

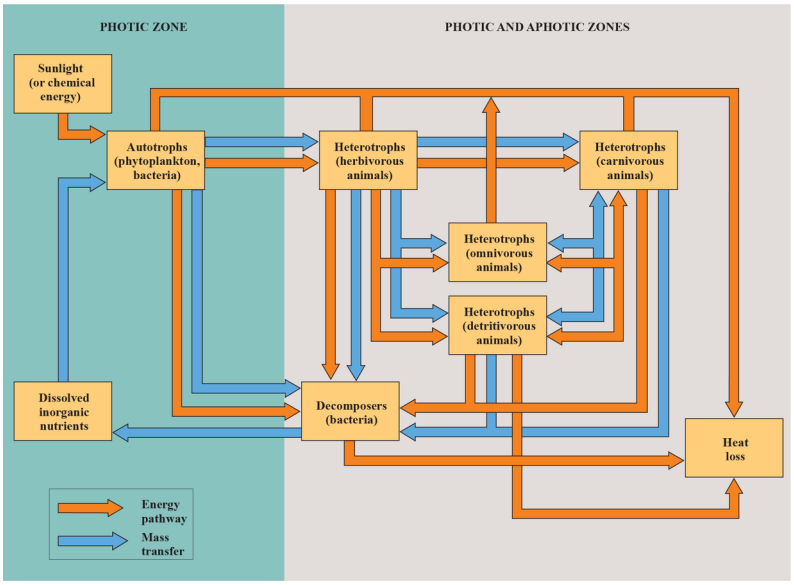


Figure 12-3. Simplified model of the energy and mass (organic matter and nutrient element) cycles in a marine ecosystem. Sunlight (or chemical energy) and dissolved nutrients are essential for autotrophs to perform the phototrophic (or chemosynthetic) primary production of organic matter from carbon dioxide. Energy collected and mass produced by autotrophs is cycled through various heterotrophs, followed by decomposers. The mass is transferred back to solution by decomposers, and the energy is eventually released as heat. Primary production of organic matter takes place only in the photic zone (except for limited areas where chemosynthesis occurs). All other steps in the ecosystem take place partly in the photic zone and partly in the aphotic zone.

Nutrients and other elements are recycled through ecosystems (**Fig. 12-3**). Recycling is an important factor in the distribution and abundance of ocean life. In parts of the oceans, life is limited because nutrients are not recycled fast enough to support photosynthesis.

This page titled 12.2: Production, Consumption, and Decomposition is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.3: Microbes in Charge

At the end of the twentieth century and the beginning of this century, marine science came to recognize that ocean ecosystems were dominated by microscopic organisms in a manner that would also change our perception of what constitutes the true foundation of life on our planet. To understand this, we first need to review just a few of the startling findings that have now revealed the dominance of **microbial** organisms on our planet. Note that microbe and microbial are terms to identify those groups of tiny ocean organisms that are commonly less than about 60 μm, which is smaller than the approximately 100 μm smallest object that a normal human eye can see clearly. First, we have learned that there are about 10^6 bacteria in each milliliter of ocean water (or $5 \bullet 10^6$ per teaspoon) and viruses are about 10 times more abundant. This means that Earth's ocean is estimated to contain 10^{29} bacteria which is a number one hundred million times greater than the total estimated number of stars in the universe (10^{21}). The total mass of bacteria and other single-celled organisms in the ocean exceeds the combined mass of all living organisms that are non-microbial by 5 to 10 times.

Microbial organisms perform essentially all of the primary production in the oceans, a role played by plants in terrestrial ecosystems and despite their small size, total primary production in the oceans is estimated to approximately equal total primary production on land. One type of primary producer, a cyanobacteria belonging to the genus *Prochlorococcus* is among the smallest primary producers (about 0.5 to 0.7 μm) but, despite its small size it is estimated to be responsible for about 20% of the oxygen in Earth's atmosphere. *Prochlorococcus* is able to do this because it is believed it may be the most plentiful genus of species on our planet, with an estimated 10^{27} individuals.

If it were not for microbes, no animals, including humans, would ever have lived on the planet. Earth's early atmosphere and oceans had no free oxygen. Microbial photosynthesis is thought to have developed in cyanobacteria during an event known as The Great Oxygenation Event that began about 2.45 billion years ago and lasted between 200 and 400 million years. During this event, the abundant free oxygen in the atmosphere and oceans that all animals depend on was generated by microbes (cyanobacteria) in the oceans and the atmospheric and ocean oxygen concentrations reached levels comparable to the current concentrations.

The facts about ocean microbes listed above demonstrate the dominance of microbial organisms in the oceans. However, the dominance of microbial organisms is probably much greater still because they have the ability to metabolize at rates far exceeding those of larger organisms. This is because the ratio of surface area to volume generally is greater for smaller than for larger organisms. We can visualize why if we consider an apple cut into two halves. The apple's total volume does not change when it is cut, but the total surface area is increased by the area of the newly exposed interior surface of each apple half. If the apple is cut again, the total volume still remains the same, but the total surface area again increases. All organic and inorganic nutrients, oxygen, and waste products have to pass through the cell surface. Because of its much greater ratio of surface area to volume a microorganism 1 micrometer (one thousandth of a millimeter) in diameter can metabolize about 100,000 times faster than a human. To put this in context, for one oceanic bacterium that divides every 10 min. (not unusually fast for bacteria if appropriate nutrients are available) the energy throughput of a mass of the bacteria equal to a human would be about 20 megawatts, enough to provide power and heat for about 10,000 homes in the U.S.

Because they are a large fraction of the total biomass and because they have high metabolic rates when nutrients are available, microbes dominate the movement of energy and biologically important elements in the oceans. Because the microbial community includes primary producers, consumers, and decomposers all with fast metabolic rates, most of the total primary production in the oceans is recycled through microbes and only a small fraction moves far enough up the food chain to reach non-microscopic eukaryotes. It is estimated that less than 1 to 2% of the total production is available for non-microscopic eukaryotes.

The rapid rate of division of microbial species also suggests that these microbial species are far more adept at adjusting to changing environmental conditions than larger longer lived organisms. It has been found, through genetic studies like metagenomics (**CC19**), that ocean microbial **biomes** are as small as a single drop of water and that complex ecological and evolutionary processes take place in each such biome. Additionally, recent studies have shown that viruses, especially those that infect bacteria and archaea, play a much larger role in microbial biomes than previously thought. Thus, although global climate change may cause many species to become extinct, ocean microbes will almost certainly evolve, adapt, and survive as they have done for billions of years.

This page titled 12.3: Microbes in Charge is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.4: Studying Microbes in the Ocean

The study of how microbes exist within their intricate communities and the impact they have on biogeochemical cycling continue to reveal how much we still have to learn about the vast array of life living hidden beneath the ocean's surface. Advances in science, specifically DNA technology, has led to affordable and accurate methods of determining the exact DNA sequence (or genetic code) of an organism. **Metagenomics**, a type of DNA sequencing, is used to reconstruct the genetic makeup of all of the organisms in a sample (**CC19**). Metagenomic studies can tell us what organisms are there and what they are capable of doing but can't tell us if they are actually doing it. Metagenomic studies have revealed a vast array of microbes in marine environments, many of which have never been identified let alone cultured in the lab. The Tara Oceans Expedition took place from September 2009 to March 2012 and is one of the largest efforts to capture microbial life to date. Samples were collected from all major oceans across 210 sampling stations and over 35,000 oceanic samples. The data collected from this study has provided an immense amount of data resulting in over 100 scientific publications, and new information is still being gleaned from this research even today. The study cataloged over 150 million genes from microbes and classified over 100,000 plankton species. Through studies like this and the use of technology like metagenomics and other sequencing technologies (e.g., amplicon sequencing, metatranscriptomics, proteomics, metabolomics), an open-source database of ocean microbes named KMAP Global Ocean Gene Catalog 1.0 was established in 2024 with the goal of creating a comprehensive collection of the microbes living in the ocean.

While we have learned a lot from metagenomics and other sequencing approaches (e.g., amplicon sequencing, metatranscriptomics, proteomics, metabolomics; see **CC19**), there is still a lot to be learned. This will require additional sampling efforts, cross-disciplinary research projects, and modeling efforts.

This page titled 12.4: Studying Microbes in the Ocean is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.5: Phototropic Primary Production and Light

Seawater absorbs sunlight, so no light at all reaches ocean depths greater than a few hundred meters, even in the clearest open-ocean waters (**Chap. 5**). In coastal waters with high concentrations of **suspended sediments**, **turbidity** is high and light is more effectively absorbed and **scattered**. Turbidity is generally highest in shallow coastal waters where bottom sediments are **resuspended** by waves and in estuaries fed by rivers with high suspended sediment loads. In such waters, light penetrates a few meters at most, and only centimeters in extreme cases.

Phototrophy, including photosynthesis, cannot take place without light. Light is absorbed by the water, and its intensity is reduced with depth. Light of sufficient intensity to support photosynthesis, usually thought to be about 1% of the intensity at the surface, rarely penetrates more than about 100 m, and photosynthesis is generally restricted to the water column above this depth. This upper layer is called the **photic zone**.

Terrestrial plants must obtain nutrients from soil or, in some cases, from raindrops. In contrast, marine organisms can extract nutrients from the surrounding seawater. As a result, they do not have to be attached or rooted to the seafloor. Furthermore, attached marine photosynthesizers can receive enough light for photosynthesis only if they are attached to the limited areas of seafloor that are shallower than the depth of the photic zone. Consequently, most marine photosynthetic species live in the water column and are not attached to the seafloor.

Because they need light and therefore must remain near the surface, even where the water is deep, most marine phototrophs must avoid sinking. Pockets of gas or air contained within some species of algae enable them to float. The bubblelike lumps on some seaweeds (**Fig. 12-4a**) are such pockets. Some species that live attached to the seafloor also use these devices to keep their upper fronds in the light, including the giant kelp in California coastal waters (**Fig. 12-4b**).

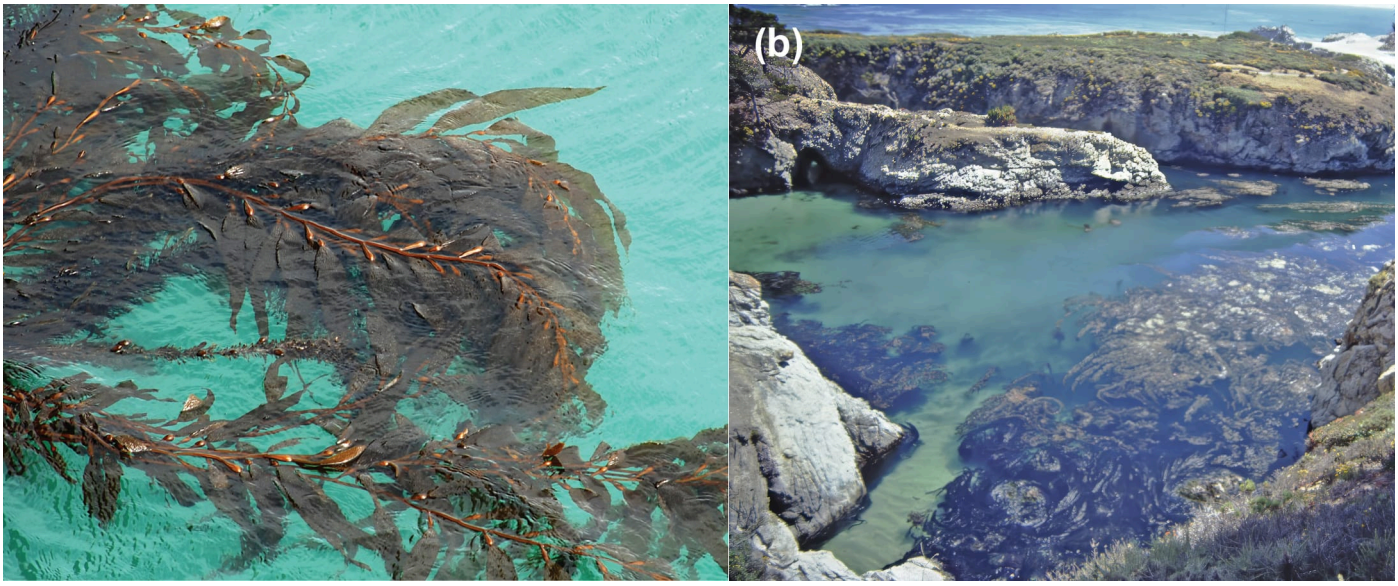


Figure 12-4. Many species of macroalgae have gas enclosures within their structure that enable them to float and stay in the surface waters where there is ample light for photosynthesis. (a) The round ball-like structures in this kelp frond (*Macrocystis sp.*) are gas-filled. (b) Masses of kelp (*Macrocystis sp.*) grow in this cove near Monterey, California. The plants are anchored to the seafloor, but you can see the upper ends of the kelp fronds covering much of the water surface, held up by their gas enclosures.

Phytoplankton include algae and many species of bacteria and archaea that are autotrophs (either phototrophic or chemosynthetic). Phytoplankton are the major primary producers in the oceans and are abundant especially in productive ocean waters. Most phytoplankton species are microscopic and drift freely in the photic zone. Due their small size and the relatively high viscosity of water (**Chap. 5**), phytoplankton cells sink very slowly (**CC4**), even if their **density** is higher than that of seawater, and this helps them to remain in the photic zone. As discussed elsewhere in this chapter, the microscopic size and, therefore large ratio of surface area to volume, of most phytoplankton species also aids them in taking up dissolved substances.

In addition to sinking very slowly due to their small size, **turbulence** due to waves and currents counteracts the tendency of

small phytoplankton to sink (**CC4**). Some species of **diatoms**, which are among the largest phytoplankton, store oils that increase their **buoyancy**, and many also have spines or aggregate in chains, which also reduces their tendency to sink because this increases drag. **Dinoflagellates**, another group of phytoplankton species, are able to counteract sinking because they have whiplike **flagella** that provide them with limited motility.

Phototrophy, Light, and Depth

Until near the end of the 20th century, it was thought that almost all ocean life ultimately depends on photosynthetic primary production for food. Other food sources, other phototrophic and chemosynthetic primary production (especially in hydrothermal vent communities) and the dissolved or particulate organic matter supplied by rivers, were previously considered insignificant in comparison. Research in hydrothermal vent areas and studies in estuarine and coastal waters have shown that these sources are probably much more important than previously thought. Because the relative importance of these other sources is not yet well known, we will focus heavily on photosynthetic primary production in this chapter.

Because photosynthesis requires light energy, most of the primary production in the oceans takes place in the upper layer of the water column, down to the depth that sunlight penetrates. Light intensity at any given depth is determined by the intensity of sunlight reaching the sea surface, the **angle of incidence** of the sun's light to the sea surface, and the water turbidity. The intensity of sunlight reaching the sea surface varies with the seasons, throughout the day as the angle of incidence of the sun changes, and as cloud cover varies. Light intensity at a given depth changes continuously. Fortunately, at any given location and depth, the variation of daily mean light intensity within a season is relatively small in comparison with the variation between seasons. Hence, the daily average intensity is a reasonable measure of the amount of light available for photosynthesis at any location and depth.

Light intensity decreases rapidly with depth (**Fig. 5-18**). In the clearest ocean waters, light intensity is reduced to 1% of the surface intensity at a depth of about 100 m. In coastal-ocean waters, the corresponding depth is variable, but it is often about 5 to 10 m. The rate at which photosynthesis can occur is reduced as light intensity decreases (**Fig. 12-5**). In contrast, the rate of respiration remains virtually constant at all depths because the energy needed to fuel the life processes of all species including photosynthetic autotrophs do not change significantly with depth. At relatively shallow depths, the rate of photosynthesis exceeds the rate of respiration, and photosynthetic autotrophs can grow larger and reproduce. At depths where the respiration rate exceeds the photosynthetic rate, autotrophs must metabolize more organic matter than they create by photosynthesis in order to survive. At least some species of photosynthetic autotrophs can survive for a limited time at these depths, but they cannot grow or reproduce. Although many species die if they do not return to conditions in which photosynthesis exceeds respiration, some species are capable of entering a resting phase, much as terrestrial plant seeds do, and can survive prolonged periods below the photic zone. Some microbial photosynthesizers are mixotrophs, meaning they are capable of switching from autotrophy to heterotrophy when insufficient light energy is available.

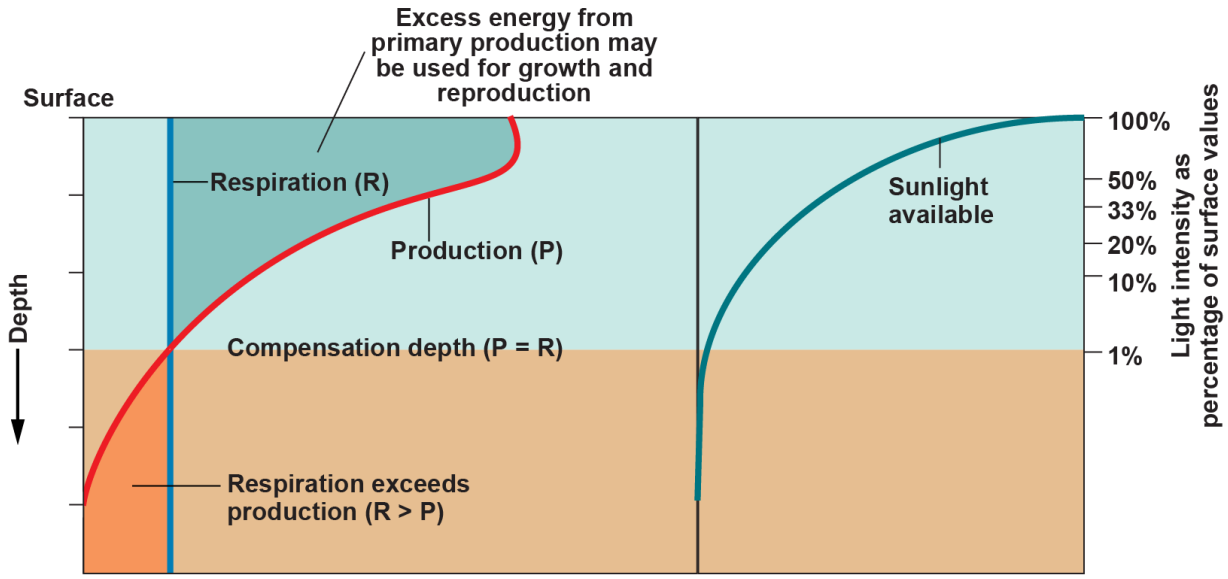


Figure 12-5. Photosynthesis depends on the availability of light energy. Therefore, the rate of primary production by phytoplankton decreases with decreasing light intensity and, thus, with depth. The rate of metabolic energy use and, thus, respiration by phytoplankton remains almost constant regardless of depth. The depth at which the rate of phytoplankton primary production equals the rate of phytoplankton respiration is called the "compensation depth." Below this depth, phytoplankton use more organic matter for respiration than they can produce by photosynthesis. The water column above the compensation depth is the photic zone, and everything below this depth is in the aphotic zone.

The depth at which photosynthesis equals respiration is called the **compensation depth**. The water column between the ocean surface and the compensation depth is the photic zone, and depths below this region constitute the **aphotic zone**. The compensation depth identifies the depth range within which phytoplankton and other photosynthesizers can produce more organic carbon than they consume. The compensation depth lies approximately where light intensity is reduced to 1% of that at the surface. Hence, the photic zone is rarely deeper than 100 m and is much shallower in turbid coastal waters (**Fig. 5-18**). Except in tropical regions, where the day length does not vary much, the compensation depth generally increases during local summer because the angle of incidence of the sun is smaller and the days are longer. However, in some regions where phytoplankton reproduce more rapidly, the additional phytoplankton cells and increased numbers of **zooplankton** that eat them contribute additional "particles" that absorb and scatter light. Consequently, in these regions the **plankton** reduce light penetration and the compensation depth, offsetting the effect of greater light intensity in summer.

Primary Production and the Ozone Hole

Although the rate of photosynthesis usually decreases with increasing depth (**Fig. 12-5**), in many areas it is often lower in water immediately below the surface than at a depth of several meters. One reason for the lower photosynthetic rate near the surface is that high light intensity, especially ultraviolet light, appears to interfere with photosynthesis. Because ultraviolet light is absorbed rapidly in the first few meters of seawater (**Chap. 5**) the maximum photosynthesis often occurs slightly below the surface.

Chapter 7 discussed depletion of the Earth's **ozone layer** by synthetic chemicals that may cause an increase in ultraviolet light intensity at the Earth's surface. Higher ultraviolet light intensity may significantly increase near-surface inhibition of photosynthesis, reduce the total food supply, and adversely affect ocean ecosystems.

This page titled 12.5: Phototropic Primary Production and Light is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.6: Primary Production and Nutrients

Phototrophic organisms require many elements for growth. Carbon, hydrogen, and oxygen are obtained from dissolved carbon dioxide and water. Because carbon dioxide is abundant in seawater (**Chap. 5**), carbon is plentiful in the ocean. Several other elements are required, each in different amounts. Some, such as magnesium and sulfur, are present in high concentrations in seawater. As with carbon dioxide and water, they are always available in plentiful amounts. Other elements, such as nitrogen, phosphorus, iron, zinc, cobalt, and, for some species, silicon, are needed sometimes in substantial quantities but are present in low concentrations in seawater. In some areas, uptake by organisms or by processes such as **adsorption** on **lithogenous** particles can remove enough of these elements from solution that the remaining concentrations are too low to meet the needs of some species of autotrophs. Growth of phytoplankton, marine algae, and other species can be limited by the unavailability of a single essential element. In this condition, the system is said to be **nutrient-limited**.

Where growth is slowed or prevented because the concentration of a particular element is too low, that element is called a **limiting nutrient**. The most important limiting nutrients are nitrogen as dissolved inorganic ions nitrate (NO^{3-}), nitrite (NO^{2-}), and ammonia (ammonium ion NH^{4+} and NH^{3+}); phosphorus (phosphate ion, $(PO_4)^{3-}$); iron; and silica (probably present in seawater as silicic acid: H_2SiO_4). When inorganic nitrogen and phosphorus ions are at very low concentrations, phytoplankton will often use organic nitrogen and/or phosphorus compounds. Although other nutrient elements, such as cobalt and zinc, are present only as trace elements, biological requirements for trace elements are so small that these elements generally do not limit phototrophic growth. Some marine species that are otherwise autotrophs may need certain organic compounds that they are unable to synthesize for themselves. In certain circumstances, including some newly upwelled water in **coastal upwelling** zones (**Chap. 13**), **primary productivity** may be limited by lack of such organic nutrients.

Nutrient Uptake

Phytoplankton and other autotrophs obtain nutrients from seawater through their outer membrane, relying on the random movements of nutrient ions to that membrane. At lower nutrient concentrations the nutrient ions are farther apart and the probability that an ion can diffuse to the cell outer membrane is lower. More nutrient ions will be in contact with a cell that has a larger surface area. Hence, autotrophs improve their chances of capturing low concentrations of nutrient ions if they have a large surface area. Because phytoplankton and other autotrophs are composed of similar organic compounds, their nutrient needs are approximately proportional to their volume. Recall that, the ratio of surface area to volume generally is greater for smaller than for larger organisms. Thus, phytoplankton compete more effectively for nutrients at low concentrations if they are small (their surface area is large in relation to their volume). This is one reason that most marine autotrophs are small, particularly where nutrient concentrations are low.

Small size is not the only adaptation that allows phytoplankton to compete for a limited supply of nutrients. When phytoplankton grow in seawater with relatively high nutrient concentrations, they take up more nutrients than are immediately needed. The excess nutrients are stored and can be used as a reserve to support growth and reproduction after nutrients have been depleted to concentrations that would otherwise limit growth. If concentrations remain growth-limiting for more than a few days, even the stored nutrients will be used up and primary productivity will slow. However, phototrophy is not completely halted, because some nutrients are continuously made available by recycling (**Fig. 12-3**).

Nutrient Recycling

Nutrients are recycled to solution when organic matter is decomposed during animal and decomposer respiration. Animals excrete nutrients in fecal material and in dissolved form, either in the equivalent of urine or by **diffusion** losses through external membranes. Bacteria and other decomposers act as recyclers and release nutrients as they extract energy from dead organisms, fecal material, and dissolved organic compounds. The four most important limiting nutrients are recycled at different rates: phosphorus is recycled very rapidly, nitrogen more slowly, and iron and silica more slowly yet.

Phosphorus

Phosphorus recycling is relatively simple (**Fig. 12-6**). Phosphate is returned to solution rapidly after the death of any organism by the action of **enzymes** within the organisms that break phosphorus away from organic molecules with which it is combined.

Phosphorus is converted directly to phosphate and released to solution. Enzyme-mediated release of phosphate also occurs within living zooplankton and other animals as they digest their food. Certain zooplankton species are known to excrete as much as 60% of the phosphorus taken in with their food. More than half of what remains is disposed of in fecal pellets, from which it is rapidly released by continued enzyme activity as the fecal material is attacked by decomposers. While much phosphorus is recycled in the water, some sinks to the seafloor sediments. Once released to solution, this phosphorus can migrate into the water column through sediment pore water movements or hydrothermal vents where water movements can eventually bring it back to the surface where it can fuel new life.

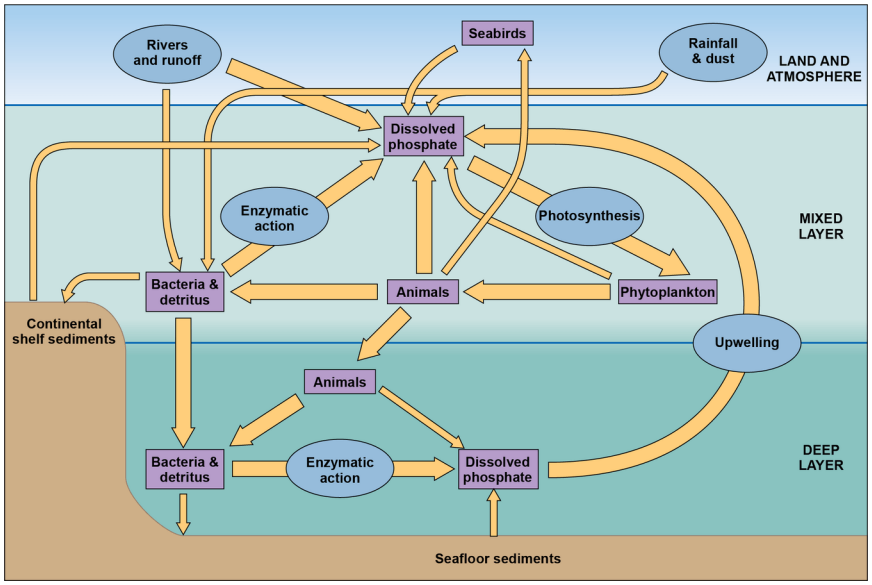


Figure 12-6. Simplified representation of the phosphorus cycle in the oceans. The boxes represent compartments (**CC10**) or components of the system among which the phosphorus is distributed. The ovals represent processes that transport or convert the phosphorus between these compartments. Thicker arrows show more important pathways. Phosphorus is recycled rapidly between solution and organisms in the photic zone. Thus, although some phosphorus is progressively transported below the thermocline, the low concentrations present in some ocean surface waters are still usually sufficient to support phytoplankton growth.

Because it is recycled rapidly, phosphorus is usually not a limiting nutrient in the oceans. In contrast, phosphorus is often the limiting nutrient in lakes. As a result, phosphate-free detergents are required by law in many areas, primarily to prevent **eutrophication** in lakes and rivers.

Nitrogen

Nitrogen, which occurs in living tissue mainly in amino acids, is released during the decomposition of organic matter, primarily as the ammonium ion (**Fig. 12-7**). Animals also release nitrogen in their liquid excretions as **soluble** organic compounds, such as urea and uric acid, which are broken down by decomposers to the ammonium ion. Decomposition of particulate organic matter to release ammonium is much slower than processes that release phosphorus. Consequently, nitrogen often becomes the limiting nutrient in the marine environment.

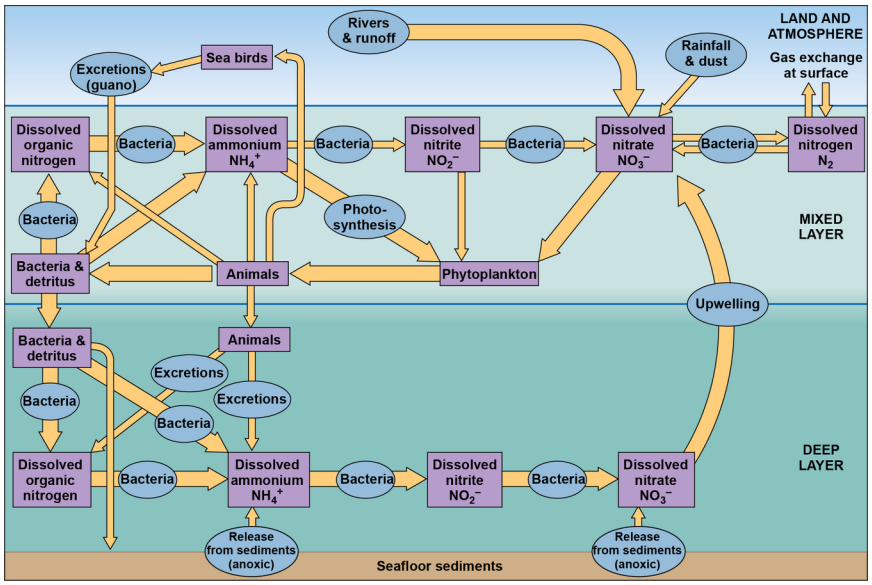


Figure 12-7. Simplified representation of the nitrogen cycle in the oceans. The boxes represent compartments (**CC10**) or components of the system among which the nitrogen is distributed. The ovals represent processes that transport the nitrogen or convert its chemical form between these compartments. Thicker arrows show more important pathways. Unlike phosphorus, nitrogen is not recycled rapidly in the photic zone, and it tends to be transported below the thermocline. Thus, nitrogen is often the limiting nutrient. Nitrogen can be recycled from shallow seafloor sediments (as depicted for phosphorus in **Figure 12-6**), but these pathways have been omitted in this figure to reduce its complexity.

The ammonium ion is used as a source of energy by specialized bacteria and, as discovered more recently, vast numbers of archaea that oxidize ammonium to nitrite. In turn, nitrite is oxidized to nitrate by another group of bacteria. Conversions from ammonium to nitrite and nitrate take place relatively slowly. Nevertheless, in waters below the photic zone this process can be completed, and nitrate is the principal form of dissolved nitrogen other than molecular nitrogen in the aphotic zone. Although the ammonium ion is the preferred source because it requires less energy to capture and use, it is taken up so rapidly that the more abundant nitrate is often the primary nitrogen source, even in the photic zone. Conversion to nitrate may also be completed in temperate **latitudes** during winter, when reduced light availability limits primary productivity. Such seasonal variations are discussed in **Chapter 13**.

Because nitrogen may be predominantly ammonium, nitrite, or nitrate ions at different times and locations, primary producers are generally able to utilize nitrogen in each of these forms. Nitrogen and phosphorus are both supplied to the oceans by river flow, **erosion**, and **weathering** from the land, and, in small amounts, in rain and dust (**Figs. 12-6, 12-7**). The **biogeochemical cycle** (**Chap. 5**) is balanced in each case by the removal of equivalent amounts of nitrogen and phosphorus to the ocean sediment in detritus particles. This balance has been disturbed by human activities, particularly through the use of agricultural fertilizers and the discharge of sewage (**Chaps. 13, 16**). The supply to and removal from the oceans of biologically available nitrogen are complicated by the exchange of molecular nitrogen gas between the atmosphere and ocean waters. Molecular nitrogen dissolved in ocean waters is biologically available only to nitrogen-fixing bacteria and archaea that are able to convert molecular nitrogen to nitrate. Other types of bacteria and archaea, called “denitrifying,” can convert nitrate to molecular nitrogen. Although nitrogen fixation and denitrification are very limited, nitrogen fixation in particular is thought to be extremely important in some nutrient-poor environments, such as **coral reefs** and the centers of **subtropical gyres**. Phosphorus does not usually become the limiting nutrient because it is readily released from detritus back into solution in the photic zone and, as a result, only a small fraction of the phosphorus remains in the detritus that sinks below the photic zone. However, if the supply of nitrogen through nitrogen fixation is continuous, even if very slow, phosphorus can eventually be depleted and become the limiting nutrient.

Silicon

Silicon is essential for the construction of **hard parts** (the shells or skeletons) of a variety of marine organisms, of which diatoms (**Fig. 6-7**) are the most important. Diatoms are larger than many other types of phytoplankton and are covered by an external silica (SiO_2) **frustule**.

Silica is not soluble, so organisms must create their hard parts from silicate ions ($(SiO_4)^{4-}$) which are soluble. Silica is not utilized or released to solution by decomposers, so silica hard parts must dissolve chemically in seawater before silicate ions are again biologically available. Silica dissolves very slowly (**Fig. 6-10**). Consequently, once silicate is depleted in the photic zone,

silicate limitation inhibits the growth and reproduction of organisms with silica hard parts. Because phytoplankton that do not have silica hard parts are not affected, silicate depletion can lead to a change in the dominant species, but it does not limit primary productivity. Silicate is supplied to the oceans in large quantities by rivers and is slowly dissolved from ocean sediment. Therefore, silicate limitation is generally only a temporary situation that develops in locations where diatoms undergo an explosive population growth called a bloom.

Iron

Iron is essential to certain reactions within the process of photosynthesis. Iron is supplied to the surface waters of the oceans from the land, primarily in **runoff**, but also through the transport and **deposition** of airborne dust. Iron is also continuously transported below the photic zone in detritus. As a result, iron can become the limiting nutrient in areas of the oceans where inputs from the continents are too small to replace the iron that is lost from the surface layers by sinking of detritus.

Iron is now thought to be the primary nutrient that limits primary production in large areas of the open oceans where river and dust inputs of iron are low.

High Nutrient Low Chlorophyll (HNLC) Areas

In approximately one third of the area of the oceans there is always sufficient biologically available nitrogen (and phosphorus) but chlorophyll concentrations are very low. These are known as high nutrient, low-chlorophyll (HNLC) areas. Chlorophyll concentration and primary production are closely related so chlorophyll is a proxy measurement of primary production. HNLC areas have plentiful nitrogen and phosphorus but low chlorophyll concentrations indicating low primary productivity so some factor other than nitrogen and phosphorus must limit primary production in these regions.

There are at least three different possible reasons for the low primary productivity of HNLC areas.

- In some areas, wind mixing can be strong enough to transport phytoplankton well below the depth of the photic zone. Phytoplankton cells can then spend too large a percentage of their time without sufficient light for phototrophy. This mechanism appears to be important in limited areas and only during strong wind events but it is likely to apply in some areas, including parts of the Southern Ocean.
- Phytoplankton biomass is kept extremely low due to very high rates of predation by zooplankton. High rates of predation can reduce the average length of time during which each phytoplankton cell lives and can photosynthesize. There is strong evidence for the effect of this mechanism in parts of the subarctic North Pacific Ocean and of the eastern equatorial Pacific Ocean.
- Phytoplankton growth is limited by the lack of availability of nutrients other than nitrogen and phosphorus such as iron or zinc.

Although a lack of other nutrients such as zinc may be possible in some areas, a lack of available iron is now thought to be the primary mechanism limiting productivity in most of the world's HNLC areas. Iron is supplied to the oceans primarily from rivers and wind blown dust and it is highly insoluble in oxygenated sea water. HNLC areas are, therefore, in the open ocean far from the continental shelf and river inputs, and in areas that do not receive large inputs of iron containing wind blown dust. These areas include the sub-Arctic Pacific Ocean, large areas of the equatorial Pacific Ocean, and much of the Southern Ocean.

Nutrient Transport and Supply

Substantial quantities of nutrients are supplied to the oceans by rivers, and smaller quantities are supplied in rainfall (and for some elements wind blown dust). In the open ocean remote from river influences, nutrients in the photic zone are depleted rapidly by growing phytoplankton and other autotrophs. Therefore, if primary production is to continue, nutrients must be resupplied. Some nutrients are continuously recycled and resupplied directly to the photic zone by decomposition. However, there must also be nutrient resupply by upwelling from below the photic zone because a proportion of the nutrients taken up by primary producers is released back to solution in waters below this zone.

Nutrients are transported from the photic zone to the aphotic zone by two major mechanisms. Almost all phytoplankton are eaten by **grazers** (herbivores or omnivores). Although these grazers are mainly small zooplankton, most are much larger than phytoplankton. Much of these animals' undigested waste material is excreted as fecal pellets. Fecal pellets from even the smallest zooplankton are packages of partially digested remains of numerous phytoplankton cells. Consequently, fecal pellets are larger

than phytoplankton and sink much faster (**CC4**). As they sink, fecal pellets provide food for other heterotrophs, bacteria, and fungi and are progressively decomposed with some or all of the nutrient elements being released to solution. Much of the decomposition and release of nutrients occurs below the photic zone, including on the ocean floor.

The second major mechanism of nutrient removal from the photic zone is the vertical migration of zooplankton and other animals. Phytoplankton generally remain in the photic zone unless carried out of it by turbulence or **downwelling**. However, many species of zooplankton and other animals migrate vertically between the photic and aphotic zones each day, rising into the photic zone at night to feed, and then sinking or swimming down to waters below the photic zone during the day. This **diel** vertical migration is probably a defensive mechanism that prevents many potential predators from feeding on the zooplankton when it is light enough to see them easily, while also allowing phytoplankton communities to recover and multiply during the day, ensuring a sustainable food supply for zooplankton when they return to the surface layer to feed. Zooplankton and other animals that practice diel vertical migration continue to digest food and excrete fecal pellets and liquid wastes during the day when they are below the photic zone. Consequently, much of the organic matter that they consume is transported below the photic zone, where it is released, with its nutrients, to the decomposer community.

Vertical Distribution of Nutrients

The photic zone is usually less than 100 m deep in the open oceans and normally restricted to water above the permanent **thermocline** (Chap. 8). Phytoplankton and nutrients are distributed throughout the **mixed layer**, above the thermocline, by turbulent motions of wind and wave mixing. Where light is sufficient to support primary production in the open oceans away from terrestrial sources of supply, nutrients are continuously transported below the thermocline by the mechanisms already described. Below the thermocline, these nutrients are released by decomposers, but the water into which they are released does not mix with the mixed-layer or photic-zone water above, because vertical mixing is inhibited by the density difference across the thermocline. Consequently, the mixed layer becomes nutrient-depleted, while deeper water becomes more nutrient-rich (Fig. 12-8).

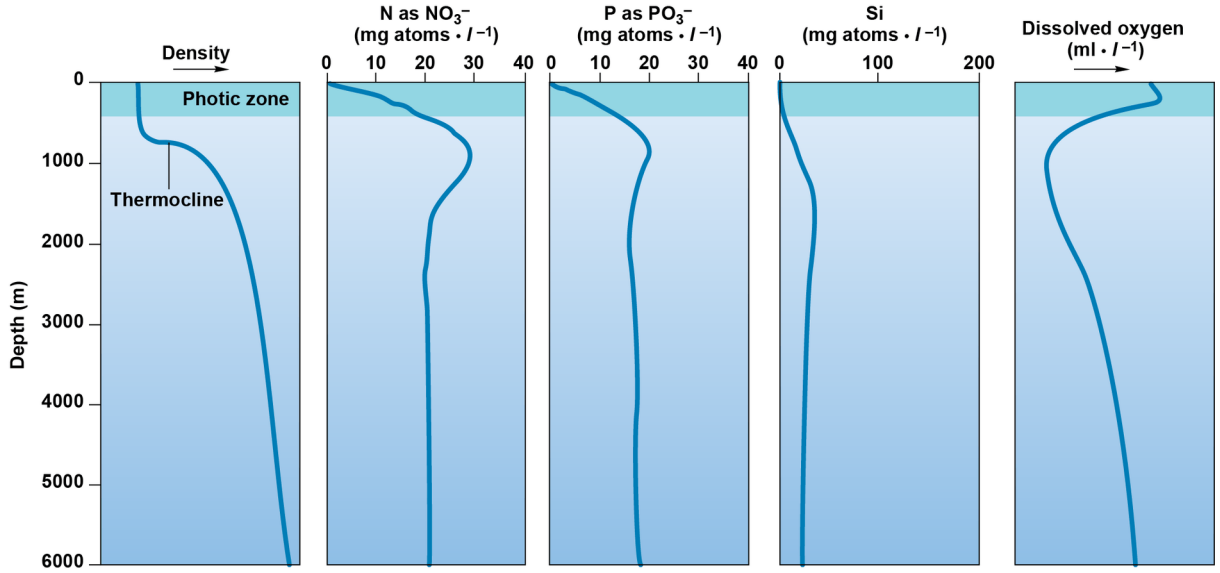


Figure 12-8. General representation of the vertical profiles of water density, nitrate, phosphate, silicate, and dissolved oxygen concentrations in the center of the Atlantic Ocean basin. Dissolved oxygen concentration is high above the thermocline, reaches a minimum just under the thermocline, and then increases to somewhat higher levels in the deep water masses formed by cooling of surface water at high latitudes. Nutrients (nitrate, phosphate, silicate) are generally depleted in the photic zone, increase rapidly in concentration until they reach a maximum at about the same depth as the dissolved oxygen minimum, then remain at relatively constant high levels in the deep water masses.

Organic matter is decomposed continuously at all depths in the oceans, although the rate of decomposition varies with depth. Hence, water that leaves the surface to form deep water masses progressively accumulates recycled dissolved nutrients. Nutrient concentrations are usually highest in water immediately below the thermocline layer. The primary reason for this nutrient maximum is that this is the “oldest” deep water because it is formed by slow upward mixing of bottom waters that were formed near the poles (Chap. 8). The effect of water-mass “age” in determining nutrient concentrations is illustrated by the nutrient concentration differences between the deep waters of the Atlantic Ocean and the much older deep waters of the Pacific Ocean (Fig. 12-9).

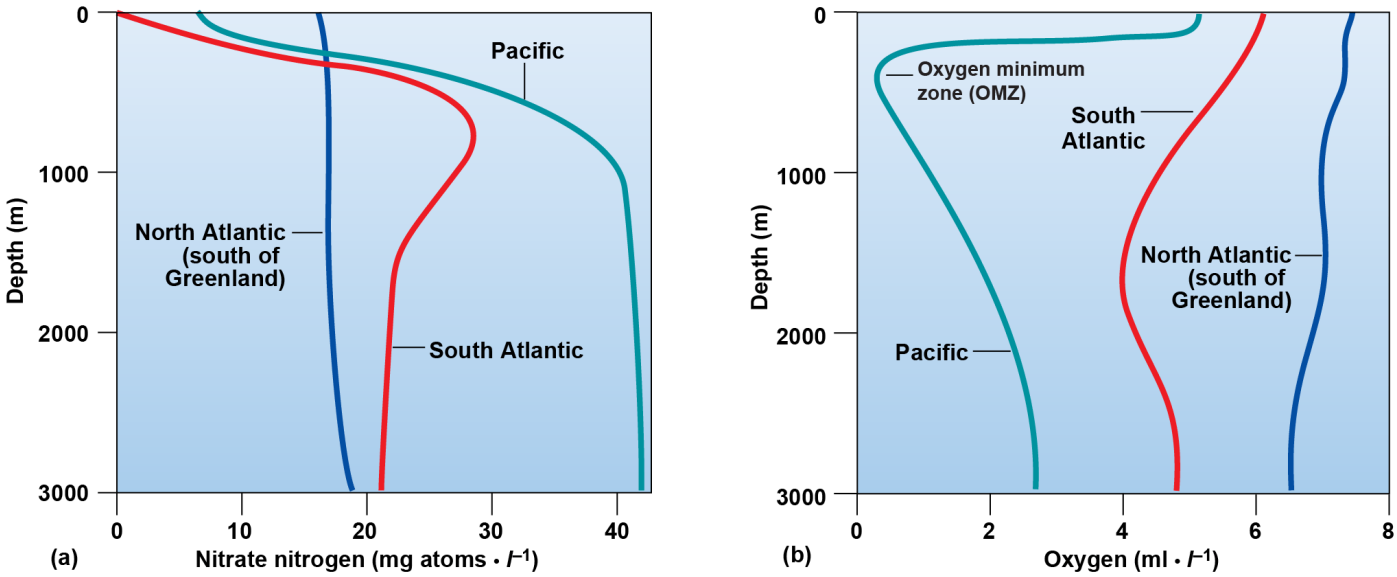


Figure 12-9. The vertical profiles of nitrate nitrogen and dissolved oxygen in the North Atlantic (south of Greenland), South Atlantic, and North Pacific Oceans are very different. In the North Atlantic, where cooled surface water is sinking to form North Atlantic Deep Water (NADW), the concentrations vary little with depth. During its passage southward to the South Atlantic, this water mass loses oxygen and gains dissolved nitrate as respiration by animals and decomposers uses oxygen and releases dissolved nitrate. This process continues as the water mass moves around Antarctica and northward into the deep water of the Pacific Ocean. Thus, the ratio of nitrate concentration to oxygen concentration is an indication of the "age" of the water mass since it left the surface layer. Higher ratios indicate "older" water. The "oldest" water has too little oxygen for animal life to survive in and areas with this low level of oxygen are called oxygen minimum zones (OMZs).

Several additional factors cause concentrations of nutrients to be higher in waters just below the bottom of the thermocline than in bottom waters. First, bacterial decomposition processes are slowed at the low temperatures and high pressures near the deep-ocean floor. Second, the most easily oxidizable organic matter is decomposed long before it reaches the ocean floor. Third, animals that migrate vertically generally do not descend far below the permanent thermocline. However, these influences on the vertical distribution of nutrients, even taken together, are less important than the age of the water and the length of time during which recycled dissolved nutrients have accumulated in it.

The permanent thermocline is a persistent and widely distributed feature. Primary productivity above the thermocline is nutrient-limited except in areas where nutrients are transported into the mixed-layer waters by inputs from the continents or by upwelling of nutrient-rich water from below the thermocline. Major zones of high primary productivity and highly productive fisheries are found in coastal upwelling regions, off the mouths of rivers (which discharge nutrients), and in shallow areas where nutrients in sediments are returned to solution by decomposition and then released directly into the photic zone.

This page titled 12.6: Primary Production and Nutrients is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.7: Food Webs

Heterotrophic species from the smallest microbe to the largest whale depend on autotrophs to produce organic matter for food. Herbivores eat primary producers directly, whereas carnivores eat herbivores and other carnivores.

In the simplified food chain in **Figure 12-10**, photosynthetic organisms are eaten by herbivorous zooplankton, which are eaten by carnivores, which are eaten by larger carnivores, and so on. Each step in this food chain is a **trophic level**. As organisms at one trophic level are consumed by those at the next-higher trophic level, the ingested food biomass is not used entirely to create biomass of the consumer species at the higher trophic level. An average of only about 10% (varying from 1% to 40%) of food consumed at each level is used for growth. The remaining 90% is used during respiration to provide the consumer with energy or is excreted as waste. Clearly, the transfer of food energy in food chains is inefficient (**CC15**).

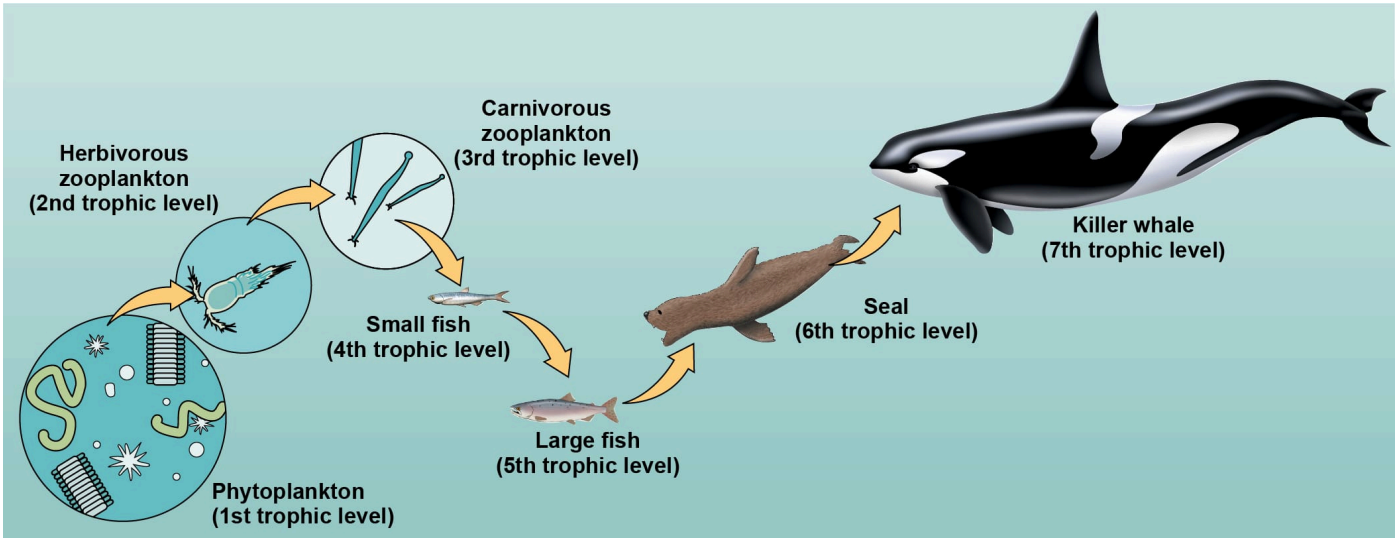


Figure 12-10. A simple food chain. Phytoplankton are eaten by herbivorous zooplankton, which are eaten by carnivorous zooplankton, which are eaten by small fishes, which are eaten by large fishes, which are eaten by seals, which are eaten by killer whales. Each step in this chain is one trophic level higher.

The consequences of food chain inefficiency are illustrated by the food chains in **Figure 12-11**. A good-sized tuna sandwich may contain about 100 g of tuna. For each 100 g of tuna produced, a tuna must eat about 1000 g of a large fish (e.g., mackerel). In turn, mackerel must eat 10 times their weight of smaller fishes (e.g., **herrings**), herrings must eat 10 times their weight of zooplankton (e.g., copepods), and copepods must eat 10 times their weight of phytoplankton (e.g., diatoms). The next time you eat a tuna sandwich, remember that phytoplankton synthesized about 1 million g (1 metric ton, or tonne) of phytoplankton biomass to produce your 100 g of tuna. In contrast, 100 g of sardines represent only 10 kg of primary production (**Fig. 12-11b**). We use ocean food resources 100 times more efficiently when we eat sardines instead of tuna.

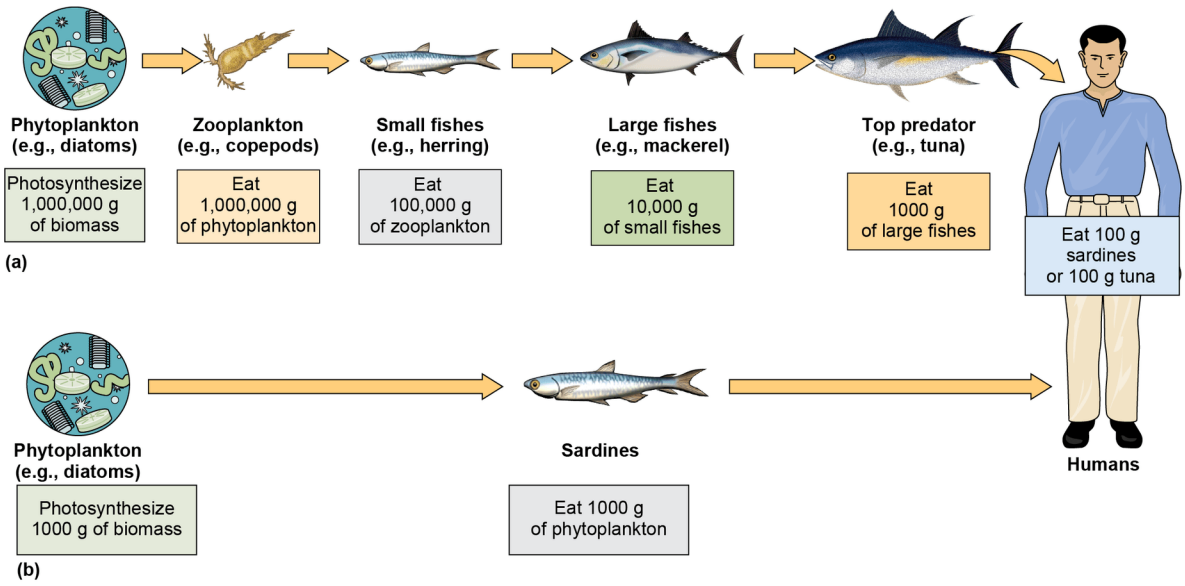


Figure 12-11. The food chains leading to (a) tuna and humans, and (b) sardines and humans, assuming that the trophic efficiency at each trophic level is 10%. Phytoplankton must photosynthesize 1 kg of organic matter to feed the human 100 g of sardines, but must photosynthesize 1000 kg (1 tonne) of organic matter to feed the human 100 g of tuna. However, because sardine species feed on both phytoplankton and zooplankton, their overall trophic efficiency is somewhat less than depicted. In addition, some tuna feed on both small and large fishes, so their overall trophic efficiency is somewhat better than depicted.

Marine animals usually eat organisms that are not much smaller than themselves. Therefore, most large marine animals are high-trophic-level predators at the top of long food chains. **Baleen** whales are spectacular exceptions. The blue whale, the largest known animal, can be 30 m long and weigh more than 100 tonnes (Fig 12-25b). This magnificent creature eats only shrimplike **crustaceans** called **krill** (Fig 12-19b), each of which is only a few centimeters long. Thus, blue whales are at the same trophic level as herrings (Figs. 12-11a, 12-12).

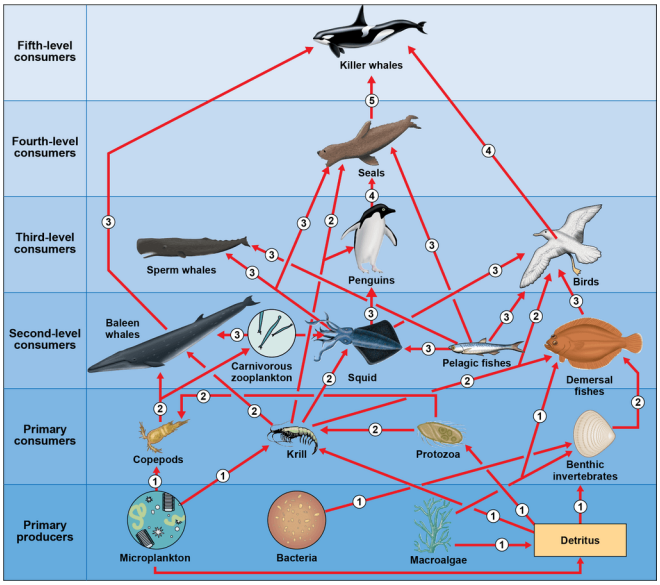


Figure 12-12. Simplified representation of the pelagic food web of the Southern Ocean. The number beside each arrow represents the trophic level at which the consuming organism is feeding. Note that many species feed at more than one level. For simplicity in this diagram, a number of pathways have been left out. For example, pelagic fishes feed on various prey including phytoplankton, zooplankton, and krill.

Food relationships in the ocean are far more complex than the simplified food chains depicted in Figure 12-11 because they almost always include opportunistic carnivores that eat animals from several different trophic levels. The killer whale in the Antarctic marine ecosystem is a good example (Fig. 12-11). Complex food relationships, such as that shown in Figure 12-12, are called **food webs**. Food webs are further complicated by omnivorous species that will eat almost any living organism and sometimes even detritus. Detritus feeders eat particulate organic matter produced as waste products and/or dead tissues of

organisms of all trophic levels. Certain detritus-based food webs are somewhat independent of phytoplankton-based food webs, although they depend ultimately on primary producers to synthesize organic matter that becomes their detrital food.

There are also microbial food loops. By some estimates, about one-half of the primary production in the oceans is performed by microscopically small phototrophic eukaryotes and bacteria. The microbial loop also uses dissolved and particulate organic matter. Heterotrophic bacteria and archaea can absorb this material and utilize it to grow. These organisms may then be consumed by **protozoa** and by other organisms that are themselves consumed by organisms in the higher levels of the food chain. Free-living heterotrophic bacteria in the water column help to recycle elements lost from the food web through the microbial loop, or by releasing them as they consume organic matter for energy and growth.

This page titled 12.7: Food Webs is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.8: Geographic Variation in Primary Production

All marine animals ultimately depend for food on phototrophic or chemosynthetic primary producers, which, in the water column, are mostly phytoplankton. Consequently, zooplankton, fishes, and other **pelagic** animals are most abundant in areas of high primary productivity. Most **benthos** rely on the rain of detritus from above. Exceptions are some species that live in limited areas where the seafloor is shallower than the photic-zone depth and in hydrothermal vents and other chemosynthetic communities. The quantity of detritus depends on the abundance of organisms in the overlying water, so **benthic** communities are also more abundant in areas of high primary productivity.

Phytoplankton biomass (**standing stock**) is determined by phytoplankton growth and reproduction rate versus consumption rate. If the productivity is high and the rate of grazing of phytoplankton by zooplankton is low, phytoplankton biomass increases, and vice versa.

Zooplankton populations and biomass adjust to changes in food supply, but the changes lag days or weeks behind changes in phytoplankton biomass. As a result, phytoplankton biomass initially increases when phytoplankton productivity increases, but it may fall when zooplankton biomass increases, even if there is no change in phytoplankton productivity. Thus, phytoplankton biomass does not vary in concert with phytoplankton productivity. Fortunately, averaged over several months, phytoplankton biomass is reasonably related to the average primary productivity. **Figure 12-13** shows the seasonal variation in the distribution of chlorophyll (**CC14**) in ocean surface waters measured by satellites. The concentration of chlorophyll is a good indicator of the abundance or biomass of phytoplankton and also provides a reasonable approximation of the distribution of primary productivity.

The most productive parts of the ocean are coastal regions (**Fig. 12-13**), particularly along the western margins of continents, where coastal upwelling brings nutrients from deep waters into the photic zone (**Chap. 13**). Most of the open ocean has low productivity, the exceptions being certain high-latitude regions and the equatorial upwelling band across the eastern Pacific and, to a lesser degree, across other oceans (**Fig. 12-13**).

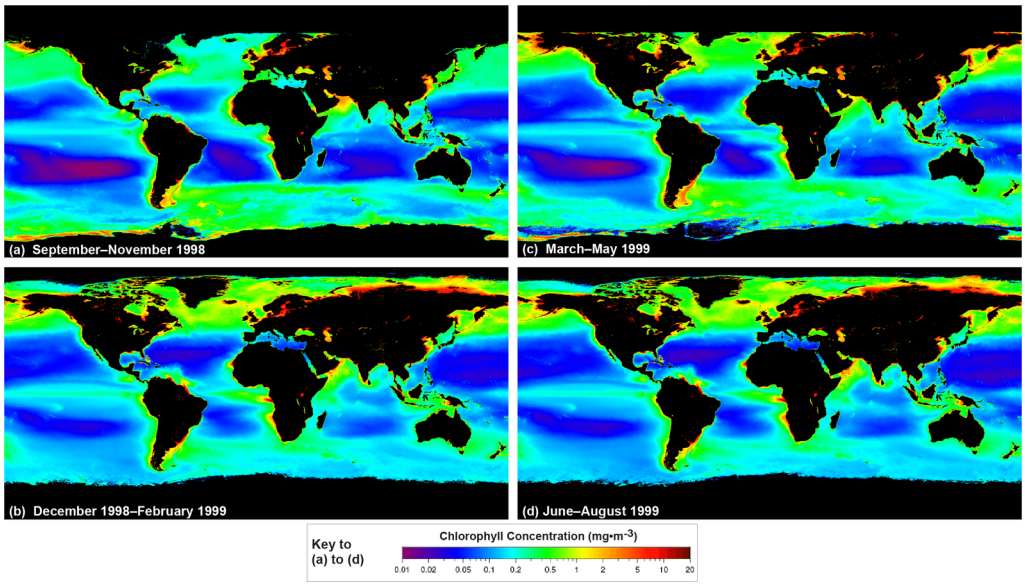


Figure 12-13. Seasonal variations in the distribution of chlorophyll in the surface waters of the world oceans for (a) September to November, (b) December to February, (c) March to May, and (d) June to August. Chlorophyll concentration is a reasonable measure of primary production. The data were obtained by a satellite sensor that measures ocean surface color within several wavelength bands. Black areas are land or where there were no data, because of cloud or ice cover or the limited orbital coverage of the satellite near the poles. Note the high productivity of many coastal regions and of the high-latitude seas in the Northern Hemisphere.

Throughout most of the tropical and subtropical open oceans, a permanent thermocline begins at a depth of about 100 to 200 m. Light intensity is relatively high, and the photic zone extends throughout most or all of the mixed layer above the thermocline. Nutrients are depleted in the photic zone, and the steep thermocline inhibits vertical mixing that would be necessary to resupply nutrients from the nutrient-rich water below the thermocline. High productivity in tropical and subtropical open oceans is limited to areas of upwelling. The band of high productivity across the equator, particularly in the eastern Pacific, coincides with upwelling

at the tropical **convergence** along the equator (**Fig. 8-3**). During **El Niño**, primary productivity is dramatically reduced in this region because upwelling is inhibited (**Fig. 7-19**).

Upwelling at the Antarctic Divergence (**Fig. 8-3**) is responsible for high productivity in the ocean around Antarctica. The Northern Hemisphere has no comparable **divergence**, primarily because of the presence of continents. At high latitudes in the North Atlantic and North Pacific Oceans, productivity is seasonally high because cooling of surface waters, strong **westerly** winds, and **extratropical cyclones** effectively mix surface and subsurface waters during winter, when light levels are low. Nutrients supplied by winter mixing support high productivity when light intensity increases in spring.

Productivity is lowest in the interior of subtropical gyres in each ocean. These regions are remote from nutrient inputs in runoff, have low rainfall (which can carry small amounts of nutrients), and are characterized by downwelling and deep thermoclines (**Fig. 8-13**). Light is plentiful, but lack of nutrients limits phytoplankton growth. Ocean waters are a brilliant blue in these regions because of the lack of suspended particles. However, areas of the Sargasso Sea (the interior of the North Atlantic subtropical gyre) are covered by vast rafts of *Sargassum* seaweed, despite the lack of nutrients (**Chap. 15**).

This page titled 12.8: Geographic Variation in Primary Production is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.9: Dissolved Oxygen and Carbon Dioxide

Oxygen is released to solution during photosynthesis and consumed during respiration. The distribution of dissolved oxygen is controlled by these processes and by exchanges between atmosphere and ocean.

Oxygen is exchanged continuously between the atmosphere and ocean. Hence, dissolved oxygen concentrations in the upper few meters of ocean water almost always equal the **saturation solubility** for the water temperature (**Fig. 5-7**). However, oxygen is produced by photosynthesis much faster than it is consumed by respiration in this layer, particularly when primary productivity is intense. As a result, oxygen concentration increases and the water becomes **supersaturated**. The depth and concentration of the oxygen maximum in the photic zone depend on the depth of maximum primary productivity and the depth and intensity of wind mixing that brings water to the surface, where excess oxygen can be released to the atmosphere. By many estimates, marine photosynthesizers contribute more oxygen to our atmosphere than land plants.

As they release oxygen to ocean waters and atmosphere, phytoplankton also remove dissolved carbon dioxide. Carbon dioxide is highly soluble in seawater as it reacts to form carbonate and bicarbonate ions (**Chap. 5**). The carbon dioxide concentration in seawater is so high that its removal by photosynthesis does not produce a marked minimum concentration of carbon dioxide at the oxygen maximum. Nevertheless, much of the carbon dioxide that primary producers remove from ocean waters is carried down into the deep ocean, where it is removed from contact with the atmosphere for hundreds of years.

It has been estimated that almost half of all carbon dioxide released to the atmosphere by human activity since the Industrial Revolution has entered the oceans. Substantial research is now focused on identifying the fate of this carbon dioxide. Predictions of future **climate** change depend on the rate at which carbon dioxide is transported into the deep ocean. Global climate models are limited by a lack of understanding of this rate and of various other critical ocean processes. For example, we do not know whether and to what degree increased carbon dioxide concentrations might increase ocean primary productivity, or how much of the additional organic matter that might be produced by such an increase would be transported below the thermocline. Proposals have been made to lessen or avoid global climate changes due to the increasing carbon dioxide concentrations in the atmosphere by injecting some of the carbon dioxide into the deep. The effectiveness of placing more carbon dioxide in the deep oceans would depend on the mode and location of introduction and would need to ensure that the carbon dioxide was stored in sediments or rocks as, otherwise ocean circulation would simply return the carbon dioxide to the atmosphere. The likely effectiveness and possible adverse side effects of implementing such carbon dioxide storage in the deep oceans cannot be properly assessed, given the present state of knowledge of the biogeochemical cycles of the oceans.

Organic matter transported below the thermocline as detritus or through animal migration is decomposed, releasing nutrients and carbon dioxide. At the same time, the processes that decompose this organic matter consume dissolved oxygen. The continuous decomposition of organic matter progressively depletes dissolved oxygen beneath the thermocline, where it reaches a minimum. Below the minimum, oxygen increases slightly at greater depths due to the influx of cold, oxygenated deep water masses originating from polar regions. The oxygen concentration minimum below the bottom of the thermocline coincides with the nutrient maximum (**Fig. 12-8**).

At present, oxygen is fully depleted only in limited areas of the oceans, mainly where the **residence time** of subthermocline water is extremely long and/or where mixed-layer primary productivity is exceptionally high. Such conditions are present, for example, in the Baltic Sea, where water residence time is long and primary productivity high because of large **anthropogenic** nutrient inputs. Oxygen depletion also occurs in many fjords, where water residence time is very long. Once dissolved oxygen has been completely depleted, the water is **anoxic**. Most marine species cannot survive in anoxic water. However, certain bacteria and archaea in anoxic water can obtain energy by reducing molecules that contain oxygen. First, bacteria that reduce nitrate (NO^{3-}) to ammonium (NH^{4+}) thrive, and then, when all nitrogen compounds are reduced, they are replaced by other species that reduce sulfate (SO_4^{2-}) to sulfide (S^{2-}). Sulfide is highly toxic and can be released from anoxic bottom waters into the overlying photic zone if vertical mixing is temporarily enhanced. In some cases, water containing hydrogen sulfide reaches the surface, and its foul smell is released to the atmosphere. Anoxic conditions are becoming more common and widespread in some coastal-ocean regions and estuaries because of inputs of nutrients from sewage and agricultural chemicals. In addition, there is a large area of the North Pacific Ocean where there is a strong oxygen minimum zone below the **pycnocline**. Similar strong oxygen minimum zones have also been observed in the Atlantic Ocean off the coast of Southern Africa (Angola and Namibia) and in the Indian Ocean off both coasts of India (Arabian Sea and Bay of Bengal).

There is evidence that the North Pacific oxygen minimum zone is expanding and deepening. Oxygen depleted water from this

water mass now seasonally moves onto the Oregon continental shelf in most years to form a dead zone (**Chap. 13**). Expansion of the oxygen minimum zone in the North Pacific could be related to a slowdown of the ocean conveyor belt circulation (**Chap. 8**) since this would increase the age of this water mass and, therefore, the length of time during which oxygen was consumed by respiration and decomposition. Alternatively, the expansion could be related to an increase in overall ocean primary productivity that would increase the rate of transport of oxidizable organic matter below the thermocline. Both a slowdown of the ocean conveyor belt circulation and an increase in primary production are effects of anthropogenic carbon dioxide and nutrient releases that are predicted by some of the mathematical climate models.

It should also be noted that one proposed approach to mitigation of climate change is that we should fertilize the oceans (primarily with iron in areas where it is the limiting nutrient) so that some of our released carbon dioxide would be transported into the deep oceans as organic detritus. If this were to be done successfully one of the undesirable side effects would be deoxygenation. The oxygen minimum zone would almost certainly deepen and expand further, perhaps eventually causing mass extinctions of ocean species as has occurred in the past when oxygen has become depleted in the deep oceans. In any event, ocean fertilization is a temporary solution at best. The deep oceans would only be a temporary storage location for our excess carbon dioxide since the excess carbon dioxide would be released back to the atmosphere within the next few generations as the deep water was mixed back to the surface.

Anoxic conditions have been more widespread in the oceans in the past. During anoxia, detritus that falls into the anoxic layer is no longer subject to decomposition by **aerobic** respiration, because there is no oxygen. Also, animals and other species that require dissolved oxygen can no longer live in the deep oceans. In such circumstances, organic matter can accumulate in large quantities in sediments. Oil and gas deposits are probably the result of **diagenetic** changes to such sediments over the millennia.

This page titled 12.9: Dissolved Oxygen and Carbon Dioxide is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.10: Organic Carbon

Most organic matter in the oceans is nonliving and exists as both dissolved compounds and particulate matter (detritus). The reason most organic matter is nonliving is that organic matter created by phytoplankton and other marine algae undergoes a series of transfers between organisms, and conversions among physical and chemical forms, before finally being converted back to dissolved inorganic constituents by decomposers.

Under normal conditions, about 10% of the organic carbon that phytoplankton create by photosynthesis is excreted to seawater as dissolved substances. When phytoplankton are stressed by low nutrient or low light levels, nonoptimal temperatures, or other unfavorable factors, they may excrete 50% or more of the organic matter they create by photosynthesis. Although most phytoplankton are eaten before they die, those that are not eaten excrete substantial amounts of dissolved and particulate organic material as they reach the end of the life cycle. In addition, dissolved and particulate organic matter are released when phytoplankton cells are ruptured by **viruses** and during inefficient feeding by herbivores.

The distinction between dissolved and particulate matter is arbitrary. Oceanographers usually consider compounds to be dissolved if they pass through a filter with a pore size of 0.5 μm (human hair is about 100 μm in diameter), although filters used to separate dissolved and organic matter often have pore sizes somewhat larger or smaller than this value. Material that collects on such a filter is called “particulate.” Much of the “dissolved” material that passes through a 0.5-μm filter is not truly dissolved, because it consists of **colloidal**-sized (**CC4**) inorganic or organic particles in addition to living organisms (such as microbacteria and archaea) and viruses. Nevertheless, the distinction between dissolved and particulate is useful. “Particles” less than 0.5 μm in diameter are believed to be too small to be eaten by most species in food chains that lead to higher animals, such as fishes.

Both particulate and dissolved organic matter are produced continuously wherever organisms are present in the oceans. Most organic matter is released in the photic zone, where marine organisms are concentrated. Many organic particles, such as fecal pellets and phytoplankton or zooplankton fragments, are relatively large (ranging from about 1 μm to more than 1 mm in diameter) and sink relatively fast (**Chap. 6**, **CC4**).

Marine bacteria, archaea, and fungi utilize organic particles and dissolved organic matter that is present throughout the ocean depths. These decomposers rapidly break down the more easily oxidized organic compounds in solution and in detritus. As particles are decomposed, smaller particles and more dissolved organic matter are released. Decomposition continues until all organic matter is converted to carbon dioxide and water, which may take many years or even centuries. There are three reasons for the long-term persistence of some dissolved and particulate organic matter. First, many organic compounds are extremely resistant to oxidation. Second, the rate of bacterial decomposition is very slow in the cold temperatures and high pressures of the deep oceans. Finally, dissolved organic molecules and microbial decomposers are so small and well dispersed in seawater that encounters between particles and decomposers are rare events.

The mass of dissolved and particulate matter in the oceans far exceeds the mass of living organisms because dissolved and particulate organic matter is produced continuously but much of it is decomposed only very slowly. In the photic zone, dissolved and particulate nonliving organic matter usually makes up 95% or more of the total organic carbon. An unknown proportion of the “dissolved” organic matter is, in fact, living microbes (including microbacteria and archaea) and viruses. Although the number of these microorganisms is extremely large, they are very small and their total biomass is thought to be only a small fraction of the nonliving organic matter. The declines in total mass of carbon from phytoplankton to zooplankton and from zooplankton to fishes reflect the low efficiency of food transfer between trophic levels (**CC15**).

Relatively little is known about the composition of dissolved or particulate organic matter. These materials are extremely difficult to study because they consist of hundreds of thousands or millions of chemical compounds, each present in extremely small concentrations.

The concentration of particulate organic carbon is generally higher in the upper layers of the oceans than in deep waters. Within the surface layer, the concentration is lowest in areas where phytoplankton productivity is high. This seemingly paradoxical situation is created by the abundant populations of zooplankton and other **filter-feeding** organisms that thrive in productive waters because of the continuously abundant food supply. Particulate organic matter concentrations are high in deep waters below such productive areas because large numbers of fecal particles fall from the abundant filter-feeding organisms.

Organic particles generally account for about one-quarter of the suspended particles in ocean waters. In most areas, the majority of suspended particles are fragments of algae and animal hard parts with lesser amounts of lithogenous particles (**Chap. 6**).

This page titled 12.10: Organic Carbon is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.11: Biological Provinces and Zones

The communities of organisms in different parts of the oceans are as distinct and different from one another as the species that live in tropical rainforests, deserts, and Arctic tundra. The oceans are separated into distinct biological zones delineated by the availability of light, nutrients, and food, and by the temperature and **salinity** characteristics of the water.

Differences between communities that live in or on the seafloor and those that inhabit the water column are so great that the two locations are considered to be separate environments. The seafloor is the benthic environment, and the water column is the pelagic environment. In contrast to pelagic organisms, benthic organisms do not have to swim or control their density and/or size to avoid sinking, which can be energetically advantageous, but they must compete with each other much more intensely for living space and food. Although the pelagic and benthic environments support different communities, many species live part of the life cycle in one environment, then change form (in some species as dramatically as that of caterpillars turning into butterflies) and live in the other stage of **metamorphosis**.

Benthic Environment

The benthic environment is separated into zones by depth. The deepest zone, the **hadal zone** (Fig. 12-14a), which occupies less than 1% of the ocean floor, is limited to the seafloor of deep-ocean **trenches** at depths greater than 6000 m. The seafloor between 2000 and 6000 m is the **abyssal zone**, and corresponds roughly to the **abyssal plains**. The hadal and abyssal zones are generally covered with soft, fine-grained muds. Much of these zones is below the **carbonate compensation depth** (**CCD**; Chap. 6), and the surface muds generally contain little or no calcium carbonate.

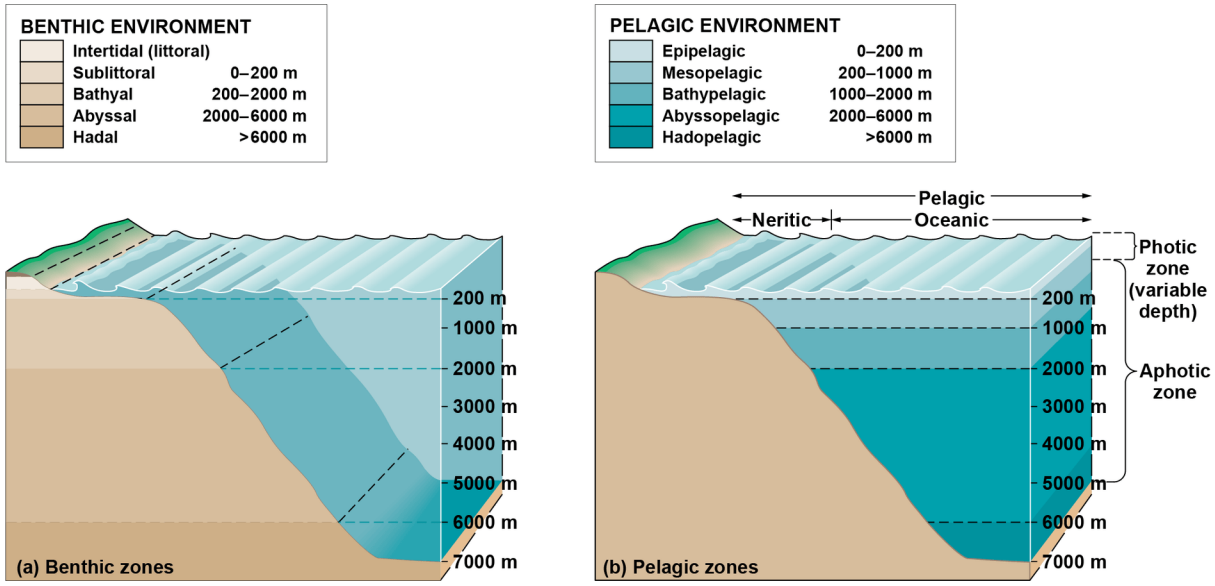


Figure 12-14. The oceans are separated into zones. The assemblage of species in each zone is different from that found in the other zones. (a) The benthic environment is separated into a number of zones, each of which spans a different depth range. (b) The pelagic environment (water column) is separated into a neritic province (water over the continental shelf) and an oceanic province. The oceanic province is divided into several zones that each span a different depth range.

Sediment characteristics are important to the species composition of the benthos. For example, some polychaete worms feed by passing sediment through the gut to digest microbes and organic matter in the sediment, just as earthworms do. Such polychaetes are more successful in soft, fine-grained muds than in coarse-grained sediment, particularly if the muds contain a significant amount of detritus. Where sediments are coarse-grained with little organic matter, many benthic animals live on the sediment surface and obtain food from particles or other animals in the water column. Benthic animals are called **infauna** if they live much or all their life in sediments, and **epifauna** if they live on or attached to the seafloor.

The seafloor environment between 2000 m and the approximate depth of the continental **shelf break** (about 200 m) is the **bathyal zone** (Fig. 12-14a). Because sediments in the bathyal zone are more variable than those in the abyssal and hadal zones,

the benthos is more variable in composition in the bathyal zone.

The hadal and abyssal zones and almost all of the bathyal zone are below the photic zone, so no photosynthetic organisms are present in the benthos. Instead, benthos in these zones are dependent on food transported to them from the overlying photic zone, primarily as detritus. All benthic communities in these zones thus belong to the detrital food web, with some exceptions. One is the unusual benthic communities around hydrothermal vents. These organisms are supported by a chemosynthesis-based food web (**Chap. 15**). A second exception occurs in locations where methane and other hydrocarbons are abundant, and methane based chemosynthesis occurs, generally at seeps of oil and gas (most of which are natural seeps). A third, and minor, exception occurs in the shallowest parts of the bathyal zone in areas of the clearest waters, where light levels are very low but sufficient to support limited photosynthesis by species adapted to very low light levels. Such conditions are generally present only on some **seamounts** far from land.

The benthic environment between the continental shelf break (200 m depth) and the land is divided into three major zones. The **sublittoral zone** (subtidal zone) is the continental shelf floor that is permanently covered by water. It extends from the low-tide line to the depth of the continental shelf break (about 200 m). The **intertidal zone** (**littoral zone**) is the region between the **low-tide line** and the high-tide line and is covered with water during only a part of each tidal cycle. The **supralittoral zone**, often called the "splash zone," is the region above the high-tide line that is covered by water only when large storm waves or **tsunamis** reach the **coast** or during extremely high **tides** or **storm surges**.

Organisms that live in the supralittoral and intertidal zones must endure much more extreme conditions than other marine organisms. Such conditions include exposure to the temperature extremes of the atmosphere, loss of body fluids by evaporation while exposed to the atmosphere, salinity variations due to rainfall, predation by both marine and terrestrial species, and mechanical shock and turbulence created by breaking waves. In intertidal zones with sand or mud sediments, most organisms bury themselves to avoid extreme conditions. This is why **beaches** and mudflats appear to be unpopulated. Most sediment-covered **shores** support an abundance of infauna living a few centimeters beneath the sediment surface. Organisms that live in **rocky intertidal zones** adapt to their changeable environment in other ways (**Chap. 15**).

Pelagic Environment

The pelagic environment consists of the neritic province and the oceanic province. The neritic province is the water column between the sea surface and seafloor in water depths to about 200 m. Thus, it consists of the water overlying the continental shelf and shallow banks (**Fig. 12-14b**). The neritic province experiences greater variability in salinity, temperature, and suspended sediment concentrations than the oceanic province (**Chap. 13**). Further, the mixed layer reaches the seafloor in most of the neritic zone, at least for part of the year, which is important to nutrient cycles (**Chap. 13**).

The oceanic province comprises the entire water column of the open oceans beyond the continental shelf. It consists of several zones, each of which is a water layer distinct from the layer above or below because of differences in salinity, temperature, and light intensity. The **epipelagic zone** extends between the surface and 200 m, the approximate depth where light intensity becomes too low for photosynthesis. Note that this depth is not the same as the shallower (usually less than 100 m) compensation depth, where photosynthesis occurs just rapidly enough to match respiration (**Fig. 12-5**). The epipelagic zone is the only zone in the deep oceans in which food can be produced directly by phototrophy. In all deeper zones, the original source of food is primarily detritus falling from the epipelagic zone. Minor additional food sources include vertically migrating animals and chemosynthesis, especially at hydrothermal vents (**Chap. 15**).

Below the epipelagic zone, between 200 and 1000 m is the mesopelagic zone (**Fig. 12-14b**). Although light intensity is too low to support phototrophy, many organisms in this zone have photoreceptors or eyes adapted to detect faint ambient light or **bioluminescence** (**Chap. 15**). Some of these species migrate upward into the epipelagic zone at night to feed and return to the mesopelagic zone during the day. Below the mesopelagic zone are the bathypelagic zone (1000 to 2000 m), the abyssopelagic zone (2000 to 6000 m), and the hadopelagic zone (>6000 m). These are zones of perpetual darkness, high pressure, and low temperature, and their inhabitants feed on detritus or on each other (**Chap. 15**). The boundary between the bathypelagic and abyssopelagic zones is essentially the boundary between relatively young bottom water masses of the ocean basins and older overlying deep water masses (**Chap. 8**). The hadopelagic zone is restricted to the deep trenches, in which water movements are generally very slow and water residence time is long (**CC8**).

Latitudinal Zones

Most benthic and pelagic environments below the photic zone are relatively uniform at a given depth. Latitudinal variations in

temperature and salinity are less in waters or sediment below the photic zone than in the epipelagic zone and in the pelagic and benthic environments of the continental shelf. Most marine organisms are adapted to live successfully within a narrow range of environmental conditions, particularly a narrow range of temperatures.

Variations in water temperature with latitude (**Figs. 7-14, 7-24**) act as effective barriers to latitudinal dispersal of many marine species. Hence, the marine **biota** of each biological zone varies with latitude (although the variation is small in the abyssal benthos and the abyssopelagic zone). For example, coral reefs are present in warm-water areas in all oceans, but not at cold, high latitudes. The **fauna** and **flora** of the Arctic and Antarctic regions are also very different. Polar bears and most penguin species live only at high latitudes, but polar bears live only in the Arctic region, and most penguin species live only in the Antarctic.

In some cases, continents act as geographic barriers to the distribution of marine species. The tropical Atlantic Ocean is effectively separated from the tropical Pacific and Indian Oceans by the continents. To move between the Atlantic Ocean and either the Pacific Ocean or the Indian Ocean, marine organisms would have to pass around the southern tip of Africa or South America or through the Arctic Ocean. Many tropical species could not tolerate the low water temperatures they would encounter in such journeys. Therefore, almost all tropical Atlantic species are different from those in the tropical Pacific and Indian Oceans. In contrast, the tropical Pacific and Indian Oceans (the Indo-Pacific) have many species in common because they are connected at tropical latitudes.

Tropical Atlantic and Indo-Pacific species, although different, are more closely related than Arctic and Antarctic species. The reason is that the continents have drifted to their present positions in the latter stages of the present **spreading cycle** (**Chap. 4**). Previously, all the oceans were connected at tropical latitudes, and tropical species could move freely around the globe during this relatively recent period in evolutionary history.

This page titled 12.11: Biological Provinces and Zones is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.12: Plankton

The term plankton includes all marine organisms and viruses that do not swim or are very weak swimmers and that do not live on or attached to the seafloor (**Fig. 12-15**). Plankton generally do not settle to the seafloor, and they have very limited or no control of their horizontal movements, so they drift with the ocean currents. Phytoplankton are the phototrophic autotrophs that produce more than 99% of the food used by marine animals. Zooplankton are planktonic herbivores, carnivores, or omnivores.

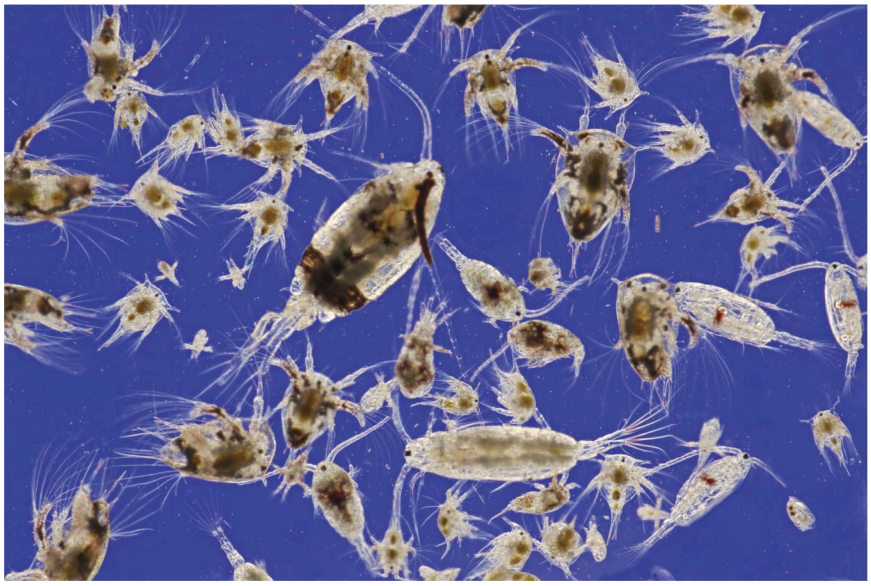


Figure 12-15. A typical plankton sample (photographed at 3× magnification) contains many species of both phytoplankton and zooplankton. This image shows mostly zooplankton.

Plankton are often categorized by size. The largest, which are almost exclusively zooplankton, are the macroplankton (>2 mm). Most plankton are microplankton (20 µm to 2 mm), nanoplankton (5 to 20 µm), or ultraplankton (2 to 5 µm), and these size ranges are dominated by phytoplankton. The smallest plankton are picoplankton (0.2–2 µm), thought to be predominantly bacteria and archaea, and femtoplankton (<0.2 µm), thought to be primarily viruses. Because even the finest mesh nets used by biologists are too coarse to collect the smaller species, they have not been well studied. Less is known about nanoplankton than about microplankton and macroplankton, and very little is known about the even smaller plankton or their **ecological** importance. For this reason, in the following sections we discuss what is known about the microbial part of the food web that consists of bacteria, archaea, and viruses in the nanoplankton size range or smaller and then describe those groups of phytoplankton and zooplankton that are better studied. Most species among these better studied plankton groups are larger than nanoplankton and fall into the macroplankton or microplankton size range. However, it is important to remember that the distinction we make between the very small, less well understood plankton and the larger better-known plankton is a necessity driven by the relative lack of knowledge of the very small plankton. There are autotrophic and heterotrophic plankton in all size ranges and they act together in an ocean ecosystem that is dominated in most areas by microbial communities that are nanoplankton or smaller. As our understanding of these microbes improves (aided by modern techniques such as flow cytometry and genomic sequencing), the long-standing but outdated view of a large-phytoplankton food chain operating independently of a microbial food chain is being replaced by an integrated model of the microbial loop linked with the better known food web of larger species.

The better-known large marine plants and **macroalgae**, including kelp, seaweeds, and **sea grasses**, constitute only a tiny fraction of the photosynthetic species in the oceans and so are not described in this text. Macroalgae normally grow attached to the seafloor and are found on the beach only when they have been broken loose by storm waves or animals.

Bacteria and Archaea

The majority of plankton biomass is now known to be outnumbered by plankton species so tiny that they escaped detection until the 1990s. These poorly studied, microbial plankton are dominated by species of bacteria and archaea. Viruses are also included in this group although viruses are not considered by many to be living organisms. Extremely sophisticated techniques are needed

to isolate and identify bacteria and especially archaea and viruses, but they exist in vast numbers in all ocean water. As stated above, concentrations of bacteria in ocean water average around 1 million per milliliter. Archaea and virus concentrations of millions per milliliter are also believed to be not unusual.

One group of photosynthetic bacteria (also known as cyanobacteria), *Prochlorococcus*, absorbs blue light efficiently at low light intensities, so it can grow throughout the depth of the photic zone. *Prochlorococcus* may be the most abundant component of the phytoplankton, especially in the tropical and subtropical oceans, and it is estimated to contribute 30% to 80% of the total photosynthesis in areas of the oceans where nutrients are scarce and in HNLC areas. Until recently it was thought that archaea existed primarily in extreme environments, but it is now known that archaeal species are found throughout the ocean environment and that some are phototrophic primary producers

We are only just beginning to study the microbial populations of the oceans and the role that they play in life on our planet. However, what we do know is that these microbes are responsible for regenerating dissolved nutrients and metals previously removed from solution by living organisms. Thus, without microbes, the ocean productivity and the entire ocean food web would quickly stop entirely.

Although there are millions of microbes in each milliliter of seawater, which might suggest that ocean water is a thick biological soup, in reality microbes are so small that their environment is 99.99999% water or dissolved ions. As a result, from the microbe's perspective life consists of being separated from its neighbors by what to it are long distances (hundreds of body lengths), and of waiting around for an occasional encounter with an organic particle or an occasional plume of nutrient from the waste liquids of larger organisms. In response to this, microbial bacteria and archaea have developed the ability to slow their metabolism, sometimes even essentially stop their metabolism and form a resting phase until they encounter food or the chemical energy source they need to grow and divide. Microbes have also developed abilities to rapidly adapt to changing opportunities and environments. They do so by transferring genes among themselves. Although this particular mechanism has not been fully demonstrated yet, the concept can be illustrated by the example that a species adapted to live in deep water might be able to pick up the gene for photosynthesis if it is transported into the upper layers where there is light. Interestingly much of this gene transfer may take place through virus-like gene-containing fragments of the host species chromosome. We do not know what all of the roles are that viruses play in ocean ecosystems. Mounting evidence suggests that they cause diseases in marine organisms spanning the entire range from the smallest bacteria to the largest whales but likely they also have a major role in gene transfer and adaptation in bacteria and archaea and even in animals and plants and they are now thought to have been the ultimate ancestors of all other life.

The discovery of the abundance and central role of bacteria and archaea, together with the discovery that viruses are even more abundant throughout the oceans and perform essential roles in the microbial effects on biogeochemical cycles and likely also in evolutionary processes, has revolutionized ocean sciences. Extensive research efforts are currently directed toward gathering a better understanding of these organisms and viruses and details of their role in ocean ecosystems. A billion or more years ago, the microbial food web in the oceans became a self-sustaining community of organisms, likely even before photosynthesis developed. The non-microbial world, including all eukaryotes, are late comers on the planet and are completely dependent on the microbes.

Phytoplankton

There are only three known planktonic species of macroalgae, all of the genus *Sargassum*. *Sargassum* grows in dense rafts, often many square kilometers in area, floating at the surface of the Sargasso Sea (**Chap. 15**).

Phytoplankton are generally much smaller than 1 mm in diameter—mostly too small to be seen clearly by the naked eye. The oceans contain an abundance of phytoplankton dispersed throughout the surface waters at concentrations that may exceed 1 billion individuals per liter. There are tens of thousands of species of phytoplankton. A sampled phytoplankton assemblage always consists of many species, but in any individual sample one species is often dominant and far outnumbers all others (as pine trees do in a pine forest).

Phytoplankton communities consist of different species in different climatic regions, and concentrations range from very low in some areas (equivalent to deserts) to very high in others (equivalent to rain forests). Unlike most land plant communities, phytoplankton communities can vary dramatically in composition and concentration within hours or days. Such variability occurs in part because phytoplankton are often concentrated in patches (tens to hundreds of meters across) that drift with ocean currents. Patches of phytoplankton can develop when rapid reproduction exceeds the rate of dispersal by mixing processes. Because phytoplankton can double their population within a day, or in even less time under favorable conditions, patches may develop quickly in calm seas. Phytoplankton are also concentrated by **Langmuir circulation** (**Chap. 8**). Smaller or less dense cells

tend to concentrate at the surface convergence between Langmuir cells. Larger or higher-density cells, which tend to sink, may be concentrated below the surface where subsurface Langmuir cell currents converge (beneath surface divergences). Phytoplankton can also be concentrated within mesoscale eddies.

Individual phytoplankton species respond favorably to slightly different light intensity levels, temperatures, and nutrient concentrations. Under favorable conditions, one or more species can reproduce rapidly and become the dominant species. If conditions change and another phytoplankton species prospers in the changed conditions, the second species can become dominant within a few days because grazing zooplankton can rapidly remove the previously dominant species.

Diatoms

Diatoms (**Fig. 12-16a**) are among the most abundant types of phytoplankton in many of the more productive areas of the oceans. They are relatively large single cells (up to about 1 mm in diameter) with a hard, organically coated external **siliceous** casing called a "frustule," which is made of two halves much like a pillbox. The frustule is porous, allowing dissolved substances to diffuse through it and to be taken into or excreted from the cell. Because silica is denser than seawater, most diatoms contain a tiny droplet of a lighter-than-water natural oil to reduce their density and thus their sinking rate. Many diatoms have protruding, threadlike appendages and may link to form chains (**Fig. 12-16a**). These adaptations also reduce the cell's tendency to sink and may decrease predation by small zooplankton. The threadlike appendages also increase the surface area over which nutrients can be taken up and light collected for photosynthesis.

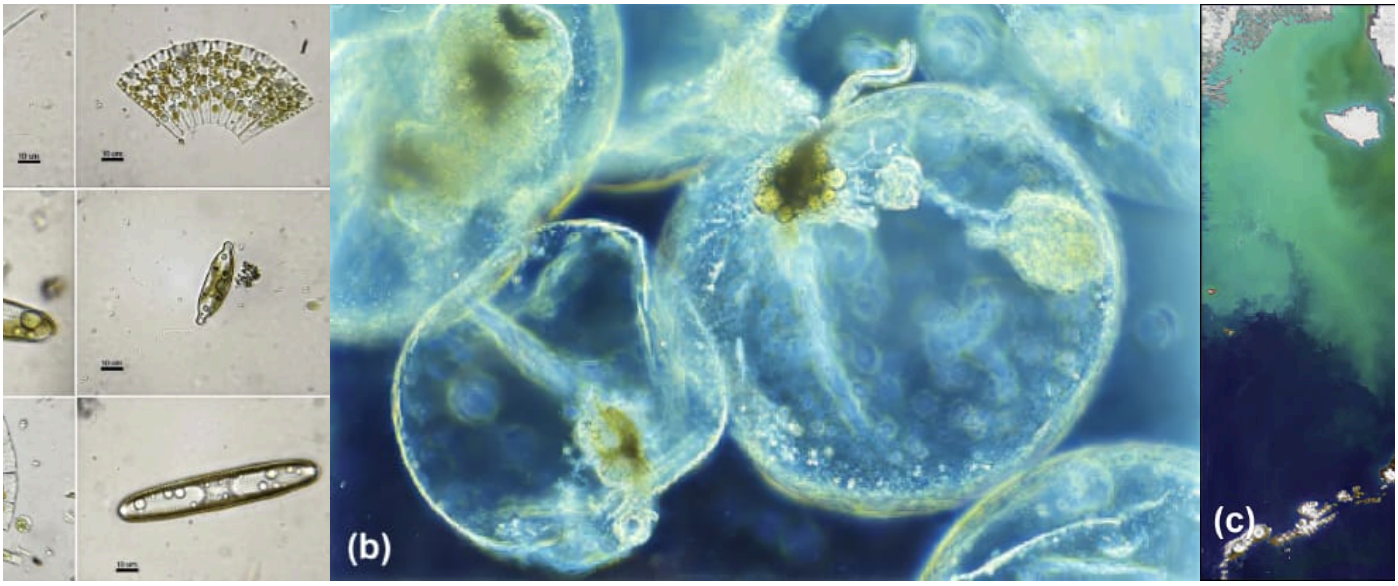


Figure 12-16. There are many varieties and species of phytoplankton (a) Various species of diatoms - approximately 20x magnification. (b) Bioluminescent dinoflagellates (*Noctiluca scintillans*, each individual is about 200-2000 micrometers across. (c) A bloom of coccolithophores seen from space. Since each individual coccolithophore is only about 40 micrometers across (about 100 times too small to see with the naked eye) there must be an almost unimaginable number of individuals that make up this bloom.

Their oil droplet and relatively large size make diatoms a favored food source for many species of juvenile fishes and zooplankton. Consequently, food chains based on diatoms are generally shorter than those based on smaller classes of phytoplankton. Smaller phytoplankton must usually be eaten by small zooplankton before they can provide food for juvenile fishes and larger herbivorous and omnivorous zooplankton.

Diatoms reproduce asexually by cell division. At each division, which may follow the previous one by less than a day, the frustule separates, with one-half taken by each of the two new cells, and each cell then manufactures a new half. Consequently, one of the new cells is smaller than its parent (**Fig. 12-17**). After several such divisions, the now much smaller cells may reproduce sexually, restoring the cell size to its maximum.

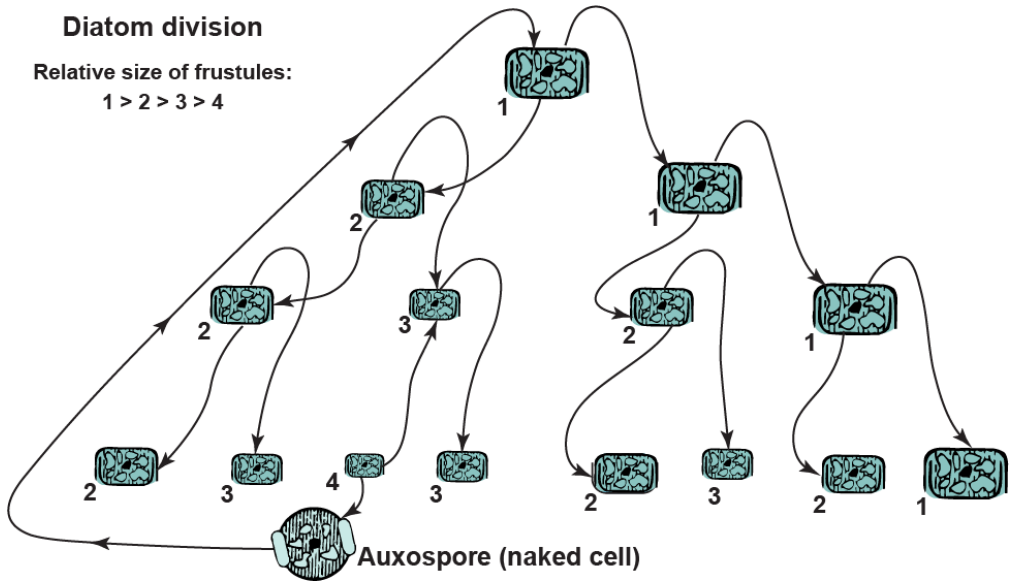


Figure 12-17. Most diatoms reproduce primarily by cell division. At each division, the two halves of the diatom frustule separate, and each half secretes a new half to its frustule. However, because one of the new halves is smaller than its parents, the mean size of the diatom is reduced with progressive divisions. Eventually, a small diatom reproduces sexually by producing an auxospore that at first has no frustule. After some period of growth, the auxospore secretes a new frustule.

Dinoflagellates

Dinoflagellates range widely in size, but many are nanoplankton, smaller than most species of diatoms (**Fig. 12-16b**). Most species of dinoflagellates have two hairlike projections, called “flagella,” that they use in whiplike motions to provide a limited propulsion ability. Some dinoflagellate species use this propulsion to migrate vertically. This allows them to stay at a depth where light levels are best for photosynthesis during the day, or move deeper to reach higher concentrations of nutrients that may be unavailable at the surface. Because of this "best of both worlds" strategy, certain species tend to concentrate at a specific depth within the photic zone.

Many dinoflagellate species have an armored external cell wall made of cellulose, but others are “naked.” Because cellulose is decomposed relatively easily by bacteria and other decomposers, dinoflagellates do not contribute significantly to deep-ocean bottom sediments. Dinoflagellates are not always autotrophs. Some species are mixotrophic and are able to use dissolved or particulate organic matter as food, and some are predators. Indeed, many species can live and grow both autotrophically and heterotrophically.

Dinoflagellates are more abundant than diatoms in the open oceans far from land because the silica needed to construct diatom frustules is in short supply. However, phytoplankton biomass is substantially lower in most such open-ocean areas (**Fig. 12-18a**). Silica can also be scarce in coastal waters at certain times of year. These temporary silica-deficient conditions can lead to explosive blooms of dinoflagellates if other nutrient, temperature, and light conditions permit (**Chap. 13**).

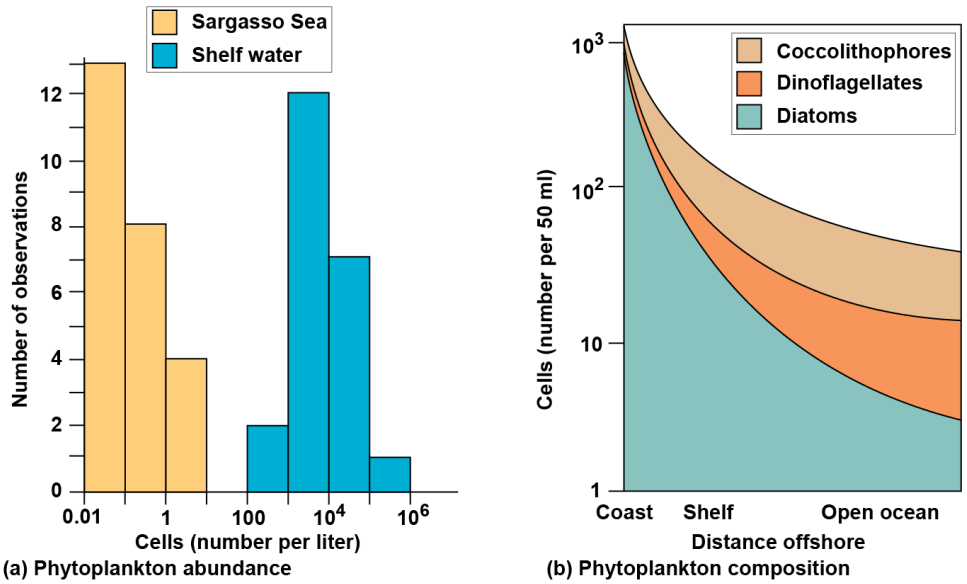


Figure 12-18. The composition of the phytoplankton community is different in coastal, offshore, and open ocean waters. (a) Phytoplankton are much more abundant (note the logarithmic scale) in the surface layer waters of the continental shelf than they are in the Sargasso Sea, where nutrients are depleted. (b) In many areas, the relative proportion of diatoms generally decreases with distance offshore, reflecting the lower availability of dissolved silicate in the offshore waters. These data were obtained from the Caribbean Sea.

Coccolithophores and Other Types of Phytoplankton

Coccolithophores (Fig. 12-16c) are generally nanoplankton, and they are smaller and less abundant than diatoms and most dinoflagellates. Coccolithophores are single-celled **flagellates** whose external cell surface is covered by a mosaic of tiny **calcareous** plates (Fig. 12-16c). In certain areas, these plates are a major component of seafloor sediments (Chap. 6). Coccolithophores make up a greater fraction of the phytoplankton biomass in relatively nonproductive temperate and tropical open ocean waters than they do in coastal waters.

In most of the oceans, phytoplankton consist primarily of diatoms, dinoflagellates, and coccolithophores (Fig. 12-16). However, there are several other types of phytoplankton, including silicoflagellates (which have an intricate silica shell), cryptomonads, chrysomonads, green algae, and **cyanobacteria** (**bluegreen algae**).

Zooplankton

Zooplankton are animals that do not swim strongly enough to overcome currents and so drift with the ocean water. As is the case for phytoplankton, and for similar reasons, the distribution of zooplankton is patchy. Dense patches of zooplankton attract fishes and other predators. In contrast, **nekton** are marine animals that swim strongly enough to move independently of ocean currents, are distributed nonuniformly because they are attracted to food sources, and they often exhibit **schooling** behavior, which is discussed in Chapter 14. The nonuniform distribution of both plankton and nekton, together with the temporal variability of their populations, makes it difficult for biologists to obtain precise population estimates, even when large numbers of samples are taken.

Zooplankton consist of a bewildering array of species from many different groups of organisms. Many zooplankton species tolerate only narrow ranges of environmental conditions, especially temperature. Consequently, zooplankton species composition changes from one water mass to another and with depth. At any one location, many species are represented, including species that are bacteriovores (bacteria eaters), herbivores, carnivores, and omnivores.

Zooplankton belong to one of two categories based on the life history of the species. Species that live their entire life cycles as plankton are known as **holoplankton**. The **larvae** (juvenile stages) of species that later become free swimmers or benthic species are **meroplankton**. Meroplankton include many species of fishes, **sea stars**, crabs, oysters, clams, **barnacles**, and other **invertebrates**. Holoplankton are the dominant zooplankton in surface ocean waters, whereas meroplankton are more numerous in continental shelf and coastal waters. In tropical coastal waters, larvae of benthic species make up as much as 80% of all zooplankton.

Many zooplankton species tend to concentrate at the same depth and collectively migrate between the photic zone and aphotic zone each day, but the depth to which this diel vertical migration takes place differs depending on the species. Zooplankton also tend to collect at density interfaces between water layers because these interfaces inhibit (but do not prevent) vertical migration and sinking and thus collect food particles. When zooplankton are present in large numbers within a thin layer below the surface, they scatter or reflect sound and are observed by echo sounders as a "deep scattering layer." This layer changes depth during the day as the zooplankton make their daily migration between the photic and aphotic zones.

There are too many important species of zooplankton to describe in this text, but the characteristics of the major groups of holoplankton and meroplankton are described briefly in the sections that follow.

Holoplankton

The most abundant holoplankton are copepods (**Fig. 12-19a**), **euphausiids** (**Fig. 12-19b**), and **amphipods** (**Fig. 12-19c**), which constitute 60% to 70% of all zooplankton in most locations. Copepods and euphausiids are both crustaceans, a class of invertebrates that includes crabs and lobsters. In the open oceans, most copepod species are herbivorous, whereas many coastal forms are omnivores. Copepods eat about half their body weight in phytoplankton or other food each day. They are abundant throughout the oceans and can double their population within a few weeks. Euphausiids are generally larger and reproduce more slowly than copepods. Euphausiid population doubling times are typically several months. Many euphausiids are omnivorous, eating smaller zooplankton, as well as their major food, phytoplankton.

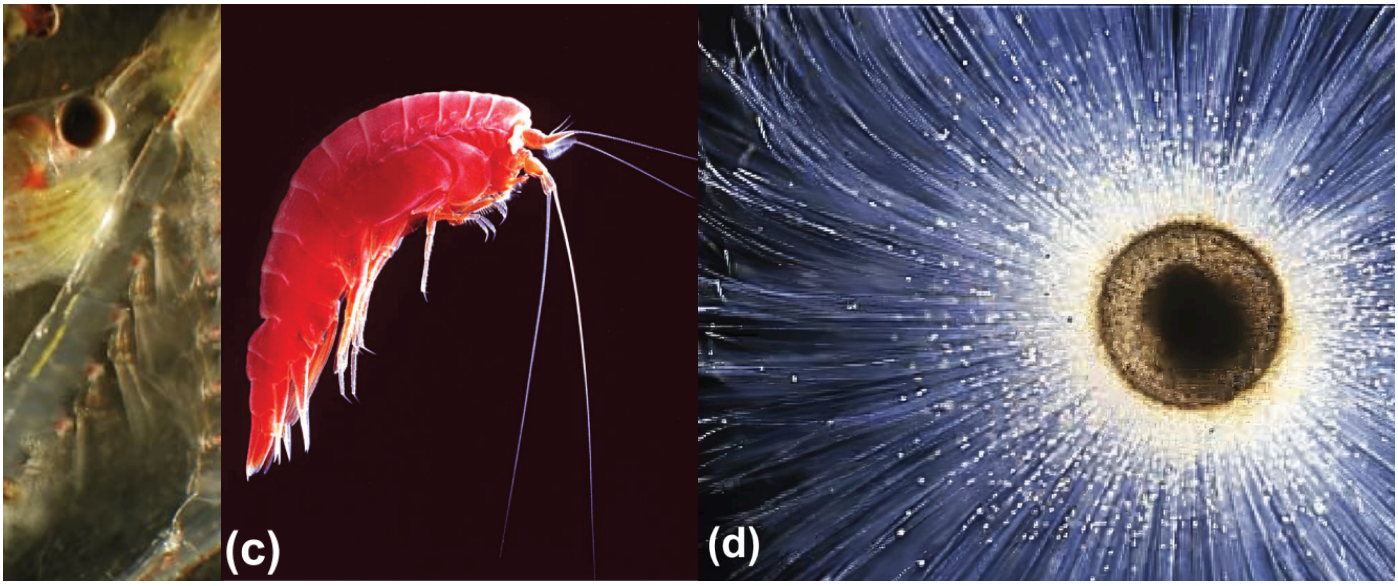


Figure 12-19. Typical holoplankton. (a) Copepods such as this *Scaphocalanus acrocephalus* are extremely abundant in some regions of high primary productivity. (b) Euphausiids. (c) Amphipods (possibly *Ampeliscidae*) are a major food source for fishes (d) Foraminifera *Orbulina universa*. Only the spherical inner shell will remain in sediments. The central spherical chamber is about 0.5 mm diameter and the many bright white spots are symbiotic dinoflagellates. (e) Radiolarian (class probably *Acantharea*), northeast Pacific Ocean, approx 160x magnification, cell body approx 50 µm across.

Euphausiids called "krill" are especially abundant in waters around Antarctica, and they constitute the principal food source of the abundant marine animals there. Baleen whales, including the blue, humpback, sei, and finback, feed directly on krill. These baleen whales gulp large volumes of water, then squeeze it out through net-like baleen plates suspended from the roofs of their mouths. Krill collect on the baleen and are removed by the tongue and ingested.

Two groups of single-celled **amoeba**-like microplankton have hard parts that become important components of sediment in some areas: **foraminifera** (**Fig. 12-19d**), with shells composed of calcium carbonate, and radiolaria (**Fig. 12-19e**), with shells composed of silica. Both are holoplankton and feed on diatoms, small protozoa, and bacteria, often capturing them on their many long, sticky, spikelike projections. Radiolaria and foraminifera are most abundant in warm waters. Individual species, especially of foraminifera, are very sensitive to small changes in water temperature and salinity. Because of the sensitivity to temperature, the species compositions of radiolaria and foraminifera in sediments are important indicators of past climates. **Isotopic** compositions of the shells of these organisms also provide a record of the temperature at the time they lived (**Chap. 6**).

The **pteropods** (Fig. 6-8b), another group of holoplankton, are also important in marine sediments. Pteropods are **mollusks** and are related to slugs and snails. In pteropods, the "foot" on which slugs or snails crawl is modified into a delicate transparent wing that undulates and propels the organism like a fin. This modified foot enables pteropods to migrate vertically hundreds of meters each day. Some pteropod species are carnivorous and do not have a shell. Others are herbivorous and have a calcareous shell that contributes to sediment, especially in tropical regions (Chap. 6), where they often occur in dense swarms.

Gelatinous Holoplankton

Many holoplankton differ from other holoplankton species because they have gelatinous bodies and are apparently not a major part of the food webs that lead to fishes and other marine animals exploited by humans. Although these species consume large amounts of other zooplankton, their gelatinous bodies provide little or no food for species at higher trophic levels.

The most familiar group of gelatinous holoplankton, the **jellies** (commonly but incorrectly called **jellyfish** or sea jellies but they are not fish and are not restricted to seawater), are **cnidarians** (phylum Cnidaria), a phylum which also include **corals** and **anemones**. All cnidarians have stinging cells, called "cnidocysts," within which they have harpoon-like structures called "nematocysts" that can fire into their prey to inject toxins. In some species, the toxins are extremely strong and can paralyze or kill large fishes or even people.

Some jelly species (Fig. 12-20a-c) are very large in comparison with most other holoplankton, perhaps because their food value is low and they are the preferred food source for relatively few predators. Some, such as the moon jelly *Aurelia* (Fig. 12-20a) and *Cyanea*, are holoplankton. Others are meroplankton that spend part of their lives in the plankton, then settle to the seafloor, where they attach with their stinging tentacles extended upward. In the benthos they resemble their close relatives: anemones and corals.

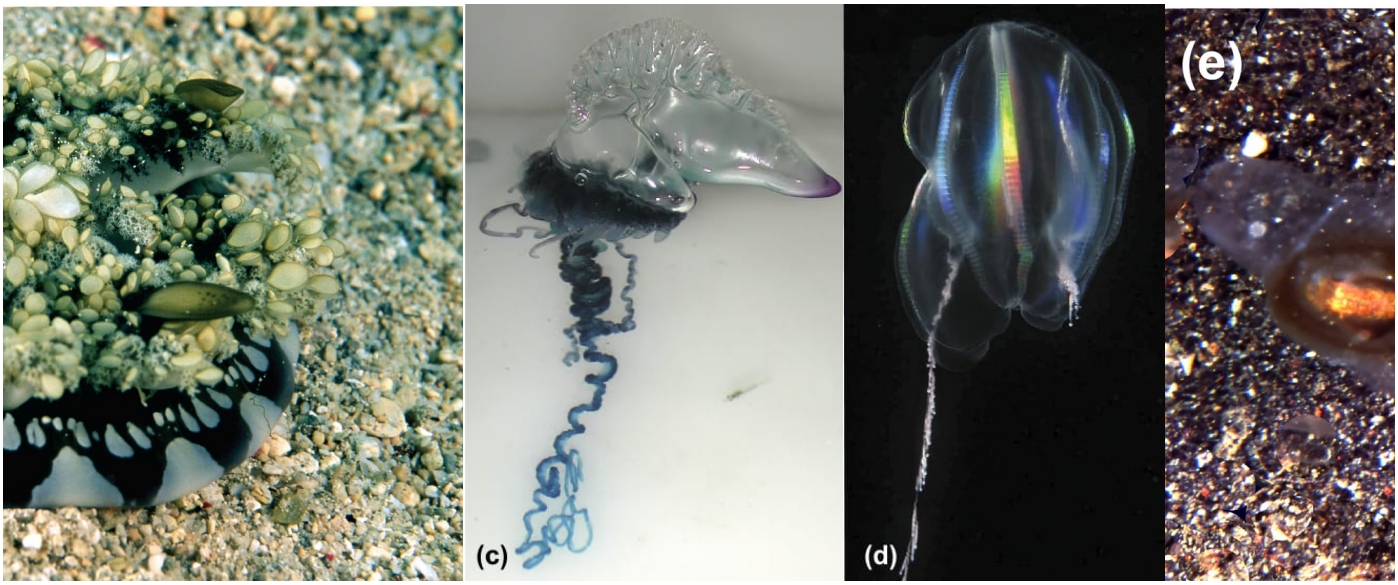


Figure 12-20. Typical gelatinous holoplankton. (a) Jelly *Aurelia sp.* This one was in a unique marine lake called Jellyfish Lake in Palau. (b) This upside-down jelly (*Cassiopeia andromeda*, Papua New Guinea, about 10–15 cm across) generally lies on its "back" on the shallow seafloor to provide light to the symbiotic algae that live in its tentacles. (c) The Portuguese man-of-war (*Physalia physalis*) is a colonial jelly whose long tentacles bear stinging cells that can severely injure a swimmer. (d) Ctenophores, like this *Mertensia ovum*, propel themselves by pulsating the columns of bright hairlike cilia visible through the body. (e) Each individual of this colonial salp species, photographed after it had floated into shallow water and settled on the sand in Indonesia, may be 10 cm or more across. The colonies, which normally drift through open-ocean surface waters, may exceed several meters in length.

Among the most unusual jellies are species, such as the Portuguese man-of-war (Fig. 12-20c), that appear to be a single organism but in fact are a colony. In colonial forms, many individuals of the same species form a cooperative group that appears to be a single organism. Each colony member has a specialized task, such as protecting the colony, gathering food, digesting food, or reproducing. In the Portuguese man-of-war, one colony member is filled with gas to provide flotation and a "sail" that can partially control the colony's drift.

Widely occurring gelatinous plankton that are not jellies include **ctenophores** and **salps**. Ctenophores are transparent,

bioluminescent organisms, some of which have long, trailing tentacles (**Fig. 12-20d**). Ctenophores propel themselves through the water by beating eight columns of hairlike cilia that are usually visible through the ctenophore's body (**Fig. 12-20d**) and that give these species their common name, "comb jellies." Small, rounded species of ctenophores are often called "sea walnuts" or "sea gooseberries."

Salps (**Fig. 12-20e**) are the holoplankton species of **tunicates**, most of which are benthos. Tunicates are among the most advanced invertebrates. They are **chordates** (phylum Chordata), a phylum which also includes the **vertebrates** (fishes and mammals). This close relationship is difficult to envision from the simple form of the adult salps or other tunicates, but tunicate larvae closely resemble vertebrate larvae because they possess a primitive backbone called a notochord. Adult tunicates have a simple baglike form with two openings. Water is pumped into one opening (**incurrent** opening) and out the other (**excurrent** opening). Food particles are removed from the water by a mucous layer spread over the interior of the tunicate's body.

Many salps are hollow and barrel-shaped, with incurrent and excurrent openings at opposite ends (**Fig. 12-20e**). They propel themselves slowly by pumping water through their bodies. Several salp species are bioluminescent. I witnessed a "magic moment" one moonless night on a research vessel sailing through an exceptionally dense patch of salps in the tropical Atlantic Ocean. For miles, the ship's wake was a brilliant, sparkling light show, as a continuous stream of salps was disturbed by the wake and emitted pulses of light.

Meroplankton

Meroplankton are the eggs, larvae, and juveniles of species that spend their adult lives as benthos or nekton. Larvae or juvenile forms of the majority of benthic species, including clams, oysters, crabs, snails, lobsters, sea stars, **sea urchins**, corals, and **sea cucumbers**, spend their first few weeks of life as meroplankton. Many fish species also release eggs to the water column. These eggs, the larvae that hatch from them, and juvenile fishes that emerge from the larvae are meroplankton until the fishes become big enough to swim actively against currents as nekton. The release of eggs and larvae as meroplankton is an effective means of distributing species over wide areas (**Chap. 14**).

Many species of meroplankton larvae do not remotely resemble their adult forms (**Fig. 12-21**). In the past, numerous meroplankton have been named and classified as separate species, even though their adult forms were already well known. Modern scientific methods, particularly genetic studies, now enable us to match meroplankton with their adult forms.

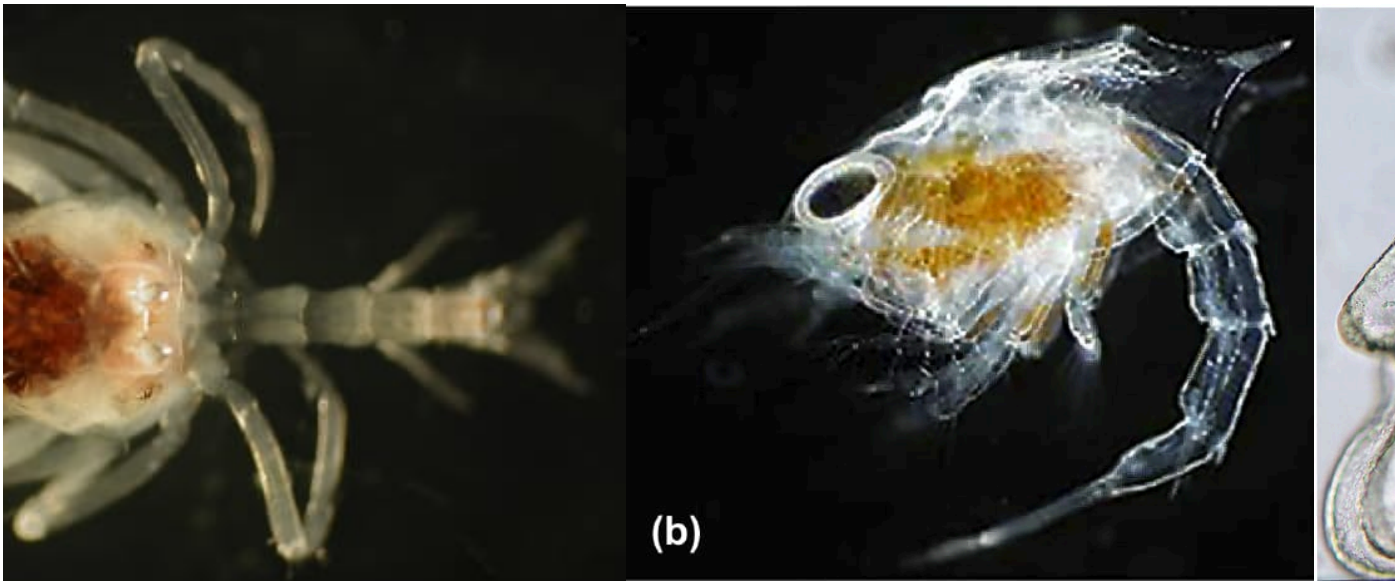


Figure 12-21. Typical meroplankton. (a) A crustacean larvae. (b) A blue king crab larvae, zoeal stage 2 (*Paralithodes platypus*) (c) The first larval stage of a starfish. This stage is called a bipinnaria.

This page titled 12.12: Plankton is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.13: Nekton

Nekton are animals that swim strongly enough to move independently of ocean currents. Most nekton are vertebrates, which are organisms that have an internal skeleton, spinal cord, and brain. Marine vertebrates include fishes, **reptiles**, birds, and **marine mammals**. Some invertebrates, such as squid, are also nekton. The nekton are dominated by the varied and abundant fish species. There are three types of fishes: primitive jawless fishes including hagfishes and lampreys; fishes that have cartilage skeletons, including sharks and rays; and bony fishes that have bone (largely calcium carbonate and calcium phosphate) skeletons. More than 95% of all living species of fishes are bony fishes.

Bony Fishes

Bony nektonic fishes have a wide variety of body forms and range in length from just a few centimeters to more than a meter. Such pelagic giants as tuna (**Fig. 12-22a**) or swordfish can exceed 3 m. Species of fishes that live on or near the seafloor are called "demersal" fishes, and those living predominantly in the water column are called "pelagic" fishes.

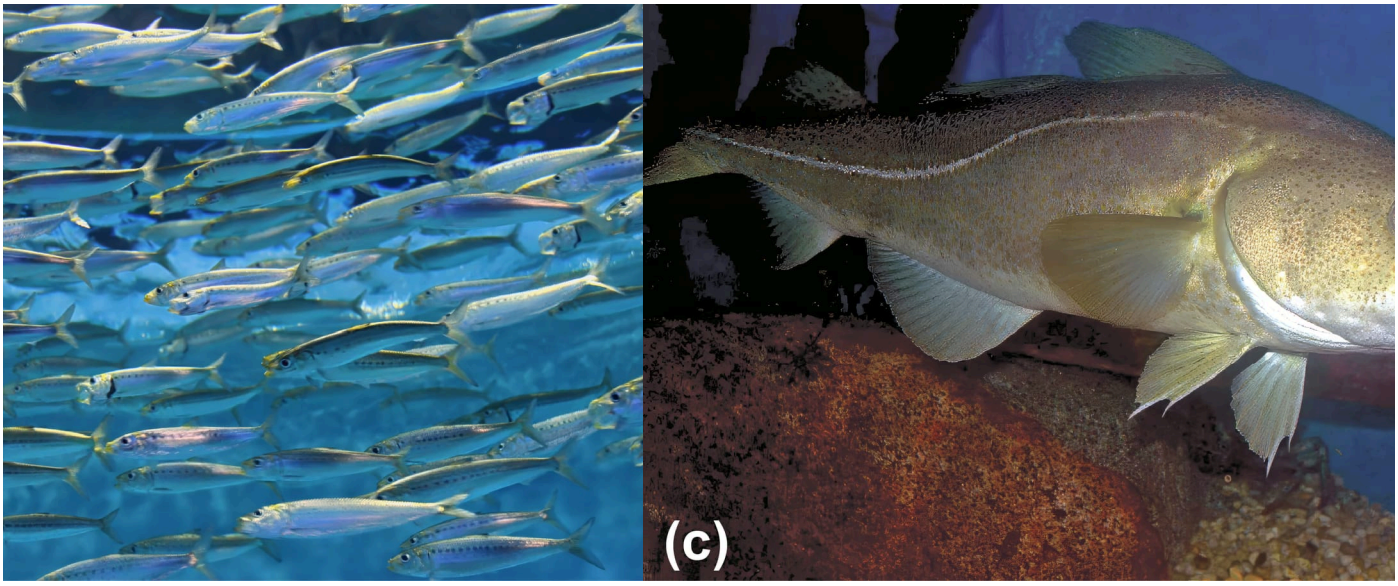


Figure 12-22. Representative species of fishes. (a) The dogtooth tuna (*Gymnosarda unicolor*, Papua New Guinea) is an oceanic predator. (b) A school of Pacific sardines (*Sardinops sagax*). (c) The Atlantic cod (*Gadus morrhua*) was once extremely abundant, but its populations have now been substantially reduced by overfishing. (d) The flowery flounder (*Bothus mancus*, Solomon Islands) spends much of its life lying camouflaged on the seafloor.

The most abundant fishes in coastal waters are various species of silver-bodied schooling fishes including herrings, anchovies, pilchards, sardines (**Fig. 12-22b**), and menhaden. Because these fishes generally do not exceed a few tens of centimeters in length and have oily tissues, they were historically undervalued in some cultures, especially for their use as fresh fish in the U.S. and Great Britain. However they are now highly valued for human consumption and industrial uses. Their numbers are so vast that the world's most productive commercial fisheries target these species. Much of the catch is processed and used as animal feed, fish oils for paints and diet supplements, soaps, and lubricants. These species have each been **overfished**, and their historical populations have been decimated in many coastal regions (**Chap. 16**). Various species of cod (**Fig. 12-22c**) are extremely abundant in middle and high latitudes, especially in the Northern Hemisphere, and they are commercially very valuable. Cod are demersal, living most of their lives on or near the ocean floor, where they feed on invertebrates and small fishes.

The body shapes of fishes are determined primarily by their swimming habits (**Chap. 14**). Species that are members of the flatfish group have bodies modified to enable them to lie flat on the seafloor, where they remain well camouflaged (**Fig. 12-22d**). Juvenile flatfishes, like other fishes, have eyes on each side of their body. As they mature, one eye migrates across the body so that both eyes are on the same side of the head (**Fig. 12-22d**). If the eye migrates to the left side of the body, the species is usually called a "flounder." If it migrates to the right side, the species is usually called a "sole." However, some species, such as the starry flounder, have both left- and right-eyed forms.

Many fishes contain **swim bladders** filled with a gas that is usually a mixture of oxygen and carbon dioxide. Swim bladders are used to maintain buoyancy, and, at least in some species, they appear to be used in the fish's "hearing" mechanism. Because gas expands or contracts as pressure changes with depth, fishes with swim bladders must add or remove gas from their swim bladders as they move vertically. Some fishes absorb the gas back into the bloodstream, and others, such as herrings, can release the gas as bubbles. Many fishes bloat and die if brought rapidly to the surface because they cannot purge their swim bladders fast enough. Fish species that migrate vertically each day often have no swim bladders, but maintain buoyancy through high concentrations of lighter-than-water oils.

Sharks and Rays

Sharks and rays are more primitive than other fishes, and many species have existed essentially unchanged for millions of years. Unlike bony fishes, sharks and rays have skeletons made of cartilage. Sharks and rays have reproductive cycles that require eggs to be fertilized inside the female. Most other fish species reproduce by the simultaneous release of eggs and sperm by female and male. In many **cartilaginous** fish species, the fertilized eggs are hatched and develop into the adult form within the mother before being released in a live birth.

The common perception of sharks is that they are fearsome predators that will hunt and kill anything, including people (**Fig. 12-23a**). Although many species are able to hunt and kill large fishes and marine mammals, only a few species are known to attack people. Because humans are not the natural prey of sharks, most shark attacks on humans are thought to be cases of mistaken identity. The fact that most carnivorous sharks are opportunistic hunters, honing in on weak and dying prey, belies their reputation.

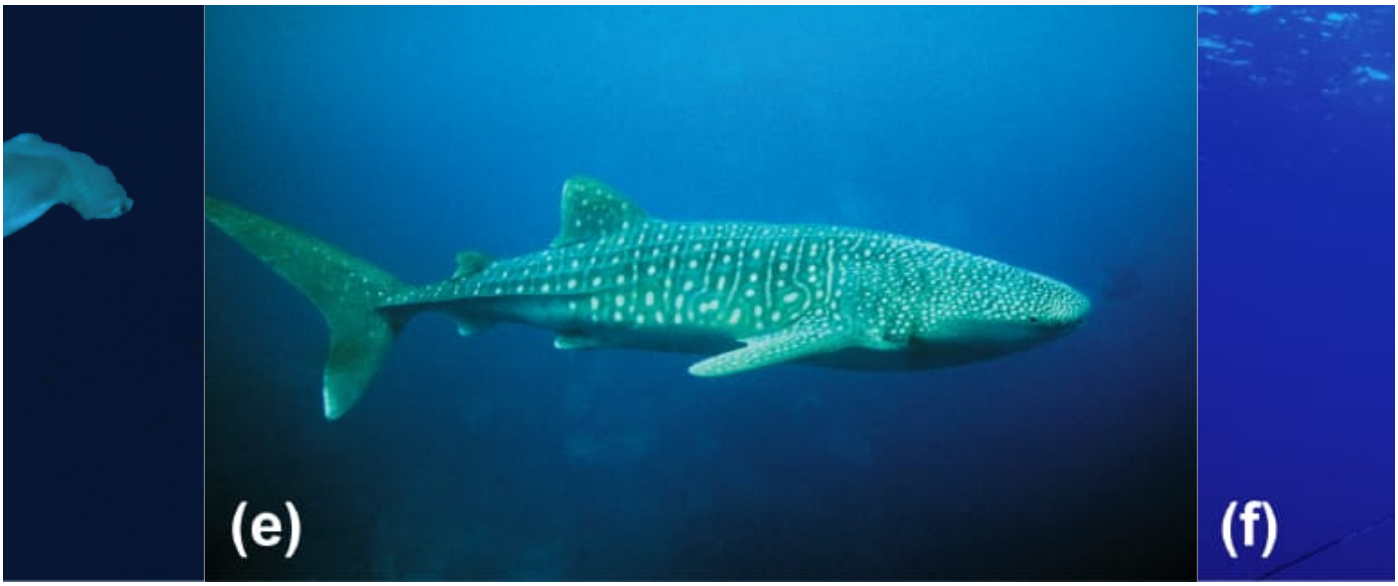


Figure 12-23. Representative sharks and rays. (a) Great white shark (*Carcharodon carcharias*), attracted to a diver's bait. (b) Reef whitetip shark (*Triaenodon obesus*, Papua New Guinea) with a sharksucker, or remora (*Echeneis naucrates*) attached. (c) Gray reef shark (*Carcharhinus amblyrhynchos*, Fiji), which is called a "bronze whaler" in Fiji. (d) Scalloped hammerhead shark (*Sphyrna lewini*, Papua New Guinea). (e) The whale shark (*Rhincodon typus*, Papua New Guinea) is the largest known shark species. This one was quite small, only about 8 to 10 m long. (f) The graceful manta ray (*Manta birostris*, Philippines). (g) Up close and personal with a nurse shark (*Nebrius concolor*, Philippines). (h) Blue-spotted rays (*Taeniura lymma*, Red Sea) have a dangerous barb on the tail. (i) The torpedo ray (family Torpedinidae, Red Sea) can inflict a powerful electric shock. The whale shark and manta ray are plankton feeders, the nurse shark eats benthic invertebrates, and the other sharks are scavengers and carnivores.

Species such as the reef whitetip, gray reef, and hammerhead (**Fig. 12-23b-d**) are regularly attracted to dead fishes set as bait by scuba divers. I have witnessed several occasions when these sharks ate such bait while swimming frantically among surrounding divers. Even when agitated, these sharks do not attack the divers and can almost always be chased away easily on the extremely rare occasions when they do attack. Other shark species, such as the tiger and great white, are more dangerous, and cautious divers will leave the water when they appear.

Although many sharks eat fishes and marine mammals, a number of shark species feed only on plankton. Plankton-eating sharks include basking sharks that grow to 12 m in length and whale sharks (**Fig. 12-23e**) that grow to more than 15 m. The beautiful

and graceful manta ray is also a plankton eater (**Fig. 12-23f**). Plankton-eating sharks and rays swim with mouths wide open. Large volumes of water pass into the mouth and are filtered to capture the plankton before flowing out of the **gills**. The nurse shark (**Fig. 12-23g**) and many species of rays (**Fig. 12-23h,i**) feed only on mollusks and other benthic animals.

Many rays, such as the blue-spotted sting ray (**Fig. 12-23h**), have a barbed stinger embedded in the tail. They use the stinger as a defensive weapon against natural predators, such as sharks, and occasionally against a human foot. Torpedo rays (**Fig. 12-23i**) deliver a strong electric shock that can be lethal to humans, as much as 200 volts, from an organ in the head.

Squid and Their Relatives

Squid (**Fig. 12-24a**), nektonic mollusks of the class Cephalopoda, are extremely abundant in the oceans. Squid normally live in schools that can contain huge numbers of individuals. They are extremely fast swimmers (using their fins) and voracious predators of small fishes. They can also jet-propel themselves by ingesting water and forcibly squirting it out through a special cavity. They can even propel themselves several meters into the air. Most squid species live below the photic zone during the day and migrate to the mixed layer at night to feed. The largest squid species, the giant squid, can grow to 16 m or more in length (6 m of body and 10 m of tentacles). It lives in the deep ocean and is rarely caught or seen, but it is known to be a favorite food of sperm whales.

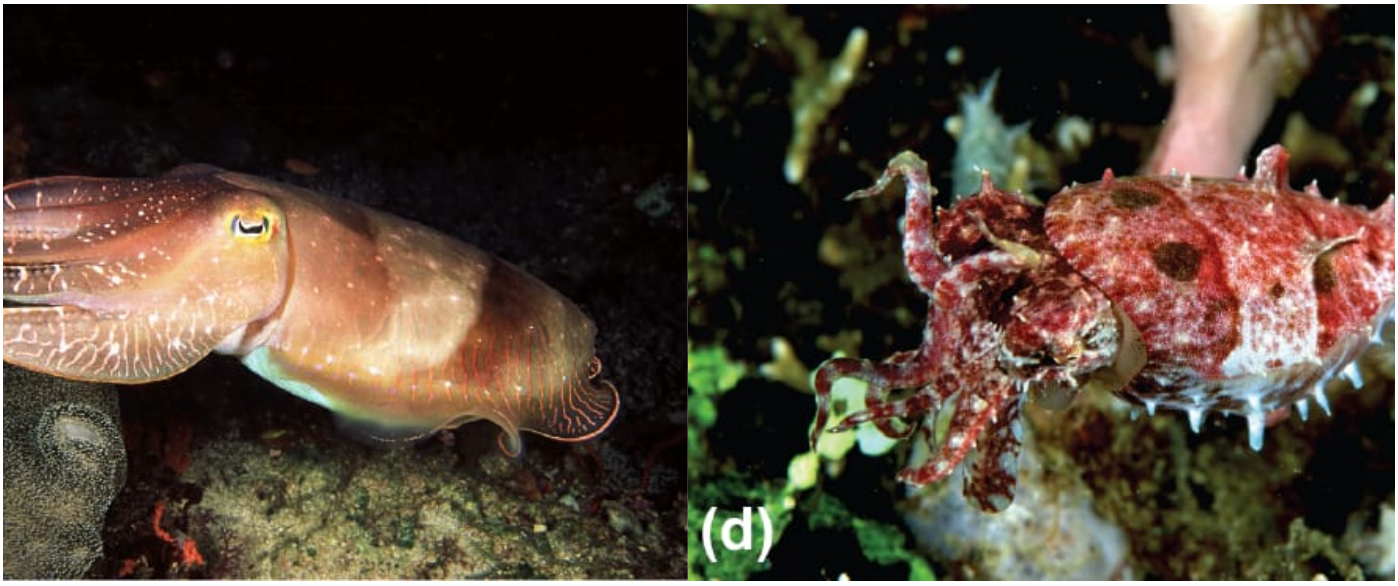


Figure 12-24. Cephalopod mollusks. (a) A common reef squid (*Sepioteuthis lessoniana*, Indonesia) photographed at night. (b,c) Broadclub cuttlefish (*Sepia latimanus*, Indonesia and Papua New Guinea). (d) Stumpy spined cuttlefish (*Sepia bandanensis*, Indonesia). (e) Nautilus (*Nautilus pompilius*, Papua New Guinea), captured at a depth of several hundred meters in a trap similar to a lobster pot, which was baited with fish. The Nautilus was released to swim back to its deepwater home after this photograph was taken. (f) This sectioned Nautilus shell shows the internal chambers that are filled with gas at low pressure to provide buoyancy. Because these chambers are incompressible, Nautilus can safely migrate great vertical distances.

Cuttlefish (**Fig. 12-24b-d**) are **cephalopods** that have an unusual internal single "bone" or shell (the "cuttlebone" that is typically hung in parakeet cages). Nautilus species (**Fig. 12-24e**) have a calcium carbonate shell with a series of internal gas-filled chambers (**Fig. 12-24f**). They are an ancient form of cephalopod that has changed little over millions of years. All cephalopods are believed to have had external shells at one time. Nautilus species, cuttlefish, and squid represent steps in cephalopod evolution in which the external shell, an excellent means of defense, has been discarded in favor of greater swimming speed, which is also an excellent means of defense, but more valuable for hunting prey.

Marine Mammals

Marine mammals, which include seals and sea lions, dolphins and other whales, and a number of other less familiar animals, are warm-blooded and breathe air. Their young are born live and nursed by their mothers, just as terrestrial mammals are.

The **cetaceans**, which include dolphins, porpoises, and other whales, live their entire lives in the water. The largest whales—the

blue, finback, right, sei, and humpback (**Fig. 12-25a-c**)—are baleen whales that feed on plankton. The smaller gray whale, also a baleen whale, is unique because it feeds mainly on small crustaceans, mollusks, and worms that it stirs up from muddy sediments with its snout. Sperm whales feed primarily on squid, whereas most small whales, including the pilot and beluga whales, dolphins, and porpoises, eat mostly fishes. However, the killer whale (**Fig. 12-25d**) lives up to its name by hunting and eating fishes, seals, sea lions, and even other whales. Killer whales often hunt in packs like wolves. On several occasions, pods of killer whales have been filmed systematically harassing a California gray whale mother and calf until they were separated. If they successfully separate the mother and calf, the killer whales, as a pack, set upon the calf, which they then kill and eat.

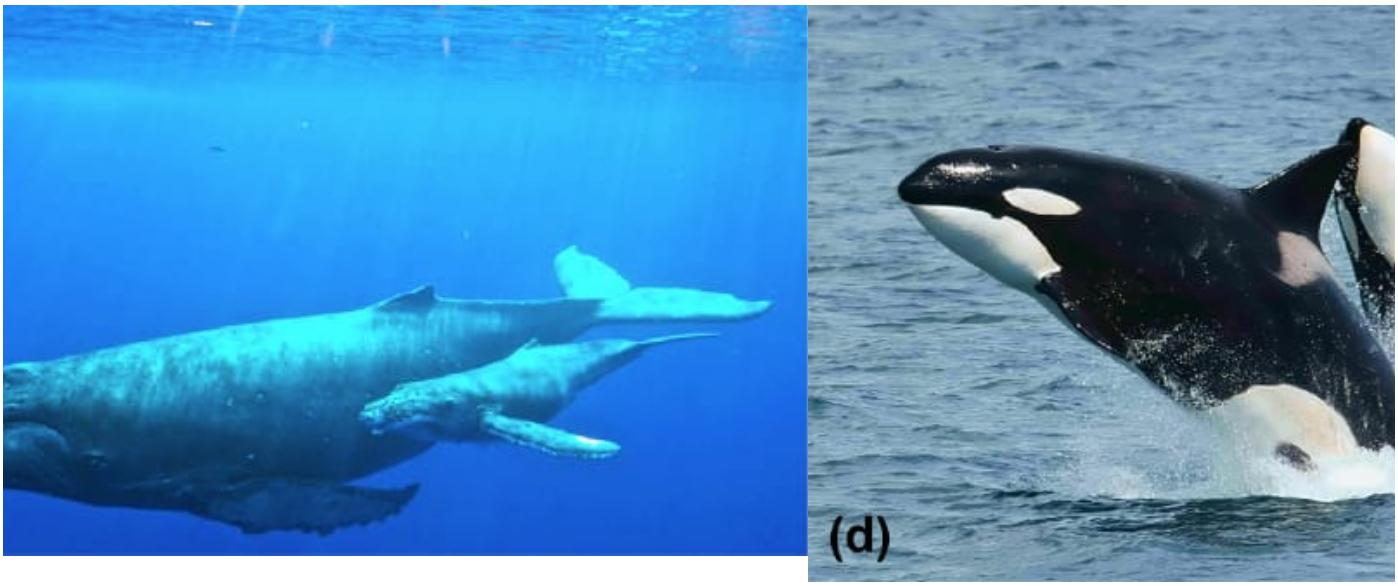


Figure 12-25. Cetaceans. (a)The relative sizes of various toothed and baleen whales, compared to humans and the largest known dinosaur, Apatosaurus. (b) Blue whales (*Balaenoptera musculus*). (c) Humpback whales (*Megaptera novaeangliae*). (d) Killer whales (*Orcinus orca*). (e) Pacific White Side dolphins (*Lagenorhynchus obliquidens*).

Manatees (**Fig. 12-26**) and dugongs, commonly called "sea cows," are herbivorous sirenians that graze on vegetation in the shallow tropical coastal waters where they live. Sea cows are considered to be the source of the mermaid (or siren) legend.

Figure 12-26. Manatees, or sea cows, (*Trichechus manatus*) in the Crystal River, Florida.

Seals (**Fig. 12-27a,b**), sea lions (**Fig. 12-27c**), and walrus (**Fig. 12-27d**) are **pinnipeds**. Pinnipeds live and feed in the oceans for most of their lives, but they must haul themselves out of the water to breed and bear young and to rest and conserve body heat. The elephant seal (**Fig. 12-27b**), whose bizarre-looking adult males weigh 2 tonnes, is the largest pinniped. Most

pinnipeds eat fishes, but walrus (**Fig. 12-27d**) eat clams and other **shellfish** from the sediment, and Antarctic leopard seals will eat invertebrates, seabirds, and other mammals.

Figure 12-27. Pinnipeds and otters. (a) Harbor seals (*Phoca vitulina*) haul out on the rocks near Monterey, California. (b) A scientist measures the length of an adult male elephant seal (*Mirounga angustirostris*) at the Año Nuevo State Reserve, California. This can be a dangerous job for the scientist because the elephant seal may weigh well over a tonne, can move surprisingly fast, and can be very aggressive when disturbed. (c) California sea lion (*Zalophus californianus*) populations are growing very quickly and becoming a nuisance in many harbors, where they aggressively compete for space when they haul out on any convenient platform, especially boat docks like this one in Monterey, California. (d) Walrus (*Odobenus rosmarus*) apparently use their tusks as the equivalent of sled runners that enable them to glide along the seafloor as they feed by sucking or slurping up sediments with their whiskered mouths. (e) Sea otters (*Enhydra lutris*, California) sometimes eat shellfish by resting the shellfish on their chests and pounding them open with a stone while floating on their backs.

The sea otter (**Fig. 12-27e**) is a unique shellfish-eating mammal that, unlike all other marine mammals, lacks an insulating layer of **blubber**. Instead, sea otters retain body heat with soft, thick fur that they must continuously groom if it is to retain its insulating properties.

Many marine mammals were once hunted intensively for their fur, meat, and oil. Because they have long life cycles, hunting quickly reduced some species to critically low population levels. Although certain species are still hunted, the hunting of marine mammals was stopped by most nations several decades ago. Most marine mammal populations that were decimated by hunting are now recovering slowly. However, some species are still threatened by the destruction of **habitat** (particularly pinniped breeding areas), by some fishing methods, such as the use of **drift nets** that kill mammals as well as targeted fishes, and by **pollution** (**Chap. 16**).

Reptiles

Only a small number of reptiles, which are air-breathing animals, live in the oceans. The best known are sea turtles, of which four large species are widely distributed in the oceans: the green (**Fig. 12-28a**), hawksbill (**Fig. 12-28b**), leatherback, and loggerhead. Green turtles are herbivorous and graze on sea grasses that grow in shallow **lagoons**. One common species of sea grass is known as “turtle grass” because it is a favorite food of these turtles. The hawksbill lives in tropical waters and eats mostly **sponges**. The loggerhead also eats sponges, but it adds crabs and mollusks to its diet. The biggest turtle, the leatherback, can weigh as much as 650 kg and is among the few large animals in the ocean that eat jellies as their preferred food.

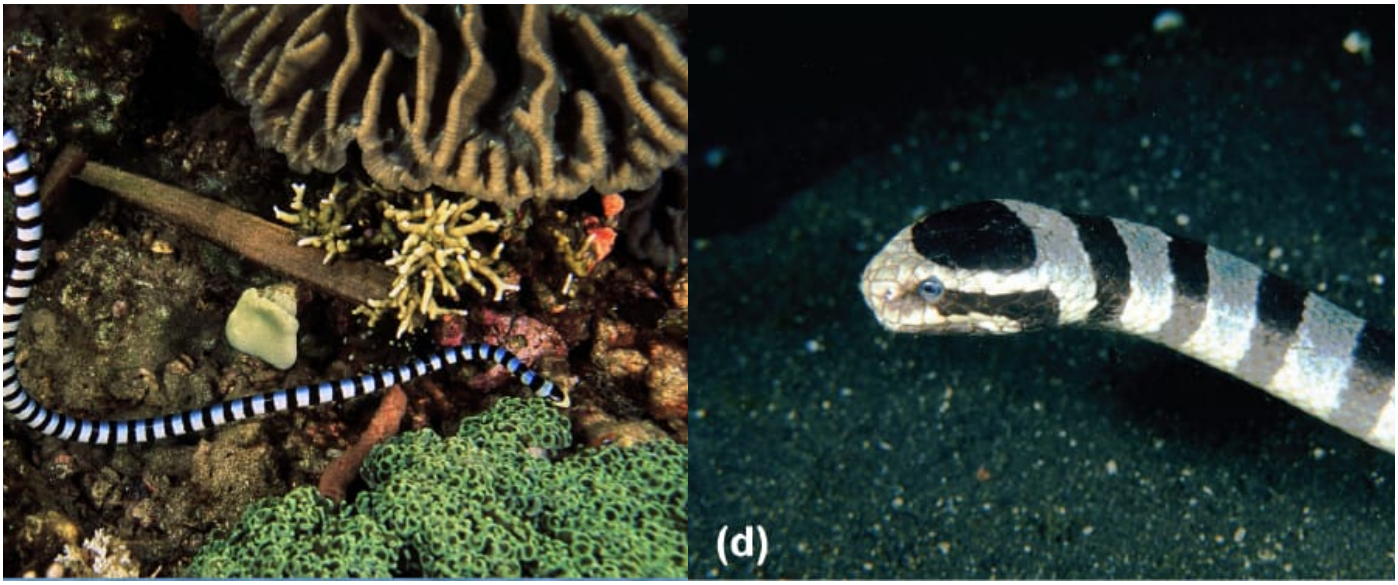


Figure 12-28. Examples of the relatively few marine reptile species. (a) Green sea turtle (*Chelonia mydas*, Hawaii). (b) Hawksbill turtle (*Eretmochelys imbricata*, Fiji). (c,d) Banded sea snake (*Laticauda colubrine*, Indonesia). (e,f) Marine iguana (*Amblyrhynchus cristatus*), which lives only in the Galápagos Islands.

Turtles have been hunted for centuries. They are easy to catch and, if placed on their backs out of water, live for many weeks but cannot escape. For this reason, turtles were an ideal source of fresh meat on sailing vessels before refrigeration was developed. Turtles were common just 200 years ago, but all turtle species have been decimated by humans, and most are threatened or endangered by extinction. The Kemp's ridley turtle, a relatively small species that lives only in the Gulf of Mexico and South Atlantic, is especially in danger of extinction.

Turtles lay their eggs in beach sands above the high-tide line. The eggs hatch in several weeks, and the young turtles immediately head to the sea. Turtle eggs are considered a delicacy in many areas, and the fact that they are easily dug out of the sand by human and other predators has contributed to the decline of turtle populations. In addition, turtles are easy for predators to catch when they climb onto the beach to lay their eggs and when they emerge as hatchlings.

Turtles use relatively few beaches throughout the world for breeding. They almost always return to lay eggs on the same beach where they hatched and usually will not lay eggs on any other beach. Because they are reluctant to lay eggs where human development has altered a beach, they have abandoned many traditional breeding beaches. The reduction in the number of breeding beaches that are suitably protected has seriously affected turtle populations.

The Pacific and Indian Oceans have about 50 species of sea snakes (**Fig. 12-28c,d**). Most are highly venomous, and their bite is generally lethal to people. Fortunately, sea snakes have small mouths and small fangs that cannot easily break human skin, unless they catch a finger or other small appendage. Sea snakes are also relatively shy and do not attack unless threatened. Some species live near the shore so that they can slither out of the water onto rocks to sun themselves, though most are fully aquatic. They eat small fishes, can dive as deep as 100 m, and can stay submerged for more than 2 h before they must come up for air.

In certain areas of Asia and the Pacific Islands, saltwater species of crocodiles feed on fishes in coastal waters. American alligators are also known to enter coastal-ocean waters, but normally they are confined to freshwater. The only marine species of lizard is the Galápagos marine iguana (**Fig. 12-28e,f**), which lives on land but frequently enters shallow coastal waters to graze on marine algae.

Birds

Penguins (**Fig. 12-29a**) are flightless birds that live in the oceans and feed on fishes. They leave the water only to lay eggs and brood their young. Most species are restricted to the cold waters near Antarctica, and none are present in the Northern Hemisphere.

Figure 12-29. Swimming seabirds that feed in the marine environment. (a) King penguins (*Aptenodytes patagonicus*), Fortuna Bay, South Georgia. Adults and molting juveniles. (b) Herring gull (*Larus argentatus*, California), one of the many species of seagulls. (c) Brown pelican (*Pelecanus occidentalis*, California). (d) Double-crested cormorants (*Phalacrocorax auritus*, Monterey, California), Cormorants are excellent underwater swimmers and dry their wings by holding them out in the sun.

Many species of seabirds feed exclusively on ocean fishes. Some, including most seagulls (**Fig. 12-29b**), feed at the ocean surface, but many, like the brown pelican (**Fig. 12-29c**), dive into the water to catch their prey. Others, such as the cormorant (**Fig. 12-29d**), are excellent swimmers that can dive tens of meters deep and stay submerged for as much as several minutes in pursuit of prey.

This page titled 12.13: Nekton is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.14: Benthos

The challenges and opportunities for marine animals in the benthic environment are different from those in the pelagic environment. The physical nature of the seafloor can change dramatically over relatively short distances, especially in coastal areas. The seafloor can be covered with soft mud into which organisms can easily burrow to live, hide, or feed, or it can be solid rock that provides a stable surface for organisms to attach to or bore into. Benthic communities have evolved to take advantage of each different type of seafloor between these two extremes. Consequently, benthic communities are more varied than pelagic communities, particularly in the **coastal zone**, where seafloor character varies the most. Most deep-ocean benthos must obtain food from particulate organic matter that rains down from above or by preying on other organisms.

Chapter 14 discusses the constraints and strategies of life on the seafloor and the approaches that various species use to meet the challenges. **Chapter 15** describes benthos in some specific regions of the seafloor, including those of the deep-ocean seafloor, the unique hydrothermal vent communities of the deep sea, and the remarkably different coastal benthos communities of the coral reef, kelp forest, and rocky intertidal zone.

This page titled 12.14: Benthos is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.15: CHAPTER SUMMARY

Production, Consumption, Decomposition

Almost all organic matter on which marine life depends for food is synthesized from carbon dioxide and nutrients by phototrophy or chemosynthesis—a process called “primary production.” Organisms are autotrophs if they synthesize their own food, and heterotrophs if they do not. Autotrophs and heterotrophs convert some organic matter to carbon dioxide and water in aerobic respiration. Heterotrophs can be herbivores, carnivores, omnivores, detritivores, or decomposers.

Microbes in Charge.

The majority of the biomass in the oceans is now known to be microbial—consisting primarily of bacteria and archaea. Microbial species are responsible for much of the primary production and most of the decomposition of organic matter in the oceans and dominate the movement of energy and biologically important elements in the oceans. It is important to understand how these microbes interact with one another and with the environment around them. Metagenomic studies provide information on what microbes are there and what their metabolic capabilities are.

Primary Production and Light.

Light is needed to provide energy for phototrophy (which includes photosynthesis). Light is absorbed by seawater and scattered by suspended particles. It penetrates to a depth of a few hundred meters in clear waters and much less in high-turbidity waters. The photic zone is the upper layer of the oceans where light is sufficient to support phototrophy. Most primary production is carried out by phytoplankton, which are floating microscopic algae and cyanobacteria. Benthic phototrophs are present only where the seafloor is within the photic zone. As light decreases with depth, phototrophy is slowed but respiration is not. The depth at which photosynthesis equals respiration is the compensation depth. This depth is shallower in winter and in high-turbidity water.

Primary Production and Nutrients.

Phytoplankton require nitrogen, phosphorus, iron, and other nutrients. If the concentration of a particular nutrient is too low to support growth, it is a limiting nutrient. Nutrients are taken up through the cell membrane. The small size of phytoplankton provides a large surface area (per volume) for this uptake and to slow sinking. Phosphorus is recycled rapidly to solution in liquid excretions of animals and during the decay of dead organisms. Nitrogen is recycled more slowly through several chemical forms and is more often the limiting nutrient in the ocean. In some areas where river runoff and atmospheric dustfall are both low, iron can become the limiting nutrient. Silica is an important nutrient for diatoms and other plankton that have silica-based hard parts and is recycled very slowly. Nutrients are carried below the mixed layer by the sinking of dead organisms and fecal pellets. Nutrients are provided to the photic zone primarily by rivers, recycling, and upwelling. They are relatively abundant in waters below the permanent thermocline.

Food Webs.

Phytoplankton are eaten by herbivores or omnivores, which are eaten by small carnivores, which in turn are eaten by larger carnivores. Each organism in this food chain is at a higher trophic level. The conversion efficiency of food to biomass is about 10% between each trophic level. Many carnivores and omnivores feed on organisms from more than one trophic level, forming a more complex food web.

Geographic Variation in Primary Production.

The highest primary productivity is in upwelling regions, most of which are in coastal waters along the western margins of continents. Upwelling also occurs around Antarctica and across parts of the equatorial regions. Primary productivity is lowest in the centers of the subtropical gyres.

Dissolved Oxygen and Carbon Dioxide.

Dissolved oxygen is released during photosynthesis and consumed during respiration. Gases are exchanged between ocean and atmosphere across the sea surface. Oxygen and carbon dioxide are saturated in surface waters. Oxygen can be supersaturated in shallow parts of the photic zone where photosynthesis exceeds respiration. Below the thermocline, oxygen is consumed in respiration and decomposition, and its concentration is below saturation. In some areas where the photic zone has high primary productivity and bottom waters have long residence time, bottom water is anoxic and may contain toxic hydrogen sulfide.

Organic Carbon.

Organic carbon is present in dissolved and suspended particulate forms and in phytoplankton and animal tissues. More than 95% is nonliving dissolved organic matter, and most of the rest is particulate detritus. Most of the dissolved and particulate organic matter is difficult to decompose and has limited nutritional value. Organic particles are more abundant in the photic zone and in areas of high primary productivity.

Biological Provinces and Zones.

The benthic and pelagic environments support fundamentally different biological communities. The benthic environment is divided into the hadal zone, which is the deep seafloor below 6000 m, the abyssal zone between 6000 and 2000 m, the bathyal zone between 2000 m and the continental shelf break, the sublittoral zone (water-covered continental shelf), and the intertidal zone. Benthic organisms in the hadal, abyssal, and most of the bathyal and sublittoral zones depend on detritus for food, with the exception of isolated chemosynthetic communities. The pelagic environment is separated into the neritic province in water less than 200 m deep, and the oceanic province in water deeper than 200 m. The oceanic province consists of the epipelagic zone, in which phototrophy (including photosynthesis) can occur; and the mesopelagic, bathypelagic, abyssopelagic, and hadopelagic zones, which are all below the photic zone. Communities of each of the benthic and pelagic zones are different, but there is some overlap.

In the sublittoral zone, the neritic province, and the epipelagic zone, latitudinal variations of temperature and salinity cause latitudinal zonation in biological communities. For example, tropical and polar regions support different species.

Plankton.

Plankton are organisms and viruses that drift freely with currents and include phytoplankton (which are microscopic photosynthetic algae and cyanobacteria) and zooplankton (which are herbivorous, carnivorous, or omnivorous animals). Phytoplankton are abundant in productive photic-zone waters, exceeding 1 billion individuals per liter, and have patchy distributions. Diatoms are relatively large phytoplankton with silica frustules. They are important in food chains that lead to commercially valuable species. Dinoflagellates are smaller than diatoms, and most lack hard parts. They dominate where silica is depleted, particularly in the open oceans, are not always autotrophic, and often bloom explosively. Generally less abundant types of phytoplankton include coccolithophores that are covered in tiny calcareous plates, and silicoflagellates that have a silica shell. Microscopically small ultraplankton, which include bacteria, archaea, and viruses, are extremely abundant but are not yet well characterized.

Zooplankton are either holoplankton that live their entire life cycles as plankton, or meroplankton that spend only their larval stages as zooplankton. Many zooplankton migrate to the surface layer at night to feed and return below the photic zone by day. Holoplankton include copepods, euphausiids, and other crustaceans, foraminifera, radiolaria, and pteropods, as well as gelatinous forms such as jellies, ctenophores, and salps. Meroplankton include eggs, larvae, and juveniles of many invertebrates and fishes.

Nekton.

Nekton are organisms that swim actively. They include many fish species. Sharks and rays are fishes that have cartilaginous skeletons. Many fishes have gas- or oil-filled swim bladders to maintain buoyancy. The nekton also include squid, which are extremely abundant, feed mostly near the surface at night, and migrate below the photic zone by day. Marine mammals, including dolphins, porpoises, other whales, seals, and sea lions, are also nekton. Sea turtles and sea snakes are nektonic reptiles.

Benthos.

Benthic organisms comprise a profusion of species adapted to live on or in the sediment or on hard substrates such as rocks and coral reefs.

12.15.1: Key Terms

12.15.2: Chapter Questions

12.15.3: References

This page titled 12.15: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.15.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- abyssal plains
- abyssal zone
- adsorption
- aerobic
- algae
- amoeba
- amphipods
- anemones
- angle of incidence
- anoxic
- anthropogenic
- aphotic zone
- archaea
- autotrophs
- bacteria
- baleen
- barnacles
- bathyal zone
- beaches
- benthic
- benthos
- biochemical
- biogeochemical cycle
- bioluminescent
- biomass
- biota
- blubber
- bluegreen algae
- buoyancy
- calcareous
- carbon compensation depth (CCD)
- carnivores
- cartilaginous
- cephalopods
- cetaceans
- chemosynthesis
- chlorophyll
- chordates
- climate
- cnidarians
- coast
- coastal upwelling
- coastal zone
- coccolithophores
- colloidal
- community
- compensation depth
- continental shelf
- convergence
- coral reefs

- corals
- crust
- crustaceans

- ctenophores
- currents
- cyanobacteria
- decomposers
- density
- deposition
- detritivores
- detritus
- diagenetic
- diatoms
- diel
- diffusion
- dinoflagellates
- divergence
- downwelling
- drift nets
- ecological
- ecosystem
- El Niño
- environment
- enzymes
- epifauna
- epipelagic zone
- erosion
- estuarine
- eukaryotes
- euphausiids
- eutrophication
- excrete
- excurrent
- extratrophical cyclones
- fauna
- fecal pellets
- fermentation
- filter-feeding
- fjords
- flagella
- flagellates
- flora
- food chain
- food webs
- foraminifera
- frustule
- fungi
- genus
- genome sequencing
- gills
- grazers
- ground water
- habitat
- hadal zone
- hard parts
- herbivores

- herrings
- heterotrophs
- holoplankton
- hydrothermal vents
- incurrent
- infauna
- intertidal zone
- invertebrates
- ions
- isotopic
- jellies
- jellyfish
- krill
- lagoons
- Langmuir circulation
- larvae
- latitudes
- limiting nutrient
- lithogenous
- littoral zone
- low-tide line
- macroalgae
- marine mammals
- meroplankton
- metagenomics
- metamorphosis
- microbial
- mixed layer
- mixotrophs
- mollusks
- nekton
- nutrient-limited
- nutrients
- oceanic ridges
- omnivores
- overfished
- ozone layer
- pelagic
- photic zone
- photoautotrophs
- photoheterotrophs
- photosynthesis
- phototroph
- phototrophy
- phytoplankton
- pinnipeds
- plankton
- plate tectonics
- pollution
- primary production
- primary productivity
- prokaryotes
- protists

- pteropods
- pycnocline
- reptiles
- residence time

- respiration
- resuspended
- rhodopsins
- rocky intertidal zones
- runoff
- salinity
- salps
- saturation solubility
- scattered
- schooling
- sea cucumbers
- sea grasses
- sea stars
- sea urchins
- seamounts
- sediment
- shelf break
- shellfish
- shores
- siliceous
- species
- sponges
- spreading cycle
- standing stock
- storm surges
- subducted
- subduction zones
- sublittoral zone
- subtropical gyres
- supersaturated
- supralittoral zone
- suspended sediments
- swim bladders
- thermocline
- tides
- trenches
- trophic level
- tsunamis
- turbulence
- tunicates
- turbidity
- vertebrates
- viruses
- weathering
- westerly
- zooplankton

This page titled 12.15.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.15.2: Chapter Questions

Study Questions

1. Why is it necessary to have primary producers, animals, and decomposers in a marine ecosystem?
2. What is metagenomics and how can it be applied to oceanic studies? What type of genetic material is typically extracted in metagenomic studies?
3. Dissolved oxygen concentration varies substantially with depth. Describe and explain the distribution of dissolved oxygen in the deep ocean.
4. Why is nitrogen generally the limiting nutrient in the oceans? List the sources of biologically available nitrogen in the photic zone.
5. Of all the organic carbon in the oceans, 99% is in nonliving particulate or dissolved form. Why?
6. Why are most of the world's major fisheries in coastal waters? Why are many major fisheries off the west coasts of the continents? Why are the waters around Antarctica highly productive?
7. List and describe the principal differences between the pelagic and benthic environments. Why are the boundaries between zones of the benthic and pelagic environments defined by depth ranges?
8. At what depths in the pelagic and benthic environments would you expect to find substantial variation in the species composition of the fauna and flora between tropical, mid-, and high latitudes? Why?
9. Describe the major types of plankton.
10. Why are phytoplankton small?
11. List the major categories of organisms that constitute the nekton. What are the principal distinguishing characteristics of each category?

Critical Thinking Questions

1. Very high productivity is found in the circumpolar ocean around Antarctica. This is also the region most often under the hole in the ozone.
   (a) Does the high productivity indicate that the ozone hole does not have a significant adverse effect on the ocean ecosystem? Explain why or why not.
   (b) What other information did you use to reach your answer?
   (c) What other studies or data would you like to have to make sure that your answer is correct?
2. The enhanced greenhouse effect is causing climate changes that will lead to warming of the mixed layer of the oceans by as much as 2–3°C.
   (a) Hypothesize how this change might affect winds, the characteristics of the pycnocline, and ocean currents.
   (b) Would these changes affect the amount of nutrients available to phytoplankton? If so, how?
3. A number of years ago it was suggested that nuclear reactors should be placed on the seafloor in tropical deep-water areas and other areas of the oceans where the primary production in the mixed layer is nutrient-limited. The intention was that the heat generated by the reactor would cause artificial convection-driven upwelling, increase the supply of nutrients to the mixed layer, and increase primary production.
   (a) Would this proposal work? If not, why not?
   (b) If so, can you suggest reasons (other than the potential for the release of radioactivity) to explain why the idea was not adopted?
4. If you measured the dissolved oxygen concentration carefully in the photic zone during a spring bloom of phytoplankton, describe how you would expect it to change over 24 h and why. Would the changes in concentration during the 24 h be different at different depths within the photic zone? If so, describe how and why.
5. Pelagic food chains that lead to large predatory fishes in the deep oceans far from land tend to be much longer than in coastal regions. What might account for this difference?
6. What is a detritus-based food web? Would you expect detritus based food chains that lead to large fishes that feed on benthic animals to be longer than the pelagic food chains in the overlying water that lead to large carnivorous fishes? Why? Would your answer be different for coastal and deep-ocean regions? Why?
7. Many fewer species of reptiles and mammals live in the oceans than on land, and many more species of fishes and invertebrates live in the oceans than live on land and in freshwater. What do you think are the reasons for these differences?

This page titled 12.15.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

12.15.3: References

Figure 12-8 – Data from H.U. Sverdrup, H. U., M. W. Johnson and R. H. Fleming, 1942: *The oceans, their physics, chemistry, and general biology*. Prentice Hall, 1087 pp.

Figure 12-9 – Data from H.U. Sverdrup, H. U., M. W. Johnson and R. H. Fleming, 1942: *The oceans, their physics, chemistry, and general biology*. Prentice Hall, 1087 pp.

Figure 12-12 – Adapted from Everson, I., 1977: *The living resources of the Southern Ocean*. Food and Agriculture Organization of the United Nations, United Nations Development Programme, 156 pp.

Figure 12-13 – Ocean Biology DAAC, 2026: Aqua-MODIS Level-3 Chlorophyll Concentration Seasonal Climatology. NASA GSFC, accessed 16 March 2026, Source.

Figure 12-15 – Walker, M.I., 2013: Live Marine Zooplankton. Pixels, accessed 16 March 2026, Source.

Figure 12-16a – Zanette, D.I., 2002: Diatoms: eukaryotic algae, accessed 16 March 2026, Source.

Figure 12-16b – Sampayo, M. A., 2007: Dinoflagellate that exhibits bioluminescence. Instituto de Oceanografia, Faculdade Ciências da Universidade de Lisboa, accessed 16 March 2026, Source.

Figure 12-16c – Kuring. N., 2026: A bloom of coccolithophores seen from space, accessed 16 March 2026, Source.

Figure 12-18a – Adapted from Valiela, I., 1984: *Marine Ecological Processes*. Springer-Verlag, 546 pp.

Figure 12-18b – Adapted from Hulburt, E. M., 2003: Phytoplankton in the southwestern Sargasso Sea and North Equatorial Current, February 1961. *Limnol. Oceanogr.*, **7**, 307–315, Source.

Figure 12-19a – Hopcroft, R., 2005: Male copopod. Hidden Ocean 2005 NOAA, accessed 16 March 2026, Source.

Figure 12-19b – Wilson M., and J. Clark, 2013: Zooplankton. Euphausiid shrimp, popularly known as krill. NOAA NMFS AFSC, accessed 16 March 2026, Source.

Figure 12-19c – Uwe kils, 2023: Amphipod image (possibly Ampeliscidae). Accessed 16 March 2026, Source, CC BY-SA 3.0.

Figure 12-19d – Zeebe, R.E. and coauthors, 2008: Vital effects and beyond: a modelling perspective on developing palaeoceanographical proxy relationships in foraminifera. *Spec. Publ. - Geol. Soc. London*, **303**, 45–58, Source.

Figure 12-19e – OOI Regional Cabled Array, 2019: Radiolarian, accessed 16 March 2026, Source.

Figure 12-20a – Segar, D., and E. Stamman Segar, 1997: Jelly *Aurelia sp*. in Jellyfish Lake in Palau.

Figure 12-20b – Segar, D., and E. Stamman Segar, 1997: Jelly, *Cassiopeia andromeda*, in Papua New Guinea, about 10–15 cm across.

Figure 12-20c – Islands in the Sea 2002, 2002: This tiny and very dangerous Portugese Man-O-War jellyfish was collected during one evening's "night-lighting" samplings. NOAA OER, accessed 16 March 2026, Source.

Figure 12-20d – Raskoff, K.: Light refracts off the comb-rows of the ctenophore *Mertensia ovum*. NOAA OER Arctic Exploration 2002, accessed 16 March 2026, Source.

Figure 12-20e – Segar, D., and E. Stamman Segar, 1997: Colonial salp species in Indonesia.

Figure 12-21a – Wilson, M. and J. Clark, 2013: Zooplankton. Crustacean larva. NOAA NMFS AFSC, accessed 16 March 2026, Source.

Figure 12-21b – Alaska Sea Grant, 2015: Figure 7. King Crab Larvae, accessed 16 March 2026, Source.

Figure 12-21c – Bruno C. Vellutini, 2005: Larva de estrela-do-mar!, accessed 16 March 2026, Source, CC BY-SA 2.0.

Figure 12-22a – Segar, D., and E. Stamman Segar, 1997: Dogtooth tuna (*Gymnosarda unicolor*) in Papua New Guinea.

Figure 12-22b – Fagen, A, 2013. Schooling Pacific Sardine (*Sardinops sagax*). Flickr, accessed 16 March 2026, Source.

Figure 12-22c – NOAA Fisheries, 2009: Atlantic cod fish, accessed 16 March 2026, Source.

Figure 12-22d – Segar, D., and E. Stamman Segar, 1997: Flowery flounder (*Bothus mancus*) in Solomon Islands.

Figure 12-23a – Gross, T., 2006: Great white shark at Isla Guadalupe, Mexico, August 2006, accessed 16 March 2026, Source.

Figure 12-23b – Segar, D., and E. Stamman Segar, 1997: Reef whitetip shark (*Triaenodon obesus*) with a sharksucker, or remora (*Echeneis naucrates*) attached in Papua New Guinea.

Figure 12-23c – Segar, D., and E. Stamman Segar, 1997: Gray reef shark (*Carcharhinus amblyrhynchos*) in Fiji.

Figure 12-23d – Segar, D., and E. Stamman Segar, 1997: Scalloped hammerhead shark (*Sphyrna lewini*), in Papua New Guinea.

Figure 12-23e – Segar, D., and E. Stamman Segar, 1997: Whale shark (*Rhincodon typus*) in Papua New Guinea.

Figure 12-23f – Segar, D., and E. Stamman Segar, 1997: Manta ray *(Manta birostris*) in the Philippines.

Figure 12-23g – Segar, D., and E. Stamman Segar, 1997: Nurse shark (*Nebrius concolor*) in the Philippines.

Figure 12-23h – Segar, D., and E. Stamman Segar, 1997: Blue-spotted rays (*Taeniura lymma*) in the Red Sea.

Figure 12-23i – Segar, D., and E. Stamman Segar, 1997: Torpedo ray (family Torpedinidae) in the Red Sea.

Figure 12-24a – Segar, D., and E. Stamman Segar, 1997: Common reef squid (*Sepioteuthis lessoniana*) in Indonesia and photographed at night.

Figure 12-24b – Segar, D., and E. Stamman Segar, 1997: Broadclub cuttlefish (*Sepia latimanus*) in Indonesia.

Figure 12-24c – Segar, D., and E. Stamman Segar, 1997: Broadclub cuttlefish (*Sepia latimanus*) in Papua New Guinea.

Figure 12-24d – Segar, D., and E. Stamman Segar, 1997: Stumpy spined cuttlefish (*Sepia bandanensis*) in Indonesia.

Figure 12-24e – Segar, D., and E. Stamman Segar, 1997: Nautilus (*Nautilus pompilius*) in Papua New Guinea, captured at a depth of several hundred meters.

Figure 12-24f – Segar, D., and E. Stamman Segar, 1997: Sectioned Nautilus shell.

Figure 12-25a – Adapted from Friends of the Earth, 1978: *The Whale Manual*. Brick House Pub Co.

Figure 12-25b – NOAA, 2007: CINMS blue whales, accessed 16 March 2026, Source.

Figure 12-25c – National Marine Sanctuaries, 2013: Humpback whale with her calf, accessed 16 March 2026, Source.

Figure 12-25d – Pitman, R., 2005: Two mammal-eating "transient" killer whales, NOAA. Accessed 16 March 2026, Source.

Figure 12-25e – Christman, B., 2003: Pacific white-sided dolphin leaping. *Lagenorhynchus obliquidens*. NOAA Corps, accessed 16 March 2026, Source.

Figure 12-26 – Hinkel, D., 2016: Endangered Florida manatee (*Trichechus manatus*), Crystal River National Wildlife Refuge, Florida. USFWS Endangered Species Flickr, accessed 16 March 2026, Source.

Figure 12-27a – Segar, D., and E. Stamman Segar, 1997: Harbor seals (*Phoca vitulina*) haul out on the rocks near Monterey, California

Figure 12-27b – Segar, D., and E. Stamman Segar, 1997: A scientist measures the length of an adult male elephant seal (*Mirounga angustirostris*) at the Año Nuevo State Reserve, California.

Figure 12-27c – Segar, D., and E. Stamman Segar, 1997: California sea lion (*Zalophus californianus*).

Figure 12-27d – Christman, B., 1978: Attentive walrus - *Odobenus rosmarus divergens* - inspecting photographer. NOAA. Accessed 16 March 2026, Source.

Figure 12-27e – Segar, D., and E. Stamman Segar, 1997: Sea otters (*Enhydra lutris*) in California.

Figure 12-28a – Segar, D., and E. Stamman Segar, 1997: Green sea turtle (*Chelonia mydas*) in Hawaii.

Figure 12-28b – Segar, D., and E. Stamman Segar, 1997: Hawksbill turtle (*Eretmochelys imbricata*) in Fiji.

Figure 12-28c – Segar, D., and E. Stamman Segar, 1997: Banded sea snake (*Laticauda colubrine*) in Indonesia.

Figure 12-28d – Segar, D., and E. Stamman Segar, 1997: Banded sea snake (*Laticauda colubrine*) in Indonesia.

Figure 12-28e – Hemminger, C. S., 1997: Marine iguana (*Amblyrhynchus cristatus*). With permission.

Figure 12-28f – Hemminger, C. S., 1997: Marine iguana (*Amblyrhynchus cristatus*). With permission.

Figure 12-29a – Segar, D., and E. Stamman Segar, 1997: King penguins (*Aptenodytes patagonicus*), Fortuna Bay, South Georgia. Adults and molting juveniles

Figure 12-29b – Segar, D., and E. Stamman Segar, 1997: Herring gulls (*Larus argentatus*), California.

Figure 12-29c – Vinson, A., 1997: Brown pelican (*Pelecanus occidentalis*) in California. With permission.

Figure 12-29d – Segar, D., and E. Stamman Segar, 1997: Double-crested cormorants (*Phalacrocorax auritus*) in Monterey, California.

This page titled 12.15.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, Michelle Valkanas, and Elaine Stamman.

13: COASTAL OCEANS AND ESTUARIES

Critical Concepts

- **CC5 Transfer and Storage of Heat by Water:** Water's high heat capacity allows large amounts of heat to be stored in the oceans and released to the atmosphere without much change in the ocean water temperature. Water's high latent heat of fusion allows ice to act as a heat buffer, which keeps the ocean surface water layer temperatures in high latitudes relatively uniform and near the freezing point.
- **CC8 Residence Time:** The residence time of seawater in a given segment of the oceans is the average length of time the water spends in that segment. The residence times of some coastal water masses are long, and therefore, some contaminants discharged to the coastal ocean can accumulate to higher levels in these regions than in areas with shorter residence times.
- **CC11 Chaos:** The nonlinear nature of many environmental interactions, including some of those that control annual fluctuations in fish stocks, mean that fish stocks change in sometimes unpredictable ways.
- **CC12 The Coriolis Effect:** Water masses move freely over the Earth and ocean surface, while objects on the Earth's surface, including the solid Earth itself, are constrained to move with the Earth in its rotation. This causes moving water masses to be deflected as they flow. The apparent deflection is at its maximum at the poles, is reduced at lower latitudes, and becomes zero at the equator. In Northern Hemisphere estuaries, the Coriolis effect tends to concentrate the lower salinity water that flows down the estuary toward the left side of the estuary as viewed from the ocean, while the high salinity water flowing up the estuary tends to be concentrated toward the right side (In the Southern Hemisphere these directions of concentration are reversed).
- **CC15 Food Chain Efficiency:** All organisms use some of their food as an energy source in respiration and for reproduction. They also lose some of their food in excretions (including wastes). On average, at each level in a food chain, only about 10% of food consumed is converted to growth and biomass of the consumer species.
- **CC16 Maximum Sustainable Yield:** The maximum sustainable yield is the maximum biomass of a fish species that can be depleted annually by fishing, but that can still be replaced by reproduction. This yield changes unpredictably from year to year in response to the climate and other factors. The populations of many fish species worldwide have declined drastically when they have been overfished (beyond their maximum sustainable yield) in one or more years when that yield was lower than the average annual yield on which most fisheries management is based.

Figure 13-1. The oceans and the land are intimately related where they meet. The vegetation, rocks, beach, waves, clouds, kelp seen floating on the water surface, and all the unseen plants and animals living within the coastal waters in this California bay are the result of processes that take place in the coastal ocean and estuaries.

In the coastal oceans, **salinity**, temperature, water **density**, **turbidity**, chemical composition, and water movements are affected by freshwater discharges from rivers, and by the physical boundaries of land and shallow seafloor. These influences vary in **spatial**

extent as seasonal changes in rainfall affect **runoff** and river input of freshwater and as **weather** and seasonal changes in winds affect wind-driven **currents**. Hence, the boundaries of the **coastal zone** are not defined precisely. On **passive margins**, where the **continental shelf** is wide, the coastal zone is generally the area between the **coastline** and the continental **shelf break**. Along tectonically active margins, this zone is less easy to define. Where large river discharges occur on tectonically active margins with narrow continental shelves, the coastal zone may extend beyond the continental shelf.

The continental shelf comprises approximately 8% of the total area of the oceans. The coastal zone, which includes only limited areas beyond the shelf, comprises less than 10%. Nevertheless, the coastal ocean is disproportionately important for humanity because more than 99% of the total world fishery catch is taken from coastal waters. The coastal ocean is also important for ports and shipping; recreational activities; mining of sand, gravel, and other minerals; oil exploration and extraction; energy production or potential production from winds, tides and currents, thermal differences between surface water and deep water; and waste disposal. This chapter examines the factors that make the coastal zone and its **ecosystem** different from the open ocean. It also describes the characteristics of **estuaries** and **lagoons** and their relationships to ocean ecosystems.

13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.5: Fisheries
13.6: Estuaries
13.7: Chapter Summary
13.7.1: Key Terms
13.7.2: Chapter Questions
13.7.3: References

This page titled 13: Coastal Oceans and Estuaries is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.1: Special Characteristics of Coastal Oceans

Because the coastal oceans are shallow, water movements are affected by the seafloor and the coastline. In addition, rivers discharge freshwater into coastal waters. As a result:

- The salinity and temperature of coastal waters are more variable than those of open-ocean waters.
- Coastal currents are generally independent of the open-ocean gyre currents.
- The mixed layer has more sources of nutrients in coastal waters than in the open ocean.
- Benthos of the coastal oceans are more diverse, abundant, and commercially valuable than open-ocean benthos.

These factors make the coastal oceans different and generally more variable in physical and biological characteristics, than the open oceans. Some of the ways in which these factors affect the coastal oceans are discussed in this section. However, you should remember that each coastal area has its own unique characteristics, and this chapter provides only a general overview and some examples of coastal and estuarine processes in some specific locations.

Salinity

Freshwater from rivers reduces the salinity of coastal waters. Where river discharges are large, low-salinity water spreads as a surface layer over higher-salinity ocean water unless vertical mixing is intense. Consequently, there is often a **halocline** a few meters below the surface near mouths of major rivers. The area affected by the halocline and the halocline's strength depend on the rate of discharge of low-salinity water from the river and on the extent of wind-, wave-, and **tide**-induced vertical mixing of the water column (Fig. 13-2). Haloclines inhibit **upwelling** and vertical mixing.

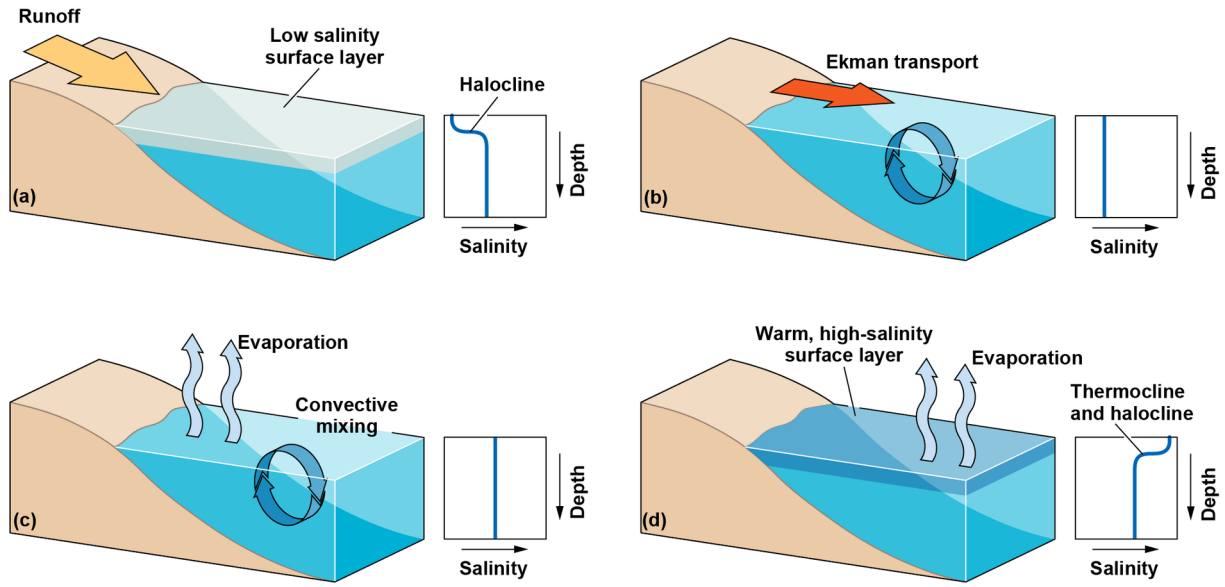


Figure 13-2. Salinity distributions in the coastal-ocean water column. (a) Near major rivers, discharges of low-salinity river water over higher-salinity ocean water create a halocline. Where winds transport surface waters offshore, the layer of low-salinity water may spread farther and be thinner. Lower-volume river discharges produce thinner, low-salinity layers that do not extend far from the river mouth. (b) Where winds are strong and river discharge is relatively small, the low-salinity river water may be completely mixed with ocean water, resulting in no halocline. (c) In areas of strong evaporation and low rainfall, warm, high-salinity surface water is produced that may be dense enough to sink all the way to the seafloor. (d) Conversely, this warm, high-salinity water may be warm enough that its density is lower than the lower-salinity but cooler offshore water. In this case, the warm, high-salinity water forms a higher-salinity surface layer overlying a halocline in which salinity decreases with depth.

Low-salinity water discharged by rivers mixes with ocean water primarily by wind action and is transported away from the river mouth by coastal currents. Chapter 8 explained that coastal currents generally flow parallel to the coastline and are often separated from the open-ocean gyre currents by sharply defined **fronts**. Thus, mixing of coastal water with water from the open oceans is slow, and **residence times** of water in the coastal ocean can be long.

The residence time of freshwater in the coastal water mass can be calculated if the volume input of freshwater from rivers, the average salinity in the coastal waters, and the volume of the coastal water mass are known (**CC8**). For example, the residence time of freshwater in the coastal oceans of the northeastern U.S. from Cape Cod, Massachusetts, to Cape Hatteras, North Carolina, is about 2½ years. The long residence time has implications for the use of this area for ocean waste disposal. Although the volume of coastal water is large, the quantities of potential **contaminants** released into the northeastern U.S. coastal ocean in discharges from cities (primarily treated sewage wastewater and storm water runoff) and industries are also large. For potential contaminants not removed by **sedimentation** or decomposition (**Chap. 16**), an amount equivalent to 2½ years of discharges is retained within this region.

Salinity varies seasonally with river flow in many coastal areas. In middle and low **latitudes**, salinity is usually lowest in the local rainy season. At high latitudes, salinity is always lowest in late spring or summer when snowmelt swells rivers and seasonal sea ice melts.

Salinity extremes occur in **marginal seas**, lagoons, or other bays with restricted connections to the oceans. In locations such as the eastern Mediterranean and the Red Sea, evaporation is high and rainfall low, so surface salinity is high—in some areas exceeding 40. In these regions, the mixed layer is usually thicker because vertical circulation is enhanced by the formation of surface waters with relatively high density due to high salinity. The mixed layer often extends to the seafloor and consists of warm, high-salinity water. The vertical circulation helps to recycle nutrients from relatively deep waters into the **photic zone**. Consequently, many of these regions have moderately high productivity despite their limited inputs of nutrients in runoff from land. However, if this body of water is not completely separated from the ocean, lower-salinity surface water from the open ocean may flow as a surface layer over the high-salinity water created by evaporation in another part of the basin (**Fig. 13-2c**). In some basins with high evaporation and low rainfall, higher-salinity but warmer surface water has lower density than the lower-salinity but colder open-ocean water. Thus, a warm, high-salinity surface layer is formed (**Fig. 13-2d**).

Salinity is lowest where river outflow is large and evaporation is low. Such conditions generally occur in middle- and high-latitude locations. These include many **fjords** and the Baltic Sea. The huge **monsoon** flows of the Ganges and Brahmaputra Rivers lower the surface salinity of the entire Bay of Bengal (**Fig. 7-15**).

Temperature

Seasonal changes in solar intensity and air temperature cause larger temperature changes in shallow coastal waters than in the open ocean. Temperature variations are particularly large in enclosed marginal seas, fjords, and other embayments that have restricted exchange with the open ocean.

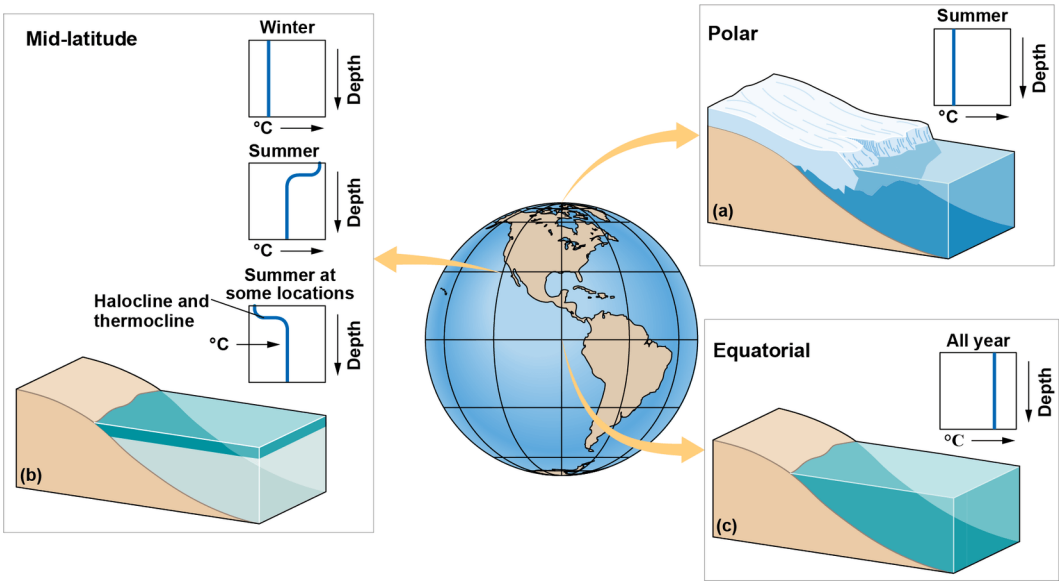


Figure 13-3. Temperature distributions in the coastal-ocean water column. (a) In high-latitude regions where ice is forming or melting throughout the year, the water column is uniformly at or near the freezing point. (b) In mid-latitude regions, surface waters are warmed in summer, producing a shallow seasonal thermocline. In winter, the surface water is cooled and sinks, and the water column is generally well mixed by this convection and by winds. In some areas near freshwater inputs, low-salinity runoff may produce cool but low-salinity surface water and a shallow pycnocline that is both a halocline and a thermocline in which temperature increases with depth. (c) In low-latitude regions where freshwater inputs are limited, the temperature of the water column above the permanent ocean thermocline depth is usually uniform and high.

In high latitudes, coastal water temperatures vary little between seasons (**Fig. 13-3a**). Sea ice in these regions melts during summer and re-forms during winter. This process maintains the water temperature at the freezing point of seawater, usually above –2°C (**Chap. 5**, **CC5**).

Seasonal temperature changes in coastal waters are most evident in mid latitudes (**Fig. 13-3b**). In these regions, the surface water layer becomes warmer and less dense as solar heating increases in the spring, and reaches a temperature maximum in the summer. In addition, in many locations, mixing by wind and waves decreases during summer. As a result, a shallow **thermocline** is formed and the mixed-layer depth is reduced to about 15 m (**Fig. 13-3b**). This seasonal thermocline breaks down in winter when surface water is cooled and **convection** and wind mixing increase. A uniformly cold mixed layer is produced that extends to the top of the permanent thermocline (about 100 m) or to the seafloor, if it is shallower. Because most coastal waters are less than 100 m deep, the mixed layer extends to the seafloor in most of the temperate coastal region during winter.

In tropical regions, coastal water temperature is uniformly high year-round, and the water column is often **isothermal** to the seafloor, particularly in shallow lagoons (**Fig. 13-3c**). In deeper coastal waters, the depth of the mixed (or isothermal) layer is determined by physical processes, including wind-, tide-, and wave-induced mixing.

Coastal-ocean water temperature in middle latitudes is usually lower in summer and higher in winter than air temperatures over the adjacent coastal land. This difference affects the **climate** of coastal regions (**Chap. 7**) by moderating temperature variations.

Waves and Tides

Tidal-current speed increases as the tide wave enters shallow coastal waters and its energy is compressed into a shallower depth (**Chap. 10**). Consequently, tides are an important contributor to vertical and horizontal mixing in the coastal zone.

As tides **ebb** and **flood**, they cause **turbulence** that enhances vertical mixing. Tides also mix water across the shelf. Near-bottom tidal currents are slowed by **friction** with the seafloor. In addition, tidal currents may be different above and below a **pycnocline** when one is present. Thus, water in different layers may move different distances across the shelf during a tidal cycle. Turbulent mixing between layers transfers water from one layer to another.

Wind-driven waves create turbulence and vertical mixing. In the coastal zone, vertical mixing is enhanced where waves move into shallow waters and break. **Langmuir circulation** and **internal waves** (**Chap. 8**) also contribute to vertical mixing.

Turbidity and the Photic Zone

The photic zone extends to the seafloor in many areas of the coastal oceans. In these areas, **kelp** and other **macroalgae** (**Fig. 13-4**) can live and **photosynthesize** on the seafloor. However, the photosynthetic **community** of the shallow seafloor is often dominated by **species** of single-celled **algae** similar to phytoplankton. These algae **encrust** rocks, dead **coral**, and other organisms that live on the seafloor. A number of such species live inside the tissues of corals, clams, and other **benthic** animals in a **symbiotic** relationship (**Chap. 14**).

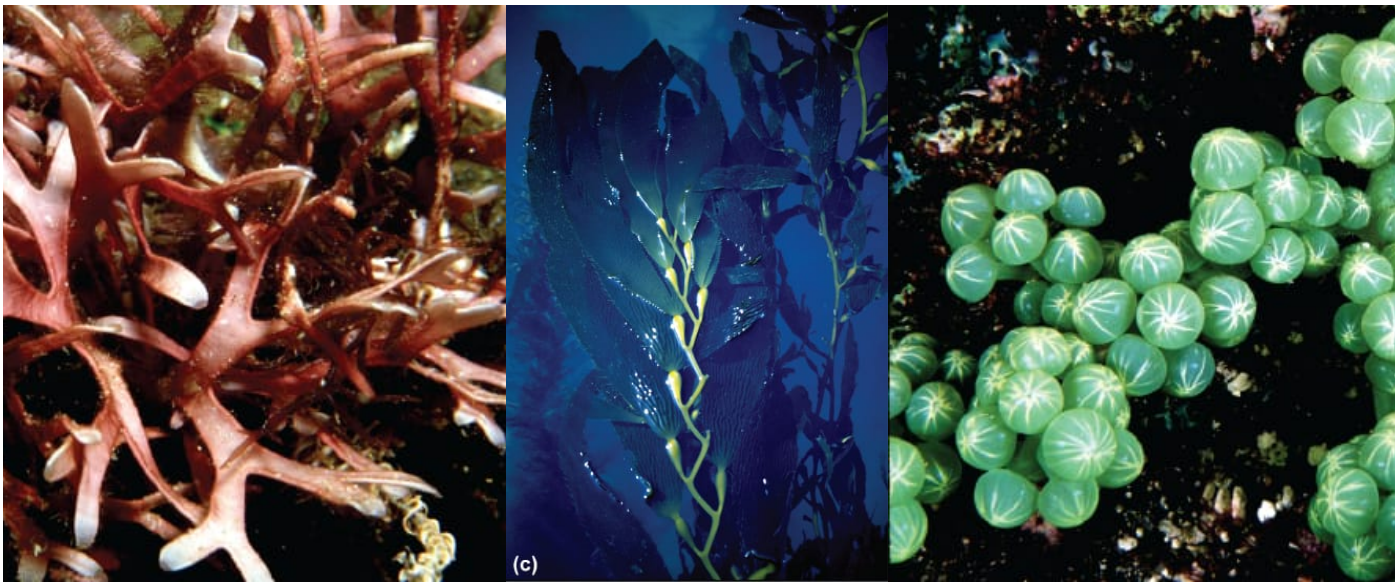


Figure 13-4. Benthic algae. (a) Various species of green, brown, and red macroalgae in a tide pool on the California coast near San Francisco. (b) A red calcareous macroalga (phylum Rhodophyta, Philippines). (c) The upper fronds of kelp (*Macrocystis sp.*) floating on the water surface in Monterey Bay, California. (d) A coral reef green macroalga called "grapeweed" (*Caulerpa racemosa*, Papua New Guinea). (e) An encrusting red alga (phylum Rhodophyta, Hawaii) covering a platelike hard coral.

The photic zone is generally shallower and more variable in depth in the coastal oceans than in open-ocean waters because the concentration of suspended particles, and hence turbidity, is higher. Turbidity is higher primarily because of the proximity of river discharges that can carry large quantities of **suspended sediment**. The finest particles are continuously **resuspended** and transported in the coastal zone by waves and currents, so their concentration remains high. Once they have been transported to the lower-energy deep oceans, these particles sink below the photic zone.

The concentration of suspended sediment in coastal waters is variable because it depends on such factors as location in relation to river discharges, river discharge rates, intensity of wave resuspension activity, and the speed of coastal currents. Each of these factors varies spatially and temporally. Similarly, phytoplankton concentrations are highly variable in the coastal zone. As a result, the photic-zone depth can vary substantially over short distances and time periods. Populations of benthic algae, particularly in the relatively deep waters of the mid-shelf region, vary accordingly. High turbidity also affects the health and distribution of **coral reefs** by reducing the light available for photosynthesis by their symbiotic algae, the **zooxanthellae**.

Currents

Because coastal currents (**Chap. 8**) are dominated by local winds, they vary seasonally and on shorter timescales in response to weather systems. Coastal currents are generally steered by their interaction with the seafloor, so they flow parallel to the coastline. However, interactions between currents and the seafloor or **shoreline** can also cause permanent or semipermanent **eddies** to form.

Nutrients

The supply of nutrients to the coastal photic zone is generally greater and more variable than the supply to open-ocean photic zones. The sources of, and variations in, nutrient supply are important to the biological characteristics of coastal waters and are

discussed in detail in the next section.

This page titled 13.1: Special Characteristics of Coastal Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.2: Nutrient Supply to the Coastal Photic Zone

The most productive areas of the oceans are in coastal regions (**Fig. 12-13**). With only a few exceptions, coastal waters have higher **primary productivity** than the adjacent open-ocean waters because a variety of mechanisms supply or resupply nutrients to the photic zone. The most important nutrient supplies are provided by river inputs and coastal upwelling.

Some rivers discharge large quantities of nutrients into coastal waters, where currents carry water parallel to the coastline so that it mixes only slowly with offshore waters. Consequently, nutrients discharged by rivers tend to be distributed along the **coast** in the direction of the prevailing coastal current. The influence of river-borne nutrients and coastal-current transport can be seen in the high primary productivity near the Amazon River mouth in Brazil (**Fig. 13-5**).

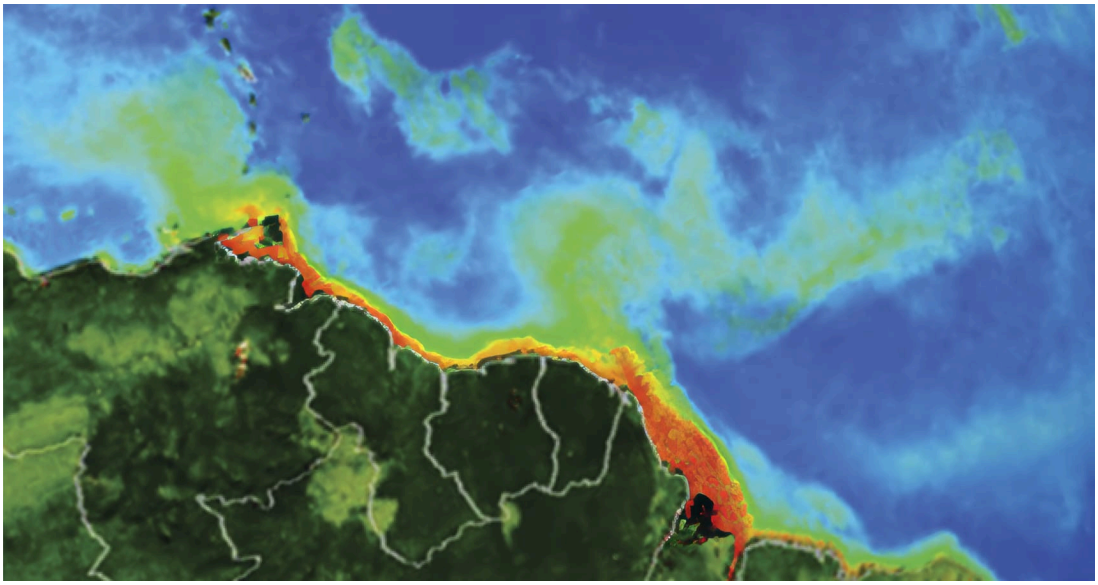


Figure 13-5. This false color–enhanced satellite image of the Atlantic coast of South America depicts the elevated (red colors) chlorophyll concentrations associated with the nutrient inputs in the outflow of the Amazon and Orinoco Rivers. The river outflow is transported north along the coast by the coastal currents.

River-borne nutrients are important in many areas, but in other areas, much larger quantities of nutrients are supplied to the coastal photic zone by wind-driven coastal upwelling (**Chap. 8**). In addition, various other physical mechanisms mix nutrients into photic-zone waters of the continental shelf.

Wind-Driven Coastal Upwelling

Coastal upwelling occurs when wind-driven **Ekman transport** (**Figs. 8-4, 8-6**) moves the surface water layer offshore. Coastal upwelling brings offshore water from depths of 100 m or more to the surface to replace the mixed-layer water that is transported offshore. If the water raised to the surface has high nutrient concentrations and replaces nutrient-depleted surface water, and if the upwelling occurs where photosynthesis is not **light-limited**, the nutrients enhance primary productivity. In coastal-ocean regions where these conditions are met and where winds that cause the upwelling are persistent, primary productivity is high and remains high while upwelling persists (**Fig. 13-6**). The term "**coastal upwelling** region" is usually applied to such areas. Ekman transport also causes upwelling in other areas where the upwelled water originates from above the permanent thermocline. This water is typically poor in nutrients, and the upwelling has little effect on productivity.

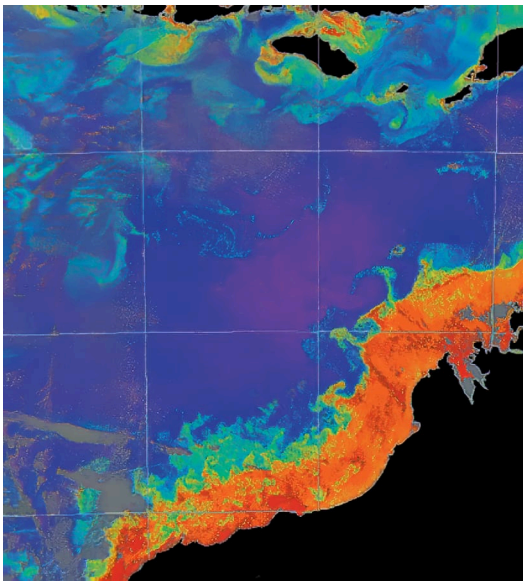


Figure 13-6. Surface water chlorophyll concentrations in the Indian Ocean between Indonesia and northwestern Australia (June 1981). High concentrations are shown in red and are associated with high rates of primary production. The high productivity off northwestern Australia occurs because winds cause offshore Ekman transport and coastal upwelling. Note the complex local variations along the coast. The productivity is much lower near Indonesia, where there is no major coastal upwelling.

Nutrient-rich water from below the permanent thermocline generally does not penetrate far onto the continental shelf off middle latitude east coasts because of the swiftly flowing, deep **western boundary currents** of the ocean gyres (**Fig. 8-10**). In contrast, nutrient-rich water does penetrate onto the continental shelf off the west coasts of the continents. Hence, high-productivity coastal upwelling regions are concentrated off the west coasts of the continents, notably off California, Peru, and the western Africa coast both north and south of the equator (**Fig. 12-13**).

The intensity of upwelling varies locally in coastal upwelling regions (**Fig. 13-6**). It is especially strong and persistent off coasts where prevailing winds produce offshore Ekman transport in regions with dry or desert climates, such as California and Peru. In some other coastal regions, winds cause offshore Ekman transport, but river inputs of freshwater are substantial. In these areas, the low-salinity water discharged by the rivers spreads to form a low-density, relatively shallow surface layer. Because this layer "slides" relatively easily over the higher-density water below it, offshore Ekman transport tends to spread it farther offshore to be replaced by additional low-salinity water from the river. The low-salinity, low-density surface layer inhibits upwelling. If Ekman transport is strong and river flow rate relatively low, upwelling does occur, but the upwelled water may be low-nutrient shelf water from below the shallow halocline rather than high-nutrient water from below the deeper permanent thermocline. Upwelling of nutrient- rich water from below the permanent thermocline can occur only if the winds and offshore Ekman transport are persistent and strong, and if river input remains relatively low.

For coastal upwelling to occur, winds must blow in the appropriate direction long enough to initiate Ekman transport, which then must move sufficient surface water offshore to bring nutrient-rich waters to the surface. This process may take many hours or several days, depending on the wind strength and direction and the depth of the thermocline. In locations where winds are variable in strength and direction, upwelling may be sporadic and highly variable from day to day and from year to year. Coastal upwelling is generally most consistent in the **trade wind** region because trade winds are less variable than winds at other latitudes (**Chap. 7**).

In many regions, such as off the California coast, climatic winds and coastal currents change direction seasonally (**Fig. 8-15**) and upwelling is also seasonal. For example, upwelling occurs during summer and fall off the California coast, when winds blow from the north. Seasonal upwelling may occur each year, but the timing, intensity, and persistence of upwelling vary from year to year. These variations are important to major fisheries that depend on the upwelling.

Coastal upwelling often extends farther offshore and is stronger and more persistent near capes (a headland, peninsula, or promontory that extends out into the ocean) (**Fig. 13-7**). Fishers have long known about this phenomenon because capes have especially abundant fish, seabird, and often **marine mammal** populations. The reason is still not totally understood, but the phenomenon appears to be the result of an interaction of coastal currents with the undersea **topographic** ridge that usually extends onto the continental shelf as an extension of the cape. The coastal current flows parallel to the coastline, and as it meets the ridge, it is deflected first offshore and then back toward the **shore** as it passes up and over the ridge. This deflection sets up

a complex eddy circulation similar to the mesoscale eddies discussed in **Chapter 8**. Upwelling is especially intense immediately downstream of the ridge, but upwelled water is also entrained in the eddy circulation and carried farther offshore.

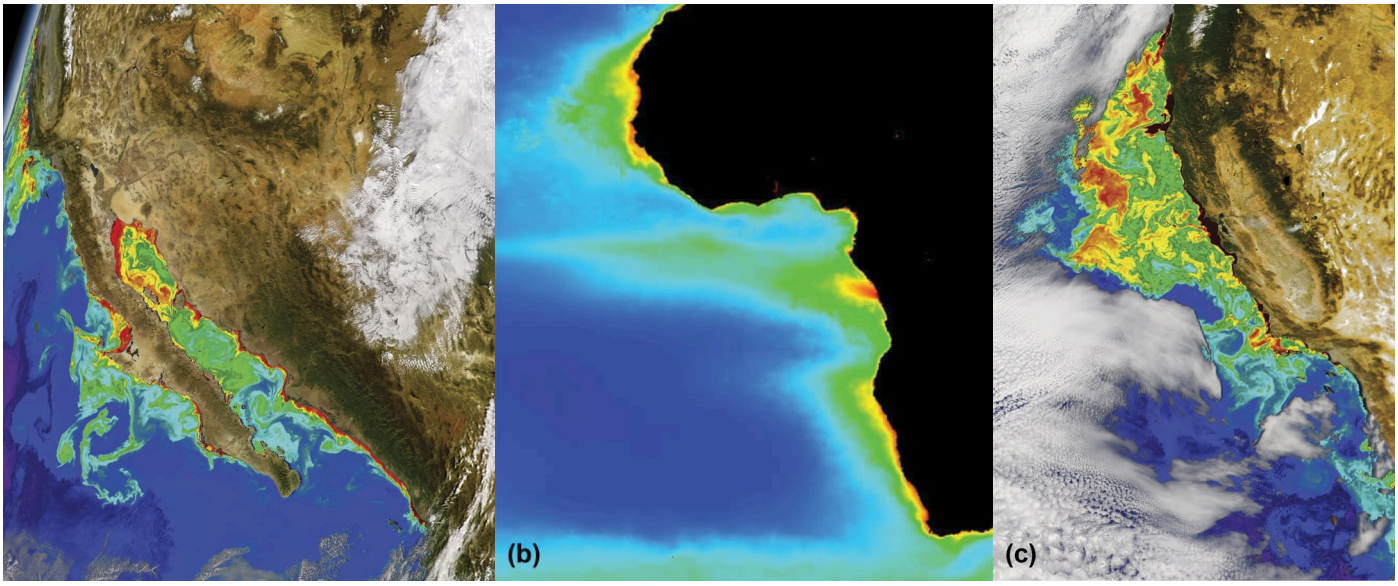


Figure 13-7. These false color–enhanced satellite images show the elevated (red colors) chlorophyll concentrations associated with the interaction of coastal currents and capes. (a) The complex eddies that form at capes and that support enhanced primary productivity are clearly seen in this monthly composite image of the coastal region off Baja California. (b) Upwelling is especially strong and extends farther offshore at each of the several capes on the western African coast. (c) This image shows the intense upwelling that sustains the abundant marine species populations of the California coastal zone. This, and other eastern boundary current coastal upwelling zones, are difficult to capture from satellites since they are often obscured by the marine layer clouds and fog that you can see offshore in the image and that characterize these zones.

Biology of Coastal Upwelling Zones

In most highly productive coastal upwelling regions, upwelling persists for several months or more each year. During this period, continuous, although variable, onshore–offshore transport processes are established (**Fig. 13-8**). Surface layer water flows offshore due to Ekman transport and is replaced by cold, nutrient-rich water that flows inshore near the seafloor. As the upwelled cold water is transported offshore, it is progressively warmed by solar heating and mixed with warmer offshore surface water.

Table 13-1. *Biology in upwelling regions.*

	Upwelling Region	Mixing Region	Offshore Region
Primary Productivity	Low	High	Low
Plankton	Few. Mostly small diatoms and flagellates. Few zooplankton.	Abundant. Mostly large diatoms. Abundant zooplankton. Mostly copepods.	Few. Mostly flagellates. Zooplankton abundance decreasing seaward. Copepods progressively replaced by larger euphausiids.

	Upwelling Region	Mixing Region	Offshore Region
Fish	Few. Mostly small omnivores.	Abundant. Mostly small herbivores (anchovy, sardine).	Abundance decreasing seaward. Small herbivores progressively replaced by larger carnivores (mackerel, hake).

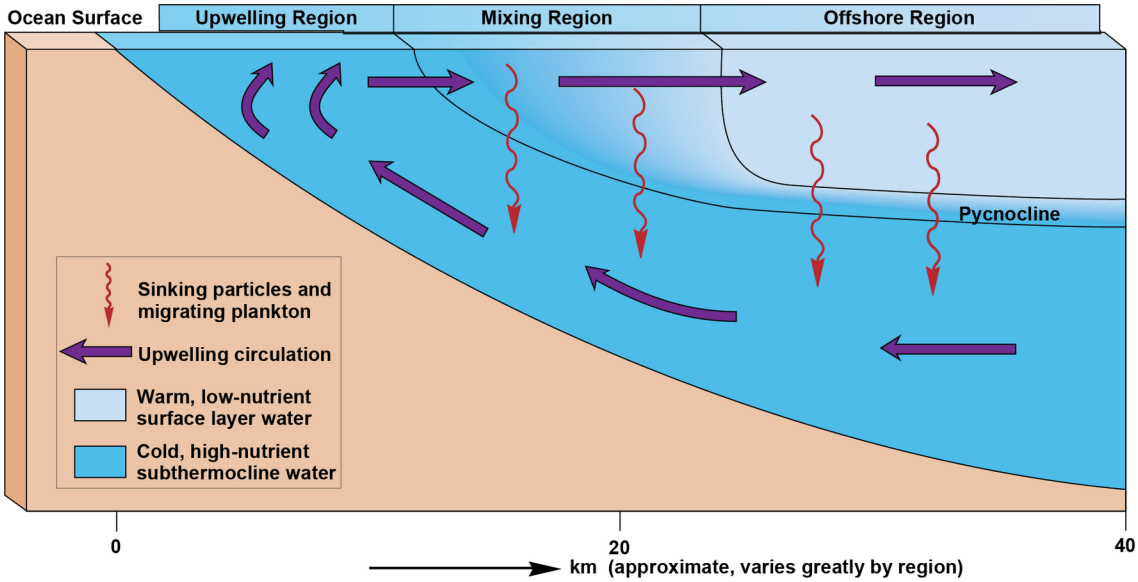


Figure 13-8. In coastal upwelling regions, the distribution of water properties and biota is related to the upwelling circulation. Upwelled water, high in nutrients, supports only low levels of primary production at first, but as it moves offshore as the surface layer, it supports a diatom bloom that depletes the nutrients, after which flagellates become dominant. Zooplankton and fish species are distributed to take advantage of this distribution of food. Many species of plankton use the upwelling circulation by changing their depth to be transported either onshore or offshore. Many fish and invertebrate species also use the circulation by placing their eggs and larvae at the right place and depth so that they are transported to a location where the appropriate food source is available at the different stages of their life cycle.

Although the newly upwelled water near the coast has high nutrient concentrations and adequate light, phytoplankton growth does not begin immediately. There may be several reasons for the time lag, including the following:

- Newly upwelled water contains few viable phytoplankton or other primary producer cells.
- Dissolved **micronutrient** organic compounds may be lacking in newly upwelled water.
- Dissolved trace metals in the newly upwelled water may be in a more toxic ionic form because the concentration of dissolved organic compounds is low.
- Dissolved nutrient trace metals, such as iron, may be in an ionic form unavailable to some species of primary producers, again because the concentration of dissolved organic compounds is low.

The toxicity and biological availability of metals in solution are known to be changed by chelation, a process in which the metal atom or **ion** is surrounded by an organic molecule or molecules and stable bonds are formed with part of the organic molecule structure. Generally, toxicity is decreased by chelation, and the metal may become more readily available for uptake by living organisms. In laboratory experiments, the addition of a simple organic compound known to be a chelating agent to newly upwelled water reduces the lag time. Therefore, either toxicity or unavailability of metals must account for at least some of the lag time in phytoplankton blooms. In newly upwelled water, a variety of small **flagellate** phytoplankton species and cyanobacterial communities are apparently less affected by the factors that create the lag time. These species are the first to grow and reproduce actively in the newly upwelled water. In doing so, they are thought to synthesize a variety of organic compounds that they then, at least in part, release to solution. These compounds must include chelating agents and/or the micronutrients needed by other phytoplankton species, which then allows those species to grow.

As upwelled water moves offshore, larger phytoplankton, including chain-forming **diatoms**, reproduce rapidly. In the mid-shelf

region, these blooms peak, and phytoplankton **biomass** reaches more than 100 times that present in nutrient-poor surface waters farther offshore. The **blooms** consume and thus deplete nutrients (nitrogen, phosphorus, and silica) in the previously upwelled water. As the nutrient-depleted water is transported farther offshore and mixed with offshore surface waters, the phytoplankton, which are now **nutrient-limited** and no longer actively growing, are reduced by **grazers**. Consequently, the larger phytoplankton are replaced by smaller numbers of flagellate species that are able to grow at low nutrient concentrations.

Zooplankton and **nekton** populations within the coastal upwelling ecosystem are distributed to take advantage of the local food supply. Few zooplankton or nekton are present in newly upwelled water, where primary productivity is still low. The mid-shelf area where concentrations of diatoms are highest has large populations of **herbivorous** fishes and zooplankton. Farther offshore, larger **carnivorous** zooplankton and carnivorous fishes feed on the herbivores brought to them by the offshore flow. These **food webs** are short in comparison with open-ocean food webs (**CC15**). Hence, coastal upwelling fisheries use photosynthetically produced organic matter more efficiently than open-ocean fisheries do (**CC16**).

Fecal pellets and **detritus** produced by organisms in the surface water layer of the upwelling zone fall below the surface layer into the near-bottom water or to the seafloor, where they are decomposed. Nutrients released by this decomposition are added to the water that moves inshore to be upwelled. Thus, upwelling-zone circulation ensures that nutrients are rapidly recycled.

Coastal upwelling circulation is also important to the life cycles of species in the upwelling-zone ecosystem, particularly **plankton**, because they are continuously transported offshore in the mixed layer. If they are carried offshore beyond their optimal location, members of these species can return by sinking through the water column so that they are carried onshore by the upwelling circulation. Many phytoplankton species form resting phases called **spores** or **cysts**, and zooplankton species lay eggs in offshore waters, where they sink into cold subthermocline water that moves onshore. Once upwelled into an area where nutrients or food is available, the phytoplankton revive and zooplankton eggs hatch.

Nekton use a similar strategy to distribute their eggs and **larvae** to favorable locations. Many nekton species migrate inshore or to estuaries to **spawn**, while others may spawn offshore. Their eggs are carried in the near-bottom waters toward shore. They hatch into larvae that are then upwelled into nutrient-rich waters, timed with the large phytoplankton bloom. The larvae may change their physical form and feeding strategy several times to exploit different food sources as they are transported offshore as plankton. A final change to the adult nektonic form occurs when an appropriate location in the upwelling circulation is reached.

Many upwelling-zone species rely on the upwelling circulation to deliver their eggs and larvae to locations where appropriate food is available at specific times during their life cycles. Consequently, in years when upwelling occurs at a different time or differs in intensity and duration, some species may be advantaged or disadvantaged in relation to others. The result is considerable year-to-year variation in the breeding success of many species. The cause of the collapse of fisheries for many species has likely been that they were overfished during one or more years when environmental factors substantially reduced reproduction.

Other Mechanisms of Nutrient Supply

Coastal upwelling and runoff from the land are the two most important mechanisms of nutrient supply to the coastal photic zone. However, there are other mechanisms of nutrient supply that depend on waves and tides, the interaction of ocean currents with the seafloor, and thermal convection.

The mixed-layer depth is partially determined by the intensity and duration of winds. Winds cause vertical mixing due to Langmuir circulation (**Chap. 8**) and wave action. In regions where a seasonal thermocline is formed, intense storms in summer or fall may generate waves with long **wave periods** and wave heights that disrupt the thermocline. This disruption mixes water from above and below the thermocline, weakens the thermocline, and transports some nutrient-rich waters upward into the mixed layer. Nutrients are also returned to the mixed layer by convection that occurs when the surface layer water cools in fall-winter and sinks, causing the seasonal thermocline to disappear (**Fig. 13-9c**). Where the seafloor is shallower than the permanent thermocline depth, detritus falls through the photic zone and accumulates on the shallow seafloor. As this detritus is decomposed, nutrients are released into near-bottom waters, from where they can be returned to the photic zone by vertical mixing.

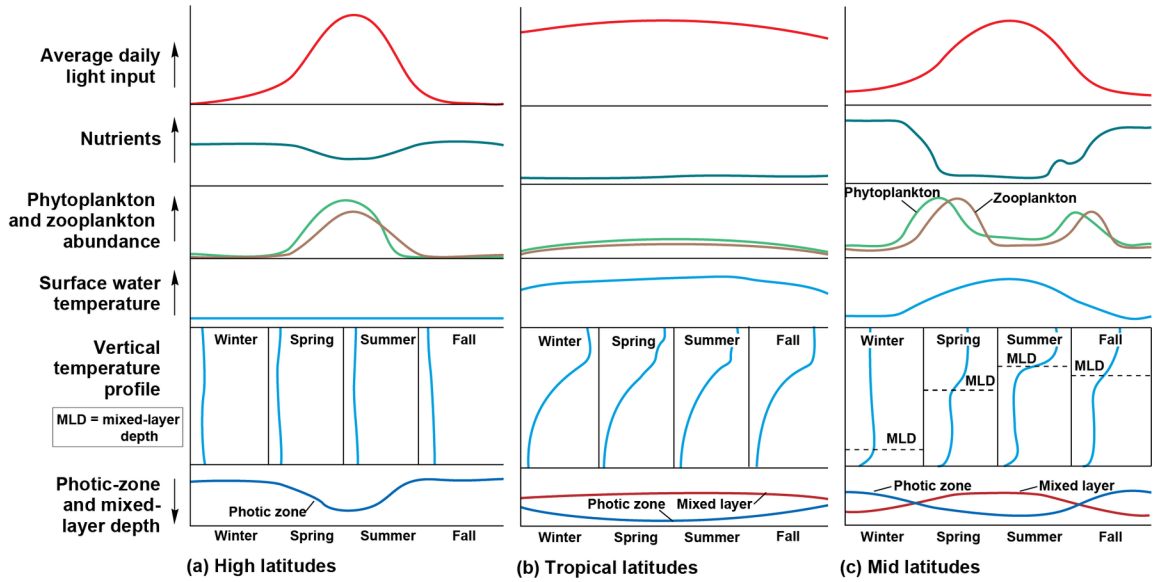


Figure 13-9. Seasonal variations in solar intensity, nutrient concentrations, phytoplankton and zooplankton biomass, and surface water temperature are different in middle (temperate), high, and tropical latitudes. (a) In high latitudes, there is an intense plankton bloom in the short summer when light intensity is sufficient for growth. Nutrients are not depleted, because strong vertical mixing occurs and this mixing prevents the formation of a thermocline. (b) In tropical latitudes, a strong permanent thermocline exists. This thermocline inhibits vertical mixing, and nutrients in the surface layer are permanently depleted. As a result, primary production is limited to uniformly low levels throughout the year. (c) In middle latitudes, there is a distinct seasonal cycle in which a shallow thermocline forms in summer, a spring bloom of plankton depletes the nutrients, and a smaller fall bloom sometimes occurs as nutrients are recycled.

Long-**wavelength** internal waves that form on the thermocline are slowed as they reach shallow water, and they eventually break on the outer portion of the continental shelf (**Chap. 9**). In some areas, such as the outer continental shelf next to Long Island and New Jersey, these waves may transport nutrient-rich subthermocline water into the mixed layer.

Tides in certain regions generate swift currents, particularly in bays, estuaries, and gulfs. Turbulence created by these currents can prevent a seasonal thermocline from forming. Thus, nutrients are recycled more effectively in such locations.

Where currents encounter seafloor topographic features, such as ridges, plateaus, **seamounts**, banks, or islands, water must flow over or around the feature, and nutrient-rich deep water can be mixed into the photic zone by complex eddies and turbulence. The increase in productivity caused by the interaction of ocean currents with islands can be seen in the increased **chlorophyll** concentrations around some such islands (**Fig. 13-10**).

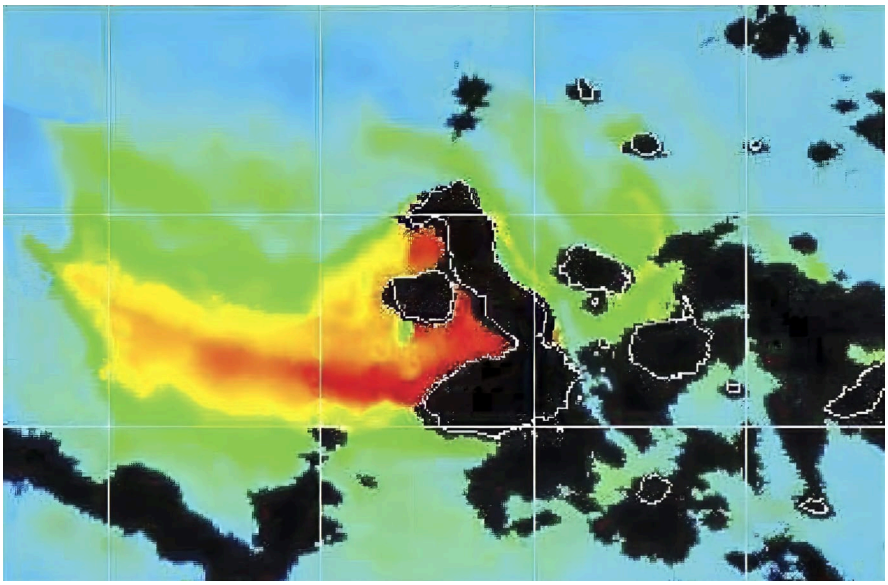


Figure 13-10. Elevated chlorophyll concentrations are found around islands, including the Galápagos Islands, which are depicted in this false color–enhanced satellite image. Primary productivity is enhanced by the vertical mixing that is caused by the interaction of ocean currents with the island topography, which brings nutrients to the surface layer, and by nutrients in freshwater runoff from the islands.

When surface mixed-layer water is cooled, it becomes dense and sinks while convective motions replace it with nutrient-rich water from below. Convective mixing is extremely important in the high-latitude oceans (**Fig. 13-9b**). In these regions, surface waters are cooled continuously except in midsummer, so there is no thermocline, and convective mixing is nearly continuous. When light intensity is sufficient, these regions sustain intense phytoplankton blooms that are rarely slowed by nutrient depletion.

Georges Bank off the New England coast is an extremely productive fishery, primarily because of a plentiful supply of nutrients. The bank is a huge shallow **shoal** in the mouth of the Gulf of Maine (**Fig. 13-11**). Cold, nutrient-rich water flows out of the Gulf of Maine and intersects the north side of the bank, where it is deflected. Some water flows over the bank, but most forms a huge eddy, and a large fraction (approximately 10% to 30%) flows all the way around the bank. Wind waves, tidal currents, and turbulence and upwelling caused by deflection of the water mass into and around the bank continuously inject nutrients into the eddy and the water on the bank.

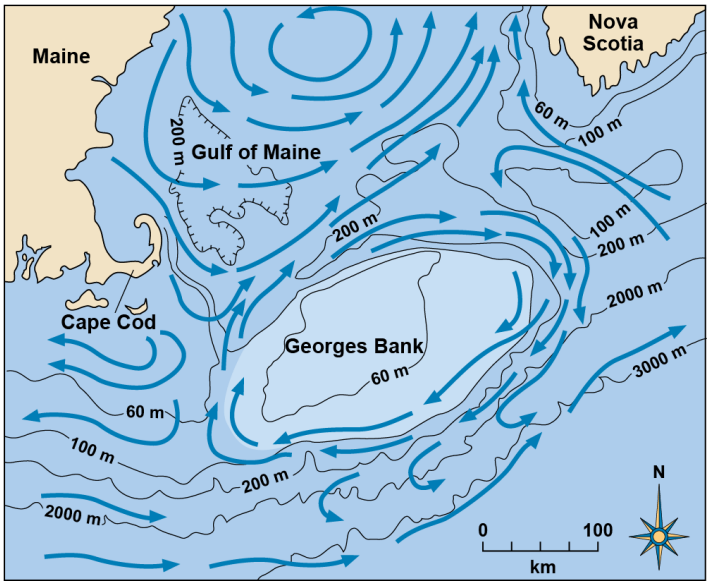


Figure 13-11. The water circulation around Georges Bank, New England, is characterized by two permanent current patterns. One is a clockwise-rotating fixed gyre that runs around the edge of Georges Bank. The other is a counterclockwise gyre in the Gulf of Maine. The circulation around Georges Bank causes upwelling of nutrients into the water over the Bank and also helps to retain plankton over the Bank so they are able to reproduce to high population densities.

Productivity is especially high because the gyre that circulates around the Bank enables phytoplankton to remain in the water on the Bank, where nutrient supplies are high. The residence time of water and plankton on the Bank is about 2 to 3 months, so phytoplankton may go through many cell divisions before being swept off the Bank. The phytoplankton provide abundant food for zooplankton, which also remain on the bank long enough to feed and reproduce, ensuring that their young have ample food. Fishes, in turn, exploit the abundant phytoplankton and zooplankton. Unfortunately, Georges Bank fisheries declined substantially because of overfishing, especially prior to 1994 when some fishing restrictions were put in place. The populations of commercially valuable species of fishes have recovered very slowly but not yet to their previous abundance.

This page titled 13.2: Nutrient Supply to the Coastal Photic Zone is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.3: Seasonal Cycles

Substantial seasonal variations in primary productivity occur in middle and high latitudes, but not in the tropics and subtropics (**Figs. 12-13, 13-9**). Seasonal variations are complex and controlled by the availability of light and nutrients and the depth of the mixed layer, each of which varies with latitude, location, and year-to-year climatic variations. As a result, seasonal cycles in high latitudes, middle latitudes, and tropical latitudes have different general characteristics that are representative of each of these broad regions, but there are no sharply defined latitudinal boundaries between areas with these characteristics, and characteristics are substantially modified by local conditions. Seasonal cycles are generally similar in coastal and open-ocean waters within each latitudinal region. However, seasonal cycles are especially important in coastal and estuarine ecosystems.

Polar and Subpolar Regions

Figure 13-9a is a simplified representation of the annual cycles of light intensity, nutrient concentrations, phytoplankton biomass, zooplankton biomass, and water temperature in subpolar seas remote from major freshwater inputs. In these regions, the water column remains well mixed year-round. Frequent storms and low light intensity (limited solar heating) prevent the formation of a seasonal thermocline, and surface cooling produces year-round convective circulation, so there is no permanent thermocline. Nutrients are plentiful because they are supplied continuously by this convective mechanism, but light intensity is sufficient to sustain phytoplankton growth only during a short period in summer, when an explosive bloom of mostly diatoms occurs. Because nutrients are continuously resupplied and do not become limiting, the bloom continues unchecked, except by predation, until light intensity declines at the end of summer.

In polar regions, subpolar regions near land, and areas of seasonal sea ice, this seasonal cycle is modified by freshwater inputs (**Fig. 13-9a**). During spring and summer, substantial quantities of freshwater mix with surface waters. The freshwater comes from river runoff fed by snowmelt and from the melting of seasonal sea ice or **glacial** ice. It is mixed with surface ocean water to form a cold but low-salinity surface layer separated from the water below by a halocline. In many areas, the surface layer has high turbidity due to suspended particles from river or glacial ice inputs. High turbidity and low sun angle restrict the photic zone to a very shallow depth. Because the halocline restricts the vertical movement of nutrients and the photic zone does not extend below this layer, primary production quickly becomes nutrient-limited after an initial bloom that occurs in early summer when light intensity first increases. Frequent and intense storms in many subpolar regions vigorously mix subpolar seas, break down the halocline, and resupply nutrients to the surface layer. Hence, primary production is nutrient-limited primarily in protected inshore regions such as fjords.

Productivity is also nutrient-limited throughout most of the Arctic Ocean that is ice-free in summer (**Fig. 8-27**). With the exception of a few areas of upwelling, these ice-free waters have a steep, almost permanent halocline a few meters below the surface for several reasons. First, freshwater input from rivers and from the melting of the permanent ice pack is large in volume and continuous during the short summer. In addition, salt is removed from the surface layer by **ice exclusion** (**Chap. 8**), which creates high salinity brines that sink below the surface layer. Water exchange between the Arctic Ocean and the Pacific and Atlantic Oceans is limited, so the low salinity Arctic Ocean surface water is transported to lower latitudes very slowly. Furthermore, because the Arctic coast is close to the polar atmospheric downwelling area (**Chap. 7**), there are relatively few storms in summer. Storms that do occur have only a limited fetch in which to build wind waves because the ice-free Arctic Ocean is restricted to the zone between the polar ice and the coast. The continuing loss of permanent sea ice in the Arctic Ocean is modifying these factors such that the nutrient-limited area in the Arctic Ocean is declining, producing significant shifts in the distribution and composition of Arctic species.

In the Southern Hemisphere, the oceans around Antarctica do not have a halocline like the Arctic Ocean halocline, because Antarctica is a desert and the continental ice sheet melts comparatively little during summer. The very limited freshwater input from the melting of sea ice and from runoff is readily and rapidly mixed and transported into the much larger volume of the Southern Ocean. Because persistent haloclines do not form in the oceans around Antarctica, the area has higher annual productivity than the Arctic Ocean ice-free regions have.

Tropical Regions

In the tropical and subtropical oceans, there is generally little seasonality of plankton growth because of the general uniformity of light intensity year-round and a very steep permanent thermocline that begins below a relatively shallow mixed layer (**Fig. 13-9b**).

Hence, light intensity is always high, and the photic zone is deep. However, nutrients transported below the relatively shallow mixed layer are removed from the photic zone. Consequently, everywhere in tropical regions other than in upwelling regions, primary productivity is low because of nutrient limitation.

Although productivity is nutrient-limited throughout the tropical oceans, most coastal areas are characterized by extremely rich and abundant coral reef communities, except where turbidity is high or salinity is variable because of runoff. Despite the extremely low nutrient concentrations in the water column above, the coral communities thrive as a result of a unique symbiotic relationship between algae and **reef**-building corals.

Reef-building corals consist of millions of individual tiny animals whose **hard parts** are cemented together to form the coral mass (**Fig. 13-12**). Each coral animal, or **polyp**, contains within its tissues a large number of dinoflagellate algae known as zooxanthellae. The relationship between coral and zooxanthellae is complex and not fully understood, but one of the principal effects of, or reasons for, this partnership is thought to be more efficient utilization of the small amounts of nutrients available in tropical waters. The zooxanthellae and coral polyps continuously and rapidly recycle nutrients between themselves, thus retaining these nutrients in the photic zone.

Figure 13-12. Reef-building corals. (a) A typical reef-building hard coral colony (probably *Diploastrea heliopora*, Papua New Guinea). (b) Another typical hard coral (possibly *Pocillopora* sp., Papua New Guinea). (c, d) Individual polyps of *Diploastrea heliopora* (Papua New Guinea) are photographed (c) during the day when the polys are not open, versus (d) during the night, when the polyps have extended their soft tentacles to feed on zooplankton or particles.

Reef-building corals can grow only in the photic zone, where their zooxanthellae can photosynthesize. Hence, these corals are present only in relatively shallow water. In high-turbidity regions, the depth of the photic zone is reduced and coral reef communities do not develop. Coral polyps are adversely affected by high concentrations of suspended particulates, as they have limited ability to remove particles that are deposited on them to avoid being smothered. The high turbidity associated with high particle concentrations also reduces the amount of light available for their zooxanthellae to photosynthesize. Thus, reef building corals die if high concentrations of suspended sediment are sustained. Increased turbidity due to human activities has destroyed or damaged many such coral reef communities (**Chap. 16**).

Mid-Latitudes

During winter in the mid-latitudes, cooling of surface waters and mixing by wind waves eliminate the seasonal thermocline and mix water that has high nutrient concentrations from below the thermocline into the mixed layer (**Fig. 13-9c**). Phytoplankton are distributed almost randomly throughout the mixed layer because of their limited swimming capabilities. Because the mixed layer is much deeper than the photic zone, phytoplankton are within the photic zone during only a small percentage of the time. Although they can still photosynthesize for some depth below the photic zone, they cannot photosynthesize enough to replace the biomass they use for respiration, and they must survive long enough to return to the surface layer for growth and reproduction to resume. Therefore, during winter, phytoplankton growth is light-limited, populations are low, and many species adopt a hibernation-like resting phase.

In spring, light intensity increases and the photic zone becomes deeper. At the same time, storms are reduced, surface waters begin to warm, the mixed-layer depth is reduced, and a seasonal thermocline begins to form. Phytoplankton are now in the photic zone for an increased percentage of their time, and nutrients remain plentiful. The phytoplankton begin to photosynthesize, grow, and reproduce rapidly.

The **standing stock** (biomass) of phytoplankton increases rapidly at the beginning of the spring bloom period, but the zooplankton population, which now has an increasing food supply, also begins to feed, grow, and reproduce rapidly. Consequently, the majority of new phytoplankton cells are eaten by zooplankton, and the rapid reproduction of phytoplankton does not lead to sustained high phytoplankton concentrations (**Fig. 13-13**). The standing stock of phytoplankton represents only a small percentage of the total production because many of them are lost to consumption by herbivores or to sinking.

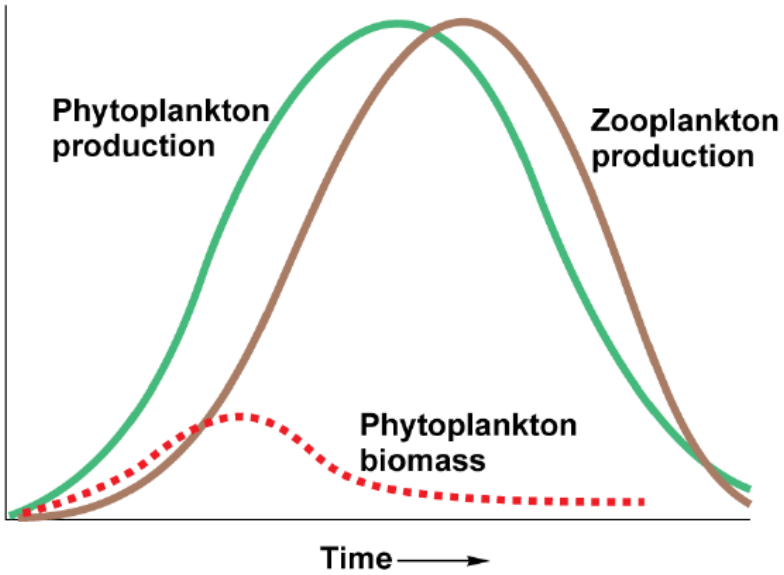


Figure 13-13. During a phytoplankton bloom, such as the spring bloom in mid-latitudes, the production rate of new biomass by phytoplankton rises rapidly and is followed, with a time lag of a few days, by a similar increase in zooplankton production. The phytoplankton biomass (standing stock) increases somewhat during the early part of the bloom, but it is rapidly reduced by zooplankton feeding, even though the phytoplankton production rate continues to increase.

As spring continues, phytoplankton rapidly deplete nutrients in the mixed layer. In addition, the mixed layer becomes shallower (about 10 to 15 m) as the seasonal thermocline forms. At this time, the phytoplankton have adequate light but are nutrient-limited. Consequently, their growth slows and the spring bloom collapses as zooplankton continue to feed on them. Soon, the zooplankton population also declines as its food supply decreases, and the zooplankton continue to be eaten by their predators.

During summer, primary productivity continues at a low level, supported by nutrients recycled by the consumers and **decomposers** in the mixed layer. However, most of these nutrients are recycled below the seasonal thermocline, where light levels are too low to support significant phytoplankton growth.

In fall, cooling and winter storms again weaken the seasonal thermocline, and nutrients are mixed into the photic zone from waters below. The addition of nutrients often supports a fall bloom of phytoplankton. However, the photic zone is now shallower because of the declining light intensity. Variations in timing, location, and year-to-year climate are particularly important in determining the magnitude of the fall bloom. If nutrients are mixed into the surface layer before the light intensity declines too much and before the mixed layer becomes too deep, a fall bloom occurs. Fall blooms are generally smaller and of shorter duration than spring blooms. If the nutrients are not mixed into the surface layer until late fall, when light intensity is already very low, no fall bloom occurs. In addition, no fall bloom occurs if intense cooling or strong storms eliminate the seasonal thermocline early and create a continuously mixed, very deep mixed layer.

In early winter, the enhanced mixing by storm winds, surface water cooling, and declining light levels combine to return the system to its winter state, in which the phytoplankton are light-limited.

Each year, at each location, the seasonal phytoplankton cycle is different from that of the previous years because of variations in weather. In addition, the details of the cycle vary in response to other influences, such as the composition of planktonic species, and the concentrations of nutrients other than nitrogen and phosphorus (particularly the concentration of silicate). The subtle

and complex interactions between ocean physics, chemistry, and biology that control the seasonal plankton cycle are further complicated in some coastal regions by other processes, such as the introduction of suspended sediment and nutrients by rivers and coastal upwelling.

Phytoplankton Species Succession

Many herbivores are selective feeders that require or prefer certain characteristics in the phytoplankton they eat. Most commonly, the requirement or preference is that the food species be within a range of sizes suited to the animals' feeding methods. Larger zooplankton and juvenile fishes generally require large phytoplankton cells that are easy to capture and provide larger amounts of food. Smaller zooplankton specialize in eating smaller phytoplankton. Most coastal **food chains** that lead to commercially valuable fish and **shellfish** species are short and based primarily on larger phytoplankton species that are eaten by zooplankton and juvenile fishes. Food chains based on the smaller phytoplankton species are generally longer and less efficient (**CC15**), and support fish and shellfish species of generally lesser commercial value. Microbial food chains may be short, but they are thought to be partially self-contained, with much of the organic matter produced being recycled by microbial species rather than consumed in food chains leading to commercially valuable species.

Many factors influence phytoplankton species composition, including temperature, salinity, light intensity, and nutrient concentrations. Because these factors vary seasonally, phytoplankton species composition also varies during the year in most locations. Many different species of phytoplankton are always present at any location, but one or, at most, a few species generally dominate and constitute the majority of phytoplankton. As environmental factors change, one dominant species is replaced by another in a progression called **species succession**.

Phytoplankton species succession can be highly complex and can vary, often greatly, from year to year. One species may become dominant at approximately the same time of year for many years, but may not become dominant at all in other years.

In high latitudes and in regions of persistent upwelling, nutrients are present in relatively high concentrations throughout the phytoplankton growth period. High nutrient concentrations favor larger phytoplankton species, which are predominantly diatoms. Diatoms dominate the phytoplankton community throughout the active growth period in these regions, which consequently support short, efficient food webs (**CC15**).

In mid-latitudes, nutrient concentrations vary during the summer growing period. In spring, nutrient concentrations are high, and, consequently, diatoms dominate. The spring bloom of diatoms removes most nutrients from the water column, causing the bloom to collapse and to be replaced by flagellates (**Fig. 13-14**). During summer, after the spring diatom bloom, flagellates (sometimes a succession of different species) dominate the phytoplankton community. The flagellates can obtain their nutrients from the very low concentrations that are sustained by nutrient recycling in the photic zone. In fall, storms return enough nutrients to the photic zone to support fall diatom blooms in some years and areas.

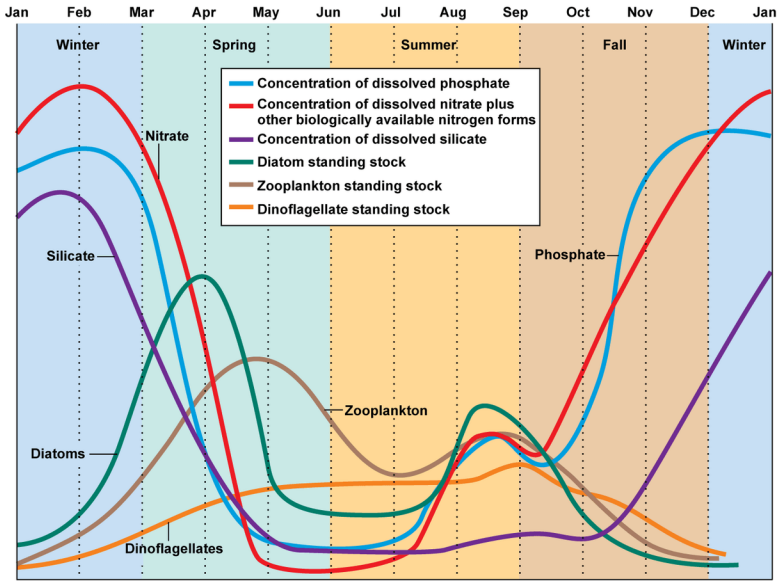


Figure 13-14. The seasonal variations of nitrogen, phosphorus, silicon, diatoms, flagellates, and zooplankton in the mid-latitude coastal oceans usually follow a similar pattern each year. The spring bloom of diatoms rapidly depletes the nutrients, and the diatoms are superseded by dinoflagellates. Because the dinoflagellates are not suitable food for many zooplankton species, the zooplankton population also declines. By utilizing nutrients recycled within the photic zone, flagellates and some diatoms continue primary production at relatively low levels throughout the summer. A fall bloom of diatoms often occurs when storms and cooling weaken the thermocline and return some nutrients to the surface layer by mixing. In late fall, the light intensity becomes too low to support substantial primary production. Thus, decomposition, cooling, and wind mixing enable the return of nutrients to the surface layer before the following spring.

As a result of the seasonal availability of diatoms, many species of zooplankton and juvenile fishes that prefer or require diatoms as food have life cycles that are attuned to the seasons. For example, some fish and **invertebrate** species spawn in spring, and their eggs hatch into larvae that feed voraciously on diatoms. The juvenile fishes reach a sufficient size by late spring, when the diatom bloom collapses, to be able to feed on zooplankton. Some species spawn twice during the year to take advantage of both spring and fall blooms.

Many mid-latitude fish and invertebrate species have life cycles that depend on the availability of abundant diatoms as food for their juveniles during only a few days or weeks at a specific time of year. However, the timing, intensity, and dominant species of seasonal diatom blooms vary from year to year. As a result, the survival rate of the larval stages of many fish and invertebrate species also varies from year to year. The number of juveniles that survive their first critical year of life is called the “year class strength.” Year class strength may be high if the species spawns at the optimal time and locations in relation to the seasonal phytoplankton cycle. If year-to-year variability causes spawning and the phytoplankton seasonal succession to be misaligned in time or location in any specific year, sufficient food is not available when needed and year class strength is low.

Phytoplankton species succession is further complicated by variations in the availability of specific nutrients, particularly silicate and nitrogen compounds. During the spring diatom bloom in the mid-latitudes, silicate may be depleted before other nutrients (**Fig. 13-14**), causing diatoms to be replaced by flagellates that do not require silica. Silica is recycled extremely slowly into solution once it has been incorporated in diatom **frustules**. As a result, silica is not returned to the photic zone until the fall, when nutrient-rich waters from below the seasonal thermocline are mixed back into the mixed layer. These deeper waters have relatively high concentrations of silica dissolved from diatom frustules and **terrigenous** particles.

All phytoplankton cells require and take up phosphorus and nitrogen in approximately the same ratio as these elements occur in seawater. Similarly, diatoms require and take up silica in a reasonably constant ratio to other nutrients. The proportions of silicate, phosphate, and nitrogen compounds in water below the permanent thermocline are remarkably constant throughout the oceans. In coastal regions, terrestrial runoff normally contains high concentrations of dissolved silicate.

Consider what might happen if there were an excess of nitrogen compounds and phosphate in relation to silicate at the onset of the spring diatom bloom. The diatom bloom would deplete silicate sooner than nitrogen and phosphorus. Flagellates would replace diatoms and could bloom explosively because nitrogen and phosphorus would still be ample. This bloom might produce very large standing stocks of flagellates if, as is often true, herbivores that are able to graze the tiny flagellate cells efficiently were not abundant and could not reproduce quickly. Flagellate blooms do occur periodically, although infrequently, in limited regions where runoff has high nitrogen and phosphorus concentrations but low silicate concentrations. Blooms are becoming

more frequent and widespread in regions near centers of human population, particularly where coastal waters have long residence times. Some of these blooms are caused by or substantially strengthened by the huge volumes of treated or untreated human sewage wastes and agricultural fertilizer discharged to the oceans (**Chap. 16**). Untreated sewage has high concentrations of nitrogen compounds and phosphate but relatively low silicate concentrations and sewage treatment employed in most locations in the U.S. and worldwide does not remove nitrogen or phosphorus.

This page titled 13.3: Seasonal Cycles is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.4: Algal Blooms

Intense dinoflagellate blooms occur naturally, but infrequently, in many parts of the coastal oceans, particularly in subtropical to temperate climates. The frequency of such blooms has been observed to be increasing, and they are occurring in areas where they have not been seen before.

Dinoflagellate blooms can be caused by a variety of mechanisms. For example, a large increase in freshwater runoff can form a shallow surface layer of low-salinity water in a previously well-mixed water column. Phytoplankton that were previously light-limited because they were frequently mixed below the photic zone may be restricted to this shallow surface layer, where they are always in the photic zone. This situation may trigger a bloom. In some instances, it is thought that runoff containing dissolved inorganic or organic compounds may trigger blooms because these substances react with toxic substances, such as copper or mercury, and reduce their toxicity to the phytoplankton. Many blooms are caused by such natural events, but increasing nutrient fertilization due to human activity (**Chap. 16**) is generally believed to be a significant causative factor in the increasing frequency and severity of blooms.

Because the nitrogen compound concentration in discharges of treated sewage wastewater and of agricultural runoff and industrial waste are high in relation to the silicate concentration, such discharges are generally believed to be responsible for the higher frequency of dinoflagellate blooms, as compared to diatom blooms. However, several **cyanobacteria**, also called “blue-green algae”, convert dissolved molecular nitrogen, a nitrogen form that most phytoplankton cannot use, to usable nitrate. In some areas, these cyanobacteria may provide the nitrogen needed to initiate or sustain dinoflagellate blooms.

Dinoflagellate blooms can appear quickly but often disappear within a few days as nutrients become limiting. However, under appropriate conditions, blooms can last for many weeks if the source of nutrients is sustained. Dinoflagellates reproduce very rapidly. In optimal conditions, dinoflagellate populations can double in as little as a few hours. Accordingly, blooms can grow within a few days to concentrations of dinoflagellate cells a million or more times greater than normal. Dinoflagellates can also concentrate themselves into blooms by using their weak swimming ability to congregate at depths where light levels are optimal.

The dense concentration of phytoplankton in blooms colors the surface water yellow, green, brown, or reddish, depending on the species. Many dinoflagellates are reddish in color, so their blooms are often called “red tides.”

Dinoflagellate blooms often have adverse effects on other marine organisms and, occasionally, on human health. These effects are caused by toxic substances synthesized by certain phytoplankton species, particularly when they are stressed in bloom situations, or by the depletion of oxygen caused by the decaying organic matter that remains when the bloom collapses.

Dinoflagellate and Other Plankton Toxins

Like many terrestrial plants, certain phytoplankton species produce toxic substances as a defense against herbivores. Some, but not all, species of dinoflagellates produce particularly virulent toxins, including saxitoxin, brevetoxin, okadaic acid, and domoic acid. These substances are complex organic compounds and often mixtures of several compounds, some of which are many times more toxic than, for example, strychnine. The toxins are selective: some are toxic to **vertebrates** only, whereas others also affect some invertebrates, particularly **crustaceans** such as shrimp and crabs. Toxins produced by dinoflagellate blooms often kill large numbers of fish.

In addition to fish kills, dinoflagellate toxins can cause serious human health problems because the toxicity is selective for vertebrates. Most **filter-feeding** organisms (organisms that strain or sift food particles from the water), including clams, mussels, and oysters, are unaffected by dinoflagellate toxins but concentrate the toxins in their body tissues. Other invertebrates, such as crabs that feed on filter-feeding invertebrates, can also become contaminated with these toxins from their food without being affected. If any of these contaminated invertebrates and shellfish are eaten by vertebrates, the toxins can still cause their deadly effects in vertebrates. Unfortunately, humans are one of the vertebrate species that feed on shellfish and other invertebrates.

Human poisoning by dinoflagellate toxins concentrated in shellfish is well known. At least four major pathologies are caused by shellfish-borne toxins. The most widespread is paralytic shellfish poisoning. The toxins that cause paralytic shellfish poisoning are strong nerve poisons that can cause permanent nerve injuries or, if the concentrations are high enough, paralysis and death. There are an estimated 1600 annual cases of paralytic shellfish poisoning worldwide, and an estimated 300 fatalities among these cases. Neurotoxic shellfish poisoning is less common and less serious than paralytic shellfish poisoning, but it produces

symptoms that are often mistaken for the more common types of bacterial food poisoning. Diarrhetic shellfish poisoning is also less serious but probably more widespread than neurotoxic shellfish poisoning, and it causes severe diarrhea. Amnesic shellfish poisoning is a much rarer pathology, in which short-term memory is destroyed, perhaps permanently.

These toxins can also affect coastal communities when winds blow ashore sea spray containing dinoflagellate cells or their detritus from nearshore blooms. When this happens, the toxins cause the local human population to suffer from allergy-like reactions, such as irritated eyes, running noses, coughs, and sneezes.

Most dinoflagellate toxins are stable compounds that remain toxic even after prolonged cooking of contaminated shellfish. Once the bloom has disappeared, some shellfish lose their toxicity within days, but other species may remain toxic for months. The only certain way to avoid human poisoning by these toxins is to identify affected shellfish and prevent them from reaching the marketplace. Many countries, including the U.S., have extensive seafood monitoring programs to ensure toxin-free shellfish. Unfortunately, these monitoring programs are extremely expensive and difficult to implement. As an alternative, in areas where dinoflagellate blooms occur with reasonable frequency, shellfishing areas are closed to harvesting during the summer and fall, when blooms are most likely. Such closures lead to substantial loss of income for fishers. Areas closed to shellfishing because of dinoflagellate blooms are expanding each year, perhaps indicating an increase in the frequency and geographic extent of such blooms. For example, while historically infrequent, blooms of the dinoflagellate *Karenia brevis* now recur annually. These blooms typically develop in late summer or fall off Florida's coast. They mainly occur off the Gulf Coast, but can also be found off the Atlantic coast.

At one time, only dinoflagellates were thought to produce the powerful toxins that affect people and other vertebrates. However, blooms of at least two different diatom species have been observed to produce domoic acid. An event that occurred in 1987 in Prince Edward Island, Canada, left several victims afflicted with amnesic shellfish poisoning. The victims suffered short-term memory loss that lasted at least 5 years. A second event, in 1991 off the west coast of California, Oregon, and Washington, did not affect people. In this case, pelicans were poisoned by domoic acid produced by a diatom species that had been consumed by anchovies, on which the pelicans were feeding. Shellfisheries throughout this huge area were closed to harvesting for many weeks.

Humans and pelicans are not the only vertebrates poisoned by phytoplankton toxins. For example, in 1986, hundreds of bottlenose dolphins died off the coasts of New Jersey and Maryland after consuming fish contaminated with a dinoflagellate toxin, and more died in the same way off the Carolina coast in 1987. In these cases, the fish were either less susceptible to the toxin than the dolphins, or the dolphins accumulated more of the toxin as they fed on large quantities and were unable to destroy or **excrete** the toxin.

Other groups of planktonic organisms also produce toxins. For example, cyanobacteria produce toxins that have caused numerous fish kills in freshwater. In 2010, a toxin from cyanobacteria was found to be responsible for the deaths of sea otters. This event was the first known occurrence of hepatotoxic shellfish poisoning in the marine environment. The cyanotoxins apparently originated in freshwater rivers but were consumed by coastal shellfish, which were, in turn, consumed by the sea otters. It has been hypothesized that these toxins may be produced by bacteria or **viruses** living in the phytoplankton cells, rather than by the phytoplankton themselves. More recently, a persistent bloom of the diatom *Pseudo-nitzschia* off the Southern California coast in 2023–2025 produced domoic acid, a neurotoxin that killed hundreds of sea lions, dolphins, and whales.

Harmful algal blooms (HABs) now occur in the waters of almost every coastal state of the U.S. and are estimated to be responsible for direct economic losses of between $10 million and $100 million annually. Costs are incurred in public health, the closure of commercial fisheries, and fisheries management and monitoring. The most frequent toxic algal blooms occur off the Atlantic coast of Florida every year. It has been estimated that the impacts of these blooms alone cost tens of millions of dollars annually, and a 2018 red tide event in Florida is estimated to have an estimated total economic impact of $2.7 billion. Algal blooms in other areas have caused deaths of many species, including pelicans, sea otters, and bottlenose dolphins, and tainted commercially valuable species, causing closures of recreational and commercial fisheries and major economic losses. In 2015, one of the largest and most severe HABs occurred along the entire West Coast of North America from southern California to the Aleutian Islands. The principal organism in the bloom was a diatom that produced domoic acid, but there were dinoflagellates that produced toxins that cause paralytic and diarrhetic shellfish poisoning in some parts of the affected area. The bloom lasted from May until October and resulted in the closure of several important fisheries. Closures imposed an estimated $48 million in economic losses for the Dungeness crab fisheries, which remained closed until March 2016, $22 million in the Pacific razor clam fishery, and major losses in other closed shellfisheries. The bloom toxins were also implicated in the deaths of many top predators, especially sea lions, and several seabird species.

Oxygen Depletion and Dead Zones

Below the permanent thermocline, oxygen concentrations are reduced by oxygen consumption during **respiration** and decomposition of detritus that has been transported through the thermocline from the photic zone (**Chap. 12**). The same processes consume oxygen below seasonal thermoclines, but the reduction in oxygen concentration is usually small for two reasons. First, the residence time of water in the layer below a seasonal thermocline (but above any permanent thermocline) is at most a few months. When the seasonal thermocline breaks down in the fall, oxygen is returned to this water when it becomes part of the mixed layer in contact with the atmosphere. In addition, in locations where a seasonal thermocline forms, the total production of phytoplankton during the summer is limited by nutrient availability in the shallow mixed layer above the seasonal thermocline. Hence, only a limited quantity of organic matter is transported through the thermocline to be decomposed. Areas with high productivity are generally areas of upwelling or other physical processes that prevent the formation of a steep thermocline. Vertical mixing in these areas continuously transports oxygen-deficient water to the surface, where it regains oxygen from the atmosphere.

Severe oxygen depletion does not generally occur below seasonal thermoclines, but it may occur in specific circumstances when the **oxygen demand** in the subthermocline layer is very large. The cumulative oxygen demand is the total amount of oxygen consumed before the water mass recontacts the atmosphere or before additional oxygen is supplied by mixing with water that has a higher oxygen concentration. Two circumstances tend to produce high cumulative oxygen demand: high productivity in the overlying water, and long residence time of water below the thermocline. If oxygen concentrations become too low, many animal species cannot survive for more than a short period of time. This low oxygen condition is called **hypoxia**. If the oxygen concentration drops to zero—a condition called **anoxia**—animals cannot survive at all.

Residence times of water below seasonal thermoclines on the continental shelf are generally short, but residence times of water below permanent thermoclines in marginal seas or ocean inlets and bays may be very long (often many decades). Consequently, despite the relatively low productivity of marginal seas that have steep thermoclines, the cumulative oxygen demand is sufficient in certain locations to deplete oxygen. The result can be anoxia and the formation and accumulation of toxic sulfide in bottom waters. Examples of such locations are many Norwegian and other fjords and in the Baltic Sea, where anoxic conditions have been present continuously in the deep waters of the central region for at least 100 years. The geographic area of anoxic bottom waters in the Baltic has grown during that time in response to increasing inputs of nutrients and oxygen-demanding organic matter by humans (**Chap. 16**).

In some marginal seas or other coastal areas, normal conditions result in periodic (often seasonal) or permanent hypoxia. In some of these regions, increased phytoplankton growth and blooms can occur as a result of nutrient inputs from sewage, agriculture, and other human sources. Blooms produce large amounts of detritus, and as a result, the naturally low oxygen concentrations below the thermocline are reduced further. This may lead to occurrences of anoxia, causing most benthic **biota** and some deep-water nekton to be either killed or deprived of these areas as **habitat**.

Anoxia may occur in any particular year in a coastal region if primary productivity is high and/or residence time is long. For example, on the continental shelf of New York and New Jersey in 1976, wind mixing was limited, and an intense bloom of phytoplankton developed. The combination of long residence time and high productivity caused widespread anoxia that extended for hundreds of miles along the coast in a band tens of miles wide across the shelf (**Fig. 13-15a**). The anoxia killed enormous numbers of many marine species, including clams and other shellfish valued at more than $500 million.

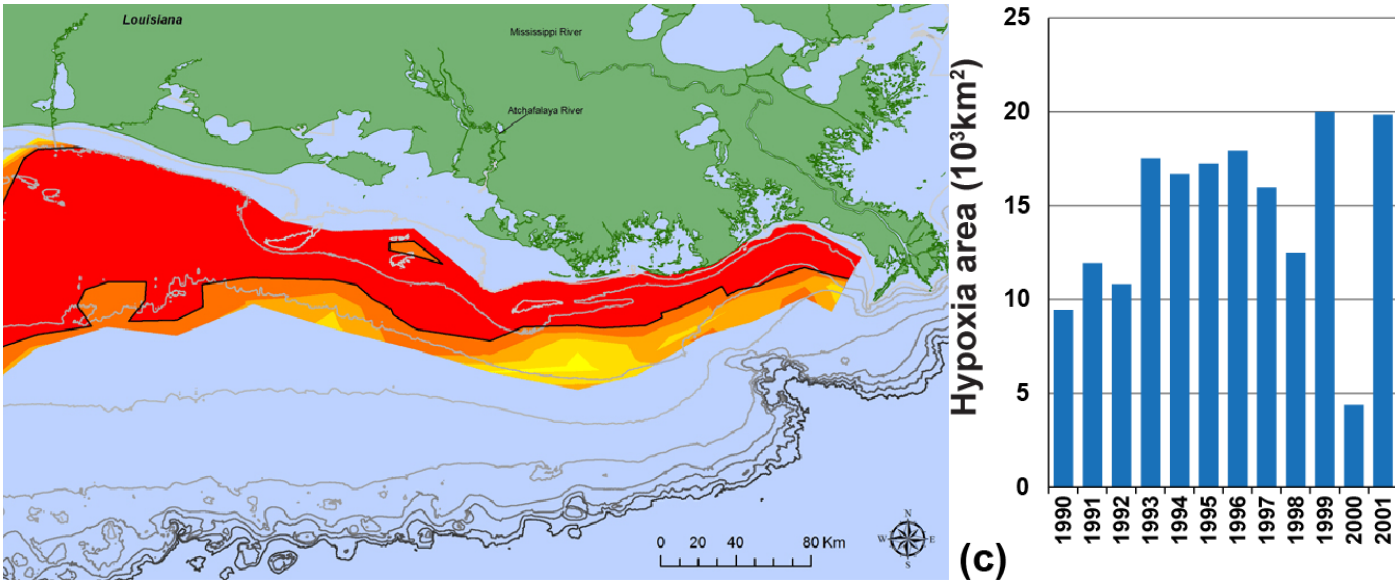


Figure 13-15. Anoxia. (a) During the summer of 1976, eutrophication and unusual climate conditions combined to produce anoxia, or conditions in which oxygen concentrations were too low to support animal life (values of less than 2 in this figure), throughout a large area of the New Jersey continental shelf. This event caused the deaths of massive numbers of clams and other benthic fauna. (b) Hypoxia and anoxia now occur annually each summer in a large area of the Gulf of Mexico continental shelf. This map shows the extent of the hypoxia, commonly referred to as a dead zone, in summer of 2017, the largest area to date. Levels of oxygen above 2.0 are considered essential for marine animals to survive. (c) The Gulf of Mexico dead zone affects an area of the Gulf that varies in extent each year. However, the area is now comparable to or greater than the maximum area of fisheries closures during the Deepwater Horizon oil spill of 2010.

The 1976 anoxia was not caused directly by excess nutrients discharged to the oceans from New York and New Jersey, because natural conditions were responsible for the unusually long residence time of that year. However, nutrient inputs to this region in sewage increase productivity and, thus, the probability of blooms. If nutrient inputs are allowed to increase, hypoxia and anoxia would occur at shorter residence times and, hence, probably more frequently.

Overfertilization of natural waters with nutrients that cause blooms and sometimes hypoxia or anoxia is called **eutrophication**. It has been well known in lakes for decades, but has only more recently become the subject of intense study in the coastal oceans.

Temporary or permanent hypoxia or anoxia has been found in a growing number of areas of the coastal zone, increasing from just a handful of locations in the mid-20th century to hundreds today and still increasing. This problem is now so widespread that a new term “dead zone” has become a common way to describe such occurrences. For example, in the Adriatic Sea, the Saronikos Gulf in Greece, the Sea of Japan, and many bays and harbors periodic massive algal blooms exceed the consumption capacity of the zooplankton population and cause anoxia in the bottom waters. The blooms have been so bad in some years in the Adriatic Sea that the rotting algae have fouled **beaches** and the air, severely affecting tourism in Italy and Croatia. In the U.S., hypoxia or anoxia has occurred in the North Atlantic coastal zone, the Gulf of Mexico, Chesapeake Bay, the Oregon coastal region and a number of other bays and estuaries. Extensive efforts have been made to reduce nutrient inputs to Chesapeake Bay to lessen the severity of anoxia, but progress has been disappointing so far. In the Gulf of Mexico, there is a seasonal dead zone that now develops annually and often extends from near the mouth of the Mississippi River to the Mexican border of Texas (**Fig. 13-15b,c**).

The anoxia that now occurs annually during the summer on the Oregon continental shelf is apparently caused by a new mechanism not previously seen. The mid-depth water mass in the Pacific Ocean has an oxygen minimum zone due to the prolonged time this water mass has been below the thermocline. This oxygen minimum zone appears to be expanding and becoming thicker since it is now shallow enough to extend onto the Oregon continental shelf when winds create seasonal upwelling in summer. The cold high-nutrient water that is upwelled normally supports abundant marine life in this region. However, the upwelled water now often contains insufficient oxygen for respiration and causes a dead zone with massive kills of marine animals. The apparent expansion and deepening of the oxygen minimum layer at mid depths in the North Pacific may be a reflection of either a slowing of the MOC circulation (**Chap. 8**) or an increase in ocean primary production due to increased temperatures and nutrient availability in ocean surface waters, or perhaps both. If the expansion continues, more coastal upwelling regions will likely experience periodic hypoxia, and the area of the oceans that has a mid-depth water mass that has too little oxygen to support animal life will expand.

This page titled 13.4: Algal Blooms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.5: Fisheries

Because they ultimately depend on primary production for their food, fishes are most abundant in areas where primary productivity is high (**Figs. 12-13**, **13-16a**). Hence, fish population density and productivity are higher in the coastal oceans, especially in upwelling regions, than in open-ocean areas where there is no upwelling.

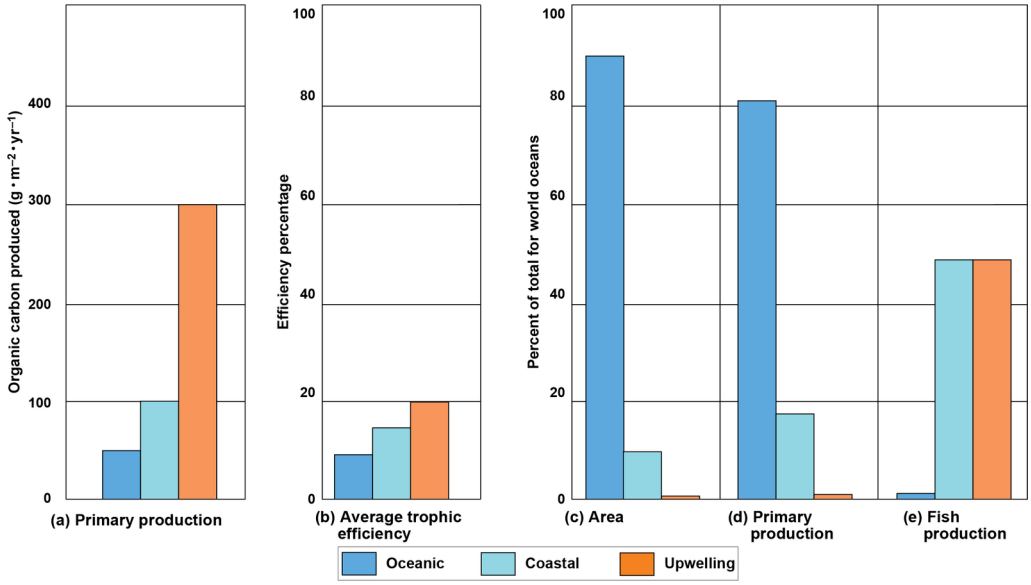


Figure 13-16. (a) Primary productivity, measured as the total amount of organic carbon produced annually in each square meter of water column, is much higher in upwelling areas than in coastal (nonupwelling) or oceanic areas. (b) The average trophic efficiency is high in upwelling regions, intermediate in coastal regions, and lowest in oceanic regions. (c) Most of the ocean is oceanic in character, whereas upwelling areas constitute only a very small percentage of the total area. (d) Because the area occupied by oceanic regions is so large, the oceanic regions are responsible for most of the worldwide primary production. (e) The total production of fish biomass in oceanic areas is small, despite the high primary production, because this primary production is performed mostly by dinoflagellates and because the average trophic efficiency is low. Dinoflagellates must pass through more trophic levels than the diatoms that dominate in coastal and upwelling regions before they are used to build fish tissue. Thus, more dinoflagellate biomass than diatom biomass is needed to sustain the same fish biomass. Fish production is high in coastal regions because of the relatively high primary productivity and trophic efficiency. Despite their very small area, upwelling regions are responsible for a large proportion of the world's fish production because of their very high primary productivity and high trophic efficiency.

The average **trophic efficiency** (**CC15**) is higher in coastal food webs than in open-ocean food webs and is highest in food webs of upwelling areas (**Fig. 13-16b**). As a result, the total fish production of upwelling areas is about half of the world's total (**Fig. 13-16e**), despite the very small area (0.1% of the total area of the oceans) in which upwelling occurs (**Fig. 13-16c**). Coastal regions constitute about 10% of the ocean area and account for about 50% of the world's fish production. Open-ocean regions constitute 90% of the oceans but sustain less than 1% of the world's fish production (**Fig. 13-16e**).

The most successful fisheries are **herring**, anchovy, and sardine fisheries in upwelling regions. These species constitute about 25% by weight of the global catch. Because they feed primarily on zooplankton at the second **trophic level**, they are extremely abundant, and their harvest represents an efficient use of ocean resources. Much of the catch of these species is used to feed animals and not directly to feed people, so we are using these resources at only about 10% of their potential efficiency. About 40% of the global fishery catch of all species is used as animal food.

Because of the distribution of fish biomass production, the world's major fisheries are almost all located in the coastal zone, especially in upwelling areas. Most of these major fisheries are being exploited at or near their **maximum sustainable yield** (**CC16**), and many are overfished. It is believed that the total world fishery catch cannot be increased substantially from its present level without causing detrimental effects on the entire ocean ecosystem, and particularly on higher carnivores, including the largest fishes, marine mammals, and seabirds, that rely on fish stocks for their food. Further, various anthropogenic influences, including overfishing and increasing ocean water temperatures and acidity, are projected to cause a decline in future fisheries food production potential. Global fishery catch data is hard to estimate due to poor accuracy in data reporting from many fishers and their nations. However, it is believed that the world fishery catch peaked somewhere during the late 1990s and is

now declining steadily. Fortunately, the total worldwide production of seafood has remained reasonably stable due to a rapid expansion in aquaculture, but it is not known whether aquaculture can continue to make up for declining wild fisheries into the future. Also, aquaculture can not yet, and perhaps will never be able to, economically produce the high trophic level species such a tuna that are of the highest value, so some shift in species used for human consumption is inevitable if aquaculture is to replace declining wild fisheries resources.

Natural variations in fish stocks are caused by complex interactions of ocean physics, chemistry, and biology. For example, many fish species release their eggs into the plankton community that drifts with the ocean currents. The eggs hatch into larvae that must have the right type of food to grow. For example, anchovy larvae, which eat only phytoplankton that are at least 40 mm in diameter throughout the first few days after hatching. Larval survival of anchovies varies from year to year because the availability of such phytoplankton varies, as do many other factors, including the concentrations of species that compete for food with or prey on the larvae. Survival and success are no less complicated for the larval fish of other species or when they become juveniles, because juveniles also have to find food and are subject to predation, disease, **parasitism**, and the direct and indirect effects of **pollution**. Each of these factors for each different fish species varies from year to year, at least in part because of year-to-year climatic variations.

Life cycles and their interactions with physical, chemical, and biological variables are so complex that studies must be conducted over many years to obtain even a limited understanding of the population variations in a single fish species. Even after such extensive studies, it is impossible to predict the future of the fish stock with any certainty because many of the influences on fish stock size and age composition are nonlinear. Consequently, fish stocks of many species are inherently **chaotic** (**CC11**) and appear to fluctuate in a random or unpredictable way. It may be impossible to predict future fish stocks accurately, just as it is impossible to forecast the weather accurately more than a few days in advance.

This page titled 13.5: Fisheries is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.6: Estuaries

As freshwater flows into the oceans, it mixes with seawater. Regions where this mixing occurs are called “estuaries.” The seaward limit of an estuary is where the dilution of ocean water by freshwater is insignificant. The landward boundary, or head, of the estuary is the maximum landward limit of saltwater movement. Estuaries are present at mouths of major rivers and in many other semi-enclosed inlets or arms of the ocean into which streams and rivers flow. If freshwater discharge is very large, the seaward boundary of the estuary can be many kilometers offshore. The inland boundary can be tens or even hundreds of kilometers inland. The boundaries are not fixed, and can move seaward if freshwater flow rate increases or farther inland if it decreases.

The fact that most major cities are located on estuaries reflects the historical importance of estuaries as harbors and ports. Estuaries are also important to marine ecosystems because many species of fishes and invertebrates spend part of their life cycle in the oceans and part in an estuary or river, and many major fisheries and shellfisheries are in estuaries.

Estuaries differ in characteristics that include their length, width, depth, **tidal range**, freshwater flow rate, shape, and coastal character. These factors affect ocean and river water movements and mixing processes in the estuary. Consequently, there is no simple description of a typical estuary, and no single classification system can capture the many variations. Every estuary is different from all others and has its own unique behavior. However, estuaries are classified into general groups according to the geological processes that formed their embayment or, alternatively, according to their water circulation and mixing characteristics.

Geological Origin of Estuaries

Almost all present-day estuaries were formed as sea level rose in the past 19,000 years from a low point approximately 130 m below current sea level (**Fig. 6-18**). As the most recent ice age ended and glacial ice melted, sea level rose. River and glacial valleys were flooded and coasts were modified by erosion, forming barrier islands in some locations (**Chap. 11**).

Estuaries are filled progressively with **sediment** supplied by the river. If sea level were to remain stable long enough, most estuaries would eventually become filled with sediment and **deltas** would form (**Chap. 11**). The inexorable process of sediment filling can be seen at locations such as the former Greek port of Ephesus, now in Turkey. This city, a bustling seaport less than 3000 years ago, is today many kilometers inland, behind a low **coastal plain** formed by the accumulation of sediment eroded from surrounding mountains (**Fig. 13-17**).

Figure 13-17. The ancient Greek city of Ephesus, now in Turkey. In ancient Greek times, Ephesus was a port city. The waterfront was just at the foot of the hill behind the temple in the center of this photograph. The ruins of the port are now several kilometers inland. In the background, you can see the wide flat coastal plain. This was once a large bay, but since ancient times it has filled with sediments.

In the past, during periods when sea level was falling, there were few estuaries. As sea level falls, a coast consists of newly exposed continental shelf. This new coastal plain is relatively featureless because the topographic lows in the shelf floor are filled with sediment transported by tides, waves, and currents. Estuaries will virtually disappear again when sea level falls in the future.

Rising sea level has created estuaries that differ according to the character of land that has been inundated. Four types are recognized: **coastal-plain estuaries, bar-built estuaries, tectonic estuaries**, and fjords. Coastal-plain estuaries were formed as sea level rose to flood river valleys (**Fig. 13-18a**). These estuaries, often called "drowned river valleys," are especially abundant on passive margins, such as the east coast of the United States. Chesapeake Bay, Delaware Bay, and the New York Harbor are examples of coastal-plain estuaries. The Mississippi River delta is an example of a former coastal-plain estuary that has been filled with river-borne sediment.

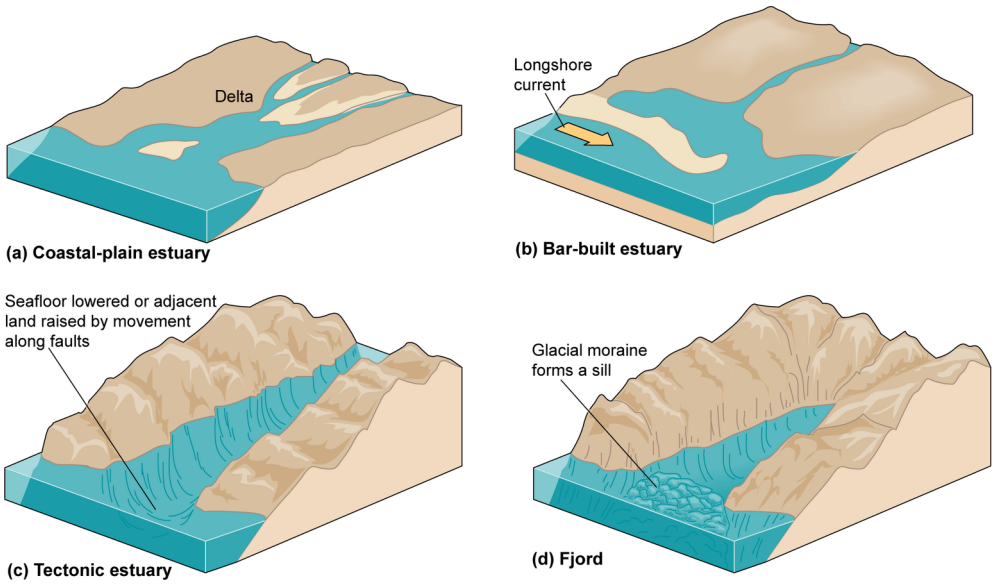


Figure 13-18. The rise of sea level during the past 19,000 years has been a major factor in the creation of all present-day estuaries. Estuaries have a variety of geological settings and origins. (a) The most common estuary is a coastal-plain estuary (also sometimes called a drowned river valley) that forms as rising sea level floods the mouth of a river. (b) Some estuaries form where longshore drift builds a bar, spit, or barrier island that isolates a bay from free exchange with the ocean. (c) Tectonic estuaries are formed where a block of the Earth's crust is lowered by tectonic processes at earthquake faults. (d) Fjords are created when rising sea level floods the steep-sided valleys that are cut by glaciers and exposed after the glaciers retreat.

Bar-built estuaries are formed when a sandbar is constructed parallel to the coastline by wave action and **longshore drift** (**Chap. 11**), and the bar separates the ocean from a shallow lagoon (**Fig. 13-18b**). Lagoons behind barrier islands are bar-built estuaries. Examples include Albemarle Sound and Pamlico Sound in North Carolina. Some estuaries have mixed characteristics. For example, the Hudson River estuary that passes through the New York Harbor and Raritan Bay is primarily a coastal-plain estuary, but the Sandy Hook spit gives Raritan Bay certain bar-built estuary characteristics (**Fig. 13-19**).

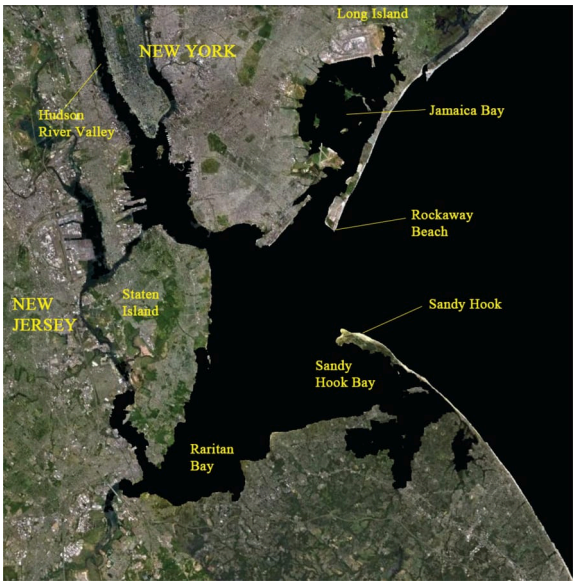


Figure 13-19. This satellite image of the Hudson River basin–New York Harbor–Raritan Bay estuary shows the complex character of this estuary. The Hudson River valley is a glacier-cut valley that was then shaped by a river and so can be considered a coastal-plain estuary. Parts of the river are steep-sided and deep enough that they may even be considered a fjord. Sandy Hook Bay and Jamaica Bay are bar-built. Raritan Bay and New York Harbor could be considered either coastal-plain estuary segments or bar-built, or both.

Tectonic estuaries are formed when a section of land drops or tilts below sea level as a result of vertical movement along a **fault** (**Fig. 13-18c**). Such estuaries are most often present on coasts along **subduction zone** or **transform fault** plate boundaries. The best-known tectonic estuary in North America is San Francisco Bay (**Fig. 4-23**), which is on a transform plate boundary.

Fjords are estuaries in valleys that were cut by glaciers when sea level was lower (**Fig. 13-18d**). These estuaries are generally steep-sided, both above and below sea level, and are often deep. Many fjords have a shallow sill near the mouth that was formed by sediment deposited at the lower end of the glacier when it flowed through the valley. Fjords are common on the Norwegian coast, the southwest coast of New Zealand, the Patagonia coast of Chile, and the Pacific coast of Canada and southern Alaska. Puget Sound in Washington State is also a fjord.

Estuarine Circulation

In estuaries, less dense river water flows over seawater, causing vertical mixing between the two layers. The movements and mixing of freshwater and seawater are affected by many factors, including tidal currents and mixing, wind-driven wave mixing, shape and depth of the estuary, rate of freshwater discharge, friction between the moving freshwater and seawater layers and between the water and seafloor, and the **Coriolis effect** (**CC12**). Because some or all of these factors vary among estuaries, between seasons within individual estuaries, and from one part of an estuary to another, estuaries have a wide variety of circulation patterns.

Estuarine circulation is extremely important because many major cities are located on estuaries. These cities discharge large quantities of wastes, particularly sewage treatment plant **effluents** and storm water runoff, into the estuaries (**Chap. 16**). Estuarine circulation patterns determine the fate of these contaminants.

In many estuaries, circulation is altered by piers and other port structures, dredging of navigation channels, filling of **wetlands**, and construction of **levees** and other coastal structures (**Chap. 11**). These alterations can affect life cycles of marine species that inhabit or transit the estuary (**Chap. 16**).

Because circulation within each estuary is complex, varies temporally, and is unique to that estuary, detailed, multiyear physical oceanography studies of each estuary are necessary to understand its circulation. Such studies are difficult and expensive, particularly in large, complex estuaries such as Chesapeake Bay or San Francisco Bay.

Types of Estuarine Circulation

Estuaries can be classified according to the major characteristics of their circulation. The major types are **salt wedge estuaries**, **partially mixed estuaries**, **well-mixed estuaries**, fjord estuaries, and inverse estuaries

Salt Wedge Estuaries

In a salt wedge estuary, freshwater flows down the estuary as a surface layer separated by a steep density interface (halocline) from seawater flowing up the estuary as a lower layer that forms a wedge-shaped intrusion (**Figs. 13-20a, 13-21a**). Vertical mixing across the steep density gradient is slow in salt wedge estuaries, but the velocity difference between seaward-flowing river water and the underlying seawater creates friction between these layers. The friction causes turbulence and internal waves at the interface. The internal waves grow, and when they break, small quantities of high-salinity water are injected and mixed into the upper layer, which progressively increases in salinity toward the estuary mouth (**Figs. 13-20a, 13-21a**). Almost no freshwater is transferred into the lower seawater layer.

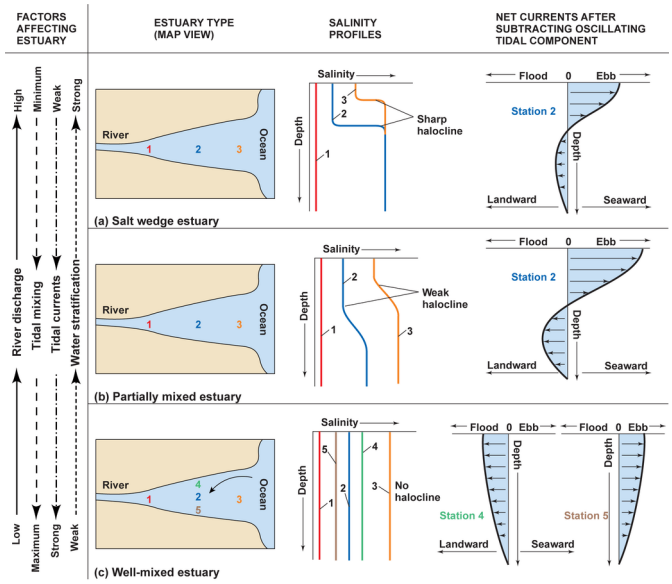


Figure 13-20. One way to classify estuaries is based on their circulation characteristics and the factors that influence their circulation. (a) Salt wedge estuaries are river-dominated and are strongly stratified with a sharp halocline. (b) Partially mixed estuaries have a weak halocline because the surface and bottom layers are partially mixed by turbulence induced by tidal currents. (c) In well-mixed estuaries, the water column is thoroughly mixed from surface to seafloor, and there is no halocline. However, salinity does increase progressively with distance from the head of the estuary to the sea and across the estuary as a result of the Coriolis effect.

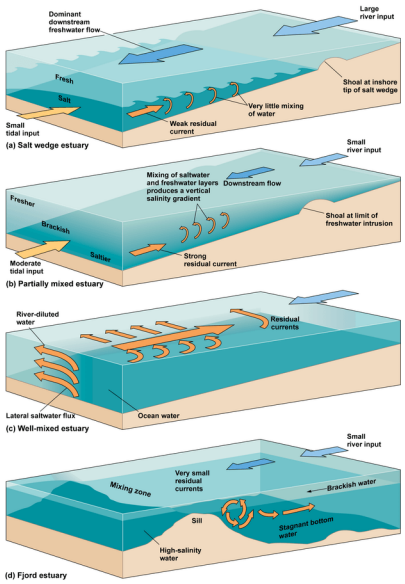


Figure 13-21. Types of estuaries categorized by circulation pattern. (a) Salt wedge estuaries are characterized by a net (after tidal motions are averaged out) landward current in the lower layer and a net seaward current in the upper layer. There is little vertical mixing, but breaking internal waves on the halocline mix some seawater up into the surface layer. (b) Partially mixed estuaries are characterized by relatively strong net landward currents near the seafloor and relatively strong net seaward currents in the lower salinity (fresher) surface layer. (c) Well-mixed estuaries have no vertical salinity variations, but they do have variations in salinity both across and down the length of the estuary. The Coriolis effect causes net estuarine currents to flow landward at all depths on the right side of the estuary in the Northern Hemisphere (viewed from the ocean), and seaward flow at all depths on the left side. These currents are on the reverse sides in the Southern Hemisphere. (d) Estuaries in fjords have little current in the deep water below the depth of any sill. In the upper layers above the sill depth, there is generally a salt wedge or sometimes a partially mixed estuary circulation pattern.

Salt wedge estuaries are most likely to be present where freshwater input is relatively large and allows a thick, freshwater surface layer to develop. Estuaries that are narrow in relation to their depth also favor the development of a thick freshwater layer. Other conditions that favor salt wedge estuaries are small tidal range, and hence limited tidal currents and tidal mixing, and limited vertical mixing due to wind-induced motions (e.g., waves, Langmuir circulation, and Ekman transport). To understand how these factors favor the two-layer salt wedge configuration, compare the estuary with a blender filled with water and cooking oil. If the blender runs at high speed, the cooking oil is quickly mixed with and dispersed in the higher-density water. If the blender runs more slowly, the oil will tend to remain in a distinct upper layer, particularly if there is enough oil to create a thick oil layer. A thin oil layer will mix and disperse at a lower blender speed than a thick oil layer.

Partially Mixed Estuaries

Vertical mixing between freshwater and seawater is greater in partially mixed estuaries than in salt wedge estuaries, and therefore the density gradient between the two layers is much less pronounced (**Fig. 13-20b**). The most important process that causes vertical mixing across an estuarine halocline is the friction and resulting turbulence created by reversing tidal currents. Hence, partially mixed estuaries are most common where tidal currents are relatively fast, river flow rate is moderate, and river current speed does not greatly exceed tidal current speed. As in a blender running at moderate speed with roughly equal volumes of water and oil, some mixing occurs at the interface, but two distinct layers persist.

In partially mixed estuaries, freshwater and seawater layers are separated by a relatively weak halocline. Seawater moves landward up the estuary as a bottom layer and is diluted progressively with freshwater from above (**Figs. 13-20b**, **13-21b**). Both layers move up and down the estuary with each tidal cycle. The distance that water moves in the estuary with each ebb and flood is the tidal excursion. The tidal excursion is usually much greater than the landward or seaward net movement that occurs as a result of the **residual currents** of the estuarine circulation (currents caused by the river flow and landward flow of seawater). **Figure 13-21** shows only these residual currents. Residual currents are often difficult to measure because they are masked by the faster reversing tidal currents that must be averaged out to reveal the residual current.

As seawater moves landward and freshwater moves seaward in an estuary, the moving water is deflected by the Coriolis effect. Consequently, if we look at an estuary from the sea in the Northern Hemisphere, seawater moving up the estuary tends to be deflected to our right, and freshwater moving down the estuary tends to be deflected to our left. In salt wedge estuaries, this deflection causes the halocline to be tilted slightly upward toward the right bank (**Fig. 13-21a**). In partially mixed estuaries, the

Coriolis deflection affects not only the mean flow, but also tidal currents. As tidal currents flow into the estuary, they are deflected to the right side (as viewed from the sea); as they ebb, they are deflected to the left side. Consequently, in a partially mixed estuary, movement of seawater landward in the lower layer is concentrated on the right side, and the lower-salinity estuarine outflow is concentrated on the left side. The halocline is strongly inclined upward to the right (as viewed from the sea) in many partially mixed estuaries. The incline is in the opposite direction in the Southern Hemisphere.

Well-Mixed Estuaries

In estuaries with swift tidal currents, mixing is very strong (like the blender at high speed), the water column is completely mixed, and no halocline is present (**Fig. 13-20c**). In these well-mixed estuaries, seawater moving landward is continuously mixed with freshwater moving seaward. Salinity decreases progressively from the ocean toward the head of the estuary. Although the salinity is uniform from surface to bottom, it varies during the tidal cycle, increasing as the tide floods and decreasing as it ebbs.

In large well-mixed estuaries, the Coriolis effect tends to separate the landward and seaward estuarine flows, so they are concentrated on opposite sides of the estuary. This separation is especially important in wide and shallow estuaries. Many such estuaries have a small residual landward movement of water along the right side (as viewed from the sea in the Northern Hemisphere), and a small residual seaward current on the left side. This is made possible by a residual current that flows from one side of the estuary to the other (**Fig. 13-21c**). This circulation causes salinity to be higher on the right side of the estuary and to gradually decrease across the estuary to the left side (**Fig. 13-20c**).

Well-mixed estuaries tend to be wide and relatively shallow. They usually have limited freshwater inputs and strong tidal currents. These factors and others that determine the circulation characteristics change with location in the estuary and sometimes with time. Consequently, an estuary's circulation can vary from well mixed, to partially mixed, and to salt wedge at different times and at different locations within the estuary. San Francisco Bay is a good example. The central portion of the bay (closest to the ocean) has relatively strong tidal currents, is wide, and is generally partially mixed. The upper part of the estuary, which is closer to the Sacramento and San Joaquin Rivers, is narrower and generally has salt wedge characteristics because tidal currents are somewhat reduced and freshwater inputs are large enough to form a thick, low-salinity surface layer. The southern part of San Francisco Bay is shallow, has very little freshwater input, and is generally well mixed. When river flow rates are low, seasonally or in drought years, the northern part of the estuary becomes partially mixed. During high spring river runoff, the southern and central parts of the bay can become partially mixed or even salt wedge estuaries.

Fjord Estuaries

Fjords are generally narrow and usually much deeper than most other estuaries. Most fjords are deep enough that vertical mixing does not reach the bottom waters, even if tidal currents are strong. Consequently, almost all fjords have a halocline separating high-salinity bottom water from lower-salinity surface water, and fjords are almost never well-mixed.

Circulation in many fjords is complicated by the presence of a shallow sill at the seaward end of the estuary that prevents the free exchange of deep water between ocean and fjord (**Fig. 13-20d**). Therefore, estuarine circulation is established only above the level of the sill. The fjord's interior above the sill depth behaves as a salt wedge or partially mixed estuary. At the sill, vertical mixing is enhanced by turbulence caused by the flow of water over the sill (**Fig. 13-21d**). The deep water below the sill depth within the fjord is stagnant and is not involved in the estuarine circulation. In many fjords, the deep water becomes anoxic because of the decomposition of organic particles that have settled from above. Periodically this deep water may be displaced by high-salinity ocean water that floods over the sill in response to some unusual set of water movements. In such circumstances, hydrogen sulfide in the fjord bottom water is displaced into surface waters of the fjord and ocean, where it may cause extensive fish mortality.

Inverse Estuaries

Shallow estuaries in arid regions can sustain sufficiently high rates of evaporation that salinity is higher than that of ocean water outside the estuary. In this situation, estuarine circulation is inverted. High-salinity estuarine water flows seaward in a bottom layer, and ocean water flows landward as a relatively lower-salinity (and less dense) surface layer. Inverse estuaries generally occur at about 30°N or 30°S, in the region of atmospheric subtropical highs, where evaporation is strong and rainfall typically is low. Some well-known examples of inverse estuaries are the Red Sea, San Diego Bay in California, and Laguna Madre in Texas. In addition, the Mediterranean Sea behaves as though it were an exceptionally large inverse estuary.

Particle and Contaminant Transport in Estuaries

Most major cities have developed next to rivers and estuaries, partly because of their convenience for the disposal of sewage and other wastes. The common belief is that wastes discharged to a river or estuary are simply swept out to the ocean, where they are diluted in the vast volume of ocean water. However, a simple explanation of water movements and suspended-particle transport within estuaries reveals that this is not what happens.

Freshwater input at the head of the estuary creates the estuarine circulation. To sustain this circulation, the amount of seawater that flows through the estuary must greatly exceed the volume of freshwater discharged during the same period of time. To understand this process, we must first observe that freshwater entering the head of most estuaries has a salinity close to zero, whereas water discharged from the estuary mouth into the ocean has a salinity almost equal to that of the seawater. Freshwater mixes with seawater, so the salinity of the seaward-moving water progressively increases along the length of the estuary (**Figs. 13-21, 13-22**).

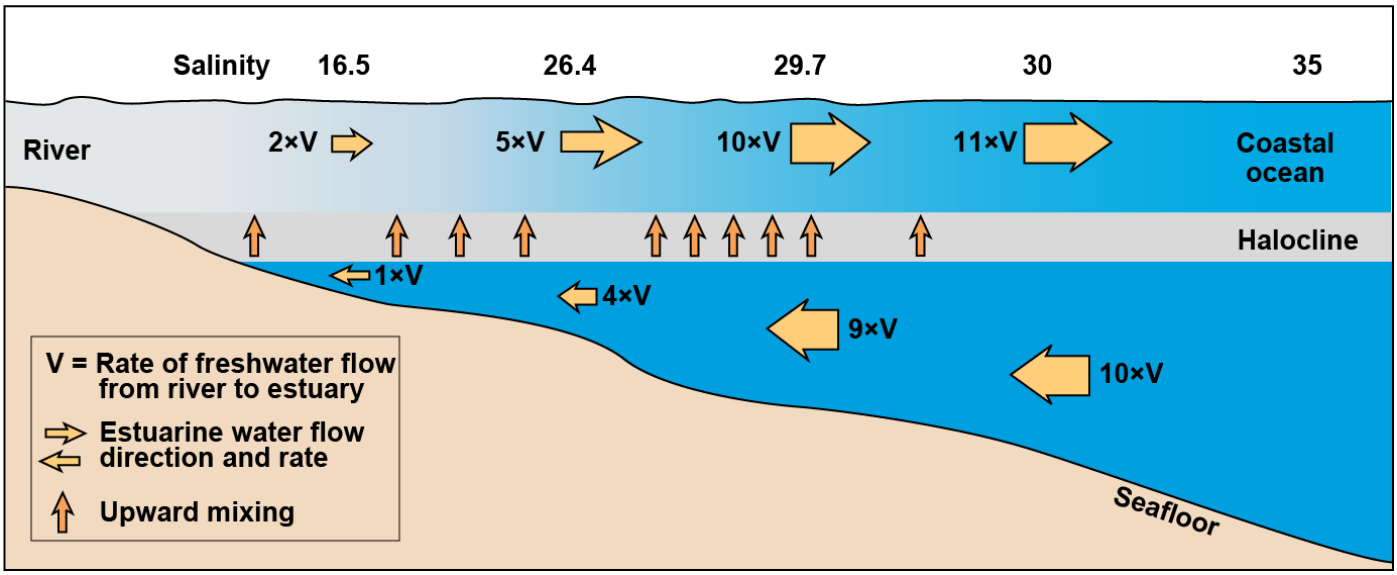


Figure 13-22. Seaward-moving surface water increases in salinity from nearly zero at the head of the estuary to 30 where it enters the ocean. Ocean water entering the estuary has a salinity of 33. Where seaward-flowing water reaches a salinity of 16.5 (half that of ocean water), it must consist of one-half river water and one-half ocean water. Seawater must flow landward to point A at a rate (volume per unit time) equal to the river flow rate, and seaward transport of low-salinity estuarine water must equal twice the river flow rate. Similarly, the landward flow rate of seawater at point B must equal 4 times the river flow rate, and the seaward flow rate of low salinity estuarine water must be 5 times the river flow rate.

Figure 13-22 depicts the relative residual flow rate (the volume of water that passes a particular point in the estuary per day) in various parts of an estuary. From the figure, we can see that the residual flow rate of seawater moving landward and the residual flow rate of estuarine water flowing seaward at the mouth of an estuary are both many times greater than the river flow rate. In addition, the seaward and landward flow rates are highest near the mouth of the estuary and decrease toward the head of the estuary. The residual flow rate is equal to the residual current speed multiplied by the cross-sectional areas of the estuary. Consequently, flow rate does not always translate directly into current speed, because the cross-sectional area of an estuary changes, sometimes in complex ways, along its length. However, the large differences in flow rate along an estuary generally cause both landward and seaward residual currents to be higher near the mouth of the estuary than near the head.

Rivers carry large quantities of suspended sediment particles, the largest of which are quickly deposited when the river reaches the flat coastal plain or, if there is no coastal plain, when the river enters the head of an estuary. The finest particles could remain in suspension as they are transported to the oceans, but the increased salinity in the estuary causes most of them to clump. Because the clumps are larger, they are then deposited.

Particles of intermediate size, which fall slowly through the water column, are usually deposited within the estuary. In well-mixed estuaries, these particles are deposited primarily on the left side (as viewed from the sea in the Northern Hemisphere). In other estuaries, seaward-moving particles fall through the water column into the landward-moving bottom layer, where they are transported back toward the head of the estuary. These particles often are deposited near the estuary's head where tidal and residual currents are low, often accumulating as a shoal (Fig. 13-21a,b). However, when elevated river flows extend the head of

the estuary seaward, the particles may be resuspended by the swifter river currents. In many estuaries, these recycled particles are eventually deposited in wetlands (**Chap. 11**).

Sediments also can be transported from the ocean into estuaries, particularly in estuaries where tidal currents are strong enough to resuspend sediments that accumulate on the nearshore continental shelf.

Substantial quantities of suspended sediment can be transported from the ocean into well-mixed estuaries, where they accumulate along the right side (as viewed from the sea in the Northern Hemisphere). Smaller quantities of suspended sediment are carried from the ocean into partially mixed and salt wedge estuaries. Even particles carried into the ocean by the seaward-moving surface layer may return to the estuary if they sink and are transported landward in the seawater layer.

The fate of particles in an estuary depends on the particle size and where the particles are introduced to the estuary. Floating and extremely small particles introduced to the surface layer (or to the left side of a well-mixed estuary) are transported to the oceans, although many will clump at higher salinities. Small or low-density particles in bottom waters are transported landward until the current slows and they are deposited. Such particles may accumulate in the estuarine sediments or in the sediments of adjacent wetlands, and they may be resuspended and carried seaward when river flow rates are high. Particles of intermediate size in the surface layers of salt wedge or partially **stratified** estuaries are carried seaward until they sink below the halocline, then are transported landward until they settle out in the upper parts of the estuary where current speeds are lower. Large particles tend to accumulate near the point of introduction.

The important point of the preceding discussion is that suspended particles tend to be retained within the estuary by the estuarine circulation and are not simply flushed into the open ocean. Because most toxic metals and organic compounds attach to particles, particularly the organic particles in treated sewage effluents, they tend to accumulate in the estuary and are not flushed out to the ocean.

A lack of understanding of estuarine circulation and its effect on particles and the fates of toxic metals and organic compounds is one reason that many estuarine ecosystems have been and continue to be severely impacted by human activities. One of many examples of the misuse of estuaries for waste disposal is in San Francisco Bay, where navigation channels, marinas, and ports must be dredged to maintain the desired depths. Most of the dredged sediments were, for decades, and some still are, dumped near the bay's mouth at a site between Alcatraz Island and San Francisco (**Fig. 13-23**). The swift tidal currents were expected to flush the material out to the ocean, where it would be dispersed. However, the dump site is near the right bank (viewed from the sea) of a partially mixed estuary, and the dumped dredged sediments instead fall to the floor of the bay and disperse in the lower part of the water column.

Figure 13-23. This image of San Francisco Bay, taken by the Aqua/MODIS satellite, shows the high turbidity in the upper estuary, where there is a salt wedge circulation, and the location of the Alcatraz dredged-material dump site near the estuary mouth, where the circulation is partially mixed.

With an understanding of estuarine circulation, we realize that much of the dredged material dumped at the Alcatraz dump site, with its toxic load, is not transported to the ocean. It is instead transported back within San Francisco Bay by estuarine circulation

and eventually is deposited in low-energy areas, including the same channels and harbors from which it was originally dredged. Perhaps more importantly, continual dredging and disposal disturbs the particle-associated toxic metals and organic compounds, which were relatively safe when buried in sediments, and resuspends them in the estuarine circulation, where they are more directly in contact with the biota (**Chap. 16**). Fortunately, the most contaminated dredged material is no longer dumped at this dump site.

Estuarine Biology

Estuaries are difficult places for organisms to inhabit because they are subject to rapid and irregular changes in environmental factors such as salinity, temperature, and current speed. The stress induced by salinity changes is particularly challenging because organisms must be able to cope with widely varying osmotic pressures (**Chap. 14**).

Because estuaries are high-stress **environments**, they support fewer species than the adjacent ocean or freshwater environments support (**Fig. 13-24**). However, there is an abundant supply of nutrients and sunlight in most estuaries. As a result, estuarine biomass is typically much greater than biomass in freshwater and ocean environments, with the exception of the most productive coastal upwelling areas.

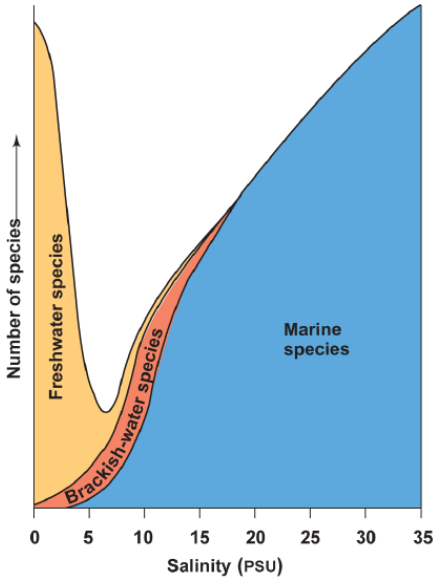


Figure 13-24. The number of species of marine, estuarine, and freshwater origin varies within the salinity gradient of a typical estuary. The fact that the total number of species is much lower in the brackish-water zone than in the freshwater or the marine zones reflects the stressful environmental conditions, particularly variable salinity, in the brackish-water region.

Although estuaries typically have relatively high turbidity, they are also generally shallow, and in many, light penetrates to the bottom. In such estuaries, primary production is dominated by benthic **microalgae**, particularly diatoms, that form mats that carpet the sediments. Macroalgae can grow on the estuary floor in areas where current speeds are low. In addition, rooted aquatic plants are abundant in the **tidal salt marshes** and **mangrove** forests along the edges of many estuaries. In contrast, because phytoplankton are continually being swept seaward with the estuarine surface layer flow, their populations are relatively limited in estuaries that have short residence times.

Nutrients are generally abundant in estuaries because they are supplied in substantial quantities in river water that has **leached** them from soils and rocks of the drainage basin. In addition, once in the estuary, nutrients tend to be retained and recycled within the estuarine circulation. Nutrient-containing detritus, which flows seaward in the surface layer, sinks below the halocline and then returns with the landward-flowing ocean water. This circulation tends to concentrate detritus in the region near the landward limit of seawater intrusion (**Fig. 13-25**), which has low residual currents. This region has high turbidity as a result of the accumulated detritus and other particles. The nutrient-rich detritus supports a varied community of fish and invertebrate larvae and juveniles.

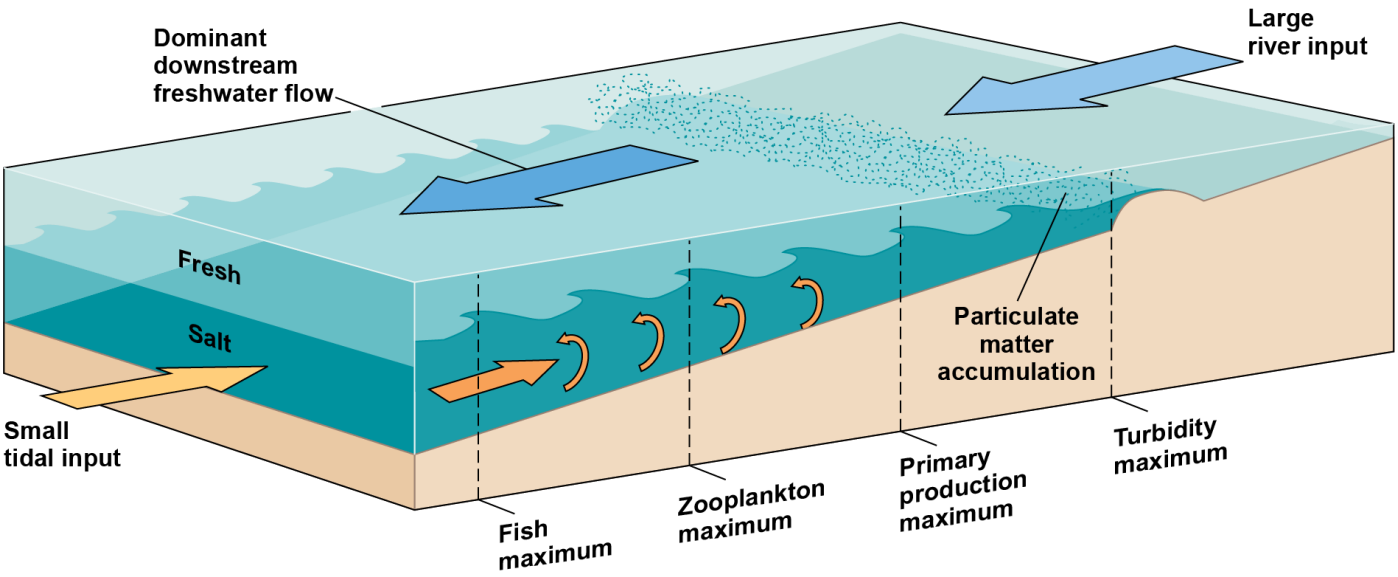


Figure 13-25. In many estuaries, an accumulation of particulate matter creates high turbidity near the landward extent of seawater intrusion. Particles transported seaward in the upper layer may sink to the lower layer before leaving the estuary and then return to the high-turbidity area in the landward-flowing lower layer. Many planktonic estuarine organisms use this pathway to stay within the estuary or to reach an appropriate point in the estuary at the right time in their life cycle. The maximum primary productivity in an estuary is usually just seaward of the high-turbidity zone. In this region, more light can penetrate. In addition, nutrients are available because they are regenerated by the decomposition of suspended detritus particles transported up the estuary in the lower layer and are then mixed into the surface water at the turbidity maximum. Maximum zooplankton populations and growth occur downstream of this area to take advantage of the phytoplankton drifting seaward in the upper layer.

In many estuaries, much of the detritus that accumulates in the upper part is derived from algal mats from the adjacent wetlands (**Chap. 11**). Where residual currents are slow, or the estuary is long, a food chain from phytoplankton to zooplankton to estuarine fishes is established that is analogous to the food chain in the coastal upwelling ecosystem described previously in this chapter. Zones of high phytoplankton abundance, high zooplankton abundance, and high fish abundance form a seaward progression from the head of the estuary (**Fig. 13-25**). Some planktonic organisms and larval fishes that are being transported seaward in the estuary's upper layer and away from the region of abundant food supply sink into the seawater layer and are transported back into the food-rich region.

Because they provide abundant food and substantial protection, estuaries are ideal habitats for the larvae and juvenile stages of marine animals. Particularly in wetlands, there are many shallow-water hiding places that are accessible to small swimming organisms but not to their larger predators. In addition, predators are less common in estuaries than in less stressful freshwater or ocean environments. Many marine fishes and invertebrates use estuaries as nursery grounds. Some species briefly visit the estuary to spawn, and others release vast numbers of eggs in the adjacent ocean so that they are transported into the estuaries by the estuarine circulation. Along the southeast coast of the U.S., more than half of the commercially important marine fish species are known to use estuarine wetlands as nursery areas or for breeding. Because of the abundant supply of detritus, estuaries also support huge populations of commercially important shellfish, including clams, oysters, mussels, and crabs.

Anadromous and Catadromous Species

Estuaries are important to two categories of migratory fishes that live most of their lives either in the oceans or in freshwater. **Anadromous** fishes, such as salmon and striped bass, live most of their lives in the ocean but return to freshwater to spawn. The stress involved in the return to freshwater and the energy lost in spawning are such that the adults die soon after they spawn and do not return to the ocean. Their offspring develop in freshwater and migrate to the estuary, which provides abundant food and relative safety from predators. Eventually, they migrate out to sea, where they live several years before returning to the rivers to spawn.

Catadromous fishes use a strategy opposite that of anadromous fishes. Catadromous species live most of their lives in freshwater and then journey to the oceans to spawn and die. Freshwater eels are the best-known and most abundant catadromous species. Sexually mature eels in rivers on both sides of the North Atlantic Ocean make a one-way journey to spawn in the southwest part of the Sargasso Sea. Their offspring return to the rivers, riding part of the way on the Gulf Stream. North Pacific

eels make a similar journey to the equivalent southwest area of the North Pacific Gyre, and their offspring use the Kuroshio (Japan) Current for the ride home.

Anadromous and catadromous species do not generally spend much time in the estuary during their reproductive migration, but transiting the estuary is especially stressful because they must make the osmotic-pressure adjustment from seawater to freshwater, or vice versa, very quickly. Striped bass, for example, rest in the outer part of the estuary to adjust to lower salinity before moving farther upstream. Many estuaries also have relatively high concentrations of toxic metals and organic compounds that are likely to cause additional stress on anadromous and catadromous species as they transit the estuary. Such contamination by toxins may account, in part, for the decline of these species in some areas.

This page titled 13.6: Estuaries is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.7: CHAPTER SUMMARY

Special Characteristics of the Coastal Oceans.

Salinity and temperature are more variable, and turbidity is generally higher, in coastal waters than in the open ocean. River discharges often lower salinity and increase turbidity near river mouths, and a shallow halocline is often present in this region. Coastal currents are driven by local winds and are generally independent of ocean gyre currents, often flowing in the opposite direction. Benthos of the coastal oceans are more diverse and abundant than open-ocean benthos are.

Nutrient Supply to the Coastal Photic Zone.

The mixed layer in many coastal regions receives nutrients from river runoff, coastal upwelling, decomposition of organic matter on the shallow seafloor, eddies formed around shallow seafloor topographic features, and breaking internal waves. Nutrients tend to be retained in coastal waters because residence times of coastal water masses are often long. Coastal upwelling is most prevalent off the west coasts of continents where Ekman transport moves the surface layer offshore. Newly upwelled water may support only limited primary productivity until needed organic compounds are synthesized by flagellates. As upwelled water moves offshore, diatoms bloom and are consumed by zooplankton, and eventually nutrients are depleted and primary productivity declines.

Nutrients may be recycled as they are transported into subthermocline waters in fecal pellets, released there by decomposers, and transported back inshore and upwelled. The eggs or spores of many species are carried inshore in subthermocline water, are then upwelled, metamorphose or hatch, and move back offshore where phytoplankton food supply increases. There, they mature and produce eggs or spores that sink and reenter the circulation.

Seasonal Cycles.

In subpolar regions, primary productivity is very low, except during a short summer when light intensity is sufficient for photosynthesis. Nutrients are plentiful in areas where strong vertical mixing prevents a pycnocline from forming. In polar and some subpolar regions, vertical mixing is restricted by a halocline formed by freshwater runoff and sea-ice melt; hence, nutrients are limited and productivity is low.

In tropical regions, there is little seasonality and nutrients are always limiting, except in upwelling areas. However, coral reef ecosystems are highly productive because of nutrient recycling, primarily between corals and their symbiotic zooxanthellae.

In mid-latitudes, the winter water column is well mixed, and primary production is light-limited. In the spring, as light intensity increases, a plankton bloom occurs and continues until nutrients are depleted. Nutrient depletion is often accelerated by the formation of a shallow seasonal thermocline. Productivity remains low until the fall, when storms provide additional nutrients by vertical mixing and a limited fall phytoplankton bloom may occur. Nutrients are returned to the photic zone by winter mixing.

There is a succession of dominant phytoplankton species during seasonal cycles of middle and high latitudes. Generally, diatoms dominate when nutrients are abundant and are replaced by flagellates as nutrients decline or as silica is depleted. Species succession is similar each year, but it varies in timing and dominant species because of differences in climate and the physical and chemical characteristics of the ocean water. Juveniles of many animal species rely on particular phytoplankton species or types at particular times and locations. When the species succession is altered, and this food is unavailable, these species' larval survival can be drastically reduced.

Algal Blooms and Dead Zones.

Intense blooms of dinoflagellates occur infrequently in coastal water, primarily at subtropical and middle latitudes. These episodic blooms can be natural. However, their frequency is increasing off many coasts, and evidence suggests that sewage-derived nutrients may favor the dominance of dinoflagellates over diatoms and contribute to some blooms. Certain dinoflagellates and diatoms produce substances toxic to vertebrates. Hence, blooms can kill fishes. They can also adversely affect human health, because the toxins do not harm shellfish but are toxic to people who eat the shellfish. The most prevalent dinoflagellate toxin

causes paralytic shellfish poisoning in humans, so many valuable shellfishing areas are closed to harvesting during months when dinoflagellate blooms may occur.

When phytoplankton blooms collapse as a result of nutrient depletion, their decomposing remains can deplete oxygen in bottom waters, particularly where residence times are long and bottom waters are separated from surface waters by a seasonal thermocline. Anoxia can decimate benthic communities. The number of coastal areas experiencing periodic or permanent anoxia has grown from just a handful in the mid-20th century to many hundreds, and the number continues to grow. There is strong evidence that nutrients, especially nitrogen, in sewage is responsible for much of the increase in the frequency and geographic extent of hypoxia and anoxia in coastal waters adjacent to major cities.

Fisheries.

Fishes are most abundant in areas of high primary productivity, especially in coastal waters, where trophic efficiency is higher because food webs are shorter. Coastal upwelling regions account for about half of the world's fishery production. Natural variations in fish stocks and reproductive success are so great that maximum sustainable yield cannot be established accurately. Many fisheries are currently exploited at or above their maximum sustainable yield, and a number have collapsed. Global fisheries production appears to have peaked and is now declining.

Estuaries.

Estuaries are regions where seawater mixes with freshwater from rivers. Each estuary is unique, so there is no simple description of a typical estuary, and no single classification system can capture the many variations. Most present-day estuaries were formed as sea level rose during the past 19,000 years. One way to classify estuaries is by their geologic setting. Coastal-plain estuaries are drowned river valleys, and are abundant on passive margins. Bar-built estuaries are located landward of sandbars or barrier islands. Tectonic estuaries are formed by uplift or subsidence of the coast at a fault. Fjords are drowned valleys previously cut by glaciers.

Another classification system categorizes estuaries by their dominant circulation pattern. In salt wedge estuaries, seawater forms a lower layer separated by a sharp halocline from a low-salinity surface layer. Seawater progressively mixes into the surface layer as this layer moves seaward, raising its salinity. In partially mixed estuaries, the halocline is much broader, usually because of greater vertical mixing by tidal currents. In well-mixed estuaries, the water column has vertically uniform salinity, but salinity progressively increases seaward. In all but inverse estuaries, residual (nontidal) currents due to the estuarine circulation flow landward in the bottom part of the water column and seaward in the surface layers. The flow rates of seawater into the estuary and of lower-salinity water into the ocean are many times greater than the flow rate of freshwater into the estuary. In the Northern Hemisphere, the Coriolis effect deflects flow into the estuary toward the right side as viewed from the ocean and deflects the outflow to the left side. Fjords have estuarine circulation only at depths above the top of their sill. Water below the sill depth is stagnant, often anoxic, and may be enriched in hydrogen sulfide. Occasional flushing of anoxic bottom water enriched in hydrogen sulfide can cause fish kills.

Particles and associated toxic metals and organics tend to be trapped and accumulate in estuaries. They can move seaward in the upper layer, but they sink to the lower layer that moves landward in estuarine circulation.

Estuaries are stressful environments for organisms because of variable temperature, salinity, turbidity, currents, and other environmental factors. However, they provide food and shelter from ocean predators, particularly in bordering wetlands. The estuarine circulation enables weakly swimming or planktonic species or their eggs and larvae to remain within the estuary. Anadromous fishes live in the ocean and transit estuaries to spawn in rivers. Catadromous fishes live in rivers and transit estuaries to spawn in the oceans.

13.7.1: Key Terms

13.7.2: Chapter Questions

13.7.3: References

This page titled 13.7: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.7.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- algae
- anadromous
- anoxia
- bar-built estuaries
- beaches
- benthic
- biomass
- biota
- blooms
- carnivorous
- catadromous
- chaotic
- chlorophyll
- climate
- coast
- coastal plain
- coastal upwelling
- coastal zone
- coastal-plain estuaries
- coastline
- community
- contaminants
- continental shelf
- convection
- coral
- coral reefs
- Coriolis effect
- crustaceans
- currents
- cyanobacteria
- cysts
- decomposers
- deltas
- density
- detritus
- diatoms
- ebb
- ecosystems
- eddies
- effluents
- Ekman transport
- encrust
- environments
- estuaries
- eutrophication
- excrete
- fault
- fecal pellets
- filter-feeding

- fjords
- flagellate
- flood
- food chains
- food webs
- friction
- fronts
- frustules
- glacial
- habitat
- halocline
- hard parts
- herbivorous
- herring
- hypoxia
- ice exclusion
- internal waves
- invertebrate
- ion
- isothermal
- kelp
- lagoons
- Langmuir circulation
- larvae
- latitudes
- leached
- levees
- light-limited
- longshore drift
- macroalgae
- mangrove
- marginal seas
- marine mammal
- maximum sustainable yield
- microalgae
- micronutrient
- monsoon
- nekton
- nutrient-limited
- oxygen demand
- parasitism
- partially mixed estuaries
- passive margins
- photic zone
- photosynthesize
- plankton
- pollution
- polyp
- primary productivity
- pycnocline
- reef
- residence times
- residual currents
- respiration
- resuspended
- runoff
- salinity

- salt marsh
- salt wedge estuaries
- seamounts
- sediment
- sedimentation
- shelf break
- shellfish
- shoal
- shore
- shoreline
- spatial
- spawn
- species
- species succession
- spores
- standing stock
- stratified
- suspended sediment
- symbiotic
- tectonic estuaries
- terrigenous
- thermocline
- tidal range
- tide
- topographic
- trade wind
- trophic efficiency
- trophic level
- turbidity
- turbulence
- upwelling
- vertebrates
- viruses
- wave periods
- wavelength
- weather
- well-mixed estuaries
- western boundary currents
- wetlands
- zooplankton
- zooxanthellae

This page titled 13.7.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.7.2: Chapter Questions

Study Questions

1. What characteristics of the coastal oceans make them different from the deep oceans?
2. How does coastal upwelling occur? How could you tell whether upwelling is likely in waters off a particular coast without making any direct observations of the oceans?
3. Describe the circulation in coastal upwelling zones. How do organisms take advantage of this circulation?
4. What are the sources of nutrients to the coastal photic zone? Discuss their distribution in the oceans.
5. Why are there seasonal cycles in populations of marine organisms in mid-latitudes and polar regions? Why is there little or no seasonality in tropical regions?
6. Why does primary production slow during midsummer in mid-latitudes, and why is there sometimes a fall bloom?
7. Why is there a succession of dominant phytoplankton species during a typical seasonal cycle? Why are diatoms usually replaced by flagellates during this succession?
8. What are the relative magnitudes of fish biomass and production in the deep oceans and coastal regions? Why are these not proportional to the relative areas of these regions?
9. What are the major types of estuarine circulation? Why do they occur in different estuaries?
10. If you were in Australia and operating a boat that had only a weak engine, on which side of an estuary would you choose to travel out to the ocean, and on which side would you choose to return? Why?
11. Why are estuaries stressful environments for marine organisms?

Critical Thinking Questions

1. Some areas of the coastal oceans have lower oxygen concentrations and a greater probability of developing periodic anoxia in the lower part of the water column than other areas. List and explain the factors that make some coastal areas more susceptible to anoxia than others.
2. Describe what would happen to the low-density organic particles in sewage if they were discharged to the surface layer at the center of a partially mixed estuary. Would the fate of these particles be different if they were discharged to the lower layer? If so, how?
3. It has been suggested that, if the Earth's climate does warm by several degrees Celsius, the seasonal cycle of primary production may be disrupted more in polar coastal regions than anywhere else on the Earth. Do you agree or disagree with this prediction? Explain the reasons for your answer.
4. Is it likely that a global climate change in which the Earth's atmosphere warms by several degrees Celsius will alter the frequency or intensity of dinoflagellate blooms? Why? Is it likely that the frequency and geographic extent of periodic anoxia in coastal waters will change? If so, how and why?
5. Many marine species are overfished. The current method of managing fisheries is to restrict the total catch of individual species to below the maximum sustainable yield of that species. However, this approach does not prevent adverse effects on other species that may depend on the targeted species for food. For example, it is hypothesized that Pribilof seal populations in the Bering Sea are declining because the fish species that they consume are being heavily fished, reducing the seals' food supply. A new approach to fishery management is developing in which ocean ecosystems are managed as a whole. For example, the growing fishery for krill in the Southern Ocean is managed by an ecosystem-based international treaty. Krill are the food supply on which a number of whale and seal species are dependent. Populations of these whale and seal species were drastically reduced by hunting in the first half of the twentieth century and have not yet recovered. Populations of penguin species that also rely on krill have increased to take advantage of the food previously eaten by whales and seals. If the krill fishery is to produce close to its maximum sustainable yield and contribute significantly to the solution of world hunger and protein deficiency, the whale and seal populations will never recover to their previous population levels, and it may be necessary to reduce penguin populations to protect whales and seals from renewed decline.
   (a) Do you think that marine ecosystems and fishing should be managed in this way? Explain your reasons.
   (b) What alternatives would you propose to solve the world hunger problem?
6. A swarm of locusts threatens the crops of a large area of North Africa. Should the locusts be eradicated to protect the food? A growing population of fur seals threatens to consume a growing percentage of fish from a regional fishery where the human population depends heavily on seafood. If the fish populations are not to crash from overfishing, the catch must be reduced, creating high prices and market shortages, or the fur seal population must be controlled. Would you kill seals to keep the population from growing so large as to damage the fishery? Did you have different reactions to these two scenarios? If so,

why?

This page titled 13.7.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

13.7.3: References

Figure 13-1 – Segar, D., and E. Stamman Segar, 1997: Coastal waters in this California bay.

Figure 13-4a – Segar, D., and E. Stamman Segar, 1997: Various species of green, brown, and red macroalgae in a tide pool on the California coast near San Francisco.

Figure 13-4b – Segar, D., and E. Stamman Segar, 1997: A red calcareous macroalga (phylum Rhodophyta) in the Philippines.

Figure 13-4c – Segar, D., and E. Stamman Segar, 1997: The upper fronds of kelp (*Macrocystis sp.*) floating on the water surface in Monterey Bay, California.

Figure 13-4d – Segar, D., and E. Stamman Segar, 1997: A coral reef green macroalga called “grapeweed” (*Caulerpa racemosa*I in Papua New Guinea).

Figure 13-4e – Segar, D., and E. Stamman Segar, 1997: An encrusting red alga (phylum Rhodophyta) covering a platelike hard coral in Hawaii.

Figure 13-8 – Adapted from Gross, M. G., and E. Gross: 1996: *Oceanography: A view of Earth*. Prentice Hall, 472 pp.

Figure 13-11 – Adapted from Bigelow, H. G., 1927: Physical oceanography of the Gulf of Maine. *Bull. U. S. Bur. Fish.*, **40**, 511–1027.

Figure 13-12a – Segar, D., and E. Stamman Segar, 1997: Typical reef-building hard coral colony (probably *Diploastrea heliopora*) in Papua New Guinea.

Figure 13-12b – Segar, D., and E. Stamman Segar, 1997: Typical hard coral (possibly *Pocillopora sp.*) in Papua New Guinea.

Figure 13-12c – Segar, D., and E. Stamman Segar, 1997: Individual polyps of *Diploastrea heliopora* (Papua New Guinea) when the polyps have extended their soft tentacles to feed on zooplankton or particles.

Figure 13-12d – Segar, D., and E. Stamman Segar, 1997: Individual polyps of *Diploastrea heliopora* (Papua New Guinea) at night.

Figure 13-15a – Steimle, F. W., and C. J. Sindermann, 1978: Review of oxygen depletion and associated mass mortalities of shellfish in the Middle Atlantic Blight in 1976. NOAA NMFS, Marine Fisheries Review Paper 1356, 17–26, Source.

Figure 13-15b – Gulf Hypoxia, 2017: Gulf of Mexico hypoxia and anoxia 2017. NOAA, accessed 16 March 2026, Source.

Figure 13-15c – Data from Gulf Hypoxia, 2026: Shelfwide Cruises., accessed 16 March 2026, Source.

Figure 13-16 – Data from Ryther, J. H., 1969: Photosynthesis and Fish Production in the Sea: The production of organic matter and its conversion to higher forms of life vary throughout the world ocean. *Science*, **166**, 72–76, Source.

Figure 13-17 – Segar, D., and E. Stamman Segar, 1997: The ancient Greek city of Ephesus, now in Turkey.

Figure 13-20 – Adapted from Pinet, P. R., 2006: *Invitation to Oceanography 4th Edition*. Jones and Bartlett Publishers, 596 pp.

Figure 13-21 – Partially adapted from Pethick, J., 1984: *An introduction to coastal geomorphology*. Edward Arnold, 260 pp.

Figure 13-X – NOAA Worldview, 2025: Satellite image of the San Francisco Bay on 9 January 2025. AQUA/Modis, accessed 16 March 2026, Source.

This page titled 13.7.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

14: MARINE ECOLOGY

Technical Editor: Emily Brownlee, Ph.D. (St. Mary's College of Maryland)

Additional contributions and edits by Elaine Stamman, M.S. (Texas A&M University, retired)

Critical Concepts

- **CC14 Phototrophy, Light, and Nutrients:** Phototrophy (which includes photosynthesis) and chemosynthesis are two processes by which simple chemical compounds are made into the organic compounds of living organisms and upon which all species are ultimately dependent. Phototrophy depends on the availability of light and can only take place in a shallow upper layer of water or on the shallow seafloor. Chemosynthesis does not use light energy but instead uses chemical energy from reduced compounds. Therefore, chemosynthesis can occur in all ocean environments in which oxygen is depleted, but these environments are very limited in extent in the present-day oceans.
- **CC17 Species Diversity and Biodiversity:** Biodiversity is an expression of the range of genetic diversity; species diversity; diversity in ecological niches and types of communities of organisms (ecosystem diversity); and diversity of feeding, reproduction and predator avoidance strategies (physiological diversity), within the ecosystem of the specified region. Species diversity is a more precisely-defined term and is a measure of the species richness (number of species) and species evenness (extent to which the community has balanced populations with no dominant species). High diversity and biodiversity are generally associated with ecosystems that are resistant to change.

Figure 14-1. Each of these images show the flamboyant cuttlefish (*Metasepia pfefferi*), which can grow to about 10 cm long. The four larger images illustrate the mastery that this species has with changing its colors. These four photos, all of the same cuttlefish, were taken in succession over a period of less than one minute. Unfortunately, the full startling range of its color cannot be captured in photographs, it can only be seen properly in person. The smaller image in the bottom left corner is a young cuttlefish, which is only about one centimeter long. It is no more than a few days old, but is already capable of the same color changing artistry that the adults of its species possess. In fact, the cuttlefish can alter their colors even before hatching from their transparent egg cases, as seen later in this chapter in **Figure 14-28**.

The oceans are populated by uncounted millions of **species**, most of which have not yet been identified. Marine ecology is the study of relationships between species and between species and their **environment**.

Because each species has a unique relationship to the ocean environment and to other species, marine ecology is a complex discipline. This chapter describes the environmental factors that govern the distribution of marine organisms, and the types of adaptations that have evolved in marine species to meet the challenges of living in the ocean.

14.1: Ecological Requirements
14.2: Habitats
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.7: Associations
14.8: Communication and Navigation
14.9: Chapter Summary
14.9.1: Key Terms
14.9.2: Chapter Questions
14.9.3: References

This page titled 14: Marine Ecology is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.1: Ecological Requirements

A fundamental imperative of life is survival of the species. Hence, all species must meet two requirements. First, an adequate food supply must be ensured for each life stage. Second, enough individuals of the species must survive until they reproduce successfully. The principal challenges are

- Where to live
- What food to eat and how to obtain it
- How to avoid predators
- Where, how, and when to reproduce

In meeting these challenges, marine species have adopted an extraordinary variety of life cycles to take advantage of the various attributes of the marine environment.

This page titled 14.1: Ecological Requirements is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.2: Habitats

The major **habitats** available to marine organisms are in the water column, on the seafloor surface, and in seafloor **sediment** (**Chap. 12**). For **photosynthetic** organisms (**CC14**), only the first two habitats are suitable, and only within the **photic zone**. Consumer and **decomposer** species are present in each of the three habitats: **pelagic** species in the water column, **benthic epifauna** and epiflora on the seafloor, and benthic **infauna** within the sediment.

The three habitats offer different advantages and challenges. Each species has selected its own unique balance between those opportunities and challenges. Many species take advantage of different balances at successive stages of their life cycle by changing their habitat between stages.

In shallow water, photosynthesis may take place at the seafloor if it is within the photic zone, and **chemosynthetic communities** are present at least in restricted areas of the seafloor below the photic zone. In both locations, **primary production** by benthic organisms creates some food. This primary production can be substantial locally, particularly in coastal waters and at **hydrothermal vents**, but in most of the ocean the food supply is produced by photosynthesis by phytoplankton that include algae and cyanobacteria. Some archaea utilize light energy to supplement their energy needs, so they are **phototrophs**, but they use a mechanism other than photosynthesis and may use simple organic compounds instead of carbon dioxide. In the deep ocean, nutrient depleted, high oxygen waters and deep sea sediments, certain species of **archaea** are the dominant “primary producers” (**Chap. 12, CC14**).

Pelagic Habitat

Phytoplankton and other phototrophic microorganisms in the pelagic environment are effectively **grazed** by **zooplankton** and single-celled **protists** that, in turn, are eaten by **carnivores** or omnivores. During this process, most of the **biomass** produced by primary producers is used by photic-zone pelagic consumers to fuel their energy needs (**Chap. 12**). Consequently, only a small fraction of primary producer biomass eventually sinks or is transported below the photic zone, and food is scarce in the environments below the photic zone.

Most pelagic organisms must actively seek food and avoid or fight off predators at the same time. Whatever strategy an organism uses for these activities, it uses up energy that must be obtained from food. Hence, one penalty for living in the pelagic habitat is that more food must be consumed than is needed for growth and reproduction. The need to avoid sinking also carries an energy penalty for pelagic organisms. **Buoyancy** can be maintained by swimming, which uses energy directly, or with pockets of gas or low-**density** oil in the body structure, each of which requires energy to produce.

All pelagic organisms must evade predators, compete for food, and counteract the tendency to sink. Below the photic zone, predators are less abundant and easier to avoid because they cannot hunt visually in the darkness. Offsetting this advantage is the scarcity of food below the photic zone. Many species, especially of zooplankton and squid, migrate between the photic zone and the **aphotic zone** to take advantage of this trade-off between the zones. These organisms feed in the photic zone at night and then descend to the darkness below in the day, in a strategy known as **diel** vertical migration. Nevertheless, even this behavior has an energy cost, because energy must be expended to move vertically between layers.

Benthic Epifaunal Habitat

Organisms that live on the seafloor do not need to expend energy to control buoyancy and generally require less energy to find and capture food than pelagic organisms require. Hunting is reduced to a two-dimensional problem, and because many benthic prey are sedentary or nearly so, hunting movement can be slow and energy-efficient. Even more energy is saved by **benthos** that remain sedentary and wait for prey to walk, swim, or drift by.

The energy savings of living in the benthic epifaunal environment may be offset by the scarcity of food and the difficulty of evading predators. Organisms that move slowly and sit on the sediment may be easy to find and capture. Hence, many benthic epifaunal species have developed defensive mechanisms such as poisons, spines, and shells or tubes into which they withdraw. Such mechanisms require energy that otherwise could be used for growth and reproduction. Where food is readily available, epifauna may be abundant and the competition for available space may be a significant disadvantage. Finding living space is a problem, particularly for epifauna that attach to rocks, because most of the seafloor is covered by sediment and rocky substrates

are very limited.

Although the benthic epifaunal environment in shallow water is similar to that in deeper water, benthic algae enhance the food supply in many shallow areas. Coral reefs sustain exceptionally diverse (**CC17**) and abundant encrusting algae communities. Also, in some shallow areas the kelp in kelp forests (**Fig. 15-8**) provides abundant hiding places for predator avoidance.

Benthic Infaunal Habitat

Even though food is scarce in most of the benthic infaunal environment, some organic **detritus** accumulates continuously in sediments. Many animals, such as worms and certain **crustaceans**, feed by sifting sediment grains to obtain organic particles or by digesting organic matter as sediment is passed through the gut. Other benthic infaunal species live in the sediment primarily for protection from predators and feed on particles in the water above or prey on passing animals.

The relatively poor supply of food available to benthic infauna is offset by the energy savings of relatively sedentary lifestyles, by the lack of need for buoyancy control, and by the reduced need for defenses against predators. In **estuarine** and other coastal environments, much of the **salinity** and temperature variation that can occur in overlying waters does not occur in the sediment, so infauna are not subject to these environmental stresses. Intertidal benthic infauna also avoid exposure to air during low **tides**.

Benthic infauna must expend energy to dig into or move through sediments or rock. They must also cope with an environment that varies because of **biochemical** and chemical processes within the substrate. The most important of these processes consume dissolved oxygen and produce toxic hydrogen sulfide, so most infaunal species can live only in the oxygenated surface layer of sediment. In areas where detritus inputs are high and water movements slow, the oxygenated layer is thin or absent, but elsewhere it ranges from a few millimeters to several meters in thickness. Where sulfide is present, the **biota** consists primarily of **bacteria,** archaea, and **fungi** adapted to the sulfide environment and animals that build burrows or tubes through which they can obtain oxygenated water from above.

Other Ocean Habitats

Special challenges and opportunities are also found in other ocean habitats, including the **surface microlayer**, the **intertidal zone**, and hydrothermal vents. The surface microlayer, which is only a few molecules thick, concentrates a variety of natural and **contaminant** organic compounds. A few species attach themselves to the surface microlayer to keep from sinking. **Surface tension** tends to prevent small particles within the microlayer from sinking, even if their density is higher than that of seawater. Many species distribute their eggs and **larvae** by placing them in the surface microlayer, where **currents** distribute them until they grow large enough to feed and swim independently. Placement in the surface microlayer reduces the need to use energy to provide the eggs and larvae with oils or other means of buoyancy control.

Although the intertidal zone and hydrothermal vents are sites of very high **primary productivity** and food is abundant, each of these habitats imposes substantial offsetting challenges due to variations in environmental conditions. These two unique environments are discussed in **Chapter 15**.

This page titled 14.2: Habitats is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.3: Feeding

With the exception of organisms that obtain their energy needs through **phototrophy** or chemosynthesis, all marine species must obtain energy from organic compounds that they obtain as food. The food can consist of another living organism, nonliving organic particles, or dissolved organic compounds.

Because concentrations of dissolved organic compounds in seawater are extremely low, few marine species are known to rely on dissolved organic matter as their principal source of sustenance. Nevertheless, this source of food may be important to some species, including bacteria and other decomposers, and a few higher animals. To use dissolved organic matter, an organism must take up individual dissolved molecules through its membrane surface. This process is very slow if dissolved compounds are transported to the membrane by **diffusion**, but it can be enhanced if water flows over the membrane.

In terrestrial **ecosystems**, animals feed in three basic ways: by grazing the abundant macroscopic (large) plants that cover much of the Earth's surface, by consuming detrital organic particles in leaf litter and soils, and by hunting and eating other animals. Similarly, in the oceans many species graze on the seafloor, eat detritus in the sediment, and hunt and eat other animals. However, because most autotrophs and detritus in the oceans consist of small particles dispersed in the water column, many marine species are **suspension feeders** that feed on particles.

Suspension Feeding

Suspension feeders must be able to efficiently gather the phytoplankton, zooplankton, and detritus particles that constitute their food. They do so in several different ways. Many suspension feeders are **filter feeders** that have a web-like or mat-like structure to capture particles as water flows through it. This simple method works efficiently for suspension feeders that filter out and eat large particles, but filter feeding becomes more difficult as the size of the filtered particles becomes smaller. Zooplankton that eat small phytoplankton and detritus particles would need a filter mesh so small that **viscosity** would restrict the flow of water through the filtering apparatus. Consequently, these zooplankton generally have evolved hairlike appendages, called "setae," on their mouths. As water passes through the mouth, particles are captured by the setae.

As an alternative or sometimes an addition to filter feeding, many suspension feeders secrete mucus to which particles adhere as water flows past. The organism then ingests the mucus and the food that it contains. Many other suspension feeders capture particles that come in contact with appendages designed to grab or grasp a particle and transfer it to the mouth. A number of zooplankton species feed on relatively large prey by capturing them in this way with setae on their appendages. When this mechanism is used to capture live prey, it can be considered a hunting method and the organism can be considered a predator.

Many suspension feeders ingest particles selectively, most often by size. Particles smaller than a certain size are not captured by the filtering or other feeding apparatus, and particles larger than a certain size cannot be captured or passed into the mouth. Different species have evolved to collect different size ranges of particles. Certain zooplankton species that are adapted for capturing large phytoplankton, such as **diatoms**, tend to be most abundant where and when diatoms are the dominant phytoplankton. Other zooplankton species are better adapted to capture smaller phytoplankton, such as **flagellates**, and therefore are most abundant when flagellates dominate. Because different species of carnivores feed on different **herbivorous** zooplankton species, the size of food particles available to suspension feeders can affect the entire food web.

Many suspension feeders selectively ingest only certain species within a selected size range of particles. For example, herbivores consume only living phytoplankton cells and reject detritus particles of similar size. Such selective feeding requires the feeding organism to expend energy to sort its food supply. The benefit gained is that the selected food has higher nutritional value than non-selected species or detritus has. In contrast, many suspension feeders are nonselective omnivores that ingest bacteria, archaea, algae, animal, and detritus particles of appropriate size without discrimination.

Suspension-feeding species inhabit both pelagic and benthic environments. Pelagic suspension feeders include zooplankton and **nekton**. Some **planktonic** suspension feeders simply drift through the water, capturing food particles brought to them by turbulence and diffusion. The chance of encountering a food particle in this way is low because of the low concentrations and slow movement of the particles in the water, so filter feeders generally have evolved methods of ensuring a flow of water through their feeding apparatus. The three basic methods are:

- Actively pumping water through the filtering apparatus

- Moving the filtering apparatus through the water
- Keeping the filtering apparatus stationary and allowing ocean currents to move water through it

Marine organisms have developed numerous physiological variations and apply one or a combination of these methods to capture food.

Pelagic Suspension Feeders

Many pelagic suspension feeders have evolved mechanisms to move or pump water past or through their feeding apparatus as they drift through water. Crustaceans such as **copepods** and **euphausiids** (**Fig. 12-19a,b**) have long, slender forelimbs or appendages surrounding their mouths (**Fig. 14-2a**), with which they grasp or direct suspended food particles toward the mouth. In contrast, the equivalent mouthparts of most shrimp, also crustaceans, are designed to cut or crush, reflecting the shrimp's hunting or **scavenging** feeding habits (**Fig. 14-2b,c**). Suspension-feeding crustaceans and many other zooplankton can swim weakly through the water to increase their chances of encountering food particles.

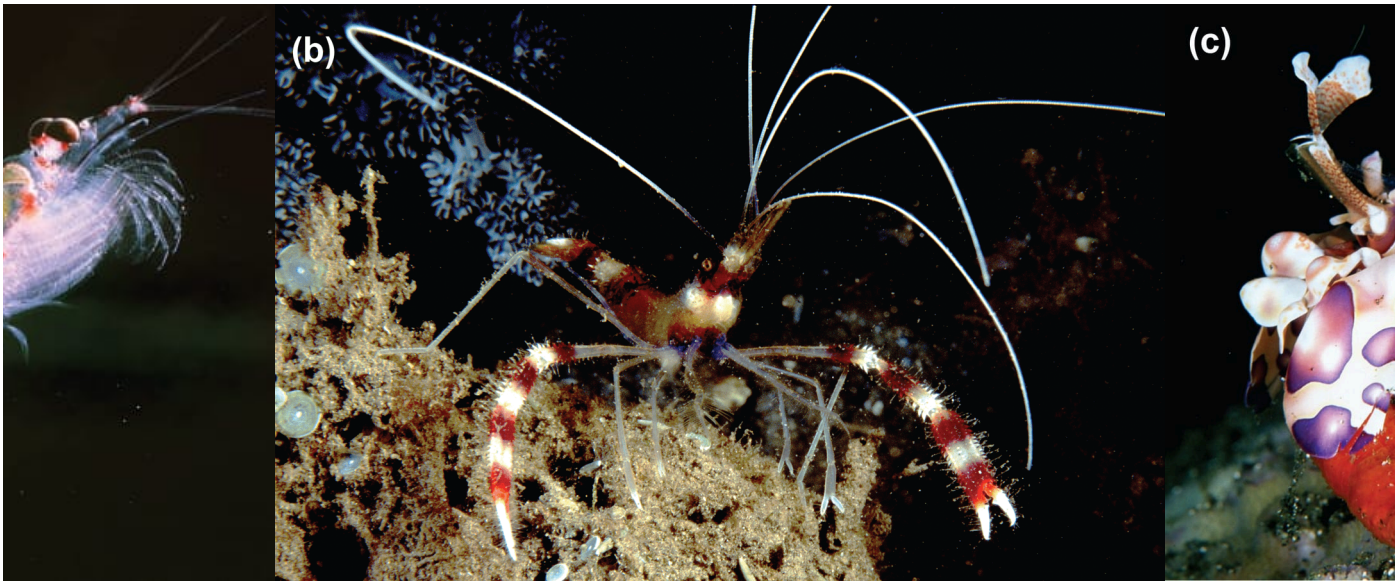


Figure 14-2. Forelimb use in crustaceans. (a) A deep-sea euphausiid or krill. (b) Banded coral shrimp (*Stenopus hispidus*, Papua New Guinea). (c) Harlequin shrimp (*Hymenocera elegans*, Papua New Guinea). The forelimbs of the krill are adapted for filter feeding. Compare these with the forelimbs of the coral shrimp, which eats larger food particles than the krill does, and the massive forelimbs of the harlequin shrimp, which feeds on tough sea star bodies.

Pteropods, which are **mollusks** related to slugs and snails, have adapted to suspension feeding by evolution of their foot (like the foot on which a snail crawls) to be a membrane that extends from the body (**Fig. 14-3**). The pteropod uses this sail-like membrane as a paddle to propel itself slowly through the water. This action moves water past the mucus-covered membrane, which captures suspended food particles that brush against it.

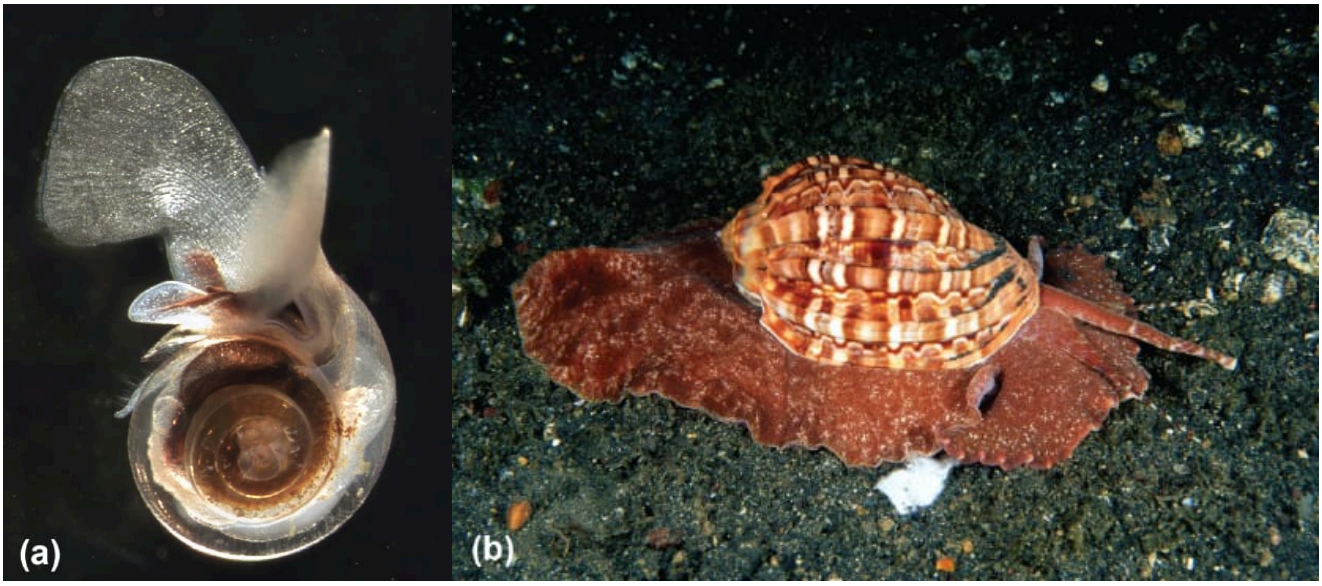


Figure 14-3. Some adaptations of the foot in gastropods. (a) Typical pteropod (*Limacina helicina*). (b) Marine gastropod, a harp shell (*Harpa major*, Indonesia). The transparent part of the pteropod is its foot, which is very different from that of other gastropods, such as the harp shell. The harp shell's foot is extended from its shell flat onto the seafloor and allows it to "walk."

Salps, a type of **tunicate**, evolved in aquatic environments to take advantage of suspended particulate food. Many salps have a simple body resembling an elongated barrel (**Fig. 12-20e**) with an opening at either end. Within the barrel, the salp is coated with a continuously moving mucous sheath, which captures food particles from water pumped continuously through the salp's body by sequenced contractions of bands of muscles. There are also colonial salps, which are composed of individual members joined to form the outer wall of a sac-like structure. Each individual's **incurrent** opening faces outward, and each **excurrent** opening discharges into the common open space inside the sac (**Fig. 14-4**). This arrangement enables the colony to "swim" as water is forced out of the relatively narrow sac opening by the combined pumping efforts of its members. As a result, each individual is more likely to encounter food particles than it would if it operated alone.

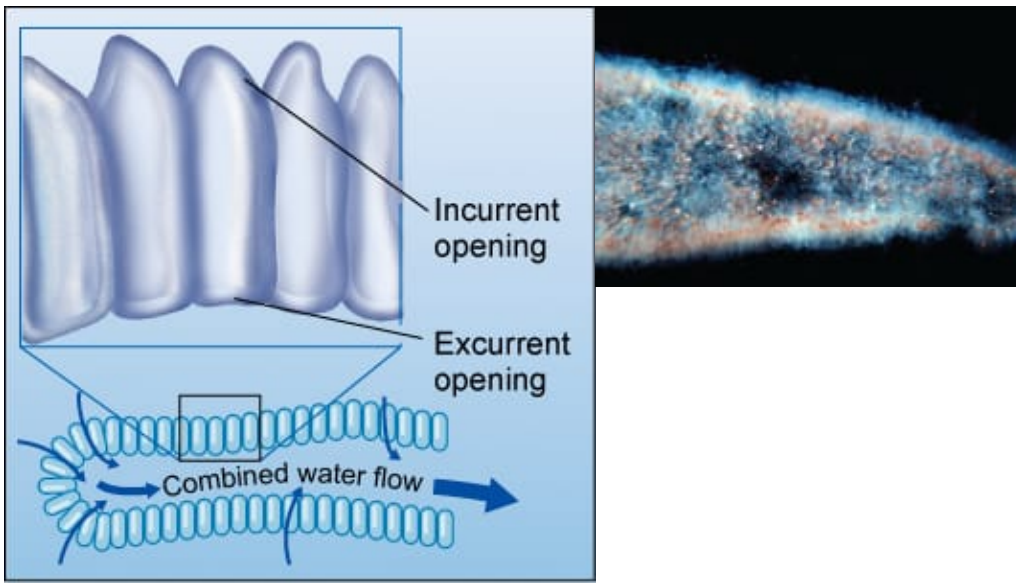


Figure 14-4. Salps. Each individual of a salp colony draws water into its body and filters it for food. The combined water flow from all individuals is passed out of the open end of the sac-like colony, which slowly propels the colony through the water. If you look carefully in the inset photograph, you will see the individual members of the salp colony. Note that they resemble their relatives, the tunicates shown in **Figure 14-9**.

Most **jellies** and **ctenophores** (**Fig. 14-20d**) feed on suspended particles by extending long trailing tentacles that, in some species, have poisonous stinging cells and, in others, are covered with a sticky substance to capture food. These organisms are

generally carnivores that feed selectively on zooplankton or small fishes, and many propel themselves by contracting rhythmically while trailing their tentacles through the water in order to increase their chances of encountering food. The Portuguese man-of-war achieves a similar increase in mobility and feeding efficiency by allowing winds to blow the colony across the ocean surface.

Although many jellies and ctenophores are suspension feeders, larger species of these organisms are clearly hunters that selectively seek, kill, and consume small fishes and other nekton. The distinction between suspension feeding and hunting is not precise. Some species are clearly suspension feeders and some are clearly hunters, but other species use elements of both approaches and feed on both suspended particles (plankton and/or detritus) and small nekton.

Benthic Suspension Feeders

Like other benthos, benthic suspension feeders benefit from the low energy needs of their sedentary lifestyle and the lack of a need to control buoyancy. They have the additional advantage that ocean currents bring food to them, which reduces the energy needed to hunt for food. Consequently, many benthic epifaunal and infaunal species are suspension feeders. Suspension feeders are especially abundant in coastal regions where suspended particulate food is plentiful.

In soft sediments that cover much of the seafloor, certain infaunal suspension feeders pump water into their feeding apparatus through tubes that they extend upward into the water (Fig. 14-5). This feeding method is particularly advantageous in intertidal mudflats because a buried mollusk can withdraw its siphon and close its shell for protection when the sediment is exposed at low tide. On mudflats or muddy sand **beaches** where cockles are abundant, buried cockles can be detected as the tide **ebbs** because they squirt a small fountain of water into the air as they close abruptly when the water recedes.

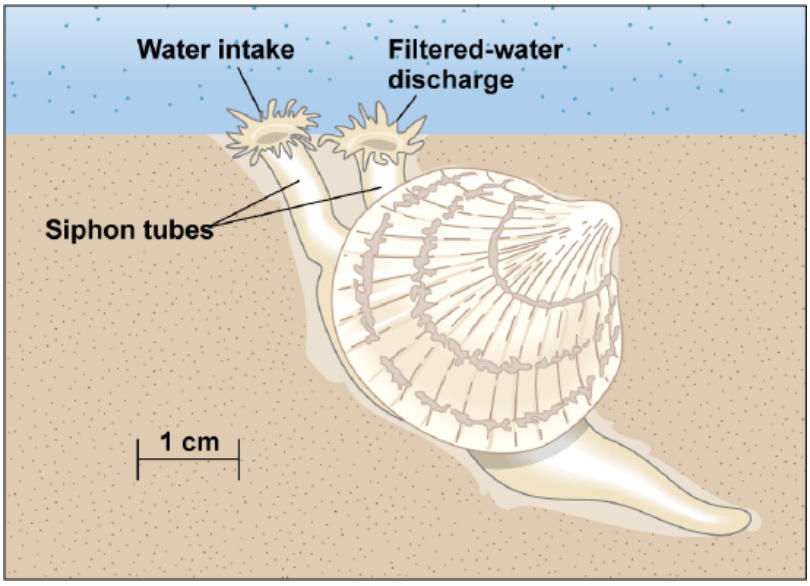


Figure 14-5. Some bivalve mollusk species, such as this cockle (family Cardiidae), live buried within the sediments but extend two tubes up into the water column just above the surface. They pump water in from one tube and out the other, and they filter food particles as the water passes through their bodies.

The sea pen is a benthic infaunal suspension feeder that extends a beautiful and intricate fanlike structure into the water column to feed (Fig. 14-6a,b). Suspended particles are captured as they drift with the current into the fan. The fan can be folded up and withdrawn into the sediment where the main body of the organism is buried. Many sea pens withdraw by day and open to feed only at night, when zooplankton are more abundant because many species migrate from below the photic zone and nocturnal zooplankton emerge from their daytime hiding places. Other types of **invertebrates**, including some species of tube worms, **sea cucumbers**, and **anemones**, live in soft sediment and extend feeding tentacles or webs into the water (Fig. 14-6).

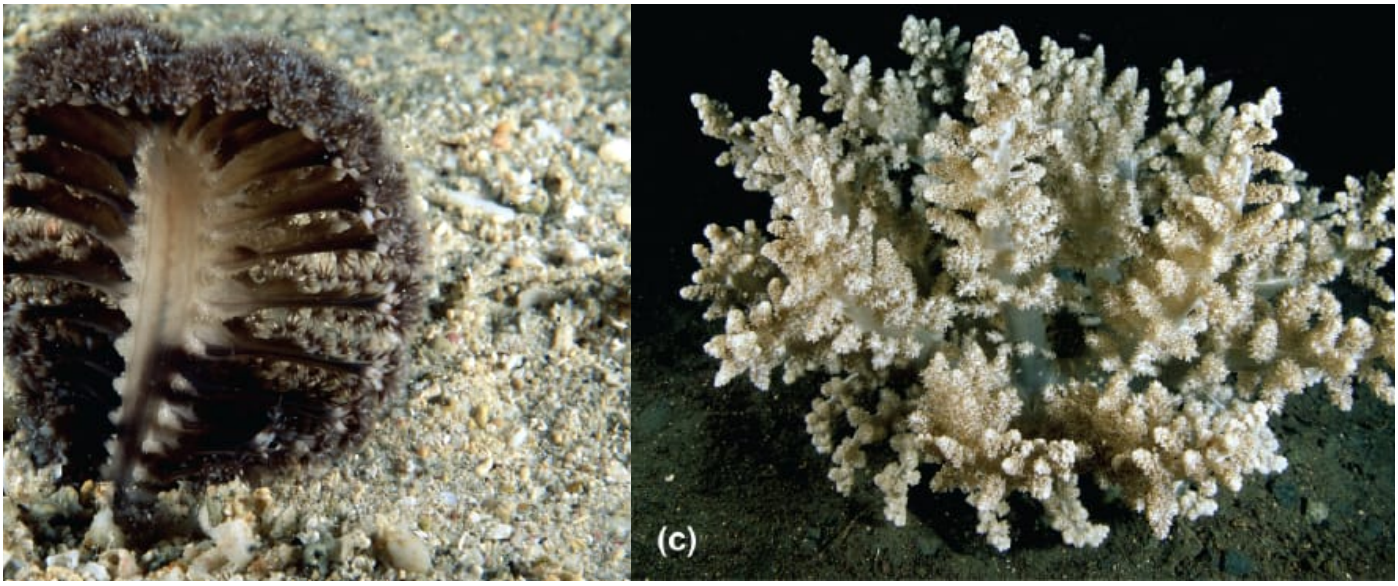


Figure 14-6. Examples of sand-dwelling benthic infaunal suspension feeders. (a) A sea pen (order Pennatulacea, Papua New Guinea) with its feeding polyps oriented into the current. (b) The reverse side of the same sea pen shown in (a). (c) This sand anemone (Papua New Guinea) feeds on large "suspended" matter such as small fishes. If it is threatened or when it captures prey, it can withdraw these tentacles almost instantly and completely into a hole in the sand, where the remainder of its body is hidden. (d) A tube worm (family Sabellidae, Papua New Guinea) with its netlike feeding apparatus extended into the water column. This worm, when it senses danger withdraws almost instantaneously and completely into its protective tube which is buried in the sand. It senses potential predators, by changes in lighting (e.g. shadows) and small pressure changes (vibrations) so getting close enough to photograph them open like in this image is not easy

Epifaunal suspension feeders are especially abundant and diverse in **continental shelf** regions, where there is a solid substrate to which they can attach without being covered by sediment. In temperate and high **latitudes** these areas are limited to a few rocky outcrops on the seafloor, but in tropical regions **coral reefs** provide extensive areas of suitable substrate. Suspension-feeding epifaunal invertebrates include species of mollusks, crustaceans, tunicates, **corals**, anemones, sea cucumbers, **sea stars**, and worms.

The most familiar epifaunal suspension-feeding mollusk is the mussel (**Fig. 14-7a**), which attaches itself to any available substrate and partially opens its shell to pump water through its body and feed. Other bivalve mollusks, such as the spiny oyster (**Fig. 14-7b**), use a similar method in coral reef communities, where they are firmly attached to the reef. Some scallop species in temperate latitudes do not attach to the seafloor, but instead live on coarse sand or gravel bottoms, from which they can "swim" up into the water column for short distances by snapping their shells shut to create propulsive jets of water.

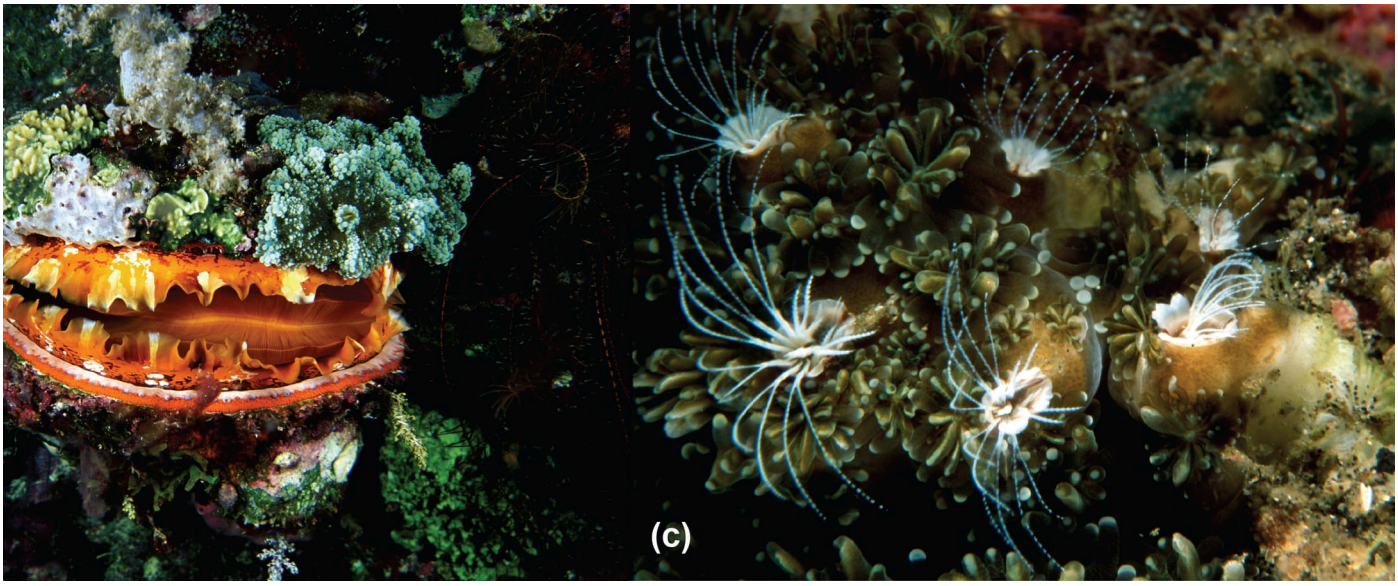


Figure 14-7. Examples of suspension-feeding invertebrates. (a) California mussels (*Mytilus californianus*, Monterey, California) grow in profusion in shallow waters. Other species of mussels are common in other areas. (b) A Pacific spiny oyster (*Spondylus varians*, Papua New Guinea) attaches to a coral reef wall and opens its shell so that it can feed by pumping water through its body. (c) A colony of barnacles in Monterey, California. The colony contains more than one species, in this case probably primarily *Balanus glandula* and *Chthamalus sp.* (d) Barnacles, including these coral barnacles (order Pyrgomatidae, Indonesia), use fanlike modified feet to sweep through the water and collect food particles.

Barnacles are suspension-feeding epifaunal crustaceans (**Fig. 14-7c**) that look nothing like their close relatives shrimp and crabs, because they have evolved an unusual feeding method. Barnacles, in essence, lie on their backs, strongly attached to rocks or other hard substrates such as ships' hulls, and are protected by hard plates and shells. They open the plates and extend their much altered legs as a weblike structure that sweeps through the water to grasp and capture suspended particles (**Fig. 14-7d**).

Most of us think of coral reefs as large, strangely shaped, hard, rocklike structures (**Fig. 14-8a,b**). However, reef-building corals are actually extremely small suspension-feeding epifaunal organisms that grow in colonies, each of which may contain millions or even billions of individuals, called **polyps**. Some of the many species of corals have **hard parts** that are left behind when individuals die and thus serve as a base on which other polyps can grow.

Figure 14-8. Polypoid species (class Anthozoa) take many different forms. (a) A hard coral known as brain coral (*Diploria sp.*, Papua New Guinea). (b) A hard coral often called staghorn coral (*Acropora sp.*, Papua New Guinea). (c) Close-up showing polyps of a hard coral species (*Tubastrea sp.*, Fiji) extended to feed at night. (d) Colony of a soft coral (*Dendronephthya sp.*, Fiji). *Dendronephthya* occur in many different colors. (e) Typical gorgonian sea fan (*Melithaea sp.*, Papua New Guinea). These fans can consist of millions of individual polyps and be several meters across.

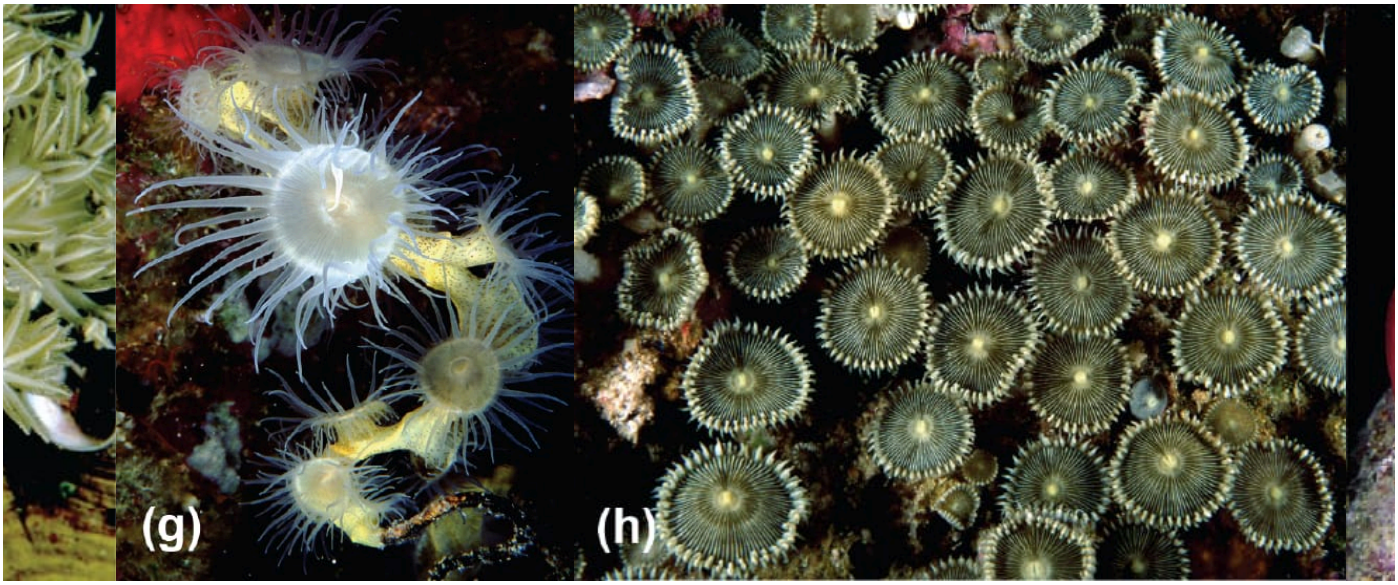


Figure 14-8. Polypoid species (class Anthozoa) take many different forms. (f) Close-up of a soft coral colony (*Xenia sp.*, Papua New Guinea) showing individual polyps at different stages of feeding. The eight fingers surrounding each polyp's mouth open wide to capture suspended particles and then close like a fist to move the particles to the mouth. (g) A small suspension-feeding colonial anemone (*Nemanthus annamensis*, Fiji). (h) A colonial zooanthid (*Protopalythoa sp.*, Papua New Guinea). (i) A "magnificent sea anemone" (*Heteractis magnifica*, Fiji), here providing a home for a pink anemonefish (*Amphiprion perideraion*).

The living coral individuals, or polyps, feed by extending tentacles that capture food particles, primarily zooplankton, and draw them down into the mouth (**Fig. 14-8c**). Most hard corals extend their tentacles to feed only at night, but many species of soft corals, which are found mostly in the Pacific and Indian Oceans, feed both day and night. Many of these spectacular soft corals (**Fig. 14-8d,f**) come out to feed (open their polyps) only when there is a strong current to bring abundant food supplies to the colony. Many corals, both soft and hard, produce colonies that extend upward or outward into the water from their attachment point (**Fig. 14-8b,d,e**). This spreading maximizes the volume of water that passes over the colony and, thus, the amount of particulate food that passes within reach. Many species of zooanthids and anemones, close relatives of corals, are also suspension feeders that eat plankton and detritus (**Fig. 14-8g,h**), but larger anemones (**Fig. 14-8i**) are primarily carnivorous hunters that eat fishes and invertebrates.

There are many species of suspension-feeding epifaunal tunicates, particularly on coral reefs. They feed in the same way as salps but are attached to the substrate and have both their incurrent and excurrent openings on their upper bodies. Tunicate species include a variety of single and colonial forms (**Fig. 14-9**).

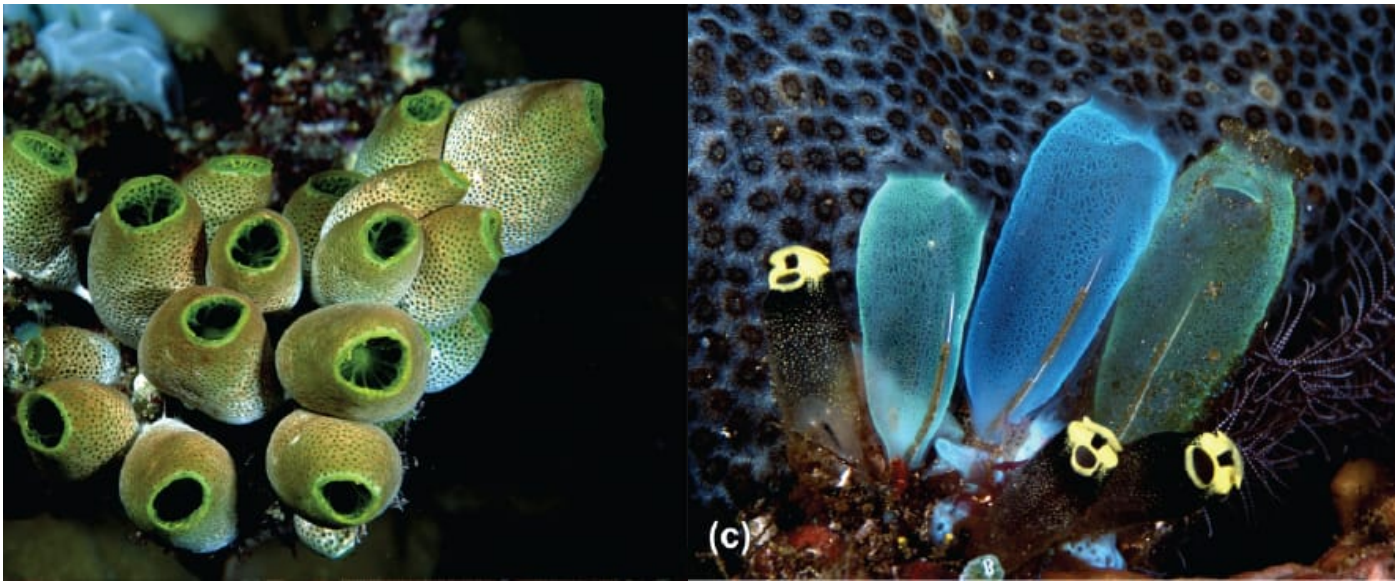


Figure 14-9. Tunicates (also called "ascidians") are among the most advanced invertebrates, although they appear very simple in structure. (a) Lightbulb tunicate (*Clavelina sp.*, Philippines). (b) The colonial tunicate species *Didemnum molle* (Papua New Guinea). Each of the rounded forms is a colony. (c) There are two species of tunicates in the photograph. The three tall, blue and blue-green individuals are *Rhopalaea sp.*, and the three smaller individuals with darker bodies and yellow markings around their incurrent and excurrent openings are *Clavelina robusta* (Indonesia). (d) A very different form of colonial tunicate that covers and encrusts the reef (*Botryllus sp.*, Indonesia).

Certain species of sea cucumbers, sea stars, and worms suspension feed by extending intricate tentacles or weblike structures into the water column (**Fig. 14-10**). The elaborate basket star (**Fig. 14-10b**) comes out to feed only at night and spreads its intricate arms, which can extend more than a meter, across the current to capture food efficiently from huge volumes of water. During the day, the basket star coils into a tiny ball and hides in crevices in the reef. Fan worms (**Fig. 14-10c**) and Christmas tree worms (**Fig. 14-10d**) live in tubes that are usually drilled into, or built with the surrounding growth of a hard coral colony. They feed by day but have an amazing ability to sense movement or shadows of moving objects and withdraw into their tubes instantaneously as divers or predators approach.

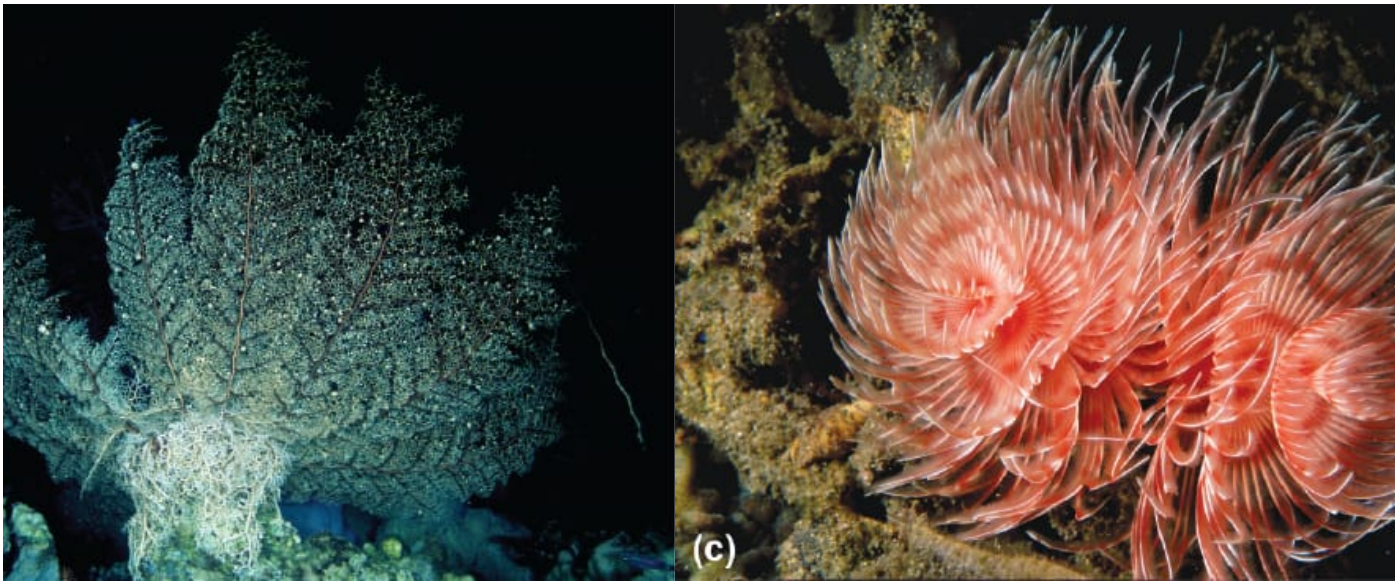


Figure 14-10. Examples of suspension feeders that employ weblike structures to capture food particles. (a) A creeping sea cucumber (*Cucumaria sp.*, Philippines) with its suspension-feeding apparatus extended. (b) A basket star (family Gorgonocephalidae, Papua New Guinea) with its arms extended into the current to feed at night. (c) A sabellid spiral fan worm (*Protula magnifica*, Indonesia). (d) A Christmas tree worm (*Spirobranchus giganteus*, Papua New Guinea).

Surface Grazing

Although shallow seafloor areas with macroscopic algae that can be grazed by herbivores are rare, much of the seafloor provides sufficient, and in some areas abundant, food for grazers. This food supply varies in composition from location to location, but on most of the ocean floor it consists primarily of detritus and the bacteria and other decomposers associated with the detritus. The concentration of these foods generally decreases with ocean depth and is higher where the seafloor is below a region of high pelagic primary productivity.

Note that, in marine ecosystems, the term grazing includes feeding on detritus and animals and is not restricted to plant eating, its common terrestrial usage. On the seafloor within the photic zone, detritus and associated decomposer biomass are supplemented by other sources of grazer food, including benthic **microalgae** and **macroalgae**. Benthic microalgae, particularly diatoms, are abundant on shallow seafloor, where light and **nutrients** are abundant and where waves or other water motions rarely **resuspend** sediments to cause abrasive scour.

Many marine animals that live on the seafloor, especially species that live on hard substrates, are colonial forms whose individuals are small and immobile. The colonies can be grazed by other animals without significant harm to the colony because grazing removes only a limited proportion of the colony's individuals, which can be replaced relatively quickly by reproduction. Many colonial animal species thus provide a renewable food supply for grazers in much the same way that plants do for terrestrial grazers.

Numerous adaptations have evolved for surface grazing. Many species of surface grazers obtain food by sifting the upper layer of sediment and therefore can also be considered **deposit feeders**. Deposit feeders, which are species that obtain food from within sediments, are described later in this chapter. Species that surface-graze may also suspension-feed or hunt larger prey.

Some surface grazers, such as **sea urchins** (Fig. 14-11a), have evolved such that their mouth is on their underside. These organisms crawl over sand and rocks, scraping off encrusting benthic **algae** and ingesting detritus as they move. Tiny allied cowries crawl across their favorite food, the **sea fan**, and ingest the individual polyps (Fig. 14-11b). Other surface grazers include **nudibranchs**, or sea slugs (Fig. 14-11c,d), that crawl across the substrate (including corals and other immobile animals that constitute or cover the hard surface of the substrate) to eat algae, detritus, or animals such as corals, **sponges**, **hydroids**, and tunicates.

Figure 14-11. Examples of surface grazers. (a) Long-spined sea urchins (*Diadema savignyi*, Papua New Guinea), browsing across the seafloor. (b) An allied, or egg, cowrie (*Primovula sp.*, Indonesia) on a gorgonian sea fan. Notice that the gorgonian's polyps have been eaten on the areas of the sea fan along which the cowrie has moved. (c) A flabellina nudibranch (*Flabellina rubrolineata*, Indonesia). These shell-less gastropod mollusks advertise their toxic nature with their flamboyant coloration and body shapes. (d) This spectacular nudibranch (*Pyllodesmium undulatum*, Papua New Guinea) has a transparent body, like a number of other nudibranch species, so its internal organs are visible.

Surface grazers do not necessarily live on the sediment surface. For example, some clams live buried in the sediment and extend long siphon tubes to the water above. They use the siphon tubes to select food particles and "vacuum" them off the sediment surface (**Fig. 14-12a**).

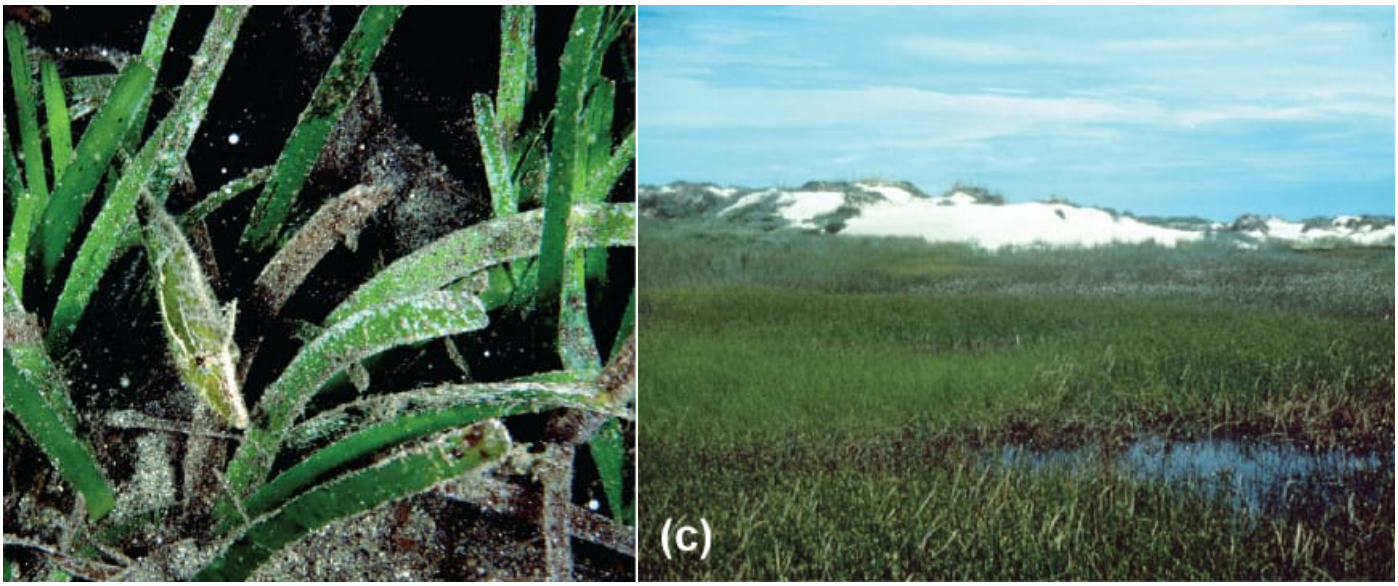


Figure 14-12. A benthic infaunal surface grazer, and seagrass communities. (a) The clam *Tellina sp*. feeds on the sediment surface. (b) Sea grass, especially turtle grass (*Thalassia sp.*), is a favorite food of the green sea turtle. Turtle grass beds are excellent places for animals such as this seagrass filefish (*Acreichthys tomentosus*, Papua New Guinea) to hide. (c) Rooted marsh grasses, primarily *Spartina sp.* (Aruba, Caribbean Sea), form dense beds in many coastal wetlands where the water is brackish. (d) Green sea turtle (*Chelonia mydas*, Hawaii).

In some shallow areas, large concentrations of macroalgae (**Fig. 13-4**) are extensively grazed by many animal species. In extremely shallow water, the seafloor may be covered with the **sea grass**, *Thalassia sp.*, (**Fig. 14-12b**) or with other rooted plants, such as the marsh grass *Spartina* (**Fig. 14-12c**). These plants are not heavily grazed, because they are difficult to digest. The plants must be broken down by bacteria and fungi to detritus before they become a desirable and widely used food source. Small

animals like the marsh periwinkle snail do not eat the tough grass itself, but instead graze on the nutritious film of diatoms, bacteria, and fungi that coat the surface of the plants. Some notable species, such as the manatee (**Fig. 12-26**) and the green sea turtle (**Fig. 14-12d**), are adapted to feed directly on sea grasses. *Thalassia*, the favorite food of green sea turtles, is commonly called "turtle grass."

Deposit Feeding

Organic detritus that falls to the sediment surface is used mostly by decomposers and surface-grazing benthic **fauna**, but some detritus is buried by **bioturbation** and by accumulating sediment, especially in areas of high **sedimentation rate** or **anoxic** bottom water.

Organic matter buried in sediments serves as food for decomposers and deposit-feeding animals (**Fig. 14-13**). As this food is depleted, the concentration of organic matter generally decreases with depth below the sediment surface, which is one reason that deposit feeders are concentrated in the upper layers of sediment. Some fish species that live in the water column hunt and feed on small animals, such as clams, that live below the surface of the sediments and are often considered to be a special category of deposit feeders.

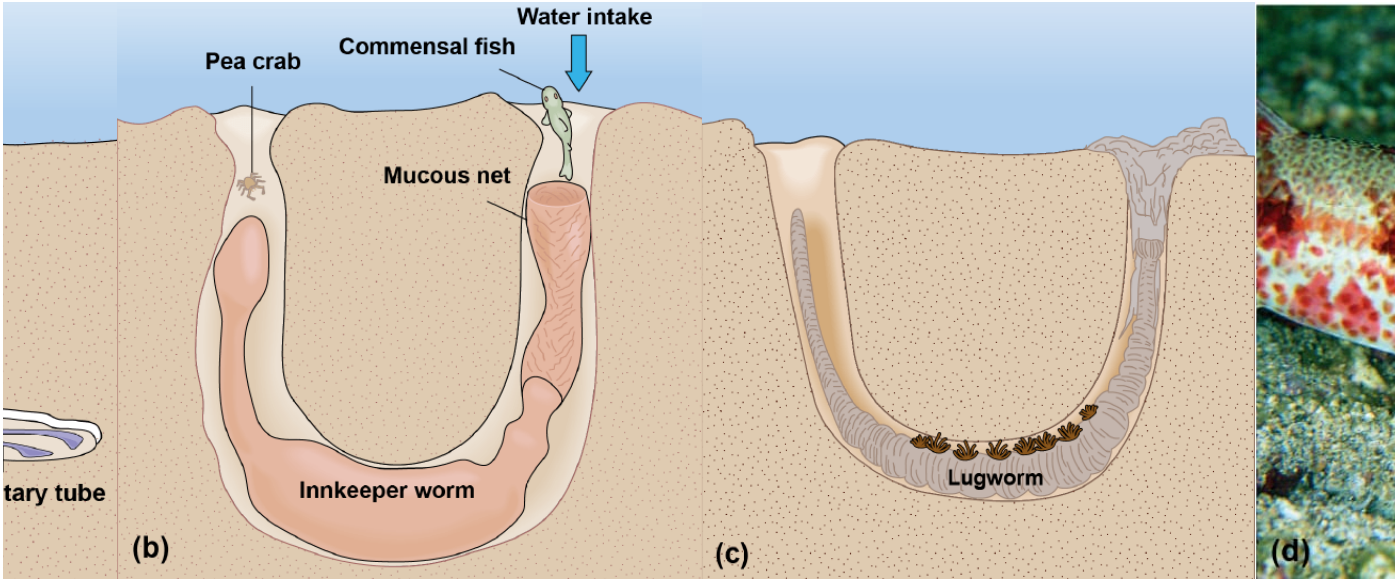


Figure 14-13. Examples of deposit feeders. (a) The heart urchin (*Echinocardium sp.*) feeds on detritus buried in the sediments and respires through long tubes extended to the surface of the sediment. (b) The innkeeper worm (*Urechis caupo*) builds a U-shaped tunnel through which it draws oxygen-containing water for respiration. The tunnel forms a perfect home for several associate species, sometimes including several gobies and two species of small crabs. Although it lives in the surrounding sediment, a small clam species also uses the burrow by extending its siphon tubes into the burrow to obtain oxygenated water. (c) Lugworms (*Arenicola brasiliensis*) use a similar U-shaped burrow. (d) Goatfishes such as this blackstriped goatfish (*Upeneus tragula*, Indonesia) use their barbells (the two yellow "whiskers" protruding from just under the mouth) to sense and smell out small invertebrates under the surface of the sediments. The goatfish digs into the sediments with the barbells to capture and eat these invertebrates, so it may be considered to be a deposit feeder, but it is also a hunter. Other fishes, especially some wrasses, often follow the goatfish to capture a free meal from among the invertebrates that the goatfish exposes as it digs.

Deposit-feeding infauna generally eat their way through sediments, digesting organic particles or the organic matter that coats some particles in much the same way that earthworms eat their way through soil. Much of the detrital organic matter that survives decomposition to be buried in sediment is refractory (difficult to decompose) and has low food value. Consequently, many deposit-feeding infauna obtain most of their food by consuming bacteria and fungi, many species of which live within the sediment. Other deposit-feeding infauna have evolved means of decomposing refractory detritus.

Because detritus particles are generally small and have lower density than inorganic sediment particles, detritus tends to settle and concentrate in low-energy areas where fine-grained sediments also accumulate. Consequently, fine mud deposits tend to have higher organic matter concentrations than coarser sediments have, and they are more favorable habitats for deposit feeders. However, to live and feed in sediment, deposit feeders must have a supply of oxygen for **respiration**. In muddy sediment where organic matter is abundant, respiration by animals and decomposers quickly uses up the oxygen dissolved in **pore waters**. Oxygen is not readily replenished from the water column above, because downward diffusion of oxygenated water through the

sediment is slow.

Many deposit feeders either are restricted to the near-surface oxygenated sediment layer or have evolved mechanisms to acquire oxygen from the water column above. For example, the heart urchin has adapted some of its tube feet to create a periscope-like breathing apparatus that it extends upward to the sediment surface (**Fig. 14-13a**). Deposit-feeding clams extend their siphons to the sediment surface for this same purpose. Many other infaunal deposit feeders form and live in a U-shaped or similar burrow and draw oxygenated water through the burrow as they feed. The innkeeper worm (**Fig. 14-13b**), lugworm (**Fig. 14-13c**), and many other species use variations of this method.

Muddy sediments tend to be dominated by **annelid** worm populations, whereas sandy substrates are more likely to be dominated by various species of deposit- or filter-feeding clams. Many deposit feeders pass sediment through the gut and discharge feces containing large volumes of processed sediment through the excurrent end of their burrow or breathing tube. In this way, food-depleted material is transported to the sediment surface, where it will not be reingested. Some species, particularly annelid worms, produce tightly packaged **fecal pellets** that accumulate on the sediment surface and protect it from **erosion**. Other species, especially certain clams, **excrete** their fecal material as a cloud of loose particles.

Clams may discharge their wastes as a cloud of particles for reasons other than to get rid of food-depleted sediment. This form of waste disposal may protect the clams' environment from invasion by other species. By working through the upper layer of sediment and continuously dispersing this material at the surface, clams prevent the silty sand in which they live from compacting and becoming firm enough for epifauna to attach to or lie on without being buried. In addition, the clouds of fine inorganic particles that the clams continuously create tend to clog the feeding apparatus of filter-feeding organisms with non-nutritious material.

This page titled 14.3: Feeding is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.4: Hunting and Defense

All marine species must ensure that enough individuals survive predation to produce the next generation. Carnivorous species must also adopt a successful hunting method that can counter the defensive methods of their prey and provide adequate food. Although there are only a few basic approaches to either offense or defense, combining these approaches makes possible an incredible variety of offensive and defensive strategies. The situation is analogous to the game of chess, in which a few simple permitted moves of the chess pieces can be combined into an almost infinite variety of offensive or defensive strategies. Each species has its own unique combination of defensive and offensive strategies. In a chess game, the most successful strategy is usually one in which individual moves are combined in such a way that each contributes to both defense and offense. Similarly, marine species often use the same approach for both purposes.

Basic offensive and defensive approaches used by marine organisms are summarized in **Table 14-1**. The following sections describe and illustrate how these approaches are used by various species, and how fish species have evolved to optimize variations of them. The unusual sensing mechanisms that some marine species use to locate prey are discussed in a subsequent section.

Table 14-1. *Some General Approaches to Hunting and Defense*

Behavior	Offensive Uses	Defensive Uses
Speed	Chase down prey	Escape predators
Lures	Attract prey	Confuse predators
Camouflage and mimicry	Ambush	Escape detection Confuse or deter predators
Concealment	Ambush	Escape detection Evade capture
Spines and armor	Overcome armor or other defenses	Deter predators
Poisons	Kill or disable prey	Reduce predator population Reduce competitor population Deter predators
Group cooperation	Overwhelm and confuse defenses	Confuse predators Minimize predation

Speed

The most familiar approach to hunting and defense in terrestrial ecosystems is speed. The predator chases the prey, and the prey tries to outrun the predator. For example, lions and tigers chase antelopes, but antelopes often escape by outrunning their

pursuers. Speed is used in much the same way in the marine environment.

Predators that hunt by using their swimming speed include a variety of sharks, bony fishes, **marine mammals**, and squid. The prey are generally other species of fishes, marine mammals, or squid. In some cases the hunt is similar to the lion–antelope chase, but it occurs in three dimensions. Prey outnumber the predator and live uneasily in the predator's presence, always carefully monitoring its movements and maintaining a respectful distance. Suddenly the predator selects a potential victim and begins the chase. Unwary prey or prey that are weak or injured fail to elude the predator and are consumed. Stronger individuals and those that are more successful in avoiding predators preferentially survive to reproduce and pass on their more successful genes to future generations. The predatory species is subject to a similar natural selection process because stronger and more skillful predators outcompete weaker and less successful members of their species for the available food resources.

Only the largest predators and prey can invest the considerable energy required to overcome water resistance and swim long distances at high speed. These species must obtain large quantities of food to replace energy reserves used in the chase. Consequently, most ocean predators use speed in ways that ensure a quick kill, and prey species seek ways to escape their predators quickly.

Predators may use a short burst of speed as the final component of a hunting strategy in which stealth or other approaches are used to initially get within striking distance of the prey. A variety of fish species, including lizardfishes (**Fig. 14-14a**), frogfishes (**Fig. 14-14b**), and hawkfishes (**Fig. 14-14c**), lie quietly in wait on the seafloor until their prey passes nearby. Then a quick burst of speed is sufficient to capture the prey. Lizardfishes rely on their prey's mistake in approaching within half a meter or so, a distance across which a lizardfish can make a very fast attack. Frogfishes are among the fastest-moving marine animals, but they can maintain their speed for only a few tens of centimeters and must be very close to their prey for a successful hunt. To ensure that its prey approaches close enough for such attacks, a frogfish uses two other approaches in its hunting strategy: camouflage and a lure, which are discussed below.

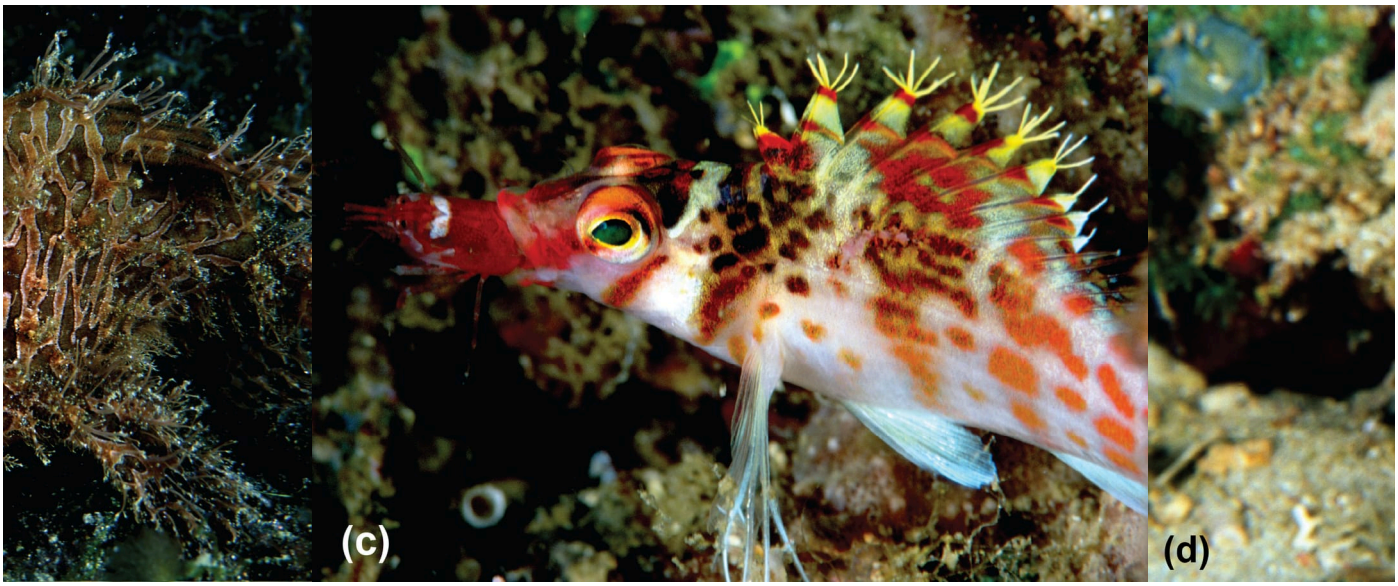


Figure 14-14. Some fishes that employ speed for offense or defense. (a) A twospot lizardfish (*Synodus binotatus*, Indonesia) lying in wait for passing prey. (b) A striated frogfish (*Antennarius striatus*, Indonesia). The frogfish wiggles the lure, called an "esca," on the end of its illicium, a modified first ray of the dorsal fin located just above the eye. When wiggled, the esca looks like a small polychaete worm to the frogfish's unsuspecting prey. (c) A Falco, or dwarf, hawkfish (*Cirrhitichthys falco*, Indonesia) that has just been rewarded for its time spent lying in wait by capturing a favorite food item, a shrimp. (d) A fire goby or dartfish (*Nemateleotris magnifica*, Papua New Guinea) hovering above its hole in the reef, into which it is ready to dart and take refuge. (e) Bennett's butterflyfish (*Chaetodon bennetti*, Papua New Guinea). Note the large false eyespot near its tail.

Speed is also used in a variety of ways as a component of defensive strategies. **Gobies** use short bursts of speed to retreat to their home, which is generally a hole in the sand or reef where they are safe from predators (**Fig. 14-14d**). Some invertebrates, such as the fan worm (**Fig. 14-10c**), Christmas tree worm (**Fig. 14-10d**), and some sand anemones (**Fig. 14-6c**), use a high-speed, almost instantaneous, retreat into a protected tube or burrow to avoid predation. Many fish species change direction quickly to evade an onrushing predator. For example, butterflyfishes (**Fig. 14-14e**) can evade predators by swimming quickly in and out of the tortuous nooks and crannies of a reef or by simply making tight turns.

Lures

Lures are used by ocean predators to attract fishes and other prey just as they are used in sportfishing. The lure is made to look like a tasty morsel of food to attract the prey species to the concealed predator (or fishhook). The best-known practitioners of this technique are frogfishes (**Fig. 14-14b**). Lying still and camouflaged, a frogfish wiggles a fleshy knob on the end of a stalk-like projection extending from above the mouth. When another fish approaches, this lure appears to be a small fish, shrimp, or other food morsel. By the time an unwary approaching fish is close enough to realize its mistake, it is within striking distance of the frogfish's lightning-quick, short-range attack. Because the predator uses little energy while lying in wait, it requires successful hunts only infrequently. In addition, the lure can attract more prey than would normally swim by. The deep-sea anglerfish (**Fig. 14-15**) uses the lure-based ambush strategy to perfection. Its **bioluminescent** lure appears as a tantalizing tiny point of light to prey that live in perpetual dark, although many potential prey species in this zone are sightless.

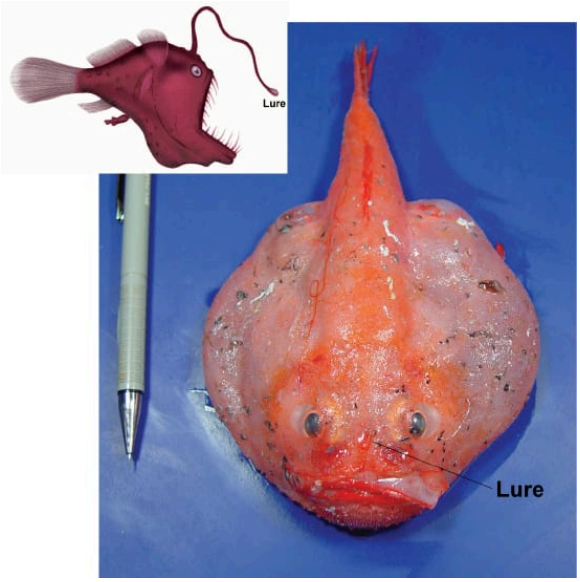


Figure 14-15. This deep-sea anglerfish (*Chaunax pictus*) is bloated from the pressure reduction as it was brought to the surface. Look carefully and you can see the angler's bioluminescent lure, a small black dot on the end of a stalk just between its eyes. The stalk is now lying limp on the fish's body. The inset shows how the lure is normally deployed.

Ironically, lures can also be used for defensive purposes. They are used by many reef fish species, especially butterflyfishes, which have a dark spot near the tail (**Fig. 14-14e**) that resembles an eye. The real eye is often camouflaged in a vertical black line. When a predator makes its initial move toward a butterflyfish, it aims its attack toward the front of the fish to block its expected escape route. If the predator is lured into orienting its attack to the false eye location, the butterflyfish gains valuable moments in which to swim in a direction other than that expected by the predator. The false eye may also be larger than the real eye to make the prey seem much bigger and thus more capable of retaliatory defense. You may have seen some of the many butterflies and moths that have false eyes for the same reason.

Camouflage and Mimicry

Camouflage (the art of making an object difficult or impossible to see against its background) and mimicry (the art of making an individual look like a completely different species or object) are practiced by many marine species. The principal use of camouflage is probably to enable an organism to escape detection by its predators, but it is also used by predators to enable them to get close to their prey without detection. In many cases, camouflage is extremely effective (**Fig. 14-16**). Countless times when my diving partner has pointed out an organism on a reef, I have needed many seconds of careful visual examination before I could suddenly see through the effective camouflage to discern the organism.

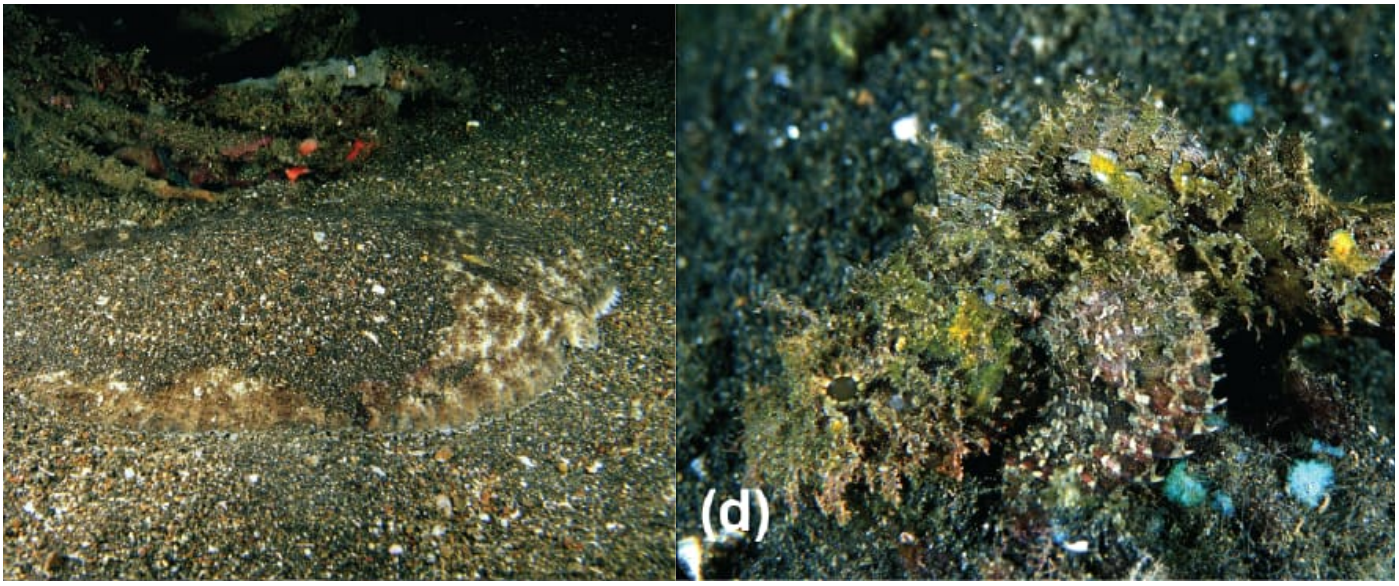


Figure 14-16. Examples of camouflage. (a,b) Crinoid shrimp (*Periclimenes amboinensis*, Papua New Guinea and Indonesia) on their host crinoid. These two photographs show how the shrimp are camouflaged to hide in crinoids of different colors. (c) A spotted sole partially buried in the sand (*Phyllichths punctatus*, Indonesia). (d) This raggy scorpionfish (*Scorpaenopsis venosa*, Indonesia) is well camouflaged against the sand and rubble seafloor, but it even enhances its natural camouflage by allowing algae to grow on its body. (e) The algae ghost pipefish (*Solenostomus halimeda*, Papua New Guinea) is well camouflaged to look just like a frond of this green calcareous alga. The ghost pipefish even spends most of its time swimming slowly in a vertical position to look even more like the alga. The pipefish is on the left-hand side of the clump of algae in this photograph. (f) There are many species of decorator crabs, each of which covers itself in sponges and other epifauna and algae. This one is a decorator spider crab (*Camposa retusa*, Papua New Guinea).

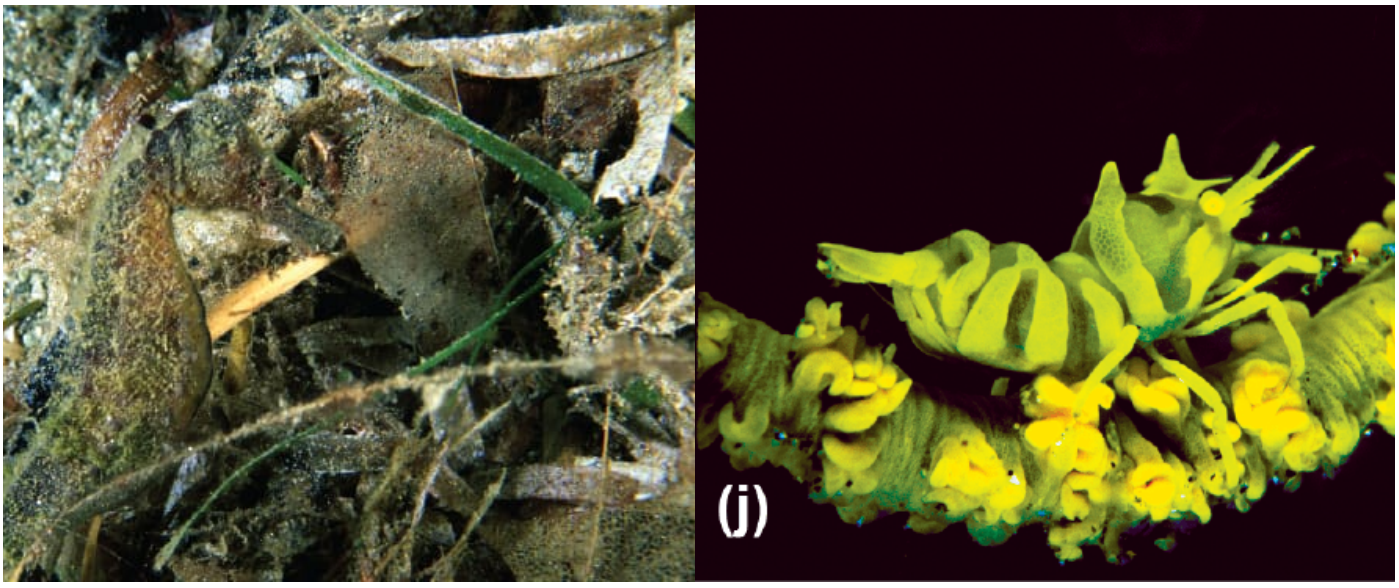


Figure 14-16. Examples of camouflage. (g) A commensal spider crab (*Xenocarcinus conicus*, Palau) on a gorgonian coral. The crab is about 1 cm in length. (h) A pygmy (or gorgonian) seahorse (*Hippocampus bargibanti*, Indonesia) blends in with its host sea fan with coloration and body "bumps" that resemble the polyps of the fan when they are closed. (i) The spotted seahorse (*Hippocampus kuda*) is found in many colors, often to blend in with its background. This individual in Papua New Guinea, with its color and coating of algae and organic detritus, blends perfectly with the dead leaves and detritus of its home near the mouth of a stream. (j) The coral shrimp (*Dasycaris zanzibarica*, Indonesia) lives on a whip coral and is camouflaged by its body protuberances and perfect color to match the whip. (k) Like the coral shrimp, this tiny spider crab (*Xenocarcinus tuberculatus*, Indonesia) is camouflaged on its whip coral. (l) A tiny (3 to 5 mm long) rounded sea star shrimp (*Zenopontonia noverca*, Indonesia) is almost transparent, but it has flecks of color to match its sea star host. It can be seen in this photograph only because the lighting was just right to leave it a little shadowed.

Marine animals take many different approaches to camouflage, but the predominant basic approach is to make texture, color, and pattern appear the same as those of the background. Some animals, such as octopuses and cuttlefish, can change their colors almost instantly to blend into their background. These species can also change their body texture to blend into a reef.

Octopuses, cuttlefish, and many other species change color by using muscles to expand or contract pigment filled cells in their outer body surface called “chromatophores” under direct control of their nervous system. However, this is not the only way to match the color of a background. For example, some species of frogfishes (**Fig. 14-14b**) have a variety of colors and color patterns, and each individual finds itself a permanent or semipermanent home where it is surrounded by similarly colored sponges or sponge-encrusted rocks. The frogfish’s body is lumpy and irregular, so its shape and texture do not instantly reveal its presence. In addition, the frogfish has small eyes surrounded by confusing decoration, again for camouflage. Scorpionfishes and many seahorses (**Fig. 14-16d,h,i**) use a very similar camouflage technique: a mixture of drab colors that blend into their background on a reef or rubble-covered bottom, and frilly appendages that disguise their body outline.

The differences in frogfish and scorpionfish camouflage reflect their different preferred habitats and habits. Frogfishes live in or on a reef and rarely move from one preferred location. Scorpionfishes live primarily on a coral rubble seafloor and move periodically from place to place to hunt. Scorpionfishes therefore need a more generalized camouflage than frogfishes do, to blend with the variety of backgrounds they encounter. Like frogfishes and scorpionfishes, the ghost pipefish (**Fig. 14-16e**) is a master of both color and body form camouflage. Some species live among algae or turtle grass (*Thalassia*) and even swim in a vertical orientation to parallel the algae or turtle grass blades.

Tiny shrimp, crabs, and other invertebrates that live on **crinoids** and soft corals make very effective use of camouflage (**Fig. 14-16a,b,g,j,k,l**). Many of these tiny and beautiful invertebrates spend their entire life cycle on a single host crinoid or soft coral that they perfectly mimic. These hosts have a wide variety of colors and color patterns, and the guest species often match the pattern. Many of the tiny invertebrate species that inhabit crinoids and soft corals have not been well studied, so we do not know how each of these species achieves its perfect camouflage. Some species of shrimp will change color within several days to match a new host perfectly when they are moved from one crinoid or soft coral to a differently colored individual or colony of the same species. The shrimp likely achieves this color change by using its chromatophores, or by eating small parts of its host and incorporating the host’s colored pigments into its own body, or possibly by a combination of these techniques.

Sometimes even the perfect color-matching capabilities of crinoid and soft coral inhabitants appear to provide insufficient camouflage. Many invertebrates, particularly several species of tiny crabs, also have modified body shapes to match the structure of their host. In some cases, the crab’s or other lodger’s body is covered with spines and protuberances that mimic the appearance of the host or its individual polyps. In other cases, the lodger plucks off some of its host’s polyps and places them on its shell or appendages (**Fig. 14-16g**). The technique of placing other organisms on one’s own body for camouflage is practiced by many invertebrate species. For example, the decorator crab (**Fig. 14-16f**) covers its entire body with algae, sponges, and other invertebrates, beneath which the crab is not easily seen, particularly if it stays motionless when a predator nears. The polyps, sponges, and other species used for camouflage can live and reproduce while attached to the camouflaged crab. As will be discussed later in the chapter, the association may even be beneficial to both species.

Pelagic fishes and marine mammals that swim in the water column away from the seafloor take advantage of somewhat different camouflage. Many of these species have countershading that minimizes the contrast of their bodies against the background as viewed by predators from above or below. Species that rely on counter-shade camouflage have a white or silver and highly reflective underside (**Fig. 14-17a,b**) that efficiently reflects the ambient light downward and reduces the contrast between the fish’s (or mammal’s) body and the water surface as seen by a predator from below. The upper side of the countershaded species is dull, nonreflective, and often mottled gray or blue-gray to reduce upward reflection of light and soften the outline by providing variations in the light reflected. A predator looking downward must distinguish the dull and confusing shape of the countershaded prey seen against the murky, confused background of the seafloor or deep water below.

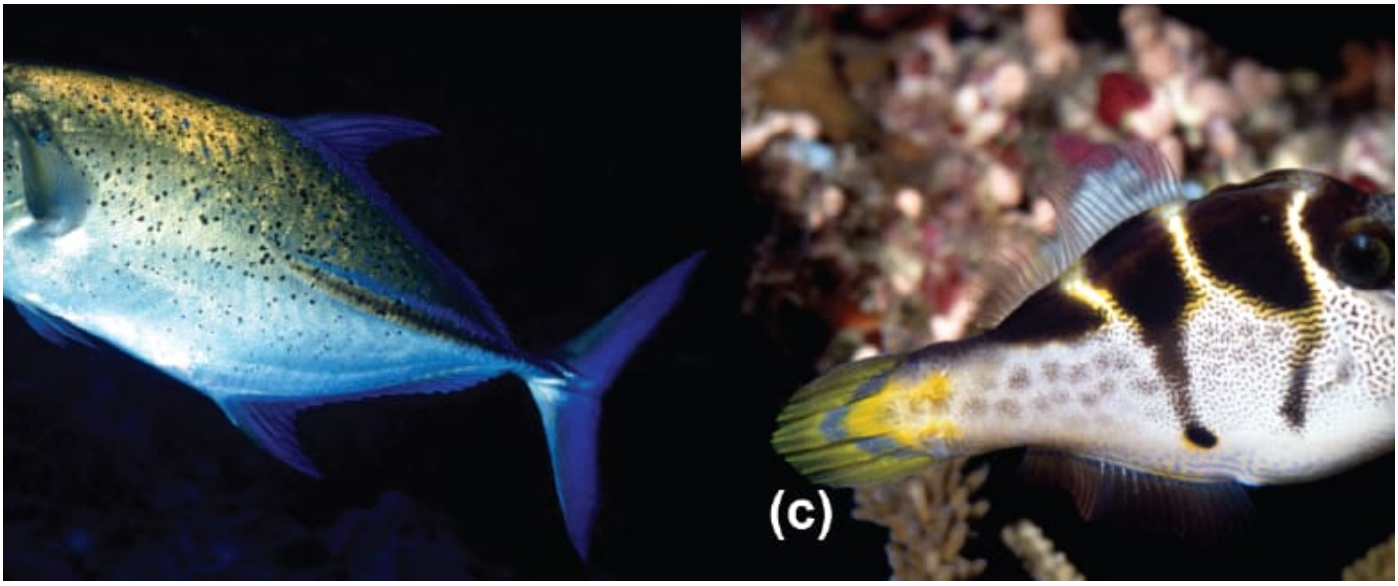


Figure 14-17. Examples of countershading and mimicry. (a) This trevally, called a "rainbow runner" (*Elagatis bipinnulata*, Solomon Islands), lives mostly in the open water column. Note the light-colored underside and nonreflective upper surface that provide countershading. (b) Another trevally species, the bluefin trevally (*Caranx melampygus*, Solomon Islands), is also countershaded with a light underside and darker upper surface, but this species spends much of its life close to the reef and seafloor. Perhaps the spotting on the trevally's upper side helps it to visually blend in better with the seafloor when seen from above. (c) Blacksaddle mimic filefish or leatherjacket (*Paraluteres prionurus*, Philippines). (d) Black-saddled toby (*Canthigaster valentini*, Papua New Guinea). Can you tell the difference between this puffer (toby) and the leatherjacket in part (c)? The puffer lacks a first dorsal fin, and it has shorter dorsal and anal fins than the leatherjacket has.

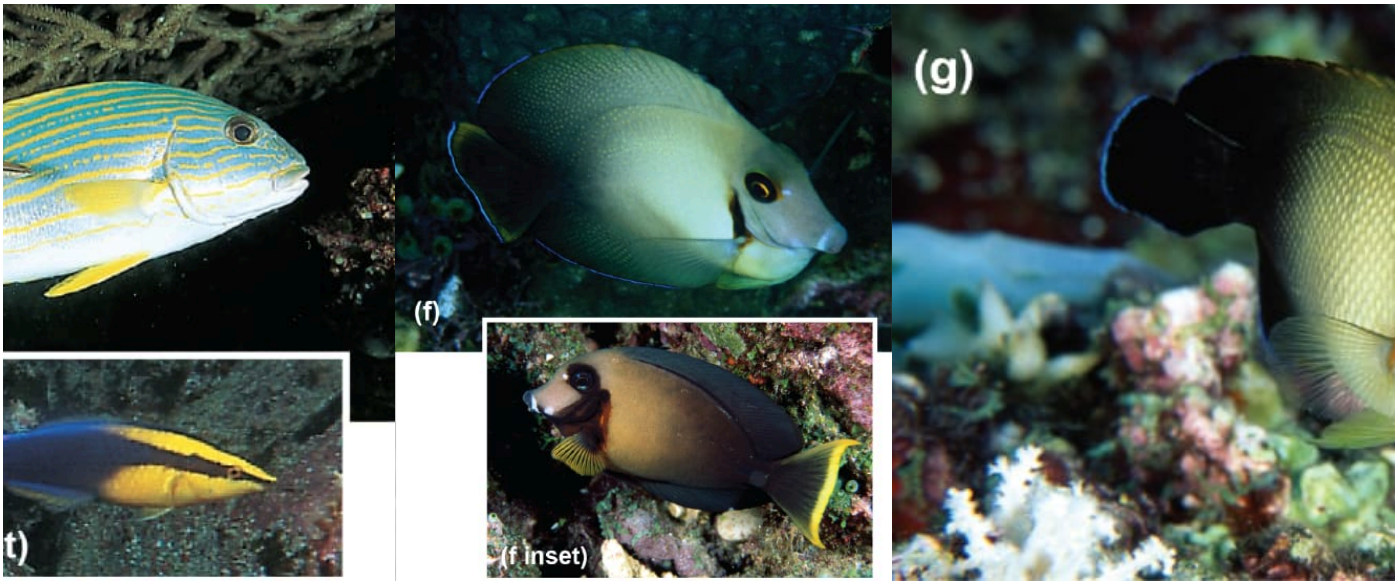


Figure 14-17. Examples of countershading and mimicry. (e) A cleaner wrasse, similar to *Labroides pcthirophagus* (Hawaii), shown in the inset, is removing parasites from a Celebes sweetlips (*Plectorhinchus celebicus*, Papua New Guinea).(f) A juvenile mimic surgeonfish (*Acanthurus pyroferus*, Indonesia). The inset shows an adult (Papua New Guinea). (g) A pearlscale angelfish (*Centropyge vrolikii*, Papua New Guinea).

Some species conceal their identity by mimicking another species. Again, this approach can be used in either hunting or defense. The false cleaner wrasse is a small carnivorous fish that mimics a species of cleaner wrasse (**Fig. 14-17e**). Cleaner wrasses establish stations where they wait to eat **parasites** off larger fishes, and they advertise their services by their bright colors and a jerky dance that is similar for all cleaner species. The cleaner benefits from the food supply of parasites, and the cleaned fish benefits by being rid of the parasites. The false cleaner mimics not only the cleaner's coloration and body shape, but also its advertising dance. The unsuspecting parasitized fish moves in and lowers its defensive guard to the false cleaner, which simply takes a bite out of the fish. Often the fish is startled but stays and allows the false cleaner a second bite before realizing its mistake and scurrying away in complete confusion.

The black-saddled mimic filefish (also known as the mimic leatherjacket) uses mimicry for defensive purposes. The filefish (**Fig. 14-17c**) is virtually indistinguishable in size, shape, and markings from a pufferfish, the black-saddled toby (**Fig. 14-17d**). Because the puffer is poisonous and predators are unable to distinguish the two species, they ignore both the puffer and its mimic. The juvenile mimic surgeonfish (**Fig. 14-17f**) mimics both the looks and the behavior of the pearlscale angelfish (**Fig. 14-17g**), and it is about the same size as the angelfish, but the larger adult mimic surgeonfish (**Fig. 14-17f**) looks completely different from the juvenile. Angelfishes have sharp spines on their cheeks to deter predators, so the predators do not attack the juvenile mimic surgeonfish even though surgeonfishes do not have such a spine.

Concealment

Although many marine predators use senses other than vision to locate their prey, most marine predators in the photic zone hunt primarily by sight. Consequently, concealment is an excellent technique that is widely used for defensive purposes and also by certain predators to ambush their prey. Camouflage and concealment are often used together to prevent detection. For example, flounders and soles (**Fig. 14-16c**) are not only camouflaged, but also may partially bury themselves in sand to conceal their presence further.

Concealment is a very effective strategy for many invertebrates and fishes that bury themselves in sand or mud; hide in cracks, crevices, or caves of a reef or rubble-covered seafloor; or build and live in holes in a reef, sediments, or rocks. Most species that use concealment for defense against visual hunters remain concealed by day and come out to feed at night. Other species emerge during the day but seldom stray far from their hiding places. These species use concealment places in two ways: to hide in to escape detection when predators approach and as a refuge into which the usually much larger would-be predator cannot follow. For example, fire gobies (**Fig. 14-14d**) live in holes excavated in coarse sand or rubble areas around coral reefs. Some will dive into these holes, of which they may have several, when a predator is in sight. Others will stay out until the predator reaches a distance of a meter or two and then dive into their holes at lightning speed. Photographing these gobies, and many other reef fishes that have similar habits, requires the patience to lie quietly in wait until the fish re-emerges.

Nocturnal species that conceal themselves by day and hunt at night are often a deep red color and have big eyes (**Fig. 14-18a,b,e**) that enable them to hunt when light levels are very low. The red color aids concealment in caves or at night because red light is strongly absorbed by seawater. Consequently, these organisms appear dark gray or black unless seen near the surface in daylight or when illuminated with artificial light. In addition, many marine species have eyes that are not sensitive to red **wavelengths**.

Figure 14-18. Some nocturnal species. (a) A nocturnal shrimp (Fiji). Note the large eyes. (b) This fish, called a crescenttailed bigeye (*Priacanthus hamrur*, Fiji), spends the daytime mostly in dark caves, emerging at night to feed. Note the large eyes and predominantly dark red coloration. (c) Most crabs are nocturnal, including this coral crab (probably *Cancer sp.*, Fiji). (d) Fire urchin (*Asthenosoma intermedium*, Papua New Guinea). This species is normally seen only at night. (e) Many octopuses, such as this one (*Octopus luteus*, Indonesia), hunt mostly at night. Note again the red color.

A wide variety of fishes, shrimp, crabs, sea urchins, octopuses, and many other invertebrate species conceal themselves by day and emerge at night (**Fig. 14-18**). In contrast, many species that hunt or feed during the day conceal themselves at night (**Fig.**

14-19a). Diving at night on a reef or rocky area reveals a community very different from the one seen during the day. Many invertebrates, including certain sea pens (Fig. 14-6a,b), shrimp, crabs, and mollusks (Fig. 14-19b), emerge only at night and bury themselves in the soft seafloor by day. Nocturnal predators use the cover of night as a form of concealment to ambush their prey.

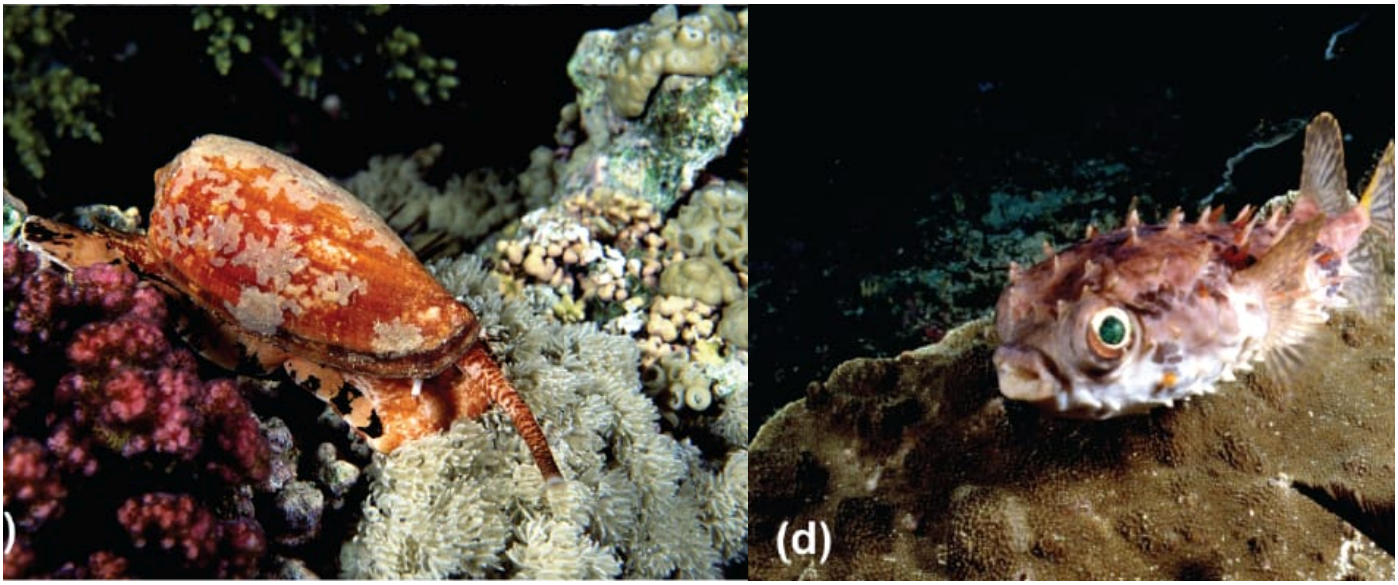


Figure 14-19. Examples of nocturnal defense and predators, and of spines and armor in fishes. (a) This black-headed parrotfish (*Scarus gibbus*, Fiji) is sleeping at night. Like most parrotfish species, it excretes a transparent mucous cocoon (visible in this photograph primarily because of the particles deposited on its surface) in order that potential predators cannot chemically sense, or "smell," it. (b) A mud snail (*Nassarius papillosus*, Hawaii) that has emerged from the sand to hunt at night. (c) This cone shell (*Conus geographicus*, Red Sea) is hunting at night across the reef. (d) This orbicular burrfish (*Cyclichthys orbicularis*, Philippines) has already partially inflated to respond to a threat from a predator. (e) When fully inflated, this same burrfish is too large for most predators to grasp. It would even be difficult to take a bite out of, because of its round shape and sharp spines. (f) Boxfishes, or trunkfishes, such as this male spotted boxfish (*Ostracion meleagris*, Papua New Guinea), have a hard, box-shaped exterior to make it difficult for predators to grasp or bite.

Concealment would seem to be impossible for species that live in the water column. Nevertheless, squid and octopuses are able to conceal themselves by speeding off behind an opaque cloud of ink that they release when threatened.

In the open ocean, where there are no reefs or backgrounds to blend into, some pelagic animals use transparency as a means of concealment. By having bodies composed largely of water, animals like jellies, comb jellies (ctenophores) and salps allow light to pass directly through them so they are nearly invisible to predators. Other organisms, like glass squid and certain larval fishes, have transparent tissues and organs, leaving only their eyes visible. This strategy makes it almost impossible to detect in the sunlit upper layers of the ocean.

Spines and Armor

Numerous invertebrates have evolved thickened shells or long spines to prevent predators from reaching their vital soft parts. Many mollusks have external shells that protect them after they retreat inside. Bivalve mollusks, including clams and scallops, have two hinged valves that they can close tightly together. Gastropod mollusks have a single shell (Fig. 14-19c) into which the animal's soft parts can be withdrawn. The entrance to the shell is closed off with a tough door called an "operculum." Some fishes, such as puffers and burrfishes (Fig. 14-19d,e), have developed special jaws to crush mollusk shells. Crabs have a large armored crushing claw for the same purpose. In turn, many mollusks have developed extremely thick and strong shells or armored spines. Spines increase the size of the shell and make it more difficult for a crab's claw or fish's jaw to grasp and crush it. Many mollusks conceal themselves during the day and hunt only at night to avoid predators, but some predators, particularly crabs, are also nocturnal. There is a perpetual contest between predator and prey as each evolves better adaptations for survival.

Many sea urchins (Fig. 14-18d) have evolved long sharp spines that cover their entire upper body. The spines are often tipped with reverse barbs or venom injectors. A direct approach by a predator to one of these sea urchins is likely to result in severe stab wounds or even paralysis and death. However, some triggerfishes will methodically pick off one spine at a time until they can attack the underlying external shell of its, then helpless, inhabitant. Other triggerfishes use a jet of water from the mouth to blow the sea urchin over, exposing the vulnerable underside that has no spines—an elegant means of circumventing the

urchin's defenses. In addition to their formidable spines and, in some species, highly toxic venom, many sea urchins have become nocturnal to avoid daytime predators.

Crustaceans, including shrimp, crabs, and lobsters, have developed hard external shells that they must periodically shed and regrow as the soft animal within becomes larger. Worms, such as the fan worm and Christmas tree worm (**Fig. 14-10c,d**), use rocks, coral, or sand to build armored burrows into which they can withdraw. Sea cucumbers do not have hard shell armor, but they have developed thick leathery outer skins containing calcium carbonate that, although pliable, are almost impossible to bite through.

Certain fish species have also developed armor and spines. Boxfishes (**Fig. 14-19f**) have a hard, boxlike body that allows only slow and awkward swimming but makes them large in cross section to prevent a predator's jaw from grasping and crushing them. The armor also prevents the predator from taking small bites off a boxfish's body. When attacked, burrfishes (**Fig. 14-19d,e**) gulp large amounts of water and swell up, extending spines that normally lie flat on their body. Burrfishes reportedly may even inflate when already caught in a predator's jaws and simply wait for the predator to weaken or die from hunger before deflating and releasing themselves.

Toxins

A variety of invertebrate species, many marine algae, and certain fishes produce or concentrate from their food a wide spectrum of toxic substances. These substances are sometimes used to kill parasites and larval stages of other organisms that may settle on the toxin-producing species, but their main use is to discourage predators. Most of the myriad toxic substances synthesized by marine species have not been identified, and many thousands of these compounds are likely to be found. Because most toxic substances discourage predators, including bacteria and **viruses**, by interfering with their biochemistry, these substances hold the potential to be used as, or to lead to the development of, drugs that can treat human illness. Several drugs derived from marine toxins are already in routine use or are undergoing clinical trials. For example, a powerful painkiller derived from the venom of cone snails (**Fig. 14-19c**) that blocks nerve channels has been found to be more potent than morphine without being addictive.

Many species that produce toxins to make themselves inedible, notably nudibranchs (**Fig. 14-11c,d**), are brightly colored. The conspicuous color schemes of nudibranchs and similarly toxic tunicates, sponges, and other invertebrates warn predators of the poison defense. In response, many predators have evolved mechanisms to detoxify the poison produced by their prey species.

Numerous species inject toxins into the predator to repel its attack. Sea urchins inject such venom with the tips of their spines. Stonefishes and scorpionfishes (**Fig. 14-16d**) inject deadly venom with sharp spines along their dorsal (upper) fin. Anemones, corals, hydroids, and other invertebrates all inject venom by firing little poisonous darts called "nematocysts" from their tentacles. Some nudibranchs even steal these defenses by consuming anemones or hydroids and assimilating the undischarged nematocysts into their own body tissues to fire at their own predators.

Toxins are used for hunting as well as defense. For example, anemones (**Fig. 14-20a**) use toxin-laden nematocysts to attack and stun or immobilize invertebrates and small fishes that blunder into their trap. Once stunned, the prey is trapped by mucus-laden anemone tentacles that fold over and pass the prey toward the mouth or open stomach at the anemone's center. Anemonefishes (often called "clownfishes" or "clown anemonefishes") live in or near the anemone (**Fig. 14-20b-f**), retreating into its protective toxic folds and tentacles when threatened. The anemonefishes avoid being stung by developing a specialized mucus coating that prevents the anemone's stinging cells from firing, effectively camouflaging the fish as part of the anemone's own body, essentially a way of gaining immunity.

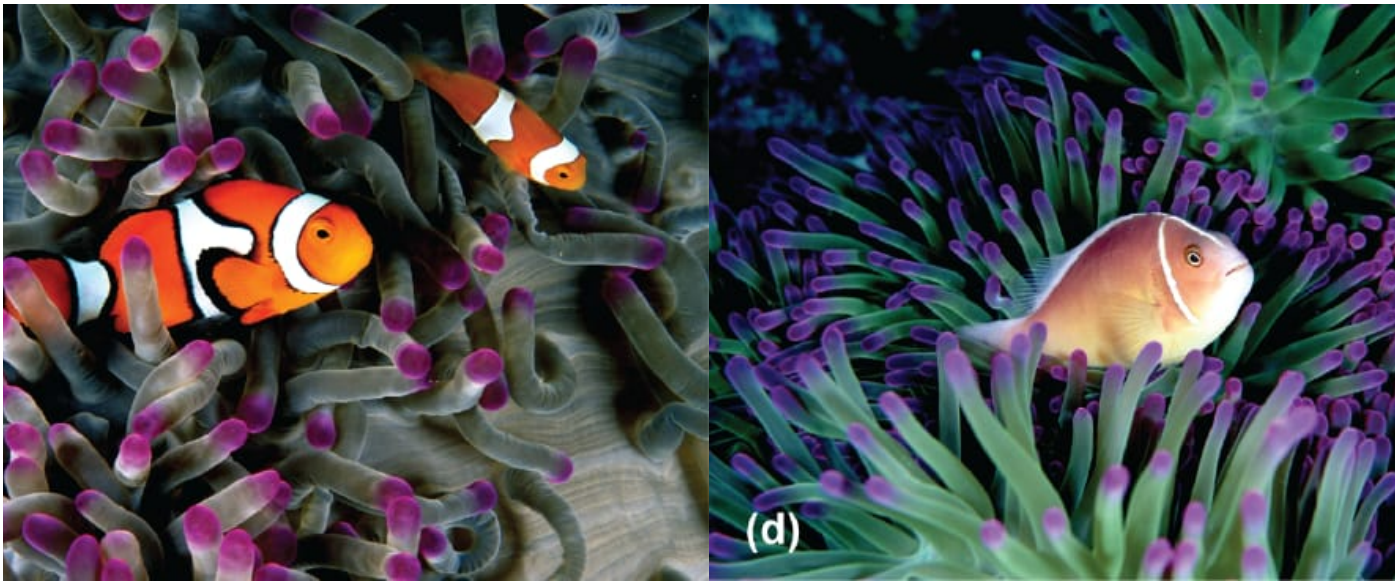


Figure 14-20. Anemonefishes, also called "clownfishes" or "clown anemonefishes," live in family groups in association with a number of different species of large carnivorous anemones, such as the Haddon's sea anemone (*Stichodactyla haddoni*, Indonesia) shown in (a). There are a number of anemonefish species, including the examples shown here. (b) Tomato anemonefish (*Amphiprion frenatus*, Papua New Guinea). (c) Clown anemonefish (*Amphiprion percula*, Papua New Guinea). (d) Pink anemonefish (*Amphiprion perideraion*, Papua New Guinea). (e) Spinecheek anemonefish (*Premnas biaculeatus*, Papua New Guinea). (f) Clark's anemonefish (*Amphiprion clarkii*, Papua New Guinea). All anemonefish species are damselfishes of the subfamily Amphiprioninae, except the spinecheek, which belongs to another subfamily.

Other notable species that use toxins to immobilize their prey include cone shells (**Fig. 14-19c**), which inject their venom through a long, thin tube extended from their body, and blue-ringed octopuses (**Fig. 12-1**), which transmit their toxin in mucus as they bite their prey. A number of human deaths have been caused by cone shell toxin injected when people inadvertently stepped on a cone shell living just buried in the sand in shallow water near the beach. One species of blue-ringed octopus, which lives in tide pools and on shallow reefs around Australia, is not aggressive but will bite if threatened or stepped on. The venom injected by this species is especially toxic, and there is no known antidote. A bite often results in death.

Numerous species excrete poisons into the surrounding water to discourage competitors or potential predators. On the West Coast of the U.S., certain species of the diatom *Pseudo-nitzschia* produce a potent neurotoxin called domoic acid during large algal events. While these diatoms are the source, the toxin accumulates in the tissues of filter-feeders like sardines and anchovies. When larger marine mammals, such as California sea lions and whales, consume these fish, they ingest concentrated doses of the toxin, which causes seizures, brain damage, and often death. Domoic acid is also the cause of amnesic shellfish poisoning which can cause persistent neurological symptoms in humans who consume filter feeding shellfish.

Group Cooperation

Both hunting and defense can be aided by cooperative approaches involving several or many individuals of the same species and sometimes between members of two different species. The Portuguese man-of-war is a good example (**Chap. 12**). Group cooperation is a common adaptation, but two special categories are of interest: colonial forms and **schooling** in pelagic animals.

Colonial species abound in the oceans and include many sponges and **cnidarians**. In some colonial forms, each individual feeds and reproduces separately, but the individuals cooperate to enhance each other's and the species' success. For example, individual polyps of sea fans and corals, by growing attached to each other, extend the colonies' reach into the water column. This extension gives each polyp a better chance of encountering food and reduces competition by other species that live on sediment or rock surfaces.

Pelagic species, including many fishes, squid, and marine mammals, congregate in schools. Similarly, phytoplankton and zooplankton often have patchy (clustered) distributions. Schooling and clustering behavior affords advantages. For example, a group of fish can often overcome the defenses of a single individual of an otherwise superior species. Damselfishes lay their eggs in clusters on a reef surface and then defend them aggressively. Wrasses and other fishes arrive in groups to feed briefly but voraciously on the eggs as the damselfishes frantically try to chase them off but fail because there are too many attackers.

Fish schools have no leader, but each fish precisely matches the seemingly random twists and turns of the school, maintaining a precise distance from each of its neighbors. Movements of the school may confuse predators by making the school appear to be a single large organism. In addition, predators may have difficulty singling out a victim from the moving school. Schooling may have benefits in reproduction as well. Finding a mate is easier in a school, and a high rate of fertilization is ensured if eggs and sperm are released to the water column simultaneously by many members of the school. Mass **spawning** can in fact be considered a form of “schooling” of the fertilized eggs and larvae.

When prey is concentrated in schools, the predator spends much of its time searching for the school. Once it encounters a school, it cannot consume the entire school, so most individuals will survive, even if the predator gorges itself. Predators are able to consume less food through periodic gorging than they could by continuous steady feeding. Schooling of the prey and mass spawning thus reduce the efficiency of predator feeding, and consequently more of the prey species survive. This outcome may be the principal advantage of schooling.

This page titled 14.4: Hunting and Defense is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.5: Selected Adaptations in Fishes

Each marine species is adapted in different ways that enable it to survive. Fishes are adapted in many different ways to respond to such challenges as swimming in a manner that best supports their hunting and defensive strategies, the high **osmotic pressure** of seawater, the variability of osmotic pressure in estuarine waters, and control of their buoyancy.

Swimming Adaptations

Water is much denser than air, so it is much more difficult to travel through water than through air. The next time you go swimming, try running through knee-deep water and you will understand just how much more difficult it is to move through water. Fishes must overcome water resistance to swim. Therefore, the body shape of a fish must be optimized to facilitate its specific swimming habits.

While swimming, fishes must overcome three types of resistance or drag: surface drag, form drag, and turbulent drag. Surface drag is the **friction** between a fish's body surface and the water. Form drag comes about because water must be pushed out of the way, and turbulent drag is related to the smoothness of water flow past the swimming object.

Surface drag increases as the surface area in contact with the water increases. Consequently, surface drag is minimized if the swimming object is spherical because a sphere has the smallest surface area per unit volume of any solid object. Form drag increases in proportion to cross-sectional area (**Fig. 14-21**). The perfect shape to minimize form drag is needlelike, but this shape has a high ratio of surface area to volume and is subject to increased surface drag. A few fish species are needle-shaped (**Fig. 14-22a**), and others have an extremely thin platelike body (**Fig. 14-22b,c**) to minimize form drag. Most fish shapes are a compromise: rounded or oval in cross section, but elongated in the swimming direction. This shape tends to minimize total (form plus surface) drag (**Fig. 14-21**).

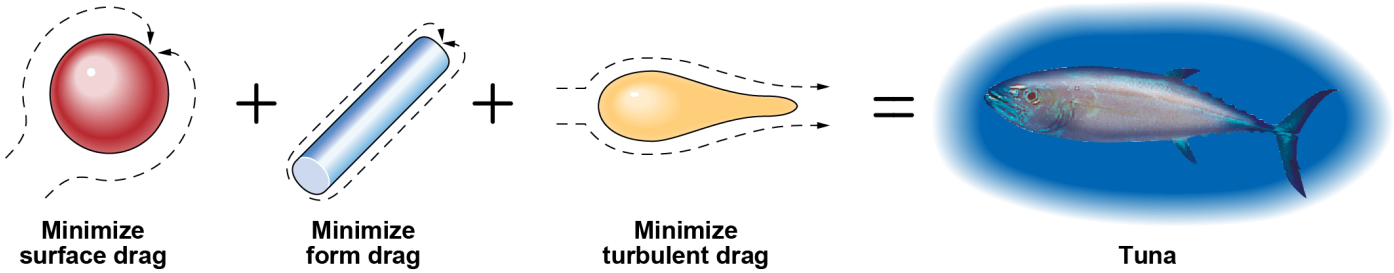


Figure 14-21. Three types of drag slow an object as it moves through water: surface drag, form drag, and turbulence drag. For each type, the magnitude of the drag varies with the shape of the moving object. This figure shows the optimum shape required to minimize each type of drag. Body shapes of most fish species are a compromise between the need to minimize drag and other needs of the animal, such as acceleration and turning speed. Fishes that swim constantly at high speed, like the tuna in this figure (*Thunnus sp.*, Fiji), have a body shape that provides the best compromise for reducing the sum of the three types of drag: a rounded body that is narrow at the tail and thicker near the front. Nuclear submarines also are built in this general shape.

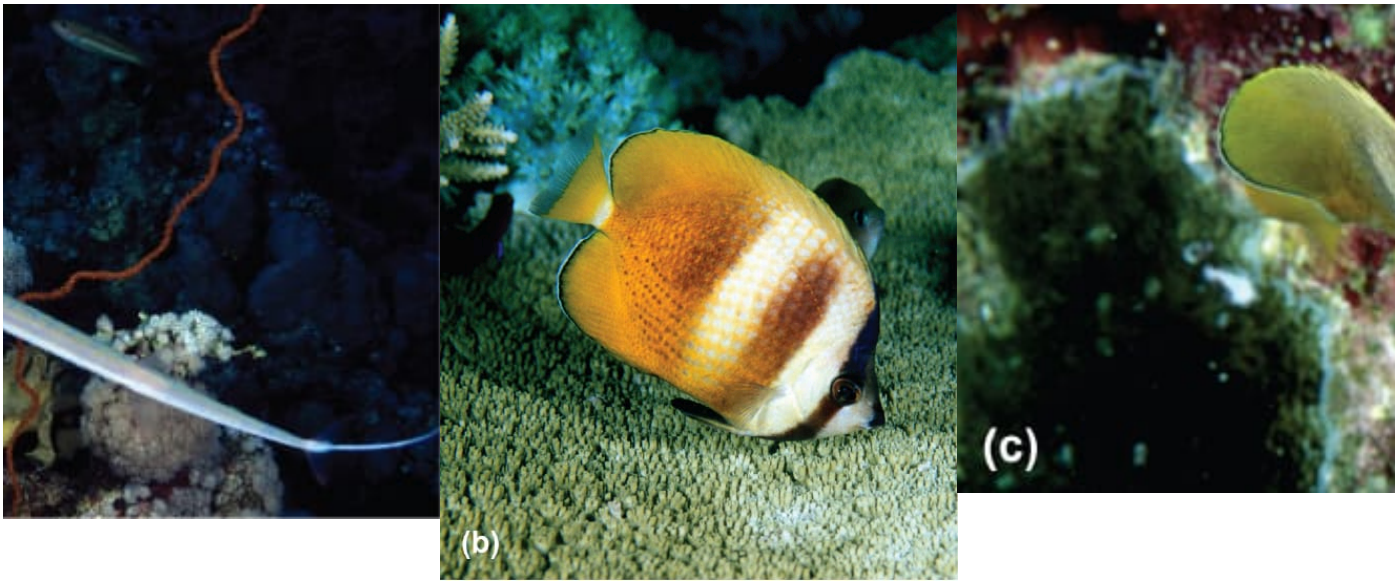


Figure 14-22. Some fish shapes that minimize form drag. (a) The cornetfish (*Fistularia commersonii*, Red Sea) has a very long, needle-shaped body to minimize form drag. (b) Side view of Klein's butterflyfish (*Chaetodon kleinii*, Papua New Guinea). (c) Front view of the same species of butterflyfish. Note the lateral compression of the body (narrower across than vertically), which reduces form drag and enables the fish to execute turns quickly.

The third form of drag, turbulent drag, dictates the final refinement in fish body shape. The turbulent flow of water around the fish is reduced if the front is rounded and blunt and the rear tapers to form a teardrop shape (**Fig. 14-21**). This shape is similar to the cross section of an airplane wing, or the form of a blimp or submarine, all of which are designed to minimize drag.

The fastest and most continuously swimming fishes, such as tuna (**Fig. 14-21**), have a generally teardrop shape to minimize total drag and thus minimize energy used in swimming. Eyes that are flush and smoothly contoured against the body and a slimy coating are other adaptations that reduce surface drag. Fishes that do not swim continuously and fishes that are adept at fast turns rather than high speed have less need to reduce drag than fast swimmers have. Thus, these fishes often have body shapes that are greatly modified from the teardrop form to conform with their own special habits. Fishes that specialize in short bursts of speed with quick accelerations generally are somewhat thickened in the middle (**Fig. 14-23**) by the heavy musculature necessary for such maneuvers (similar to the difference between long-distance runners and weight lifters). Fishes that specialize in quick turns generally are somewhat flattened (**Fig. 14-22b,c**) so that the flat sides of their body can be used in turning in much the same way that a boat's rudder is used. Fishes that hunt mostly by stealth or feed on plankton and do not need to swim quickly often have bizarre body shapes (**Fig. 14-16e**). Fin shapes are also modified to accommodate different swimming habits.

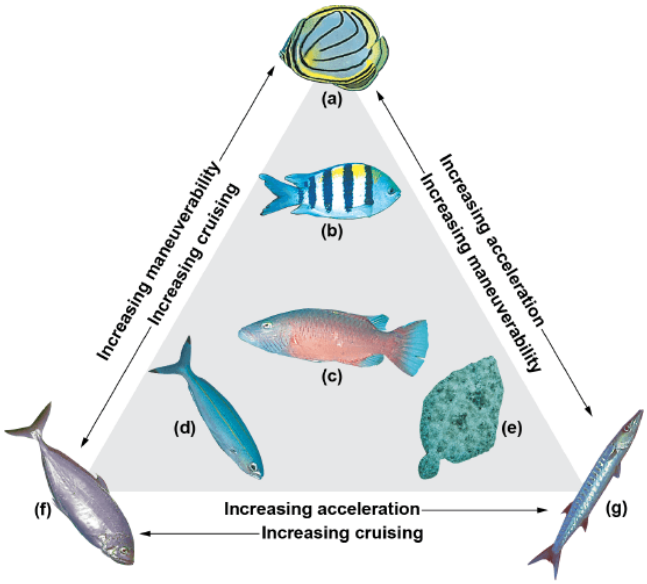


Figure 14-23. The body shapes of fish species are related to their swimming habits. Efficient cruising with minimal energy expenditure, maximum maneuverability, and rapid acceleration are all desirable but require different body shapes. Because there is no perfect shape, each species tends to specialize or compromise in a different way, as shown here. The species shown are (a) Meyer's butterflyfish (*Chaetodon meyeri*, Papua New Guinea), (b) sergeant fish (*Abudefduf sp.*, Papua New Guinea), (c) cheeklined wrasse (*Oxycheilinus digramma*, Papua New Guinea), (d) yellowstreak fusilier (*Pterocaesio lativittata*, Vanuatu), (e) flounder (*Bothus sp.*, Hawaii), (f) trevally (probably *Carangoides sp.*, Fiji), and (g) chevron barracuda (*Sphyraena putnamiae*, Solomon Islands). The relative sizes of the species are not preserved in this figure. Although most fish species follow the general relationships between shapes and behaviors depicted in this figure, some do not fit this pattern, because their behavior is highly specialized. For example, frogfishes (**Fig. 14-14b**) are sedentary, except for short bursts of rapid acceleration. Their thick bodies provide the musculature for these quick bursts, but they cannot sustain the movement as barracudas can.

Adaptations of Fins

Fish fins are extremely important in swimming and executing turns, just as the tail fin, rudder, short rear wings, and wing flaps of an airplane are important in controlling course and stability. Fishes generally have a pair of pectoral fins (one on each side of the body behind the head), two dorsal fins along the center of the back, and an anal fin beneath the rear half (**Fig. 14-24a**). They also have a caudal fin at the posterior, and a pair of pelvic fins on either side of the lower body forward of the anal fin. These fins vary greatly among species in size, shape, and location, and some fins may be absent in certain species.

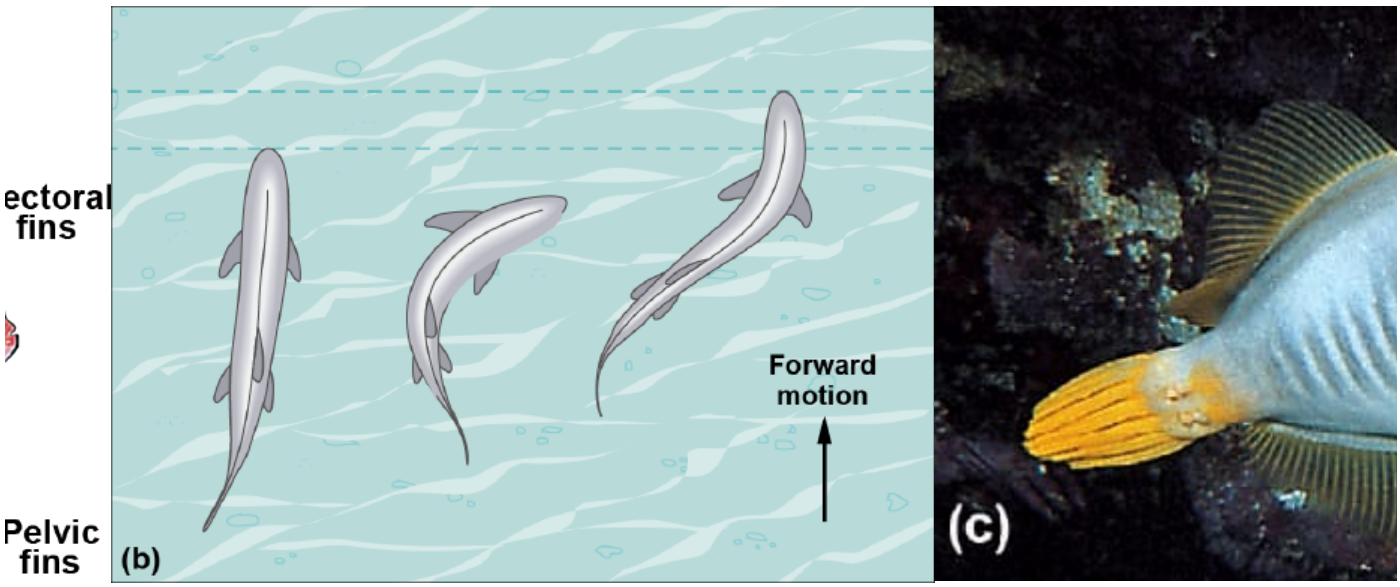


Figure 14-24. Fish fins and swimming motion. (a) Fishes have a dorsal fin or fins along the back, pectoral fins on either side of the body, a pair of pelvic fins under the forward half of the body, an anal fin under the rear half of the body, and a caudal fin at the tail. Each of these types of fins may be much modified for special purposes in different fish species. The fish shown here is a threadfin, or red-cheeked fairy basslet (*Pseudanthias huchtii*, Papua New Guinea). (b) Most fish species swim by using a series of undulating body motions. (c) Triggerfishes and their relatives, such as this barred filefish (also called an orange-fin filefish; *Cantherhines dumerilii*, Hawaii) swim not by undulating their bodies, but by back-and-forth motions of their specially adapted soft dorsal and anal fins. This movement gives them great maneuverability, but it is less efficient in providing acceleration than is the undulating body motion method of swimming.

The pairs of pelvic and pectoral fins are used primarily to execute maneuvers, including turns and stops, and usually can be folded flat against the body when not in use. The vertical dorsal and anal fins serve primarily as stabilizers during swimming and, in some species, can be folded against the body when not needed. The caudal fin is the primary provider of propulsion in most species. Most fishes also alternately contract and relax muscles along their body to create a wavelike motion that travels along the body and produces a forward thrust (**Fig. 14-24b**). The caudal fin that provides the final thrust is flared out vertically to provide a large surface area and, consequently, strong thrust as it is moved from side to side. You can experience such a thrust increase if you use swim fins.

Increasing the caudal fin surface area increases thrust but also increases surface drag. Therefore, the caudal fin is modified to reflect the swimming habits of the species. These modifications are shown by the **aspect ratio** of the fin, which is defined as

$$(\text{fin height})^2 / (\text{fin area})$$

Five major types of caudal fins are distinguished by different ranges of aspect ratio and fin shape (**Fig. 14-25**). Rounded fins (aspect ratio 1; **Fig. 14-25a**) are useful for maneuvering and quick acceleration by species such as butterflyfishes (**Figs. 14-14e, 14-22b,c, 14-23a**). Truncate fins (aspect ratio 3; **Fig. 14-25b**) and forked fins (aspect ratio 5; **Fig. 14-25c**) reduce drag in comparison with rounded fins, but they still provide substantial maneuvering assistance. Truncate and forked fins are used by many species that swim reasonably fast but also maneuver relatively quickly (**Fig. 14-23**). Lunate caudal fins (aspect ratio up to 10; **Fig. 14-25d**) are typical on fast and continuously swimming species, such as trevally (**Fig. 14-17a,b, 14-23f**), tuna (**Figs. 12-22a, 14-21**), marlin, and swordfish. Lunate caudal fins in these species are rigid and have little use in maneuvering, although they are very efficient in forward propulsion. Fast-swimming predatory species with lunate caudal fins can outrun other species with truncate, forked, or rounded caudal fins, but the prey species have an excellent chance of avoiding the predator by using their greater maneuverability.

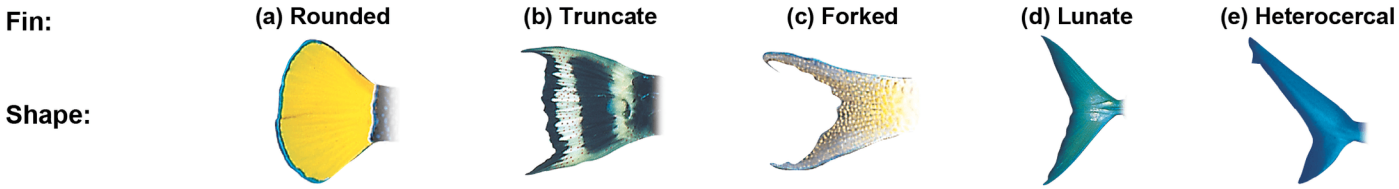


Figure 14-25. The shape of the caudal fin in fish species is related to their swimming behavior. There are five basic types of caudal fins, characterized by different aspect ratios, although caudal fins of many species have characteristics intermediate between these categories. The caudal fins shown here belong to (a) a blue-girdled angelfish (*Pomacanthus navarchus*), (b) a redbreasted Maori wrasse (*Cheilinus fasciatus*), (c) a male black-spot angelfish (*Genicanthus melanospilos*), (d) a trevally (*Carangoides sp.*)—all photographed in Papua New Guinea—and (e) a gray reef shark (*Carcharhinus amblyrhynchos*), photographed in Palau. Their relative sizes are not preserved in this figure.

Heterocercal caudal fins are asymmetrical, the upper lobe being longer and taller than the lower lobe (**Fig. 14-25e**). This type of caudal fin is used primarily by sharks (**Fig. 12-23**). It is similar to the lunate caudal fin in that it provides very efficient forward thrust but little help in maneuvering. However, its asymmetrical shape also provides upward lift, which is important to sharks because they have no swim bladder and tend to sink if they stop swimming. Lift is also provided by the sharks' pectoral fins.

Unlike those of most other fishes, sharks' pectoral fins are large, flat, and relatively inflexible (**Fig. 12-23**), providing lift like aircraft wings. The pectorals are relatively far forward on the shark's body (**Fig. 12-23**) and lift the front of the shark, while the caudal fin lifts the tail. Although sharks are powerful swimmers, their fins are designed poorly for maneuvering, and they are not adept at capturing prey that can anticipate their charge and perform evasive maneuvers. Unfortunately, this knowledge is of little use to swimmers who may be attacked by sharks, because human body shapes do not allow for quick maneuvers in the water.

Although most fish species use their fins for swimming, numerous species have fins adapted to perform highly specialized functions. A number of species have dorsal fins modified to act as defensive, and perhaps offensive, weapons. In these species, the individual rays or spines of the dorsal fins, or parts of the dorsal fins, are needle-sharp and may contain a venom that is injected into any predator that challenges the fish. Species with this type of dorsal fin include stonefishes and scorpionfishes (**Fig. 14-16d**), which lie on a reef or the seafloor and use their dorsal defenses to protect against attacks from above. They also include lionfishes (**Fig. 14-26a**), which swim close to a reef, turning their backs toward any approaching predator and their more exposed undersides toward the reef. In some species, such as frogfishes (**Fig. 14-28c**) and anglerfishes (**Fig. 14-15**), the forward-most ray or rays of the dorsal fin are adapted to become lures that can be dangled in front of the fish's mouth to entice prey.

Figure 14-26. Examples of fin adaptations. (a) The zebra lionfish (*Dendrochirus zebra*, Papua New Guinea) has long, poisonous dorsal-fin spines to protect itself. (b) The clown triggerfish (*Balistoides conspicillum*, Philippines) has a strong, sharp spine that it can raise when threatened. Like many other reef fishes, clown triggers are territorial and will chase off any intruder of their own species. (c) The speckled sandperch (*Parapercis hexophtalma*, Fiji) uses its pectoral fins for support as it sits and waits for its prey. (d) Wrasses, like this moon wrasse (*Thalassoma lunare*, Papua New Guinea), use their pectoral fins like oars to propel themselves. (e) Many fishes, especially those that live in holes in the reef or seafloor, such as this Old Glory goby (*Amblygobius rainfordi*, Papua New Guinea) have long, fused dorsal and anal fins. (f) This darkspotted moray eel (*Gymnothorax fimbriatus*, Papua New Guinea), photographed while hunting at night, showed aggression when disturbed by the diver's lights. Note the sinuous, snakelike appearance. (g) This remora, or sharksucker (*Echeneis naucrates*, Papua New Guinea), was attached to a scuba diver's leg, where it stayed for some time before leaving to try to find another ride.

In triggerfishes (**Fig. 14-26b**), the forward dorsal fin consists primarily of a single strong, rigid spine that normally lies flush against the fish's body. This spine can be extended from the body to become a fearsome weapon that gives the triggerfish its name. Certain triggerfish species may also use this spine defensively. When chased, they swim headfirst into holes in the reef and extend their trigger to lock themselves in. Because of the way the trigger is hinged, almost no amount of tugging by a predator can pull the triggerfish out. Once the predator leaves, the trigger can be relaxed, allowing the triggerfish to back out of its refuge. Triggerfishes also use modified fins for swimming. The anal and dorsal fins are enlarged and undulate back and forth in wavelike motions that replace the body undulations used by other fishes. These fins enable triggerfishes to hover, turn, and swim slowly forward or backward to enter holes in the reef just wide enough for them to fit through.

Many species that live or rest frequently on the seafloor have pectoral and sometimes caudal fins that are elongated and have strong rigid spines on which the fish can rest. Sandperches (**Fig. 14-26c**) have this type of adaptation. In certain deep-sea species that inhabit areas where currents are generally weak, both the pectoral fins and the lower lobe of the caudal fins are elongated to an extent that can exceed the fish's body length. These fishes "walk" on the seafloor as if perched on a tripod.

Like triggerfishes, wrasses (**Fig. 14-26d**) do not normally swim by using body undulations for propulsion. Instead, they propel themselves with their pectoral fins, which they stroke back and forth in much the same way that an oar is used. The fin is moved backward while spread vertically to push the water back and the fish forward. Then it is rotated and moved forward while in a horizontal orientation that minimizes drag. Wrasses can also swim by using their caudal fin and body undulation in the same way that other fishes do, but the oar-like propulsion created by using the pectoral fins provides better control of movements at slow speeds. Such control is ideally suited to the wrasses' feeding habit of picking small crustaceans, algae, and individual coral polyps from cracks and crevices in reefs.

Many fishes spend most of their lives concealed in holes in a reef and do very little swimming. These fishes generally have greatly reduced fins, and their dorsal, caudal, and anal fins are often fused into one continuous fin extending around the fish. In extreme cases, the dorsal and anal fins may be missing entirely. Such fishes usually have an elongated body and swim by using sinuous body undulations. Swimming without the use of dorsal and anal fins is slow and inefficient, but it is perfectly suited to the lifestyles of these fishes. Snakelike flexibility and the lack of protruding fins enable them to swim easily through the narrow, tortuous passages of holes in the reefs where they live. Species adapted in this way include gobies (**Fig. 14-26e**) and moray eels (**Fig. 14-26f**).

Flyingfishes have some of the most bizarre fin adaptations. These warm-water fishes have elongated pectoral (and, in some species, also pelvic) fins that can be spread out from the sides to resemble bird wings. When they sense danger, flyingfishes swim upward with a rapid burst of speed that carries them through the water surface and into the air. Once in the air, they spread their fins and, using these "wings," sail a few tens of centimeters above the waves. They can glide for as long as 30 s, and some species can prolong their glide by flailing an elongated lower lobe of the caudal fin at the sea surface as they descend close to the water. Although flying is a very effective strategy to escape from some predators, flyingfishes expose themselves to predation by seabirds.

A number of species, such as clingfishes and remoras (**Fig. 14-26g**), use fins as suction devices with which they can hold onto the seafloor or another organism. Clingfishes use modified pelvic fins to cling tenaciously to rocks in coastal areas where wave action is intense. Remoras use a modified dorsal fin as a suction cup to attach the top of their head to sharks, manta rays, other large ocean animals, and even boats and **scuba** divers. The remora's sucker looks so little like a fin that careful research was needed to identify its origin. Although remoras must swim rapidly to catch and attach to a host, they are then transported without having to expend energy. Remoras can detach themselves to feed on any nearby available food, particularly scraps of their host's meal if they are riding on a shark or other predator.

Ghost pipefishes (**Fig. 14-16e**) and seahorses (**Fig. 14-16h,i**) have perhaps the most extreme adaptation of body form and fins. Some species swim in a vertical head-down or head-up position and propel themselves slowly by rapid back-and-forth oscillations of their small dorsal fins. Because the fish is oriented vertically, the thrust from these fins is oriented perpendicular to the fish's body. Ghost pipefishes thus swim sideways, although they swim in a normal horizontal position when they need to swim rapidly to avoid danger.

Osmoregulation

The relative proportions of dissolved chemicals in body fluids of fishes, other **vertebrates**, and invertebrates are remarkably similar to their relative proportions in seawater. In most invertebrates, the salinity of internal fluids is also the same as the external seawater salinity. However, the internal fluids of fishes are less saline than seawater. The reason is that bony fishes are thought to have first evolved in freshwater. Because osmosis causes water to diffuse across cell membranes from lower salinity to higher salinity, fishes must be able to counteract osmosis.

Osmosis of water molecules across a semipermeable membrane (such as the cell surface) from lower to higher salinity is easy to understand. Because the "concentration" of water molecules is higher in the lower-salinity fluid, more water molecules are in contact with that side of the membrane. Thus, more water molecules diffuse through the membrane toward the higher-salinity fluid than diffuse in the opposite direction. In contrast, dissolved salt molecules would be more likely to diffuse from high to low salinity because of their higher concentrations, but ions of dissolved salts are generally much larger than water molecules and hence less likely to pass through the openings in a semipermeable membrane.

One way to counteract osmosis is to increase the pressure on the high-salinity side of the membrane, which forces more water molecules through the membrane to the lower-salinity fluid. The pressure needed to balance water migration across a membrane is called "osmotic pressure." Osmotic pressure increases as the difference in salinity between the fluids on either side of the membrane increases.

Marine fishes must have a mechanism for counteracting osmosis or they would continuously lose water to their surroundings and dehydrate. Fishes cannot maintain a pressure difference across their external membrane, because they would be able to do so only if they had an impermeable body surface, like the pressure hull of a submarine. With an impermeable body surface, feeding and excretion of waste products would be extremely difficult because they would have to take place through the equivalent of a submarine airlock. In addition, all aquatic organisms must exchange oxygen, carbon dioxide, and nutrients with seawater through a porous membrane. Consequently, fishes have adapted methods called **osmoregulation** to counteract osmosis in seawater.

Fishes that live in seawater osmoregulate by drinking seawater to replace water lost from their internal fluids by osmosis. The excess salt ions ingested with the seawater are excreted through specially adapted cells in the **gills** (**Fig. 14-27a**). Freshwater fishes osmoregulate by drinking almost no water and excreting large volumes of very dilute urine to discharge the water that enters their bodies by osmosis. They also must take up dissolved salts through their gills (**Fig. 14-27b**). Certain estuarine fish species or species that live in environments of variable salinity, such as **tide pools**, must be able to osmoregulate in both directions. Because few fish species have such ability, most estuarine fishes migrate within the estuary as river and tidal flows vary to remain at the same approximate external salinity.

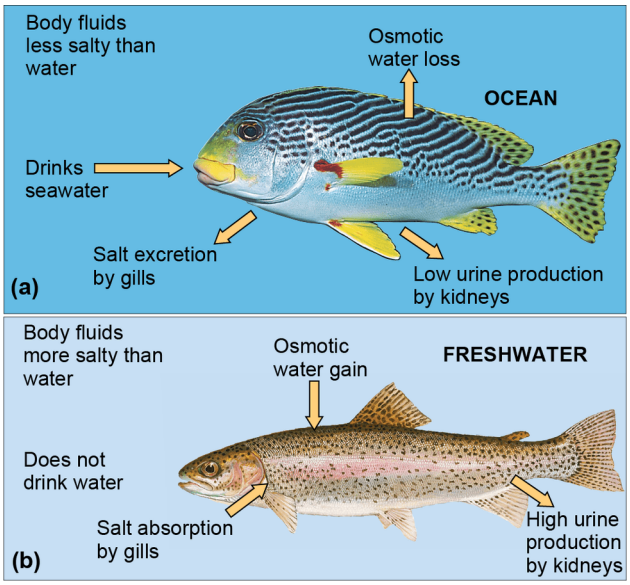


Figure 14-27. Osmoregulation. (a) Fishes that live in seawater, such as the Goldman's sweetlips (*Plectorhinchus goldmanni*, Vanuatu) illustrated here, lose water continuously by osmosis across their body surfaces and must drink seawater and excrete salt through specially adapted cells to replace the lost water. (b) Fishes that live in freshwater, such as Rainbow trout (*Oncorhynchus mykiss*), absorb water continuously by osmosis and get rid of the excess by producing copious amounts of very dilute urine.

Two special categories of fishes have evolved life cycles that require them to cross the salinity gradient between freshwater and seawater twice during their life cycles. These are the **anadromous** and **catadromous** fishes (**Chap. 13**).

Swim Bladders and Buoyancy

Some fish species that live most of the time on the ocean floor can afford to be negatively buoyant (dense enough to sink) because they expend relatively little energy swimming against **gravity** during their limited excursions above the seafloor. Examples include most species of frogfishes (**Fig. 14-28c**) and scorpionfishes (**Fig. 14-16d**). Most fishes live in the water column and cannot afford to expend energy by swimming continuously to counteract gravity and maintain their depth. Consequently, most pelagic fishes (and other vertebrates and some invertebrates) must find a way to adjust their buoyancy to be approximately equal to that of seawater. The two primary ways of doing this are to synthesize and retain low-density oils, or to develop a gas-filled bladder.

Many fish species that live in surface layers and mid depths achieve neutral buoyancy by filling an internal swim bladder with gas. The amount of gas within the swim bladder must be adjusted as the fish changes depth. Otherwise, the gas will expand or contract and change the buoyancy of the fish as it ascends or descends. Some fishes with swim bladders make only limited and slow vertical excursions because the exchange of gas between blood and swim bladders is slow. Such species often die if caught and brought rapidly to the surface, because the swim bladder expands faster than the fish can evacuate the gas (**Fig. 14-15**). Other species have a special duct that connects the swim bladder with the esophagus, and they can ascend rapidly by "burping" to release excess gas from the swim bladder. Moray eels (**Fig. 14-26f**) have this duct, allowing them to change depth rapidly while hunting prey.

In shallow-water fishes, the gases in the swim bladder are similar to the atmosphere in composition: about 20% oxygen and 80% nitrogen. Some shallow water fishes fill their swim bladder by surfacing and gulping air. Fishes that live at greater depths have higher oxygen concentrations in their swim bladders, in some species up to 90% oxygen. The oxygen concentration in swim bladder gas reflects the mechanism fishes employ to force gas from solution in their blood to a gas in the swim bladder which is more effective for oxygen than for other gases.

At a depth of 7000 m, the pressure is so great and gases so compressed that their density is approximately the same as that of fats. Consequently, many deep-water fishes have swim bladders filled with oil or fat instead of gases. These fishes do not need to adjust the amount of gas in the swim bladder as they change depth.

The largest and most active swimmers, such as mackerel and tuna (**Figs. 12-22a, 14-21**), have no swim bladder. These fishes can afford the relatively small energy penalty required to maintain their depth against their negative buoyancy because they expend

much greater amounts of energy in swimming. Many of these species, especially sharks, have fins and body shapes designed to counteract their negative buoyancy as they swim. Sharks also have large livers with high concentrations of lighter-than-water oils to provide some compensation for their negative buoyancy.

This page titled 14.5: Selected Adaptations in Fishes is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.6: Reproduction

To reproduce successfully, each species must ensure that enough of its offspring survive to reproductive age and that these progeny in turn produce enough of their own offspring to continue the cycle. Reproductive cycles are poorly documented or poorly understood for all but a few marine species, but the basic elements of their diverse strategies are known.

Separate-Sex Reproduction

The majority of marine species reproduce by sexual interaction between male and female. Sexual reproduction increases **genetic** diversity (**CC17**), which improves the species' ability to survive and adapt to environmental change.

In sexual reproduction, sperm is transferred from male to female by a copulatory organ or directly from the male sexual organ to eggs laid previously by the female. Alternatively, sperm is either discharged into the water, where it fertilizes eggs retained by the female, or eggs released into the water, where fertilization occurs.

Direct transfer of sperm from male to female requires relatively few eggs and sperm to ensure sufficient fertilizations. Hence, this strategy minimizes the energy needed to produce eggs or sperm. However, it requires that males and females locate each other for breeding, which for many species is a major problem, especially if they are permanently attached to a surface. Many motile fishes and invertebrates find mates in mating congregations that assemble at a specific time of year and place. Mating congregations enable some fishes to select mates that have desirable characteristics.

The mate location problem has been solved in an unusual way by a number of species, including some barnacles and fishes that live at depth, in the dark. In these species, one sex is relatively small and resides either near or directly on the other. The smaller sex is usually the male because producing sperm requires less energy, and hence less body weight, than producing eggs does. The deep-sea anglerfish (**Fig. 14-15**) is a good example of a species that uses this strategy.

In pelagic reproduction, sperm and often eggs are released into the water (**Fig. 14-28a,b,c,l**). Sperm are attracted to eggs or females of their own species probably by chemotaxis ("tasting" or "smelling" of chemical clues released into the water by eggs or females). If sperm are released far from the eggs or females, or at a different time than the eggs, the probability of fertilization is very small. Consequently, pelagic spawning often occurs synchronously among all males and females of a given species in a specific region. Often this synchronous spawning occurs at a specific time of year or on a specific day. The timing is usually coordinated with a particular phase of the tides, such as high **spring tides** at full or new moon. Corals (**Fig. 14-28a**), many other benthic invertebrates, and many fishes spawn synchronously. In some fish species, a male and female pair will spin around each other or swim in contact with each other (**Fig. 14-28b**), simultaneously releasing and mixing a cloud of sperm and eggs. Other species form spawning congregations in which vast clouds of sperm and eggs are released simultaneously. Often the individuals swim in a school or swarm that spirals inward and upward toward the surface as sperm and eggs are released. This movement ensures that sperm and eggs are concentrated in an intensely mixed, dense cloud; keeps the eggs away from benthic predators; and probably reduces loss to pelagic predators.

Figure 14-28. Reproduction. (a) Synchronous spawning of a hard coral colony (*Acropora sp.*, Papua New Guinea). The eggs, the pinkish white globules, are all released within minutes of each other to float up and join the plankton where they hatch. (b) Mandarinfish (*Synchiropus splendidus*, Indonesia) male and female pair photographed almost exactly at the instant that they spawned. At dusk, they swim very rapidly upward about a meter from the rubble patch where they live, release their eggs and sperm at the same instant, and then return rapidly to safety. (c) This female Lembeh frogfish (*Antennarius sp.*, Indonesia) is swollen with eggs. She and the much smaller male that was waiting close by swam upward and spawned near the surface a few hours later that night. (d) Although they are all hermaphrodites, nudibranchs exchange sperm through ducts in their side, seen connected in this image (*Hypselodoris bullocki*, Indonesia). Each individual retains and lays its own fertilized eggs. (e) Nudibranch laying eggs (*Kentrodoris rubescens*, Papua New Guinea). (f) On rare occasions, three nudibranchs may exchange eggs and sperm simultaneously (*Nembrotha rutilans*, Indonesia). (g) Nudibranch eggs are laid in a characteristic spiral ribbon pattern encased in a protective gel (species unknown, Hawaii).

Figure 14-28. Reproduction. (h) Snails, although related to nudibranchs, nearly all reproduce sexually, but they also lay eggs encased in a protective gel (*Epitonium billeeanum*, Papua New Guinea). (i) Common reef squid eggs (*Sepioteuthis lessoniana*, Papua New Guinea) laid in rubbery gel casings in a seagrass bed. (j) Flamboyant cuttlefish eggs (*Metasepia pfefferi*, Indonesia) laid under a coconut shell half. The individual near the center of the image is close to hatching, and the unhatched cuttlefish can be seen through the transparent egg case. (k) Many species of damselfishes, including this Indo-Pacific sergeant (*Abudefduf vaigiensis*, Vanuatu), lay eggs on the reef surface and then stay to defend the eggs aggressively from potential predators until they hatch. The eggs are the tiny reddish brown dots covering the reef surface behind the sergeant. (l) Sea cucumbers that normally lie flat on the substrate (see inset) rear up as high as they can before releasing clouds of eggs or sperm. Fertilization takes place in the water column, where the sperm are attracted to the eggs by chemical sensing; but the fertilization rate is low, and large numbers of eggs and sperm must be produced (*Thelenota rubralineata*, Papua New Guinea). (m) Most of this burrowing sponge (*Oceanapia sagittaria*, Indonesia) lies buried under the sediments. It reproduces vegetatively by growing the round ball-like structure in the image, which is released to the currents. (n) Banggai cardinalfish (*Pterapogon kauderni*, Papua New Guinea). The male shown here is incubating eggs (the orange spherical objects) in his mouth. If you look closely you can see the eyes of some of the young that are almost ready to hatch.

Hermaphroditism

In many species, hermaphroditic life cycles have solved the problem of finding or ensuring the presence of a mate. Hermaphrodites are individuals that have both the sexual organ (gonad) necessary to produce sperm and the sexual organ necessary to produce eggs. In certain species, including most nudibranchs, the individual may perform either the male or female function, or both, at any time in its sexually mature stage (**Fig. 14-28d,f**). The common acorn barnacle is also a hermaphrodite in which each individual has a penis that can be used to inject sperm into any other barnacle because each barnacle also has ovaries. Thus, although adult barnacles are permanently attached to a surface, a suitable mate is always present in any location where these barnacles are well established (**Fig. 14-7c**).

Sequential hermaphroditism is the ability of an individual to change from female to male or from male to female at an appropriate time in its adult life. The female-to-male sex change confers an advantage to species in which reproductive success is aided by the presence of a strong, experienced male during reproduction. For example, damselfishes (**Fig. 14-28k**), including anemonefishes (**Fig. 14-20**), lay their eggs on the reef surface (under the edge of the anemone in the case of anemonefishes). Reproductive success in these species depends on the successful defense of fertilized eggs against predators. For a given anemone, an anemonefish family may consist of a large aggressive male, a somewhat smaller female, and usually 4 to 10 smaller subadults or juveniles. The male defends the eggs from predation and defends his anemone from predation by butterflyfishes. If the male dies, the female grows and changes sex, and one juvenile grows to become a sexually mature female.

A male-to-female sex change is the preferred strategy of many species in which the female produces more eggs as it becomes larger. This strategy is common in invertebrates. For example, the common eastern oyster (*Crassostrea virginica*) is a male for several years as it grows, and then it transforms into a female. In transitional individuals, both sperm and eggs may be produced and the individual may even fertilize itself.

Asexual Reproduction

There are several different methods of asexual reproduction, including binary fission, fragmentation, and vegetative reproduction. In another form of asexual reproduction, unfertilized eggs develop into adults, but this reproductive strategy is rare in marine life.

In binary fission, a single-celled organism divides into two offspring. Diatoms reproduce by binary fission (**Fig. 12-17**). Fragmentation is a reproductive process of multicelled organisms that is similar to binary fission. For example, fragments of certain macroalgae and of some invertebrates, including various worms and sea stars, can break off and develop into new individuals. In some cases, fragmentation occurs through the production of special buds that are designed to break off the main individual. This budding process is common in macroalgae and sponges.

In vegetative reproduction, a single individual divides into many individuals that may or may not be physically connected. Vegetative reproduction is the most important marine asexual reproductive mechanism and occurs in many marine algae and invertebrate species, particularly colonial forms, including corals, sponges (**Fig. 14-28m**), and anemones. In benthic invertebrates, a single larva may settle on a substrate, **metamorphose**, and then reproduce vegetatively to cover a broad area of substrate.

Asexual reproduction produces many individuals that are genetically identical **clones**. Therefore, colonies of vegetatively reproduced anemones or encrusting sponges, for example, are all the same sex and color. Cloned colonies can occupy a substrate very densely because the cloned individuals do not have aggressive territorial responses toward each other as they would toward other individuals of their species. This lack of aggression may be important in colonial species, such as the Portuguese man-of-war, in which millions of cloned individuals must cooperate, each performing only one of a variety of separate and different functions needed to feed, defend, move, and reproduce the colony.

Asexual reproduction allows the successful colonization of suitable benthic microhabitats encountered by a single settling larva and avoids the difficulties of ensuring that sperm and eggs are united successfully. However, asexually generated populations have little genetic diversity, and hence the colony, and the species, may be poorly adapted to survive any unfavorable changes in its environment.

Egg Laying

Most marine fish and invertebrate species are oviparous and lay large numbers of eggs that hatch on the seafloor (**Fig. 14-28e-k**) or in the water (**Fig. 14-28a,b,c,l**) to become larvae that are **meroplankton**. In most cases, the vast majority of the eggs and larvae (sometimes more than 99.999%) are consumed by carnivores before they reach adulthood. For benthic species, many additional larvae may die because they settle on unsuitable substrates. Therefore, species that have pelagic eggs and/or larvae generally must produce very large numbers of fertilized eggs to ensure survival of the species. Species that spawn eggs to the water for fertilization must produce especially large numbers of eggs because many will not even be fertilized.

Production of very large numbers of pelagic eggs ensures the wide dispersal of larvae, which enables them to take advantage of the dispersed phytoplankton food resource. For benthic species, the dispersal of large numbers of larvae also facilitates the colonization of suitable substrate where this substrate is found only in isolated areas. Because this strategy carries a substantial energy cost, many species of oviparous fishes, such as anemonefishes (**Fig. 14-20**) and other damselfishes (**Fig. 14-28k**), lay many fewer eggs on the substrate and protect the eggs from predators until they hatch. Other species, including many sharks, skates, and rays, octopuses and squid (**Fig. 14-28i,j**) lay only a very few eggs that are protected from predators in tough envelopes or encased in a gel (**Fig. 14-28d-h**). Finally, **ovoviviparous** fishes retain fertilized eggs within their reproductive tracts until they hatch. Seahorses (**Fig. 14-16h,i**) and some pipefishes (**Fig. 14-16e**) have an unusual **incubation** mode whereby newly fertilized eggs are deposited in a pouch on the male's abdomen, where they remain until they hatch. Some cardinalfishes incubate eggs in the male's mouth (**Fig. 14-28n**).

Although it would seem to be energy-efficient to produce fewer eggs and protect them until they hatch, energy must be expended to protect the eggs, and this strategy does little to reduce mortality during the larval stage. Larval mortality can be reduced if larvae can evade some predators, but to do so, the larvae must be relatively large. Remember, if larvae are small, they cannot swim effectively, because of the enhanced effect of viscosity (**Chap. 5**). If larvae must be large when hatched, eggs must have large yolks to provide the energy for growth. This requirement offsets any energy gained by producing fewer eggs.

Viviparous species are at the opposite extreme from the prolific pelagic spawners in the trade-off between egg numbers and protection. In viviparous animals, including mammals, offspring are nurtured inside the mother's body until they become fully developed and assume adult or nearly adult form. Some shark and ray species are viviparous, as well as all marine mammals. Only one or two offspring are produced at a time because of the large amount of energy needed to nurture the offspring during its

prebirth development, but the large live-born offspring have a high rate of survival to reproductive age.

Timing

The timing of reproduction during the life cycle and during the year can be important. Spawning in some species takes place over a period of days or weeks at the same time each year. In other species, individuals may spawn several times a year; and in some species, spawning is almost continuous because at any given time, some individuals in the population are spawning. In middle and low latitudes, most pelagic spawning species spawn only once a year, in spring. This pattern synchronizes the production of larvae with the availability of abundant food supplies from the spring phytoplankton **bloom**.

In many species, the time of spawning may be determined by variations in light intensity and temperature, which also influence the timing of the phytoplankton bloom. Dramatic year-to-year variations in the reproductive success of many species can and do occur. They are due in part to annual variations in factors controlling the timing of the phytoplankton bloom and of spawning. In some years, spawning may occur at the wrong time or place, in which case larvae miss the phytoplankton bloom and their population incurs massive starvation losses. Omnivores and carnivores suffer in turn, and their populations decline.

In tropical latitudes and certain **upwelling** areas where plankton food supply varies relatively little during the year, pelagic spawners have a greater tendency to spawn year-round. However, this trend is not universal, because even in these areas, monthly, biannual, or annual synchronous spawning still provides many advantages for larval survival that are not afforded by continuous spawning.

In certain species, spawning is timed to provide the maximum protection of eggs from predators. The best-known users of this method are the small fish called grunion that spawn on southern and Baja California beaches, and the horseshoe crab that spawns along the Atlantic coast. These and other species deposit eggs to incubate in beach sands above the **high-tide line**. Thus, eggs are protected from the many predators that abound in coastal waters and sediments. Eggs must be placed in sand high enough up the beach to be free from wave action or they would soon be washed out and devoured.

Grunions cannot climb above the high-tide line, and both grunions and horseshoe crabs have pelagic larvae that cannot crawl across the beach to enter the water after they hatch. How do grunions and horseshoe crabs place their eggs in sand above the high-tide line and yet ensure that the sand is covered with water when the larvae hatch? Each species times its spawning to occur at or a little after some of the highest spring tides of the year. In the grunion's case, the high spring tides that occur on the Pacific coast during summer are perfect because the highest tide of the day (see the discussion of **mixed tides** in **Chapter 10**) occurs at night. At high tide in the darkness, when they are safe from air attacks by birds, grunions can move into very shallow water high up on the beach. There they lay eggs during a frenzied spawning that lasts about 30 min, until the tide recedes. After the eggs are laid, **tidal ranges** diminish as spring tides progress to **neap tides**, and then build back toward the next set of spring tides. Two weeks after the eggs are laid, the next set of spring tides occurs (**Chap. 10**). By this time, the eggs have matured. When the overlying sand is resuspended by waves of the high spring tides, eggs are released to the water and the larvae hatch, becoming meroplankton.

The factors that determine the timing of spawning within a given year are relatively simple to understand. It is more difficult to understand why some species reproduce only once and then die while others reproduce repeatedly, and why some species mature rapidly while closely related species may take much longer to reach reproductive age. The average mortality rate of larval and adult stages, energy requirements of spawning, and year-to-year variability in larval mortality rate are believed to be important and interrelated explanatory factors.

In species with high adult mortality and relatively low larval mortality, natural selection will favor early and one-time reproduction. Because adult mortality is high, individuals that mature late will be selectively removed from the population because they are more likely to die before they produce offspring. Individuals that spawn more than once will also be selectively removed because they are likely to die before reaching a second or subsequent spawning cycle. Thus, for many species it is a successful strategy to reach maturity quickly, spawn only once, and then die. Salmon and eels are prime examples of species that use this strategy. Also, these fishes make long-distance migrations to spawn during which they stop feeding and deplete their own body tissues for fuel. This metabolic cost makes it evolutionarily more efficient to invest all remaining energy in a single reproductive event rather than attempting a return journey.

If egg or larval mortality is high in relation to adult mortality, natural selection will favor individuals that mature late and spawn more than once. Individuals that mature early and divert a large fraction of their food energy resources to reproduction, but relatively little to growth and predator avoidance, will be more likely to die early. An individual that matures late, devotes most of its energy to growth and survival, and diverts only a small amount each reproductive year or cycle to reproduction will produce

more eggs in its lifetime. The late maturer is larger when reaching reproductive age, and thus it is capable of producing more eggs than a younger-maturing individual could, while expending the same amount of energy as a percentage of body weight. If this late maturer invests a relatively small proportion of its food energy in reproduction during each year, it can survive to spawn repeatedly, further increasing its lifetime egg production.

Selection for late maturity and multiple spawning cycles is further enhanced by great year-to-year variability in the survival of eggs and larvae. Survival through multiple spawning cycles enables a species collectively to outlast excessive variations in egg production or survival in successive spawnings. Because adults of several ages contribute to each year's total egg production, the effect of a small or missing year class (resulting, for example, from a poor year for larval survival) is minimized. Pelagic eggs and larvae are subject to intense predation and to ocean current and **climate** variations. Many fishes and benthic invertebrates with pelagic eggs and larvae have the relatively late-maturity and multiple-cycle reproduction that is associated with high and variable mortality of eggs and larvae.

Even closely related species can have very different reproductive timing. For example, the common mussel (*Mytilus edulis*) and California mussel (*Mytilus californianus*) are closely related and coexist in some areas. Because *M. edulis* has high mortality due to predation, it is outcompeted by *M. californianus* for living space in places where the larvae of both species settle. *M. edulis* invests a very large amount of energy in reproduction in comparison with *M. californianus* and spawns once a year in winter, whereas *M. californianus* spawns throughout the year.

Migration

Many marine animals migrate between different regions during their life cycles. The migrations can cross tens of thousands of kilometers, but typically they are much shorter in distance. Generally, one area on the species' migration route is its reproductive site, where eggs are fertilized, released, or deposited. The other end of the migration route (or other points on a complex migration pathway) is the main feeding ground for adults and/or juveniles. Hence, for most species, migration can be viewed as a strategy for ensuring appropriate habitat and food supply for each life stage of the species.

Migration may place larvae where appropriate food supplies are abundant and adults in another region where their different food supply is plentiful. However, migration is often not this simple, because adequate food supplies for each stage often are present throughout the migration route or at only a single location.

Why then do species not avoid the energy expenditure imposed by migration? There may be several answers. For example, the species might deplete available food in a specific region if it did not migrate to allow recovery of the food species populations. Migration may also place eggs, larvae, and/or the adult populations in locations where they are less subject to predation. Alternatively, migrations may enable species that are usually scattered widely across large ocean areas to congregate in one or more breeding areas, thus improving the chances of finding a mate and enhancing genetic mixing and diversity. Finally, adult migration to a spawning ground may ensure that pelagic eggs and drifting or weakly swimming larvae are carried by ocean currents to locations where food suitable for young adult stages is abundant. After drifting to such feeding grounds, larvae can metamorphose to the adult stage. Each of these possible advantages of migration and probably others are important for some species.

The North Sea **herring** has a migration–reproduction pattern typical of many abundant coastal pelagic fishes. Adults feed for much of the year in areas of rich plankton production off Norway (**Fig. 14-29**). In spring, herring migrate across the North Sea to spawning grounds near the Scottish coast. There they breed and lay eggs, which attach to stones or gravel on the seafloor. After they hatch, the larvae are transported eastward by currents to the adult feeding grounds, where they assume adult form.

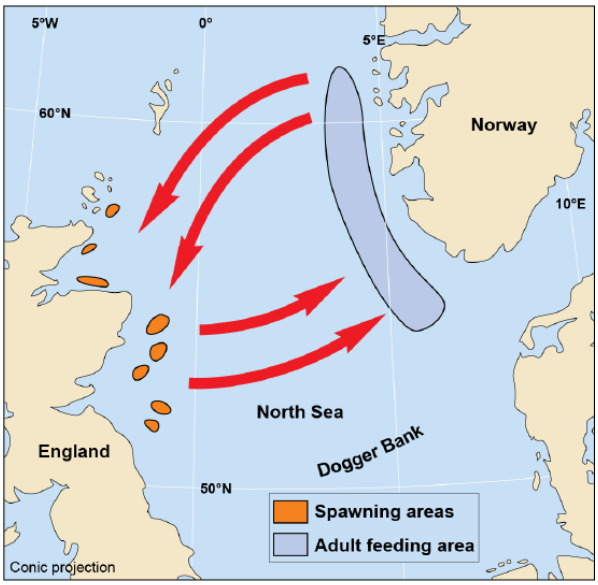


Figure 14-29. North Sea herring migrate from feeding areas off the coast of Norway to a number of localized areas off the Scottish coast to spawn.

Anadromous and catadromous fish species have perhaps the most amazing migrations. For example, the catadromous Atlantic eel, *Anguilla*, spends its approximately decade-long adult life in freshwater rivers of North America, Europe, and the Mediterranean Sea. At the end of this period, the eels undergo changes that include exchanging their dull gray color for a silver hue and growing enlarged eyes typical of pelagic fish species that live below the photic zone. The eels then migrate down their home rivers and thousands of kilometers across the oceans to the southeastern part of the Sargasso Sea (**Fig. 14-30a**). There, the eels breed and then presumably die. Their leaf-shaped larvae (called leptocephali) are a few millimeters long when they begin their journey to the streams that will become their adult homes.

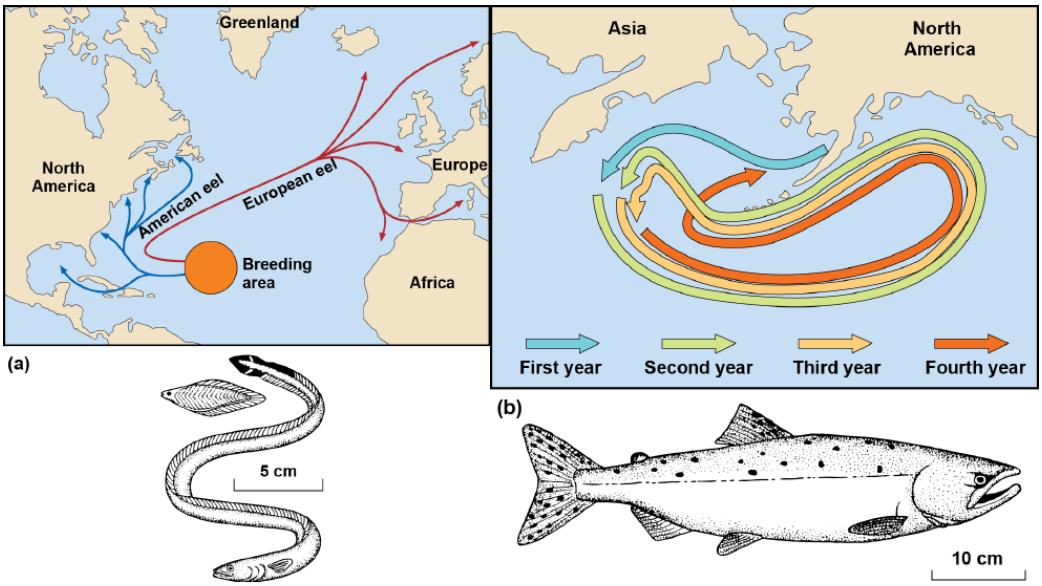


Figure 14-30. Migration routes. (a) Adult Atlantic eels migrate from their river homes in Europe and North America to the southwestern corner of the Sargasso Sea, where they spawn and die. Their larval offspring travel back to North America and Europe via the North Atlantic subtropical gyre currents and enter rivers (generally the home rivers of their parents), where they grow to maturity. (b) There are several species of North Pacific salmon, each of which has a somewhat different migration pattern. This figure shows a typical pattern in which the juvenile salmon migrate from their home streams to sea to spend 3 or more years migrating and feeding in the North Pacific Ocean between Alaska and Russia. The adults then return to rivers (usually to their home streams) to spawn and die.

The American population drifts north with the Gulf Stream for 1 or 2 years before somehow sensing the proximity of the rivers of their ancestors, at which point the eels swim upstream to reach their new freshwater homes. European eels take 2 to 3 years to drift to their European homes. Mediterranean eels take an additional year to drift north with the Gulf Stream and then back to the

south with the **eastern boundary current** of the North Atlantic Gyre until they can enter the Mediterranean. Although all Atlantic eels are considered the same species, the American, European, and Mediterranean eels are genetically distinct. These distinct populations may breed in slightly different areas of the Sargasso Sea so that larvae can join the Gulf Stream at the best point for the most efficient ride home.

In the Pacific Ocean, a similar species of catadromous eel that has adult populations in North America and Asia migrates to a spawning region in the southwestern part of the North Pacific Gyre that corresponds to the southwestern corner of the Sargasso Sea. Many details of the eel's migration, breeding behavior, and spawning locations are still unknown. In addition, as with all migrations, the chemical, magnetic, visual, or other clues that eels use to guide their migration to appropriate locations are unknown and subject to much research.

Pacific salmon are anadromous species that use a migration pattern almost opposite that of the eel (**Fig. 14-30b**). Salmon start life as eggs laid by the parent in gravel beds of rivers along the North Pacific coast. The eggs hatch 1 to 4 months later, and the resulting plankton-eating larvae grow quickly. During either their first or second year, juvenile salmon swim down the river to the ocean. By the time they reach the ocean, they are small but voracious predatory fishes. The salmon grow into adults and migrate across the Pacific Ocean to their selected feeding grounds, which range throughout the North Pacific Ocean. After remaining in the ocean up to about 5 years, they return to their home streams to breed and then die.

Salmon usually lay only several thousand eggs, far fewer than many other fish species. However, the eggs and larvae have relatively low mortality because they are protected from the abundant ocean predators through the spawning migration. Hence, the features of the salmon reproductive cycle are comparatively low egg and larval mortality, a single relatively early reproductive cycle followed by death, enormous energy costs of spawning migration, and production of a relatively small number of eggs in a favorable habitat. These features constitute a stable and successful reproductive strategy, as described earlier.

Almost all salmon return to the stream in which they hatched. They are thought to achieve this feat by "smelling" the distinctive chemical compositions of their home streams. However, contrary to popular misconception, not all salmon successfully migrate back to their home streams. As many as 10% to 20% are known to lose their way and migrate to streams other than their original home. Such "mistakes" are probably important to the maintenance of genetic diversity in the species and to the recolonization of streams where catastrophic events may have destroyed the spawning population.

The longest known ocean migrations are those of sea turtles and marine mammals, particularly the California gray whale. Sea turtles range far and wide throughout the oceans and return with great reliability to the beaches where they hatched.

All **baleen** whales, which, with the exception of the gray whale, are filter feeders, migrate seasonally. They feed in plankton-rich, high-latitude waters during spring and summer, and they return to warm tropical waters to breed in winter. California gray whales are unique among baleen whales because they feed by sifting sediment to eat small sediment-dwelling crustaceans called **amphipods** and they undertake one of the longest known seasonal migrations. In summer, they feed in the Bering and Chukchi Seas (**Fig. 14-31a**). When winter approaches, they migrate south along the North American coast, generally staying within sight of land. Their migration takes them some 11,000 km south to the shallow coastal **lagoons** of Baja California, where food is extremely limited and where adults must live off the fat reserves built up during summer. In the warm lagoon water, gray whales breed and females give birth 1 year later. Although young whales weigh about 2 tonnes, they have relatively thin layers of **blubber**, the fat that protects these warm-blooded mammals from losing body heat. If they were born in colder waters where the amphipod food supply is abundant, their mortality rate would be very high. The annual 22,000-km migration is therefore undertaken to ensure both adequate food for adults in summer and the high survival rate of offspring necessary to this species, which, like all other mammals, produces few young during its lifetime. Other whales, such as the humpback, also make long annual migrations (**Fig. 14-31b**).

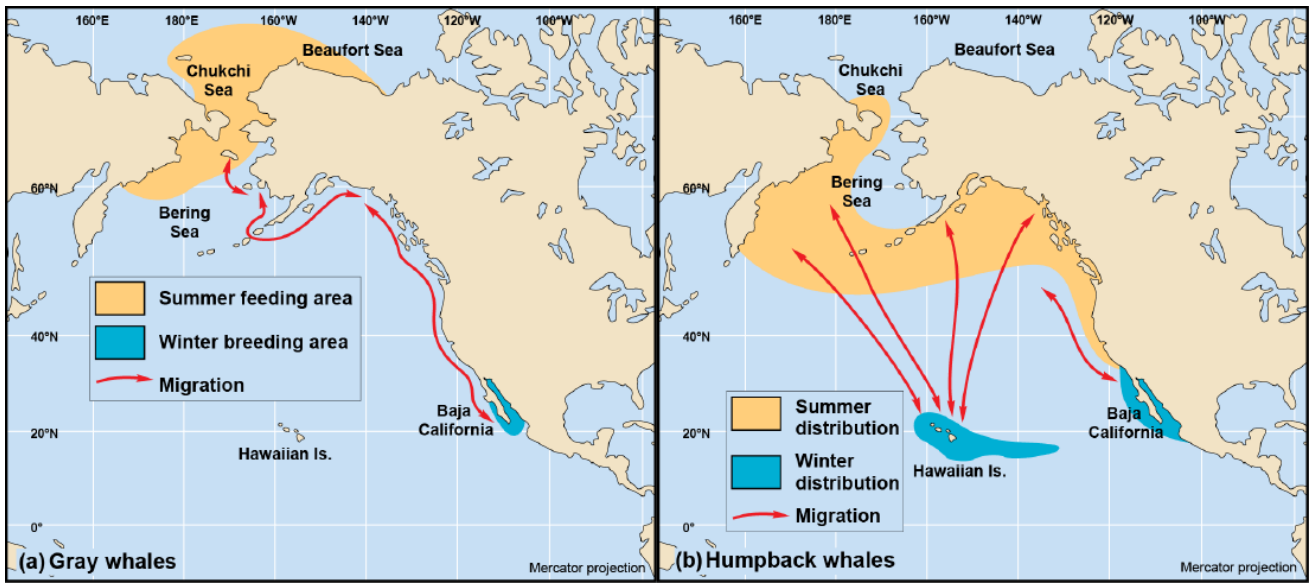


Figure 14-31. Some whale migrations. (a) California gray whales migrate annually between feeding areas in the Beaufort and Chukchi Seas and breeding areas off the coast of Baja California. (b) Humpback whales feed in the far North Pacific Ocean and Bering Sea and migrate to breeding areas around Hawaii and Baja California.

This page titled 14.6: Reproduction is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.7: Associations

Until relatively recently, competition was viewed as the dominant force shaping marine ecosystems. According to this view, ecosystems sustain a number of species, each of which competes with all others for food and living space. However, we now know that many species enter into associations with other species as an essential part of their life cycles and that these associations are critically important to the functioning of marine ecosystems and species evolution.

Associations between species, called **symbiosis**, can take a bewildering variety of forms but are of three basic types: **parasitism**, whereby one species benefits from the association and the other is disadvantaged; mutualism, whereby both species (or all species if more than two species are involved) benefit from the association; and commensalism, whereby one species benefits and the other does not benefit but suffers no disadvantage from the association. It is often difficult to determine what, if any, advantage or disadvantage the partners receive in a given association. Many microbial species, including bacteria and archaea, are associates of other larger species but little is yet known about all but a very few of such associations so the examples of associations in this chapter are but a tiny fraction of the very large number that exist in marine ecosystems. Indeed it is likely that all non-microbial species live with a variety of associated microbial species.

A parasite lives in or on, or frequently visits, another organism (the host) and feeds on the host's tissues or steals the host's food, thereby disadvantaging, without destroying, the host. A tremendous variety of parasitic species are present in the oceans, including marine worms, crustaceans, and snails. The majority are small and most live inside their host's body. Some, however, such as the fish and sea star parasites shown in **Figure 14-32**, are partly or fully external to their hosts. Fish parasites are abundant. As scuba divers can observe on a coral reef, several species of small fishes and shrimp make their living by eating parasites off larger fishes' bodies (**Fig. 14-17e**). The larger fishes seek out the cleaner, which often establishes a cleaning station to which its customers return for periodic service.

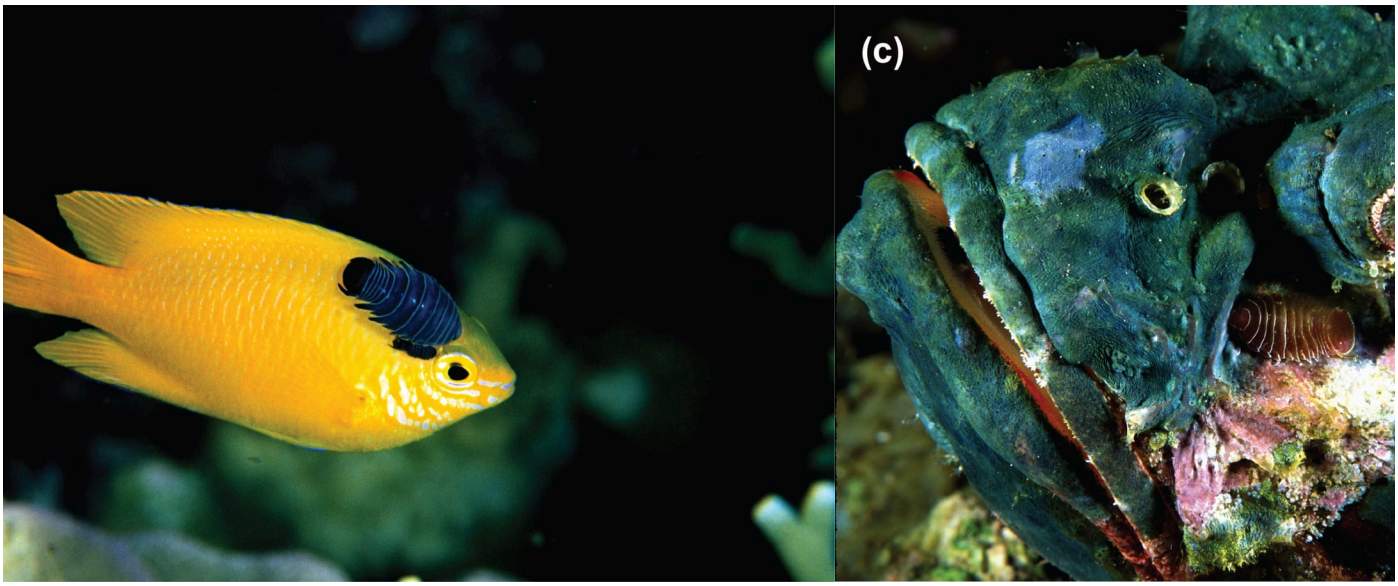


Figure 14-32. Parasites of fishes. (a) This whip goby (*Bryaninops yongei*, Indonesia) has a large parasite, a copepod crustacean, on its body. (b) A large black isopod parasite (a crustacean of the family Cymothoidae commonly known as a "fish doctor") on a lemon damsel (*Pomacentrus moluccensis*, Papua New Guinea). There are two parasites. Look carefully and you might be able to see the much smaller male just under the large female. (c) Fish doctor parasites are common on many species of fishes, including this flasher scorpionfish (*Scorpaenopsis macrochir*, Indonesia). (d) This parasitic snail (*Thyca crystallina*, Indonesia) is apparently found only on this one sea star species (*Linckia laevigata*).

Because they live either in or on another organism, many parasitic species do not need sensory, locomotory, or skeletal organs. Consequently, many have become degenerate and lack such organs. Many species are little more than a digestive and reproductive system. The reproductive system is generally large because the parasite must normally produce enormous numbers of larvae so that a few of these larvae can encounter suitable hosts. Host location is a major problem for parasites, many of which are parasitic on only one host species. Accordingly, parasite life cycles are often very complex, with one or more intermediate hosts (**Fig. 14-33**).

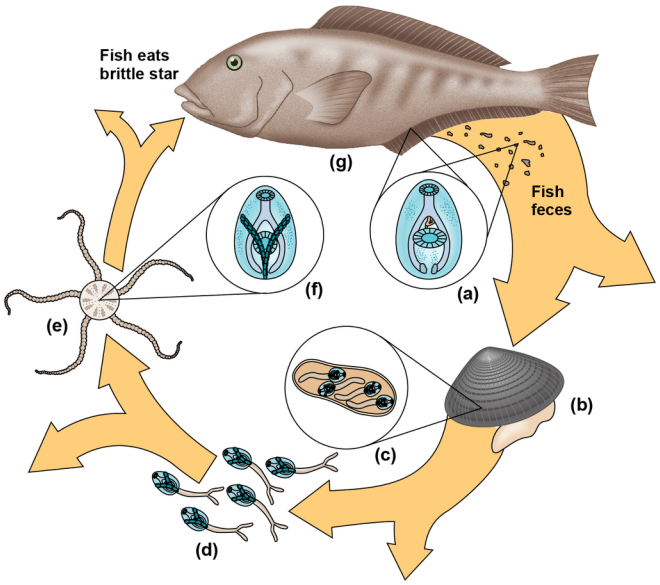


Figure 14-33. The life cycles of parasites are often complex. The adult fish fluke lives in the gut of its fish host (a). It produces very large numbers of eggs that are discharged with the host's feces and hatch into a juvenile stage called a "miracidium." Although most of the miracidia do not survive, a very small number enter a bivalve mollusk (b), which acts as an intermediate host for the parasite. While in the bivalve, the miracidium develops into a larval form (c), which in turn develops a large number of another life stage, called a "cercaria" (d). The cercariae are released to the water, and a very small number of these cercariae eventually enter a second intermediate host, a brittle star (e). Inside the brittle star, the cercariae develop into yet another life stage, the metacercaria (f). If this intermediate host is eaten by a fish (g), the metacercaria develops into an adult fish fluke parasite. At each stage in its life cycle, only an extremely small fraction of the individuals survive to reach the next host. This is true because many parasites, such as the fish fluke, can use only a single species or only a very small number of species as host for one or more of their life stages.

Commensal or mutual associations can provide various benefits to the cooperating species. These include ready availability of food, avoidance of predators or parasites, easy transportation, and suitable living surfaces.

The association between anemonefishes and their anemones (**Fig. 14-20**) has major benefits for the anemonefishes. Hiding in the anemone's stinging tentacles protects an anemonefish from potential predators. In addition, laying eggs under the anemone where they are protected from predators reduces the numbers of eggs needed. Anemonefishes may use the anemone for food in times of need, eating mucus, food scraps, and even anemone tentacles. They may also benefit by having external parasites stung to death and removed by the anemone.

Because many anemones are found without anemonefishes but anemonefishes are never found without an anemone, the benefits of the association to the anemone are less obvious. The anemone does benefit by being cleared of dead tissue, mucus, food wastes, and parasites by the anemonefish. The anemonefish also protect the anemone from its few predators. Without the anemonefish, the anemone could be consumed by certain butterflyfishes and might have to defend its delicate tentacles from such predators by closing up if the anemonefish were not present. Because the anemone cannot feed when closed, the freedom to remain open when predators are nearby is undoubtedly a benefit of the anemonefish association. In some cases, the anemone may benefit by being fed by the anemonefish, which, on occasion, have been seen to carry food to the anemone.

Many other associations also involve a mobile species that uses a nonmobile or less mobile host for protection and food. Examples are associations in which a variety of crabs (**Figs. 14-34a-e**), small fishes (**Figs. 14-34f,g**), shrimp (**Figs. 14-34h-q, 14-35e**), scale worms (**Fig. 14-34r**), and mollusks (**Fig. 14-34s**) live on crinoids (**Fig. 14-16a,b**), corals (**Fig. 14-16g,h,j,k**), sea stars (**Fig. 14-16l**), and sea fans (**Fig. 14-11b**). In most of these associations, the small lodger is well camouflaged or hidden within the host. The lodger eats mucus, dead tissue, parasites, the host's body parts or food supply, or any combination thereof. Generally, the host benefits by being cleaned of mucus, dead tissue, and/or parasites.

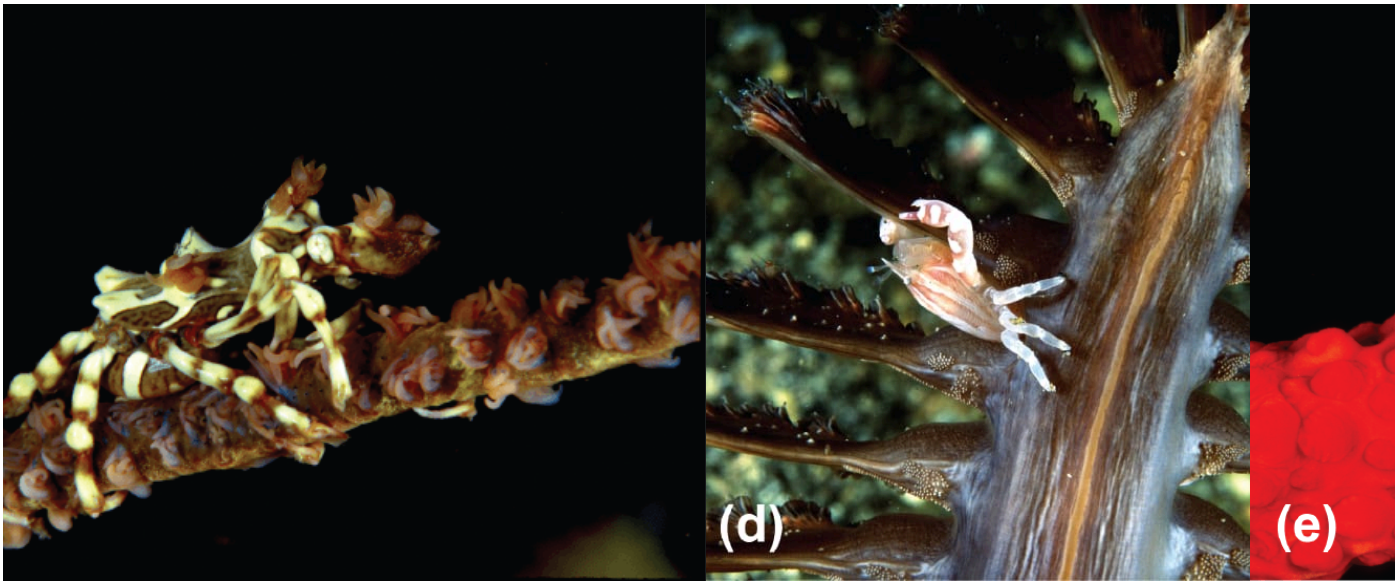


Figure 14-34. Examples of the many associations between species. (a) The arrowhead crab (*Huenia heraldica*, Papua New Guinea) shown here not only blends into the algae (*Halimeda sp.*) that it lives on, but usually carries pieces of the algae around with it, as this one is doing, to further enhance its camouflage. (b) Many small crabs live in or on a larger invertebrate for camouflage, to hide and sometimes to steal the host's food or feed on the host itself. Many species, such as the Xenia swimming crab (*Caphyra sp.*) shown here on a soft coral, are poorly studied and difficult to identify (Indonesia). (c) This tiny spider crab (*Xenocarcinus tuberculatus*, Indonesia) lives on a coral whip, usually just one pair per whip. (d) The porcelain crab (probably *Porcellanella triloba*, Indonesia) lives on a sea pen, even when the sea pen withdraws into the sand for protection. (e) Another tiny porcelain crab (probably *Porcellanella sp.*, Indonesia) lives on a whip coral. (f) The weed cardinalfish (*Foa brachygramma*, Papua New Guinea) spends its life hidden in weeds or, as in this case, in a crinoid where it cannot be seen or reached by predators. (g) The crinoid clingfish (*Discotrema crinophila*, Papua New Guinea) lives and hides in a crinoid and usually has a body color that matches the crinoid.

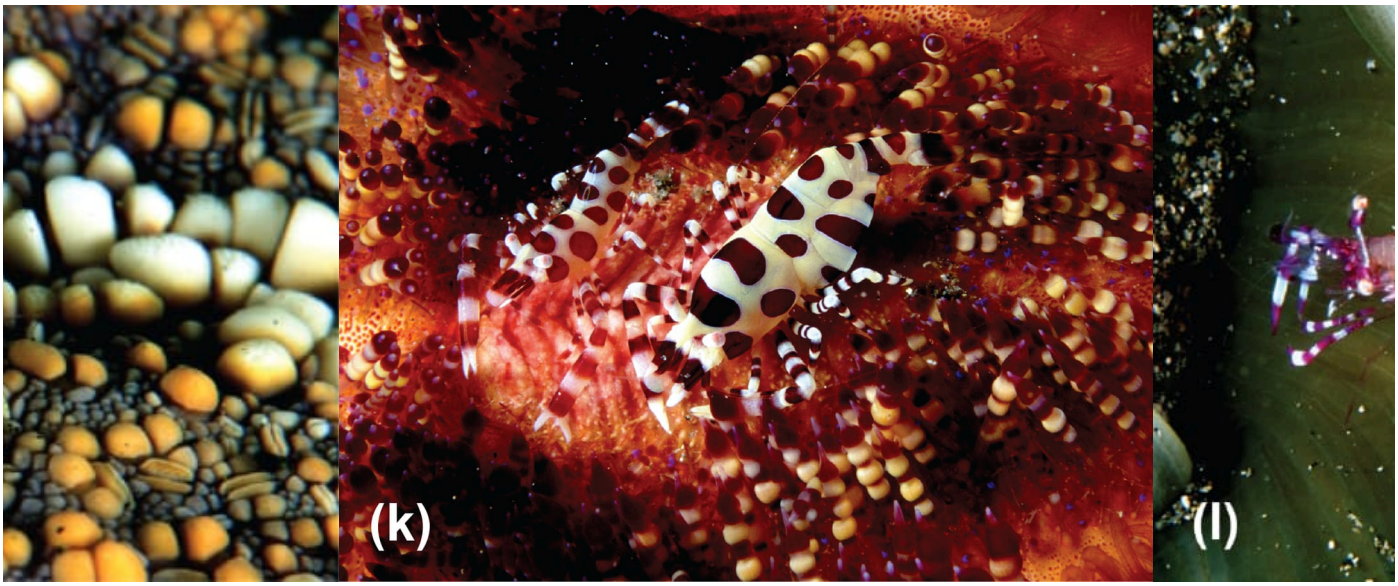


Figure 14-34. Examples of the many associations between species. (h) Emperor shrimp (*Periclimenes imperator*, Papua New Guinea) like the pair shown here live on a variety of hosts, including nudibranchs and sea cucumbers. (i) Sea star shrimp (*Periclimenes soror*, Indonesia), live on sea stars and match their body color to blend in with the host's body. (j) This sea star commensal shrimp (*Zenopontonia noverca*, Indonesia) is partially transparent, and with its spots it blends in so well with its background when it is on some parts of the sea star's body that we had to wait for the shrimp to move before getting a photograph in which the shrimp is more visible. (k) A pair of Coleman's shrimp (*Periclimenes colemani*, Papua New Guinea) living protected on their venomous fire urchin host (*Asthenosoma varium*). (l) Many species of shrimp, like this one (*Ancylomenes cf. venustus*, Indonesia), live in anemones. (m) Many small commensal shrimp species, including this one (*Periclimenes cf. tosaensis*, Indonesia), have mostly transparent bodies, so they can blend in with several different hosts. This individual is living in a bubble coral, and the eggs that it's carrying can be seen as a pink egg mass through its transparent body. (n) This species of coral shrimp (*Vir philippinensis*, Indonesia) lives only in bubble corals.

Figure 14-34. Examples of the many associations between species. (o) A humpbacked shrimp (*Hippolyte commensalis*, Indonesia) can hardly be seen when it withdraws into its soft coral host. (p) This tiny urchin shrimp (*Gnathophylloides mineri*, Indonesia) is also virtually impossible to see when it hunkers down among the spines of its urchin host where it normally hides. (q) This saw blade shrimp (*Tozeuma armatum*, Indonesia) has an unusually elongated body so that it can lie flat against the strands of black coral on which it lives. (r) Scale worms, including this species (*Gastrolepidia clavigera*, Vanuatu), have flattened bodies and coloration to match their host, usually a sea cucumber, as seen here. (s) This egg cowrie (*Pseudosimnia sp.*, Red Sea) lives on and probably eats soft corals such as the one shown here (*Dendronephthya sp.*).

Even if it is eaten by its partner species, the host may gain. Many species that feed on their host are present as only a single pair on their much larger host. This pair nourishes itself from the host's tissues but eats the host only at a rate that can be replaced by normal growth. The resident pair aggressively protect their host against colonization by others of their species, parasites, and other species that would also consume the host. Thus, although it loses some tissue to the associate species, the host is better protected.

Many associations involve a normally nonmobile species that lives on a mobile host. Examples are the associations formed by the decorator crab (**Fig. 14-16f**) and the anemone crab (**Fig. 14-35a**). The decorator crab gains camouflage and protection from its covering of algae, hydroids, and sponges, which are unpalatable and rarely attacked by predators. The immobile associates may gain by being transported through the water, which increases their chances of obtaining food or dissolved nutrients, and by using the host as a substrate on which to grow. In similar associations, the sponge may be an indifferent partner removed from its home on the reef and carried on the crab's back by a pair of specially adapted legs. The anemone **hermit crab** is well protected by the adopted mollusk shell in which it lives, but it is further protected by the stinging anemones that it attaches to the shell (**Fig. 14-35a**). The anemone gains by being transported through areas where food can be obtained, including scraps of food released as the hermit crab feeds by tearing apart its prey. This association, like many others, is highly specific. The association usually involves the same species of anemone and hermit crab, and the shell of the same species of mollusk. Some species of decorator crabs also cover themselves with sponges (**Fig. 14-16f**) or carry an urchin on their backs (**Fig. 14-35b**) to afford them protection.

Figure 14-35. Additional examples of associations. (a) This anemone hermit crab (*Dardanus sp.*, Papua New Guinea) has found a strong gastropod mollusk shell to live in and has attached several anemones on the shell to further discourage potential predators. (b) This crab (*Dorippe frascone*, Indonesia) has a pair of legs that are modified to carry an urchin, in this case *Astropyga radiata*, on its back. The urchin's toxic spines protect the crab from predators. (c) This tiger pistol shrimp (*Alpheus bellulus*, Indonesia) lives in a single burrow with the yellow shrimp goby (*Cryptocentrus cinctus*). The shrimp digs and maintains the burrow. The goby warns the shrimp, which is blind, when a predator approaches, and they both disappear into the burrow. (d) A giant clam (*Tridacna gigas*, Papua New Guinea). (e) This small pistol shrimp (*Synalpheus sp.*, Papua New Guinea) lives in its host crinoid.

The remora (**Fig. 14-26g**) is an example of a species that is mobile but hitches a ride on a larger swimming animal. In some cases, remoras gain by stealing food from their host. However, the major advantage to the remoras is probably transportation, since they often ride on manta rays, which feed on plankton. It is not clear whether sharks or manta rays gain anything from a remora's presence, but these hosts apparently are not harmed by carrying the passenger.

The association between gobies and shrimp is clearly beneficial to both partners. A goby and a shrimp live together in a common sand burrow, where they can retreat to safety from predators. The shrimp is blind, and the fish is incapable of digging its own burrow. The shrimp therefore digs and maintains the common burrow, but it is vulnerable to predators when it pushes excavated sand out of the hole. The goby repays the shrimp for its digging by sitting at the burrow entrance watching for predators (**Fig. 14-35c**). With one of its long, sensitive antennae, the shrimp maintains contact with the goby's caudal fin. If a predator approaches, the goby wiggles its fin to warn the shrimp, which immediately withdraws into the burrow. If danger increases, the goby itself darts into the burrow.

The few associations described here are all relatively easy to observe and study, but most associations are much more subtle. For example, coordinated feeding by two or more fish species that benefits both species does occur but may be infrequent and difficult to observe. In addition, many associations are difficult to study because one species is well hidden inside the other or because one or both partners are microscopically small. For example, the giant clam, *Tridacna* (**Fig. 14-35d**), has a mutually beneficial association with **zooxanthellae**. The algae live inside the soft **mantle** tissues of the clam and give this part of the clam its often brilliant coloration. The algae benefit from a secure location within the clam's tissues, where they also receive a steady supply of carbon dioxide and nitrogenous waste products from the clam, which the algae use for photosynthesis. There, the algal cells can obtain needed sunlight when the clam is open, and they are protected from settling of other benthic organisms that might otherwise overgrow them. In addition, the clam closes when threatened, protecting both itself and its algae. The clam benefits from the association by using the algae's by-products as a supplemental food supply. Thus, the algae are essentially a carefully cultivated garden that supplements the clam's food so that it does not have to filter extremely large volumes of water. This feeding arrangement is very energy-efficient and probably accounts for the giant clam's ability to grow to its huge 1- to 2-m size. Hard corals have a similar association with zooxanthellae (**Chap. 15**).

This page titled 14.7: Associations is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.8: Communication and Navigation

The majority of species must be able to communicate at least on a rudimentary level—for example, to provide an indication of sex and attract a mate. Most predators cannot rely on finding their prey by chance but must actively seek them out. Many marine species must navigate from place to place in migrations or to find suitable habitat, food, and mates. Humans and most other terrestrial animals use their five senses—vision, smell, hearing, touch, and taste—to perform these functions, with vision being the predominant sense. Light is absorbed and **scattered** in seawater, and most of the ocean is dark. In addition, when light is present, its path is distorted by **refraction**. Therefore, marine species rely much less on vision than most terrestrial species do.

In the marine environment, hearing, chemical sensing that is equivalent to taste or smell, and touch have assumed much greater importance. Marine organisms have also developed other senses, including the ability to sense electrical fields and probably the Earth's magnetic field. Our knowledge of these unusual and different senses and even of how marine organisms use their greatly enhanced hearing, chemical sensing, and touch is still extremely limited.

Although vision is much less useful in the oceans than on land, it is still widely used by many marine species, and many adaptations of vision are found in species that live in the ocean environment. The most important adaptations are greatly enlarged eyes or multiple-lens eyes that can see in very low light levels and distinguish subtle changes in light intensity (**Fig. 14-18a,b**). Most marine species do not see sharply focused images, and many, particularly some marine invertebrates, have bizarre-looking eyes (e.g., **Figs. 14-18a, 14-35a**). Some species have light-sensing organs that cannot be readily identified as eyes. For example, the giant clam (**Fig. 14-35d**) and many other benthic invertebrates are able to sense very small variations in light intensity, which enables them to withdraw into their shells, tubes, or other protective environments when predators approach.

Chemical sensing is very important in the marine environment, but very little is known about it. For example, anadromous and catadromous species are thought to use a well-developed chemical sensing capability to detect and navigate to their home streams. They apparently can sense small differences in the composition of chemicals present in very low concentrations in these home streams. Sharks also appear to have an extremely keen sense for chemicals. They are attracted to blood even when they are kilometers away from its source and the concentration of blood chemicals is exceedingly small. Adventurous scuba divers deliberately attract sharks by dumping blood and shredded fish into the water.

Sound travels through seawater with relatively little loss of intensity. Consequently, sound sensing, or hearing, is well developed in many marine species, and sound is used extensively for communication. Many species, particularly **cetaceans** such as whales and dolphins, use sound not only for communication, but also to locate objects, including prey. Most cetaceans can generate sound pulses that bounce off objects in their path. They focus and process the resulting echoes, and thus are able to "see" the objects. This **echolocation** ability is extremely sophisticated and capable of high sensitivity and precision in at least some species. For example, the dolphin echolocates by emitting low-**frequency** clicks to scan objects at distances up to hundreds of meters. As it nears the object, the dolphin uses higher-frequency clicks to get a more detailed "picture." With their echolocation abilities, dolphins can identify specific objects that differ only slightly in shape, size, thickness, or material composition. Dolphins and other cetaceans can produce and sense an extraordinary range of sound frequencies far beyond human hearing at both high and low frequencies.

We do not know exactly how cetaceans produce sound pulses or how they focus and sense returning echoes. However, some species are believed to create the sounds by moving air within hollow bony structures in the head. The bulbous brow of most cetaceans (**Fig. 12-25**) contains oil- or fat-filled structures believed to be responsible for focusing sound pulses when they are generated. Returning echoes appear to be captured by the fat-filled lower jaw, which conducts the sound directly to the inner ear, allowing for precise navigation and hunting in total darkness.

In addition to cetaceans, many other marine species produce sounds thought to be used primarily for communication. For example, certain fishes are known to generate sound by using their swim bladders, and they may sense other fishes' sounds or echoes in the same way. Scuba divers, if they listen carefully, can hear a continuous cacophony of noises in the oceans that sounds like the static on a radio. Some species use sound for more sinister purposes. The pistol (or snapping) shrimp (**Fig. 14-35e**) has one very large, overdeveloped claw, which, when snapped closed, creates a loud noise that can be heard across a room if the shrimp is in an aquarium. The concussion of the shrimp's snapping sounds stuns its invertebrate or small fish prey, sometimes into unconsciousness. The stunned prey is then easy game for the shrimp. Other species, notably killer whales, also may generate and use intense sound pulses to stun or even to kill their prey.

Fishes can sense small changes in pressure through sensory organs located in a row along each side of their bodies that form

what is called the “lateral line.” Lateral-line sensors are apparently used by schooling fish to sense motions of their neighbors. Together with visual clues, these sensors may explain how schooling fish can turn in unison. Sharks also may be able to sense pressure variations, since they are attracted from long distances by the thrashing of a sick or dying fish. The low-frequency pressure variations caused by such a fish are transmitted through the water as very low-frequency sound waves, and the shark’s ability to sense such pressure fluctuations may simply be an extension of its “hearing.”

Evidence is steadily accumulating that many marine species are able to sense magnetic fields, including the Earth’s magnetic field. This ability may be vital for migrating species such as sea turtles. In addition, some marine species are able to generate and sense electrical fields.

This page titled 14.8: Communication and Navigation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.9: CHAPTER SUMMARY

Ecological Requirements.

The fundamental needs of all species are a place to live, food, safety from predators, and successful reproduction.

Habitat.

The water column, seafloor, and sediments are three fundamentally different habitats. Pelagic organisms must avoid sinking, and most pelagic animals must actively seek food. Benthic epifauna save energy because they do not have to control buoyancy, but they have difficulty avoiding predators and, in most areas, must rely on detrital food that rains down from above or on predation. Benthic infauna have some protection from predators and do not require buoyancy control, but because food in sediments is limited, they must expend energy to move or dig through the sediment, and they must obtain oxygen that is depleted in most sediments below the upper few centimeters. The surface microlayer, intertidal zone, and hydrothermal vents are special habitats that present unique problems and opportunities.

Feeding.

Suspension feeders eat particles suspended in the water column, including living phytoplankton and zooplankton, and detritus. Many suspension feeders are filter feeders that strain water through a mesh-like structure to capture food. Other suspension feeders use mucus to capture food particles, or they are able to grab or grasp particles with arm-like appendages. Many suspension feeders increase their feeding efficiency by pumping water through their collection apparatus, moving the apparatus through the water, or placing it in currents. Many suspension feeders are present among both pelagic fauna and benthic epifauna, and a few benthic infaunal species suspension-feed using a feeding apparatus that they extend above the sediment surface or by pumping water through their tubes or burrows.

Surface grazers consume algae, small sedentary animals, and detritus from the seafloor. Food is abundant only where the seafloor is within the photic zone, where the substrate may be covered with benthic microalgae, macroalgae, and sponges, tunicates, and other animals. In the deep sea, surface grazers eat primarily detritus or bacteria. Surface grazers may feed by using specially adapted mouths on the underside of the body that can either rasp off food organisms or grasp food as they sift through the surface sediment. Others suck off and sift surface sediment to obtain food particles. Some surface grazers also suspension-feed or hunt.

Most deposit feeders move through the sediment, taking sediment into the gut, where they digest organic matter from detritus or coatings. In sediments with anoxic pore waters, deposit feeders must obtain oxygen from the water above or remain in the oxygenated upper sediment layer.

Hunting and Defense.

Speed is used by both hunter and hunted; the hunter chases, and the potential prey seeks to escape. Prolonged movement at high speed in the oceans is energy-intensive because of water resistance. Consequently, all but a few of the largest ocean animals use speed in short bursts rather than in prolonged chases. Other strategies are used to get close to the prey, and then a short burst of speed over the remaining small distance is used to capture the prey. In defense, speed generally is used for immediate escape, followed by other strategies to ensure ultimate safety.

Many marine hunters use lures to attract prey. The lure is typically an appendage on or near the mouth that resembles a bite-sized meal to the prey. Some species use false markings such as eyespots to lure predators into attacking in the wrong direction or at the wrong place as the potential prey speeds off in the opposite direction.

Hunters use camouflage to lie in wait unseen by their prey, or they mimic species that clean other species so that they can approach unsuspecting prey. Many species use camouflage to conceal themselves from predators, or they mimic fiercer or poisonous species to deter predators. Concealment in sediment, within other species, or in holes in reefs and rocks is a much-used defensive strategy. Many species conceal themselves by day and emerge to hunt only at night. Hunters also conceal

themselves to ambush prey.

Spines and armored bodies are used by many marine species as defensive mechanisms, and in some cases as weapons. Hunters use crushing claws and jaws to overcome the armor of prey species. Poisons are used by hunters that inject venom into their victims to stun or kill them, and by potential prey species that inject venom into attacking species or make themselves toxic to eat. Many species that are poisonous to predators are brightly colored to advertise their toxicity.

Schooling or cooperation can be used to overcome defenses of a confused prey species, and to confuse predators or frighten them off by creating the appearance of a target much larger than any one individual.

Selected Adaptations in Fishes.

Fish body shapes differ by species to reflect swimming habits. Species that swim continuously have teardrop shapes—a compromise that minimizes total drag, including surface, form, and turbulent drag. Fishes that make fast turns are flattened so that the flat sides can act like a rudder. Fishes that specialize in a quick burst of speed are thickened in the middle by the needed musculature.

Fin shapes also vary with swimming habits. The caudal fin is rounded in fast-maneuvering and -accelerating species, truncated and forked in faster-swimming species, and lunated in the fast, continuous swimmers. Heterocercal caudal fins and pectoral fins in sharks provide lift to maintain buoyancy. Fins are also adapted for special purposes, such as “walking,” defense and venom injection, attachment to substrates or hosts, and even gliding through the air.

Unlike invertebrates, bony fishes have internal body fluids that have lower salinity than seawater, and therefore they must prevent the continual loss of water by osmosis. Marine fishes osmoregulate by drinking seawater and excreting salts through a special gland. Freshwater fishes gain water by osmosis and osmoregulate by producing large volumes of urine. Some species that live where salinity is variable must be able to osmoregulate in both ways.

Most fish species maintain buoyancy by producing and storing low-density oils or by filling a swim bladder with gas. Gas must be released from or added to the swim bladder as the fish changes depth. Oils are favored in abyssal species because gas compressed at depths below about 7000 m is more dense than oil. Some species have no swim bladders, must swim continuously to avoid sinking, and have fins and bodies adapted to provide lift.

Reproduction.

Many species reproduce sexually, either by the direct transfer of sperm from male to female or, more often, by the release of sperm, or of eggs and sperm, to the water column for fertilization. Mating congregations and synchronous spawning by all males and females of a species maximize the probability of egg fertilization, improve egg and larval survival, and maximize genetic diversity.

Because mates may be difficult to locate, many species are hermaphroditic, and some can even function as both sexes at the same time. Asexual reproduction is common in algae and in colonial invertebrates.

To ensure egg fertilization and survival, either very large numbers of eggs must be produced and released or smaller numbers of eggs must be defended against predators. Many fishes and invertebrates release large numbers of eggs to the water column, where they hatch to become meroplankton. Other species lay eggs on the substrate and protect them until they hatch, and still others retain fertilized eggs in or on their bodies until they hatch. Spawning often occurs at times and locations that match the availability of suitable food for larvae. Species with high adult mortality and relatively low larval mortality tend to reproduce early and only once. Species with high egg and larval mortality tend to mature late and spawn more than once.

Many species migrate during their life cycle. Most often they move from adult or seasonal feeding areas to mating or spawning areas. Anadromous and catadromous fishes migrate between freshwater and the ocean.

Associations.

Symbiosis between species is very common in the marine environment. In some associations, one species benefits and the other is disadvantaged (parasitism). In other associations, both species benefit (mutualism), or one species gains and the other is neither advantaged nor disadvantaged (commensalism). Benefits gained from associations include food, living space, camouflage,

protection from predators, removal of parasites, and transportation.

Communication and Navigation.

Many marine species have and use vision, but other senses are often more important. These senses are used to locate prey, mates, and predators and for navigation. Sensing of dissolved chemicals, akin to smell or taste, is especially well developed in many species. Certain species generate sounds that are used for echolocation, to stun prey, or to communicate. Many species can sense small changes in pressure, electrical fields, and probably the Earth's magnetic field.

14.9.1: Key Terms

14.9.2: Chapter Questions

14.9.3: References

This page titled 14.9: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.9.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- algae
- amphipods
- anadromous
- anemones
- annelid
- anoxic
- aphotic zone
- archaea
- aspect ratio
- bacteria
- baleen
- barnacles
- beaches
- benthic
- benthos
- biochemical
- bioluminescent
- biomass
- biota
- bioturbation
- bloom
- buoyancy
- carnivores
- catadromous
- cetaceans
- chemosynthetic communities
- climate
- clones
- cnidarians
- contaminant
- continental shelf
- copepods
- corals
- crustaceans
- ctenophores
- currents
- decomposer
- density
- deposit feeders
- detritus
- diatoms
- diel
- diffusion
- eastern boundary current
- ebbs
- echolocation
- ecosystems
- environment
- epifauna
- erosion

- estuarine
- euphausiids
- excrete
- excurrent
- fauna
- fecal pellets
- filter feeders
- flagellates
- frequency
- friction
- fungi
- genetic
- gills
- gobies
- gravity
- grazed
- habitats
- hard parts
- herbivorous
- hermit crab
- herring
- high-tide line
- hydroids
- hydrothermal vents
- incubation
- incurrent
- infauna
- intertidal zone
- invertebrates
- jellies
- lagoons
- larvae
- latitudes
- macroalgae
- mantle
- marine mammals
- meroplankton
- metamorphose
- microalgae
- mixed tides
- mollusks
- neap tides
- nekton
- nudibranchs
- nutrients
- osmoregulation
- osmotic pressure
- ovoviviparous
- parasites
- parasitism

- pelagic
- photic zone
- photosynthetic
- phototrophs
- phototrophy
- planktonic
- polyps

- pore waters
- primary productivity
- protists
- pteropods
- reefs
- refraction
- respiration
- resuspend
- salinity
- salps
- scatter
- scavenging
- schooling
- scuba
- sea cucumbers
- sea fan
- sea grass
- sea stars
- sea urchins
- sediment
- sedimentation rate
- spawning
- species
- sponges
- spring tides
- surface microlayer
- surface tension
- suspension feeders
- symbiosis
- tidal ranges
- tide pools
- tides
- tunicate
- upwelling
- vertebrates
- viruses
- viscosity
- viviparious
- wavelengths
- zooplankton
- zooxanthellae

This page titled 14.9.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.9.2: Chapter Questions

Study Questions

1. What are the four fundamental challenges that any species must meet to survive?
2. What are the advantages and disadvantages of the benthic infaunal habitat in comparison with the benthic epifaunal and pelagic habitats? Describe how these factors change with depth and substrate type.
3. How do suspension feeders obtain food, and why are there so many suspension feeders in the oceans?
4. Why do most deposit feeders live in only the upper few tens of centimeters of the sediment?
5. Why are some pelagic predators not able to swim fast? What other capabilities do they use to capture prey?
6. Why do many fish species congregate in schools?
7. What are the physical characteristics of a fish species that swims continuously at high speeds, and of a species that waits on the seafloor for prey to swim by?
8. What are the three ways in which different fish species maintain their buoyancy? Which of these is more common in abyssal species, and why?
9. For what reasons do many species migrate at some stage in their life cycles?
10. What aspects of their environment might marine species be able to detect with senses other than sight?

Critical Thinking Questions

1. Most marine species are adapted to eat only food of a certain type, such as suspended particles, organic coatings on sediment particles, phytoplankton cells of a certain size range, zooplankton, or small fishes. Discuss the possible reasons why most species do not develop (or do not retain) the ability to feed on many different types of food.
2. Nudibranchs are snails that have no shell. They generally live on reefs and feed on algae, sponges, or corals. There are many holes in reefs that make excellent hiding places. Why do nudibranchs use poisons and bright coloration to warn off predators rather than just using the many available hiding places in the reef?
3. Discuss the advantages and disadvantages of a spawning strategy in which large numbers of eggs and sperm are released simultaneously to the water column.
   (a) Explain why you might expect this strategy to be more common in coral reef communities than in the polar regions.
   (b) Are there any terrestrial species that employ a similar strategy? If so, what are they, and how does their strategy differ from the similar strategies of ocean species?
4. Describe the advantages and disadvantages that are experienced by species that live in association with other species.
   (a) Explain why species associations are so common in the oceans.
   (b) Do species live in associations in the terrestrial environment? If so, describe some of these associations.

This page titled 14.9.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

14.9.3: References

Figure 14-1a – Segar, D., and E. Stamman Segar, 1997: Flamboyant cuttlefish (*Metasepia pfefferi*).

Figure 14-1b – Segar, D., and E. Stamman Segar, 1997: Flamboyant cuttlefish (*Metasepia pfefferi*).

Figure 14-1c – Segar, D., and E. Stamman Segar, 1997: Flamboyant cuttlefish (*Metasepia pfefferi*).

Figure 14-1d – Segar, D., and E. Stamman Segar, 1997: Flamboyant cuttlefish (*Metasepia pfefferi*).

Figure 14-1e – Segar, D., and E. Stamman Segar, 1997: Young flamboyant cuttlefish (*Metasepia pfefferi*).

Figure 14-2a – Uwe Kils, 2011: Antarctic krill (*Euphausia superba*). Accessed 20 March 2026, Source.

Figure 14-2b – Segar, D., and E. Stamman Segar, 1997: Banded coral shrimp (*Stenopus hispidus*, Papua New Guinea).

Figure 14-2c – Segar, D., and E. Stamman Segar, 1997: Harlequin shrimp (*Hymenocera elegans*, Papua New Guinea).

Figure 14-3a – Hopcroft, R., 2015: *Limacina helicina*, UAF/NOAA. Accessed 20 March 2026, Source.

Figure 14-3b – Segar, D., and E. Stamman Segar, 1997: Marine gastropod, a harp shell (*Harpa major*, Indonesia).

Figure 14-4b – Segar, D., and E. Stamman Segar, 1997: Individual members of the salp colony.

Figure 14-6a – Segar, D., and E. Stamman Segar, 1997: A sea pen (order Pennatulacea, Papua New Guinea) with its feeding polyps oriented into the current.

Figure 14-6b – Segar, D., and E. Stamman Segar, 1997: A sea pen (order Pennatulacea, Papua New Guinea) with its feeding polyps oriented into the current.

Figure 14-6c – Segar, D., and E. Stamman Segar, 1997: Sand anemone (Papua New Guinea).

Figure 14-6d – Segar, D., and E. Stamman Segar, 1997: Tube worm (family Sabellidae, Papua New Guinea) with its netlike feeding apparatus extended into the water column.

Figure 14-7a – Segar, D., and E. Stamman Segar, 1997: California mussels (*Mytilus californianus*, Monterey, California).

Figure 14-7b – Segar, D., and E. Stamman Segar, 1997: Pacific spiny oyster (*Spondylus varians,* Papua New Guinea).

Figure 14-7c – Segar, D., and E. Stamman Segar, 1997: Colony of barnacles in Monterey, California.

Figure 14-7d – Segar, D., and E. Stamman Segar, 1997: Coral barnacles (order Pyrgomatidae, Indonesia).

Figure 14-8a – Segar, D., and E. Stamman Segar, 1997: A hard coral known as brain coral (*Diploria sp.*, Papua New Guinea).

Figure 14-8b – Segar, D., and E. Stamman Segar, 1997: A hard coral often called staghorn coral (*Acropora sp.*, Papua New Guinea).

Figure 14-8c – Segar, D., and E. Stamman Segar, 1997: Close-up showing polyps of a hard coral species (*Tubastrea sp.*, Fiji) extended to feed at night.

Figure 14-8d – Segar, D., and E. Stamman Segar, 1997: Colony of a soft coral (*Dendronephthya sp.*, Fiji). Dendronephthya occur in many different colors.

Figure 14-8e – Segar, D., and E. Stamman Segar, 1997: Typical gorgonian sea fan (*Melithaea sp.*, Papua New Guinea).

Figure 14-8f – Segar, D., and E. Stamman Segar, 1997: Close-up of a soft coral colony (*Xenia sp.*, Papua New Guinea) showing individual polyps at different stages of feeding.

Figure 14-8g – Segar, D., and E. Stamman Segar, 1997: A small suspension-feeding colonial anemone (*Nemanthus annamensis*, Fiji).

Figure 14-8h – Segar, D., and E. Stamman Segar, 1997: A colonial zooanthid (*Protopalythoa sp.*, Papua New Guinea).

Figure 14-8i – Segar, D., and E. Stamman Segar, 1997: A "magnificent sea anemone" (*Heteractis magnifica*, Fiji), here providing a home for a pink anemonefish (*Amphiprion perideraion*).

Figure 14-9a – Segar, D., and E. Stamman Segar, 1997: Lightbulb tunicate (*Clavelina sp.*, Philippines).

Figure 14-9b – Segar, D., and E. Stamman Segar, 1997: The colonial tunicate species *Didemnum molle* (Papua New Guinea).

Figure 14-9c – Segar, D., and E. Stamman Segar, 1997: Three tall, blue and blue-green *Rhopalaea sp.* and three smaller *Clavelina robusta* (Indonesia) with darker bodies and yellow markings around their incurrent and excurrent openings.

Figure 14-9d – Segar, D., and E. Stamman Segar, 1997: Colonial tunicate that covers and encrusts the reef (Botryllus sp., Indonesia).

Figure 14-10a – Segar, D., and E. Stamman Segar, 1997: A creeping sea cucumber (*Cucumaria sp.*, Philippines) with its suspension-feeding apparatus extended.

Figure 14-10b – Segar, D., and E. Stamman Segar, 1997: A basket star (family Gorgonocephalidae, Papua New Guinea) with its arms extended into the current to feed at night.

Figure 14-10c – Segar, D., and E. Stamman Segar, 1997: A sabellid spiral fan worm (*Protula magnifica*, Indonesia).

Figure 14-10d – Segar, D., and E. Stamman Segar, 1997: A Christmas tree worm (*Spirobranchus giganteus*, Papua New Guinea).

Figure 14-11a – Segar, D., and E. Stamman Segar, 1997: Long-spined sea urchins (*Diadema savignyi*, Papua New Guinea), browsing across the seafloor.

Figure 14-11b – Segar, D., and E. Stamman Segar, 1997: An allied, or egg, cowrie (*Primovula sp.*, Indonesia) on a gorgonian sea fan.

Figure 14-11c – Segar, D., and E. Stamman Segar, 1997: A flabellina nudibranch *(Flabellina rubrolineata*, Indonesia).

Figure 14-11d – Segar, D., and E. Stamman Segar, 1997: Spectacular nudibranch (*Pyllodesmium undulatum*, Papua New Guinea)

Figure 14-12b – Segar, D., and E. Stamman Segar, 1997: Turtle grass (*Thalassia sp.*).

Figure 14-12c – Segar, D., and E. Stamman Segar, 1997: *Spartina sp.* (Aruba, Caribbean Sea), form dense beds in many coastal wetlands where the water is brackish.

Figure 14-12d – Segar, D., and E. Stamman Segar, 1997: Green sea turtle (*Chelonia mydas*, Hawaii).

Figure 14-13d – Segar, D., and E. Stamman Segar, 1997: Blackstriped goatfish (*Upeneus tragula*, Indonesia).

Figure 14-14a – Segar, D., and E. Stamman Segar, 1997: Twospot lizardfish (*Synodus binotatus*, Indonesia).

Figure 14-14b – Segar, D., and E. Stamman Segar, 1997: Striated frogfish (*Antennarius striatus*, Indonesia).

Figure 14-14c – Segar, D., and E. Stamman Segar, 1997: Falco, or dwarf, hawkfish (*Cirrhitichthys falco*, Indonesia).

Figure 14-14d – Segar, D., and E. Stamman Segar, 1997: Fire goby or dartfish (*Nemateleotris magnifica*, Papua New Guinea).

Figure 14-14e – Segar, D., and E. Stamman Segar, 1997: Bennett's butterflyfish (*Chaetodon bennetti*, Papua New Guinea).

Figure 14-15b – Adapted from SEFSC Pascagoula Laboratory, 2013: A large painted anglerfish (*Chaunax pictus*). NOAA NMFS SEFSC, Collection of Brandi Noble, accessed 20 March 2026, Source.

Figure 14-16a – Segar, D., and E. Stamman Segar, 1997: Crinoid shrimp (*Periclimenes amboinensis*, Papua New Guinea) on their

host crinoid.

Figure 14-16b – Segar, D., and E. Stamman Segar, 1997: Crinoid shrimp (*Periclimenes amboinensis*, Indonesia) on their host crinoid.

Figure 14-16c – Segar, D., and E. Stamman Segar, 1997: A peacockspoted partially buried in the sand (*Phyllichths punctatus*, Indonesia).

Figure 14-16d – Segar, D., and E. Stamman Segar, 1997: Raggy scorpionfish (*Scorpaenopsis venosa*, Indonesia).

Figure 14-16e – Segar, D., and E. Stamman Segar, 1997: Ghost pipefish (*Solenostomus halimeda*, Papua New Guinea)

Figure 14-16f – Segar, D., and E. Stamman Segar, 1997: Decorator spider crab (*Camposa retusa*, Papua New Guinea).

Figure 14-16g – Segar, D., and E. Stamman Segar, 1997: Commensal spider crab (*Xenocarcinus conicus*, Palau) on a gorgonian coral.

Figure 14-16h – Segar, D., and E. Stamman Segar, 1997: Pygmy (or gorgonian) seahorse (*Hippocampus bargibanti*, Indonesia).

Figure 14-16i – Segar, D., and E. Stamman Segar, 1997: Spotted seahorse (*Hippocampus kuda*) in Papua New Guinea.

Figure 14-16j – Segar, D., and E. Stamman Segar, 1997: Coral shrimp (*Dasycaris zanzibarica*, Indonesia).

Figure 14-16k – Segar, D., and E. Stamman Segar, 1997: Tiny spider crab (*Xenocarcinus tuberculatus*, Indonesia).

Figure 14-16l – Segar, D., and E. Stamman Segar, 1997: Sea star shrimp (*Zenopontonia noverca*, Indonesia).

Figure 14-17a – Segar, D., and E. Stamman Segar, 1997: Trevally, called a “rainbow runner” (*Elagatis bipinnulata*, Solomon Islands).

Figure 14-17b – Segar, D., and E. Stamman Segar, 1997: Bluefin trevally (*Caranx melampygus*, Solomon Islands).

Figure 14-17c – Segar, D., and E. Stamman Segar, 1997: Blacksaddle mimic filefish or leatherjacket (*Paraluteres prionurus*, Philippines).

Figure 14-17d – Segar, D., and E. Stamman Segar, 1997: Black-saddled toby (*Canthigaster valentini*, Papua New Guinea).

Figure 14-17e – Segar, D., and E. Stamman Segar, 1997: A cleaner wrasse.

Figure 14-17e-inset – Segar, D., and E. Stamman Segar, 1997: Cleaner wrasse (*Labroides pcthirophagus*, Hawaii).

Figure 14-17f – Segar, D., and E. Stamman Segar, 1997: Juvenile mimic surgeonfish (*Acanthurus pyroferus*, Indonesia).

Figure 14-17f-inset – Segar, D., and E. Stamman Segar, 1997: Adult mimic surgeonfish (*Acanthurus pyroferus*, Papua New Guinea).

Figure 14-17g – Segar, D., and E. Stamman Segar, 1997: Pearlscale angelfish (*Centropyge vrolikii*, Papua New Guinea).

Figure 14-18a – Segar, D., and E. Stamman Segar, 1997: Nocturnal shrimp (Fiji).

Figure 14-18b – Segar, D., and E. Stamman Segar, 1997: Crescenttailed bigeye (*Priacanthus hamrur*, Fiji).

Figure 14-18c – Segar, D., and E. Stamman Segar, 1997: Coral crab (probably *Cancer sp.*, Fiji).

Figure 14-18d – Segar, D., and E. Stamman Segar, 1997: Fire urchin (*Asthenosoma intermedium*, Papua New Guinea).

Figure 14-18e – Segar, D., and E. Stamman Segar, 1997: *Octopus luteus*, Indonesia.

Figure 14-19a – Segar, D., and E. Stamman Segar, 1997: Black-headed parrotfish (*Scarus gibbus*, Fiji).

Figure 14-19b – Segar, D., and E. Stamman Segar, 1997: Mud snail (*Nassarius papillosus*, Hawaii).

Figure 14-19c – Segar, D., and E. Stamman Segar, 1997: Cone shell (*Conus geographicus*, Red Sea).

Figure 14-19d – Segar, D., and E. Stamman Segar, 1997: Orbicular burrfish (*Cyclichthys orbicularis*, Philippines).

Figure 14-19e – Segar, D., and E. Stamman Segar, 1997: Orbicular burrfish (*Cyclichthys orbicularis*, Philippines).

Figure 14-19f – Segar, D., and E. Stamman Segar, 1997: Spotted boxfish (*Ostracion meleagris*, Papua New Guinea).

Figure 14-20a – Segar, D., and E. Stamman Segar, 1997: Haddon's sea anemone (*Stichodactyla haddoni*, Indonesia).

Figure 14-20b – Segar, D., and E. Stamman Segar, 1997: Tomato anemonefish (*Amphiprion frenatus*, Papua New Guinea).

Figure 14-20c – Segar, D., and E. Stamman Segar, 1997: Clown anemonefish (*Amphiprion percula*, Papua New Guinea).

Figure 14-20d – Segar, D., and E. Stamman Segar, 1997: Pink anemonefish (*Amphiprion perideraion*, Papua New Guinea).

Figure 14-20e – Segar, D., and E. Stamman Segar, 1997: Spinecheek anemonefish (*Premnas biaculeatus*, Papua New Guinea).

Figure 14-20f – Segar, D., and E. Stamman Segar, 1997: Clark's anemonefish (*Amphiprion clarkii*, Papua New Guinea).

Figure 14-21 – Segar, D., and E. Stamman Segar, 1997: Tuna (*Thunnus sp.*, Fiji).

Figure 14-22a – Segar, D., and E. Stamman Segar, 1997: Cornetfish (*Fistularia commersonii*, Red Sea).

Figure 14-22b – Segar, D., and E. Stamman Segar, 1997: Klein's butterflyfish (*Chaetodon kleinii*, Papua New Guinea).

Figure 14-22c – Segar, D., and E. Stamman Segar, 1997: Klein's butterflyfish (*Chaetodon kleinii*, Papua New Guinea)

Figure 14-23 – Concept from Webb, P. W. (1984). Form and Function in Fish Swimming. *Sci. Am.*, **251**, 72–83, Source.

Figure 14-23a – Segar, D., and E. Stamman Segar, 1997: Meyer's butterflyfish (*Chaetodon meyeri*, Papua New Guinea).

Figure 14-23b – Segar, D., and E. Stamman Segar, 1997: Sergeant fish (*Abudefduf sp.*, Papua New Guinea).

Figure 14-23c – Segar, D., and E. Stamman Segar, 1997: Cheeklined wrasse (*Oxycheilinus digramma*, Papua New Guinea).

Figure 14-23d – Segar, D., and E. Stamman Segar, 1997: Yellowstreak fusilier (Pterocaesio lativittata, Vanuatu).

Figure 14-23e – Segar, D., and E. Stamman Segar, 1997: Flounder (*Bothus sp.*, Hawaii).

Figure 14-23f – Segar, D., and E. Stamman Segar, 1997: Trevally (probably *Carangoides sp.*, Fiji).

Figure 14-23g – Segar, D., and E. Stamman Segar, 1997: Chevron barracuda (*Sphyraena putnamiae*, Solomon Islands).

Figure 14-24a – Segar, D., and E. Stamman Segar, 1997: Threadfin, or red-cheeked fairy basslet (*Pseudanthias huchtii*, Papua New Guinea).

Figure 14-24c – Segar, D., and E. Stamman Segar, 1997: Barred filefish, also called an orange-fin filefish (*Cantherhines dumerilii*, Hawaii).

Figure 14-25 – Concept adapted from Thruman, H. V., 1987: *Essentials of Oceanography 2nd Editions.* Charles E. Merrill, 384 pp.

Figure 14-25a – Segar, D., and E. Stamman Segar, 1997: Blue-girdled angelfish (Pomacanthus navarchus) in Papua New Guinea.

Figure 14-25b – Segar, D., and E. Stamman Segar, 1997: Redbreasted Maori wrasse (*Cheilinus fasciatus*) in Papua New Guinea.

Figure 14-25c – Segar, D., and E. Stamman Segar, 1997: Black-spot angelfish (*Genicanthus melanospilos*) in Papua New Guinea.

Figure 14-25d – Segar, D., and E. Stamman Segar, 1997: Trevally (*Carangoides sp.*) in Papua New Guinea.

Figure 14-25e – Segar, D., and E. Stamman Segar, 1997: Gray reef shark (*Carcharhinus amblyrhynchos*) in Palau.

Figure 14-26a – Segar, D., and E. Stamman Segar, 1997: Zebra lionfish (*Dendrochirus zebra*, Papua New Guinea).

Figure 14-26b – Segar, D., and E. Stamman Segar, 1997: Clown triggerfish (*Balistoides conspicillum*, Philippines).

Figure 14-26c – Segar, D., and E. Stamman Segar, 1997: Speckled sandperch (*Parapercis hexophtalma*, Fiji).

Figure 14-26d – Segar, D., and E. Stamman Segar, 1997: Moon wrasse (*Thalassoma lunare*, Papua New Guinea).

Figure 14-26e – Segar, D., and E. Stamman Segar, 1997: Old Glory goby (*Amblygobius rainfordi*, Papua New Guinea).

Figure 14-26f – Segar, D., and E. Stamman Segar, 1997: Darkspotted moray eel (*Gymnothorax fimbriatus*, Papua New Guinea).

Figure 14-26g – Segar, D., and E. Stamman Segar, 1997: Remora, or sharksucker (*Echeneis naucrates*, Papua New Guinea).

Figure 14-27a – Segar, D., and E. Stamman Segar, 1997: Goldman's sweetlips (*Plectorhinchus goldmanni*, Vanuatu).

Figure 14-27b – Raver, D., 2005: Rainbow trout illustration, USFWS. Accessed on 20 March 2026, Source.

Figure 14-28a – Segar, D., and E. Stamman Segar, 1997: Hard coral colony (*Acropora sp.*, Papua New Guinea).

Figure 14-28b – Segar, D., and E. Stamman Segar, 1997: Mandarinfish (*Synchiropus splendidus*, Indonesia) male and female.

Figure 14-28c – Segar, D., and E. Stamman Segar, 1997: Female Lembeh frogfish (*Antennarius sp.*, Indonesia).

Figure 14-28d – Segar, D., and E. Stamman Segar, 1997: *Hypselodoris bullocki*, Indonesia.

Figure 14-28e – Segar, D., and E. Stamman Segar, 1997: Nudibranch laying eggs (*Kentrodoris rubescens*, Papua New Guinea).

Figure 14-28f – Segar, D., and E. Stamman Segar, 1997: Three nudibranchs exchange eggs and sperm simultaneously (*Nembrotha rutilans*, Indonesia).

Figure 14-28g – Segar, D., and E. Stamman Segar, 1997: Nudibranch eggs are laid in a characteristic spiral ribbon pattern encased in a protective gel (species unknown, Hawaii)

Figure 14-28h – Segar, D., and E. Stamman Segar, 1997: Snails lay eggs encased in a protective gel (*Epitonium billeeanum*, Papua New Guinea).

Figure 14-28i – Segar, D., and E. Stamman Segar, 1997: Common reef squid eggs (*Sepioteuthis lessoniana*, Papua New Guinea).

Figure 14-28j – Segar, D., and E. Stamman Segar, 1997: Flamboyant cuttlefish eggs (*Metasepia pfefferi*, Indonesia).

Figure 14-28k – Segar, D., and E. Stamman Segar, 1997: Indo-Pacific sergeant (*Abudefduf vaigiensis*, Vanuatu)

Figure 14-28l – Segar, D., and E. Stamman Segar, 1997: *Thelenota rubralineata*, Papua New Guinea.

Figure 14-28l-inset – Segar, D., and E. Stamman Segar, 1997: *Thelenota rubralineata*, Papua New Guinea.

Figure 14-28m – Segar, D., and E. Stamman Segar, 1997: Burrowing sponge (*Oceanapia sagittaria*, Indonesia).

Figure 14-28n – Segar, D., and E. Stamman Segar, 1997: Banggai cardinalfish (*Pterapogon kauderni*, Papua New Guinea).

Figure 14-29 – Adapted from Jones, F. R. H., 1968: *Fish Migration. With 86 Fig.* Edward Arnold Ltd. 1968, 325 pp.

Figure 14-30a – Adapted from Bond, C. E., 1979: *Biology of Fishes*. W. B. Saunders Company, 514 pp.

– Adapted from Cushing, D. H., 1982: Recruitment. *Climate and Fisheries*. Academic Press, 61.

Figure 14-30b – Adapted from Steele, J.H., 1977: Population Regulation. *Fish Population Dynamics*. J. A. Gullard Ed. Wiley, 250–251.
– Adapted from Cushing, D. H., 1982: Recruitment. *Climate and Fisheries*. Academic Press, 68.

Figure 14-31 – Adapted from Duxbury, A. C., and A. B. Duxbury, 1994: *An Introduction to the World's Oceans, 4th Edition*. William C. Brown, 496 pp.

Figure 14-32a – Segar, D., and E. Stamman Segar, 1997: Whip goby (*Bryaninops yongei*, Indonesia).

Figure 14-32b – Segar, D., and E. Stamman Segar, 1997: Large black isopod parasite (crustacean of the family Cymothoidae commonly known as a "fish doctor") on a lemon damsel (*Pomacentrus moluccensis*, Papua New Guinea).

Figure 14-32c – Segar, D., and E. Stamman Segar, 1997: Flasher scorpionfish (*Scorpaenopsis macrochir*, Indonesia) with "fish doctor" fish.

Figure 14-32d – Segar, D., and E. Stamman Segar, 1997: Parasitic snail (*Thyca crystallina*, Indonesia) on sea star (*Linckia laevigata*).

Figure 14-33 – Adapted from Dogiel, V. A., G. K. Petrushevski and Yu. I. Polyanski, Eds, 1979: *Parasitology of fishes*. THF Publications, 384 pp.

Figure 14-34a – Segar, D., and E. Stamman Segar, 1997: Arrowhead crab (*Huenia heraldica*, Papua New Guinea).

Figure 14-34b – Segar, D., and E. Stamman Segar, 1997: Xenia swimming crab (*Caphyra sp.*, Indonesia).

Figure 14-34c – Segar, D., and E. Stamman Segar, 1997: Tiny spider crab (*Xenocarcinus tuberculatus*, Indonesia).

Figure 14-34d – Segar, D., and E. Stamman Segar, 1997: Porcelain crab (probably *Porcellanella triloba*, Indonesia).

Figure 14-34e – Segar, D., and E. Stamman Segar, 1997: Tiny porcelain crab (probably *Porcellanella sp.*, Indonesia).

Figure 14-34f – Segar, D., and E. Stamman Segar, 1997: Weed cardinalfish (*Foa brachygramma*, Papua New Guinea).

Figure 14-34g – Segar, D., and E. Stamman Segar, 1997: Crinoid clingfish (*Discotrema crinophila*, Papua New Guinea).

Figure 14-34h – Segar, D., and E. Stamman Segar, 1997: Emperor shrimp (*Periclimenes imperator*, Papua New Guinea).

Figure 14-34i – Segar, D., and E. Stamman Segar, 1997: Sea star shrimp (*Periclimenes soror*, Indonesia).

Figure 14-34j – Segar, D., and E. Stamman Segar, 1997: Sea star commensal shrimp (*Zenopontonia noverca*, Indonesia).

Figure 14-34k – Segar, D., and E. Stamman Segar, 1997: Coleman's shrimp (*Periclimenes colemani*, Papua New Guinea).

Figure 14-34l – Segar, D., and E. Stamman Segar, 1997: Shrimp (*Ancylomenes cf. venustus*, Indonesia).

Figure 14-34m – Segar, D., and E. Stamman Segar, 1997: Shrimp (*Periclimenes cf. tosaensis*, Indonesia).

Figure 14-34n – Segar, D., and E. Stamman Segar, 1997: Coral shrimp (*Vir philippinensis*, Indonesia).

Figure 14-34o – Segar, D., and E. Stamman Segar, 1997: Humpbacked shrimp (*Hippolyte commensalis*, Indonesia).

Figure 14-34p – Segar, D., and E. Stamman Segar, 1997: Tiny urchin shrimp (*Gnathophylloides mineri*, Indonesia).

Figure 14-34q – Segar, D., and E. Stamman Segar, 1997: Saw blade shrimp (*Tozeuma armatum*, Indonesia).

Figure 14-34r – Segar, D., and E. Stamman Segar, 1997: Scale worms (*Gastrolepidia clavigera*, Vanuatu).

Figure 14-34s – Segar, D., and E. Stamman Segar, 1997: Egg cowrie (Pseudosimnia sp., Red Sea).

Figure 14-35a – Segar, D., and E. Stamman Segar, 1997: Anemone hermit crab (*Dardanus sp.*, Papua New Guinea).

Figure 14-35b – Segar, D., and E. Stamman Segar, 1997: Crab (*Dorippe frascone*, Indonesia).

Figure 14-35c – Segar, D., and E. Stamman Segar, 1997: Tiger pistol shrimp (*Alpheus bellulus*, Indonesia) with a yellow shrimp goby (*Cryptocentrus cinctus*).

Figure 14-35d – Segar, D., and E. Stamman Segar, 1997: Giant clam (*Tridacna gigas*, Papua New Guinea).

Figure 14-35e – Segar, D., and E. Stamman Segar, 1997: Small pistol shrimp (*Synalpheus sp.*, Papua New Guinea).

This page titled 14.9.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15: OCEAN ECOSYSTEMS

Technical Editor: Emily Brownlee, Ph.D. (St. Mary's College of Maryland)

Additional contributions and edits by Elaine Stamman, M.S. (Texas A&M University, retired)

Learning Objectives

- **CC9 The Global Greenhouse Effect**: Perhaps the greatest environmental challenge faced by humans is the prospect that major climate changes may be an inevitable result of our burning fossil fuels. The burning of fossil fuels releases carbon dioxide and other gases into the atmosphere where they accumulate and act like the glass of a greenhouse trapping more of the sun's heat.
- **CC14 Phototrophy, Light, and Nutrients**: Phototrophy (which includes photosynthesis) and chemosynthesis are two processes by which simple chemical compounds are made into the organic compounds of living organisms. Photosynthesis depends on the availability of carbon dioxide, light, and certain dissolved nutrient elements including nitrogen, phosphorus, and iron. Chemosynthesis does not use light energy, but instead depends on the availability of chemical energy from reduced compounds, which occur only in limited environments where oxygen is depleted.
- **CC16 Maximum Sustainable Yield**: The maximum sustainable yield is the maximum biomass of a fish species that can be depleted annually by fishing but that can still be replaced by reproduction. This yield changes unpredictably from year to year in response to the climate and other factors. The populations of many fish species worldwide have declined drastically when they have been overfished (beyond their maximum sustainable yield) in one or more years when that yield was lower than the average annual yield on which most fisheries management is based.
- **CC17 Species Diversity and Biodiversity**: Biodiversity is an expression of the range of genetic diversity; species diversity; diversity in ecological niches and types of communities of organisms (ecosystem diversity); and diversity of feeding, reproduction, and predator avoidance strategies (physiological diversity), within the ecosystem of the specified region. Species diversity is a more precisely-defined term and is a measure of the species richness (number of species) and species evenness (extent to which the community has balanced populations with no dominant species). High diversity and biodiversity are generally associated with ecosystems that are resistant to change.
- **CC18 Toxicity**: Many dissolved constituents of seawater become toxic to marine life when the concentrations go above their natural amount. Some synthetic organic chemicals are especially significant because they are persistent and may be bioaccumulated or biomagnified.

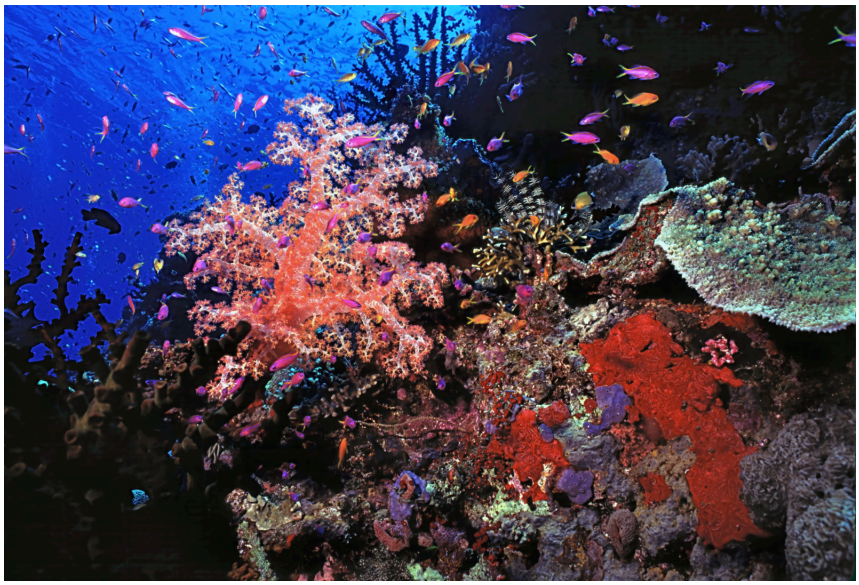


Figure 15-1. Coral reef ecosystems have a species diversity that rivals, and perhaps exceeds, the species diversity of tropical rain forests. In this photograph of Deacon's Reef in Papua New Guinea, several species of fishes are easily seen. However, the majority of species that live on this reef live in or on the reef itself, or live hidden in the seafloor substrate. The beautiful red and white coral "tree" is a colony of the soft coral *Dendronephthya sp*. Less obvious are many species of hard corals, crinoids, sponges, tunicates, hydroids, and algae. Even this profusion of species is only a tiny fraction of the species that live on this reef. Hidden from view in or among the corals and other large animals, there are literally thousands or tens of thousands of species of algae, shrimp, crabs, sea stars, brittle stars, urchins, fishes, barnacles, worms, snails and other mollusks, and other less familiar types of animals and countless microbial species.

The term **ecosystem** encompasses both living organisms and the physical **environment** of a particular volume of space. Within the ecosystem, the components and characteristics of the living and nonliving environment are interdependent, each influencing the other. Hence, conceptually the global environment can be broken down into separate ecosystems within which the **species** that are present, the physical environment, and the relationships among the species and with their environment are distinct and different from those in adjacent areas. For example, we can define each geographically distinct **coral reef** as a separate ecosystem. However, all coral reef ecosystems have common characteristics in terms of the types of organisms they sustain, their physical environment, and processes driven by biological–physical interactions. These common characteristics are different from those of other ecosystems, such as **mangrove** swamps, the **rocky intertidal zone**, and the open-ocean **photic zone**.

This chapter briefly reviews several ocean ecosystems, emphasizing the relationship between their individual physical environmental conditions and their biological **communities**. However, the classification of separate ecosystems in the oceans is not precise. In reality, all such ecosystems are linked and interdependent.

15.1: Communities and Niches
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
15.6: Polar Regions
15.7: Beyond the Sun's Light
15.8: Hydrothermal Vents
15.9: Chapter Summary
15.9.1: Key Terms
15.9.2: Chapter Questions
15.9.3: References

This page titled 15: Ocean Ecosystems is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.1: Communities and Niches

The world ocean can be considered a single ecosystem, but it is too large and complex to be studied as a whole. Consequently, marine **ecologists** often separate the oceans into ecosystems that have common biological and physical characteristics. Because even the simplest of these is complex, marine ecologists study the relationships of species with each other and their environment at a number of different levels, ranging from ocean-sized ecosystems to individual species. These levels are generally not well defined or distinct from each other.

Within an ecosystem, species are distributed in a nonuniform way because of small-scale variations in the physical environment and competition and other interactions among species. Species are often clustered in their communities. The term *community* is inexact and can refer either to all organisms that coexist in a specific location or to groups of species that are found together in many different locations. Thus, a community can be all the species within an entire coral reef, or a specific species of **sea fan** and all species that commonly live in or on it wherever this sea fan is present.

The distribution of species and their relationships with each other and the physical environment can be studied at the species level. In such studies, a very useful concept is the biological **niche**. Each species has evolved to take advantage of certain characteristics of the physical environment. Hence, each species has a range of environmental variables within which it can survive. These variables include **salinity**, temperature, **suspended sediment** concentration, substrate **grain size**, light intensity, and **nutrient** and other chemical concentrations. The combination of the ranges of each of these environmental variables within which the species can survive is the species' **survival niche**. Figure 15-2a shows how such a niche can be defined by two environmental parameters. In practice, niches are defined by many more parameters and cannot be fully represented in a two-dimensional diagram.

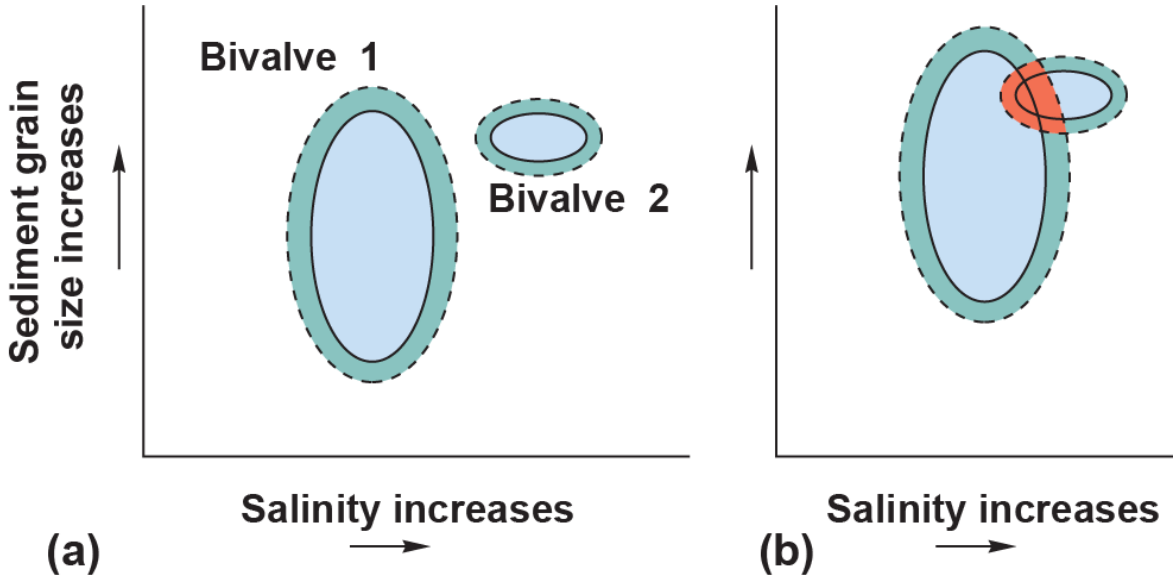


Figure 15-2. Ecological niches can be defined in several ways and by many parameters. These diagrams illustrate the concepts using just two such parameters: salinity and sediment grain size. (a) The fundamental niches of two bivalves are defined by the range of values of the two parameters defined by the blue areas. The survival niches are defined by a wider range of environmental conditions and include both the blue and green areas within each dotted line. Because the niches of these two species do not overlap, they would not occur in the same environment. (b) If the niches of the two species overlap, as they do in the red area, they may both occur in environments within this range of the two parameters, but they may also compete so that one species may be excluded from this part of its niche.

Within the survival niche of a particular species is a somewhat smaller range of environmental variables within which the species can both survive and successfully reproduce (Fig. 15-2a). This range is often referred to as the organism's **fundamental niche**. Species rarely, if ever, occupy every area within the marine environment where environmental parameters are within their fundamental niche. Two factors restrict the distribution of a species within its fundamental niche. First, each species has a smaller niche within its fundamental niche in which it finds optimal conditions. Second, species with overlapping niches may compete, and one species may be excluded from parts of its fundamental niche (Fig. 15-2b).

The interaction of species with each other and with the environment is wonderfully complex and may never be fully understood

for all species in the oceans. Thus, descriptions of ecosystems in the discussion that follows represent only a cursory view of the factors that contribute to the amazing profusion and **diversity** of species in the oceans.

This page titled 15.1: Communities and Niches is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.2: Coral Reefs

Although some types of **corals** are present in all parts of the oceans, including Arctic seas, species of corals that build coral reefs, called **hermatypic corals**, grow only in areas where the water temperature never falls below about 18°C. Coral reefs are thus restricted to a broad band of mostly tropical waters between about 30°N and 30°S (Fig. 15-3). The range of coral reef occurrence extends into somewhat higher **latitudes** on the western side of each ocean because warm **western boundary currents** flow poleward.

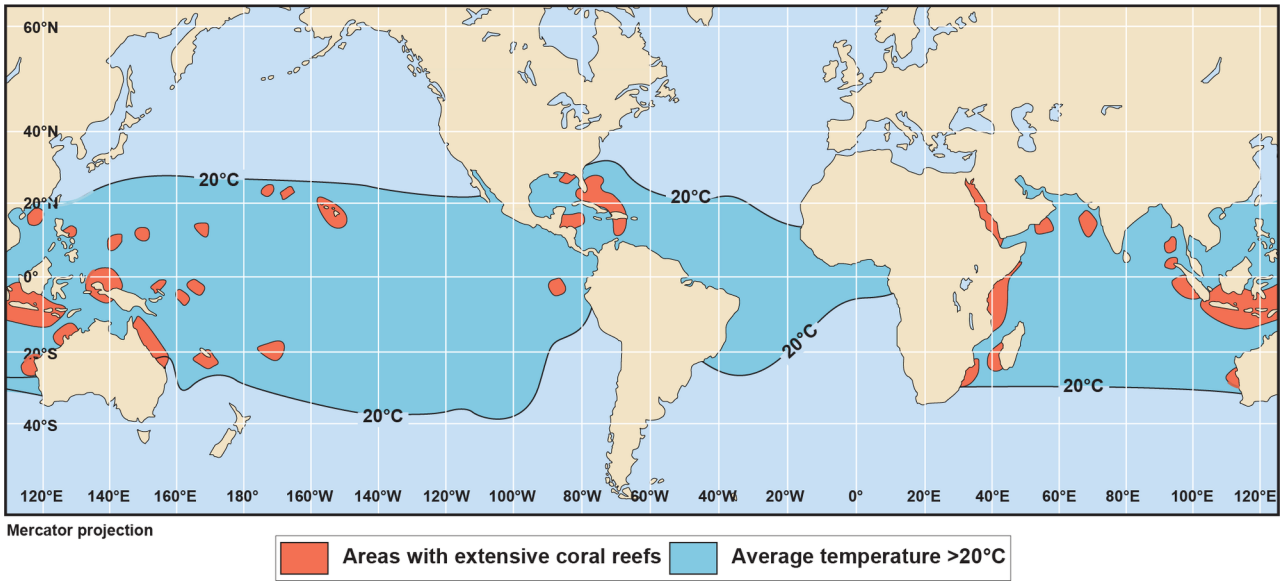


Figure 15-3. The area of the oceans with average sea surface temperatures above 20°C extends to about 30° north and south of the equator in the Pacific and Indian Oceans, and somewhat less in the Atlantic Ocean. The temperature never falls below 18°C in most of this area. Hermatypic corals grow only in these areas and only where other environmental conditions are suitable within these areas. Although many small tropical islands have extensive coral reefs, they are too small to show in this figure.

Environmental Requirements for Coral Reef Formation

Reef-building corals require an appropriate substrate on which to attach, and they have a **symbiotic** relationship with **zooxanthellae**, a nonmotile form of **dinoflagellate** that lives within the coral's tissues. Because zooxanthellae need light to **photosynthesize** (**CC14**), living coral reefs are present only in waters where the seafloor is within the photic zone. In clear waters, corals can grow to depths of about 150 m, but in high-turbidity waters coral growth is reduced or prevented by two mechanisms: First, the higher **turbidity** reduces light penetration and limits the depth at which zooxanthellae can photosynthesize and, therefore, also limits the depth at which corals can grow. Second, large quantities of suspended sediment that cause high turbidity may smother the corals. Corals can clear away a certain amount of deposited sediment material, but they must use energy to do so, and they lose additional energy because they must also stop feeding in order to do so. When these energy costs are too high, the corals cannot survive. Therefore, coral reefs do not grow in coastal areas near river mouths or near other sources of large amounts of suspended matter, such as dredging projects. Corals also grow poorly in water of low or variable salinity.

The physical conditions necessary for coral growth (clear, warm, shallow waters with relatively invariable normal ocean salinity) are present primarily between about 23.5°N and 23.5°S. This is a region that we might expect to be a biological desert because surface waters are isolated by a steep permanent **thermocline** throughout most of the tropical oceans. In contrast, instead of biological deserts, coral reefs are areas of high productivity with an amazing diversity (**CC17**) of fish and **invertebrate** species. In what may seem to be a paradox, waters surrounding reefs are clear blue and have extremely small populations of **phytoplankton**. The waters over the reef itself have somewhat higher phytoplankton populations and **primary productivity**, but these populations are extremely small in comparison with those in **upwelling**-zone ecosystems and cannot account for the high productivity of coral reefs.

Factors Affecting Coral Reef Productivity

The reasons for the anomalously high productivity of coral reefs are somewhat complex and not fully understood, but they appear to be related to a combination of physical conditions and the unique relationships among the reef organisms.

First, hermatypic corals and their associated zooxanthellae act together in a mutualistic (mutually beneficial) relationship to create a very effective mechanism for collecting, concentrating, and rapidly recycling nutrients. Zooxanthellae live embedded in the coral's tissues and use solar energy that penetrates the coral's transparent tissues to produce food by chemically recombining the coral's waste products. Carbon dioxide and nutrients, which are released by the coral through its digestive processes, are transferred directly to zooxanthellae and converted by photosynthesis into organic matter. In turn, as much as 60% of the organic matter created by the zooxanthellae through photosynthesis is released through the zooxanthellae cell wall directly into the coral tissue, providing food for the coral. In this way, nutrients are continually recycled, and food is continually produced by the **algae** and consumed by the coral. The coral–zooxanthellae association ensures very little loss of either nutrients or food to the surrounding water. Corals feed on **zooplankton** to supplement the food supplied by their internal zooxanthellae. Thus, the small amounts of nutrients that are lost from the coral–zooxanthellae partnership are continuously replaced. This efficient nutrient retention and recycling mechanism is believed to be the main reason for the high productivity of coral reefs.

The second reason for their high productivity is that most reefs are built on the sides of submarine mountains or on the fringes of landmasses with very narrow **continental shelves**. Indeed, the growth of corals off coasts where sea level has risen during the past several thousand years has created very steep drop-offs at the outer edges of many reefs (**Figs. 4-28, 15-4**). Ocean **currents** that flow past such steep continental shelves and islands form **eddies**. The eddies create vertical water movements that can bring nutrient-rich deep water, at least episodically, into the photic zone of the reef.

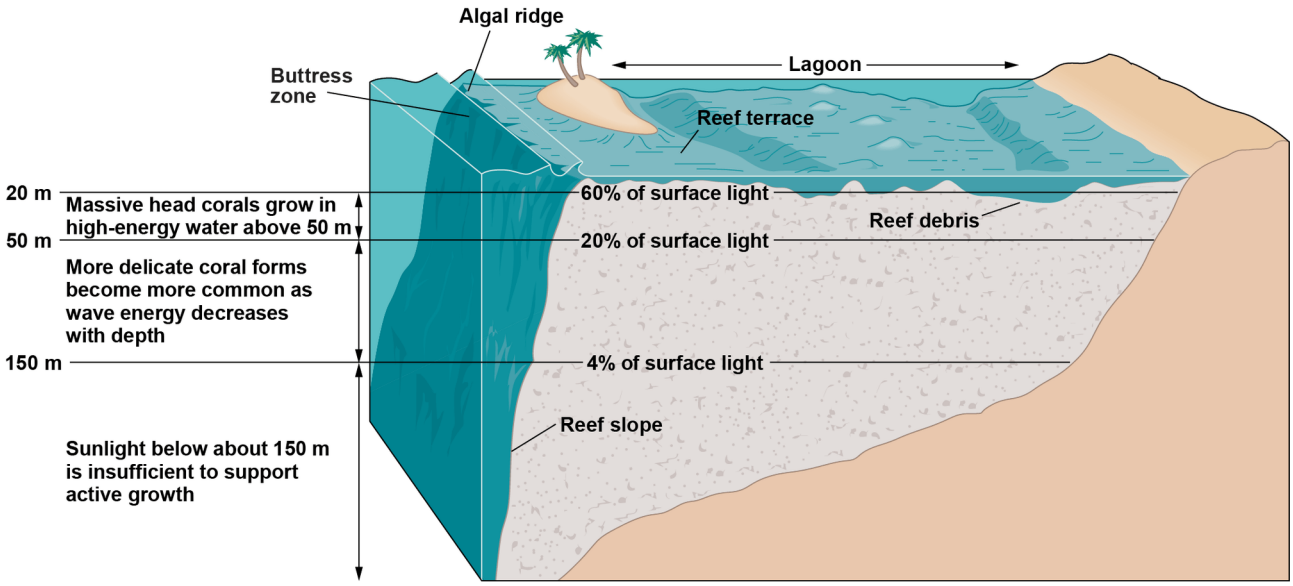


Figure 15-4. The structure of a typical coral reef.

The third reason for the high productivity of coral reefs is that many reef ecosystems are partially closed systems, within which most nutrients not retained in the coral–zooxanthellae association are continually recycled. Most of the reef floor is shallow and within the photic zone. Hence, organic **detritus** created on the reef settles largely on the reef floor, where it is consumed by **decomposers** that release nutrients to be recycled. In addition, most fishes are permanent residents of the reef. Nutrients in their urine and feces are released almost entirely back to waters of the reef, where they are rapidly recycled and taken up again by zooxanthellae or phytoplankton.

Primary Producers in the Coral Reef Community

Although zooxanthellae are important primary producers in coral reef ecosystems, they account for only a small proportion (generally less than 5%) of the **biomass** of photosynthesizers in these ecosystems. Reef ecosystems are dominated by **benthic** or **encrusting microalgae** and a variety of attached **macroalgae** (**Fig. 15-4**). The biomass of these algae generally exceeds the animal biomass in the reef ecosystem by as much as three times. In fact, the **hard parts** of **calcareous** algae are responsible for

building much of the reef structure.

Algae attached to the solid substrate of the reef are favored over phytoplankton in coral ecosystems because nutrients are recycled, made available, and rapidly reassimilated at these surfaces. Attached algae remove most of the recycled nutrients before they can **diffuse** into the water column and become available to phytoplankton. In addition, attached algae remain in the reef ecosystem, where nutrients are available, whereas phytoplankton may be transported away from the reef into nutrient-depleted adjacent deep water.

Coral Reef Niches and Topography

Coral reef ecosystems contain a bewildering variety of species. These species are distributed in niches that are defined by current and wave action, and by variations in salinity, water depth, turbidity, temperature, and other factors.

Figure 15-4 shows the general **topographic** features of a typical coral reef. In shallow waters of the **lagoon**, currents are generally weak and there is little wave action. Hence, sediments tend to accumulate in the lagoon and are removed primarily during major storms. Lagoon **sediments** sustain a wide variety of invertebrates that feed on suspended or deposited detritus. These detritus feeders include **suspension feeders** such as sea pens (Fig. 14-6a), and a variety of bivalve **mollusks** that live in the sediment and extend their feeding apparatus into the water column. Also present are many species of **deposit feeders**, including mollusks and worms that live in and sift through the sediment, **sea cucumbers** (Fig. 15-5a), urchins (Fig. 15-5c), sand dollars (Fig. 15-5d), and other species that feed on benthic algae growing on the sediment surface, and predators such as goatfishes (Fig. 14-13d) and **sea stars** (Fig. 15-5b) that hunt mollusks and other animals living in or on the sediment.

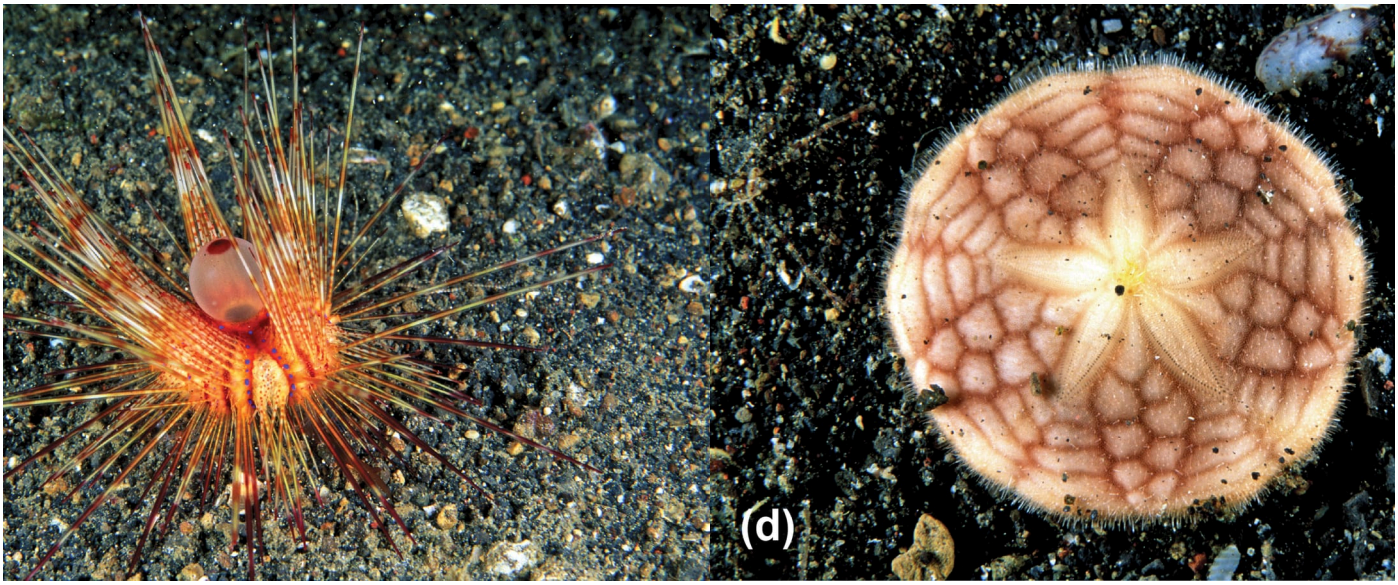


Figure 15-5. Inhabitants of the coral reef lagoon and outer reef flat. Many invertebrates—including (a) sea cucumbers (family Synaptidae, Indonesia), (b) sea stars such as this rhinoceros, or horned, sea star (*Protoreaster nodosus*, Papua New Guinea), (c) urchins (Astropyga radiata, Indonesia), and (d) the urchins' close relatives the sand dollars (*Clypeaster sp.*, Indonesia)—feed on algae, detritus, and microorganisms on sandy seafloors. (e) Spiked forms of hard corals, such as *Acropora sp.* (Papua New Guinea), grow in patches on sandy lagoon floors. (f) Elkhorn coral (*Acropora palmata*, Puerto Rico) grows in many areas on the outer reef flat of reefs that experience low wave energy.

In the lagoon, the growth of most corals is inhibited by the blanket of sediment that covers the seafloor. However, certain corals, such as *Acropora* (Fig. 15-5e,f), that grow up into the water column can prosper if **sedimentation rates** are sufficiently low and sediments are rarely **resuspended** by waves. Once established, these corals provide a sediment-free substrate for other species. Consequently, in quiet lagoons with little suspended sediment input and where salinity is not altered by freshwater input, irregularly shaped mounds of coral and associated species develop. These are called "patch reefs."

At the lagoon's outer edge is a **reef flat**, or **reef terrace**, that is relatively free of sediments because they are swept off the terrace by waves. It is an ideal location for coral growth. Coral grows upward until the reef terrace is only a few centimeters below the **low-tide line**. Further growth is inhibited because corals cannot survive for long periods out of water, although the surface of the reef terrace may be completely exposed to the atmosphere for short periods during low **spring tides** without killing the corals. Reef terrace corals are generally encrusting corals because the water is too shallow and the wave energy too high for corals that grow

in other forms (e.g., **Figs. 14-8a,b, 15-6e,f**). The surface of the reef terrace is not smooth. It has many grooves and holes created primarily by invertebrates that eat or drill into the coral to obtain food or to create safe areas to shelter from larger predators.

On some reefs, a low island is formed by sediment accumulated during storms at the landward side of the reef terrace (**Fig. 15-6**). This island may be well enough established to support palm trees. The seaward edge of the reef usually has an irregular ridge, parts of which are shallow enough to emerge from the water, especially at low **tide**. The ridge is formed by intense wave action that periodically smashes against the reef's outer edge and dislodges large chunks of the **limestone** substrate of the reef. The chunks are cemented back onto the reef by the calcareous algae that live in abundance in this region. Calcareous algae are abundant because wave energy is too intense for corals to grow effectively. The algae cement themselves to the reef surface with their calcium carbonate hard parts. The cemented algae can withstand intense wave action and can quickly colonize any new surface created by storm wave damage. These algae benefit from a continuous supply of very low concentrations of nutrients brought to the reef edge by currents and waves. Because calcareous algae are abundant and help to maintain this ridge, it is called the **algal ridge**.

Figure 15-6. Islands often form behind the outer reef flat as wave-driven debris accumulates in this area. The island can become high and stable enough to sustain vegetation, such as the palm trees and other plants on this island on a fringing reef in Fiji.

On the few reefs where wave energy is very low, no algal ridge or reef terrace is present. The outer reef may consist primarily of relatively robust branching corals, such as elkhorn coral (**Fig. 15-5f**), or the reef terrace may simply end at the reef edge. Stands of elkhorn coral characterize the seaward edge of several sheltered Caribbean reefs. Some reefs within tectonically active island chains, such as Palau in the Pacific Ocean, are protected from wave action by a **barrier reef** that surrounds groups of several or many islands. Inside some of these barrier reefs, the sheltered island shore has a **fringing reef** with a reef terrace that simply ends abruptly at the reef wall, which plunges vertically or even undercuts to depths that can exceed several hundred meters (**Fig. 15-4**).

Seaward of the algal ridge (or reef terrace if no algal ridge is present), the downward slope of a reef can range from very gradual to nearly vertical. The upper part of the outer slope (**Fig. 15-4**), called the "buttress zone," is subject to strong scouring action of waves down to about 20 m. It is usually cut across at intervals by deeper channels separated by relatively high ridges, sometimes called a "tongue and groove" formation. The channels or grooves are cut by waves and may also be called "surge channels." Sand and debris created by wave action on the reef are transported seaward through these grooves. Coral growth is limited in the grooves, but many invertebrates and fishes feed on detritus transported along them. Coral growth is more extensive on the relatively sediment-free ridges. However, because wave energy is high, these ridges sustain primarily robust massive varieties of coral, such as brain coral (**Fig. 14-8a**), and encrusting corals and algae.

Farther seaward, on the outer slope between about 20 and 50 m, is a transition from robust massive corals and encrusting species to still strong but less robust forms, such as *Acropora* (**Fig. 15-5e,f**), and then to more delicate varieties of corals, including black coral (**Fig. 15-7**), sea fans (**Fig. 14-8e**), and **soft corals** (**Fig. 14-8d,f**). The depth at which these transitions occur depends on the intensity of wave action. Delicate corals are present at shallower depths on **leeward** sides of coral-fringed islands than on **windward** sides.

Figure 15-7. Black coral colony (probably *Cirrhipathes sp.*) with a longnose hawkfish (*Oxycirrhites typus*, Papua New Guinea) hiding among its branches. The hard parts of black corals are a deep black color and are often used to make jewelry, but the polyps may be a variety of colors in the living coral.

Between about 50 and 150 m, the reef is dominated by delicate coral species that grow outward in slender fingers or arms. Such growth enables hermatypic corals to extend beyond the shadows of their neighbors in the never-ending competition to obtain light for their zooxanthellae. It also enables nonhermatypic corals, including the delicate and beautiful soft corals (**Figs. 14-8d,f**), to extend into the water that flows along the reef face to capture suspended food. Soft corals are most abundant in areas where currents are strongest and thus expose them to the greatest possible food supply.

Just as the types of coral that grow on the reef's seaward side are determined primarily by depth and the intensity of wave action, species of invertebrates and fishes that inhabit or feed on the corals change with these factors and in response to changes in coral species. The variations are too complicated to review here but can provide fascinating study for **scuba** divers when no big animals such as sharks, manta rays, or turtles are present to capture their interest.

This page titled 15.2: Coral Reefs is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.3: Kelp Forests

The name **kelp** is given to a number of species of brown macroalgae that are attached to the seafloor at depths as great as 30 to 40 m, where the water is clear enough to allow light to penetrate to that depth. Once established, kelp fronds grow toward and eventually reach the ocean surface. Where conditions are suitable, kelp grows so densely that it forms forests (Fig. 15-8).

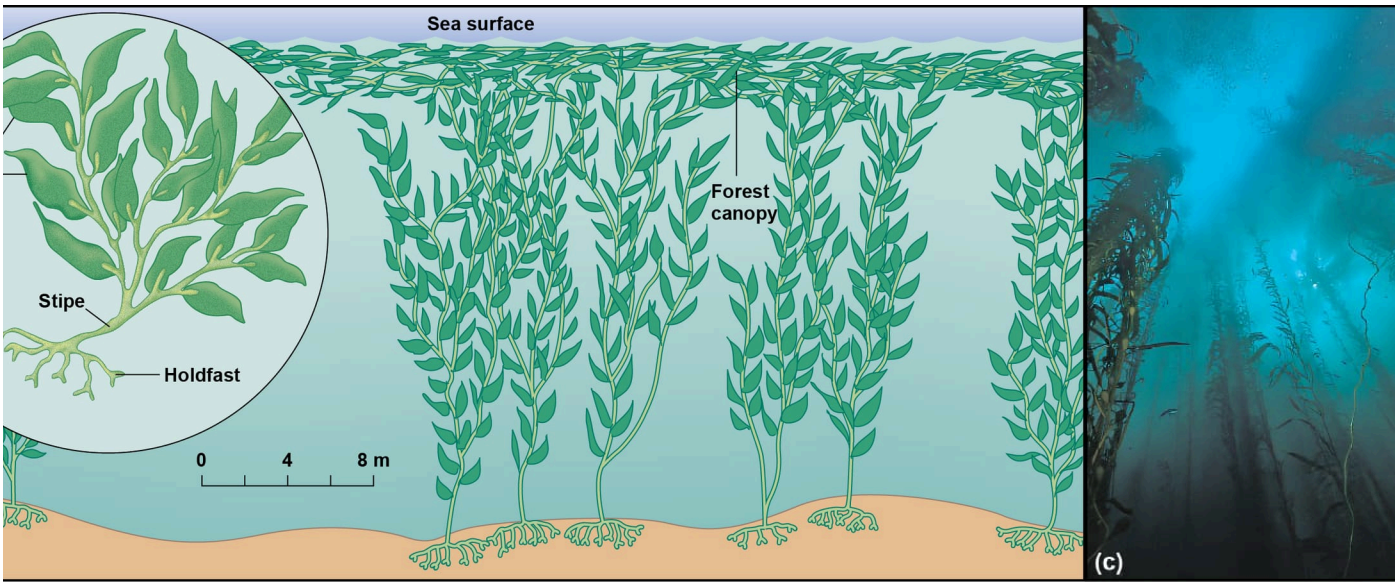


Figure 15-8. Kelp forest. (a) Kelp, such as this *Macrocystis pyrifera*, is attached to the seafloor by holdfasts and has long, flexible stems, called "stipes," that support the fronds in which photosynthesis occurs. (b) In some areas of abundant kelp, the ocean surface is covered by a canopy of kelp fronds, and the kelp growing upward from the seafloor forms an underwater forest. (c) The sunlight filters through the canopy and kelp forest in this underwater photograph from about 8 to 10 m depth in Monterey, California.

Extensive forests of one such kelp, *Macrocystis pyrifera*, are present off the west coast of North America. Kelp fronds are **buoyant**. *Macrocystis pyrifera* fronds, for example, have gas-containing sacs or bladders distributed along them (Fig. 12-4a). Once a frond reaches the surface, it continues to grow, but the new growth floats on the surface to form dense mats or a canopy. These canopies can cover the entire ocean surface within the species' depth range along substantial stretches of coast.

Kelp Community Environmental Characteristics

Kelp forests grow where the temperature and nutrient characteristics of the water column are very different from those required for coral reef formation. Kelp requires water cooler than 20°C and high nutrient concentrations. Although kelp is attached to the seafloor by holdfasts, it obtains its nutrients from the water column through the surface of its fronds. Kelp fronds grow as much as half a meter a day, and kelp forest primary productivity ranges from 500 to 1500 g of carbon per square meter per year. This rate exceeds the primary productivity in all but the most highly productive phytoplankton-based ecosystems and is approximately the same as primary production rates of terrestrial farms. Large quantities of nutrients are necessary to sustain such growth rates. Consequently, kelp forests are almost exclusively restricted to areas of intense upwelling. The fact that upwelled waters are cold probably accounts for kelp's temperature niche requirements.

Besides low temperatures and high nutrient concentrations, kelp needs a stable, generally rocky, seafloor to which it can attach. In addition, kelp can grow only in shallow waters where sufficient light penetrates to the seafloor to support the growth of newly settled kelp spores.

Kelp Life Cycle and Communities

The kelp life cycle consists of two distinctly different stages. Although the large kelp that form kelp forests grow vegetatively (asexually), they also produce microscopic swimming **spores** that enter the plankton and eventually settle on the seafloor. These

spores grow into microscopic male and female plants, which produce eggs and sperm and reproduce sexually near or on the seafloor. If they encounter suitable conditions, the resulting embryos then grow into the large, mature kelp.

Kelp forests support amazingly diverse populations of invertebrates, fishes, and **marine mammals**. The kelp provides several benefits to species that live within its forests. First, kelp produces large quantities of detritus, which forms the base of the **food web**. Surprisingly, only a few animals, including only a few fish, snail, and **sea urchin** species, eat kelp itself, but kelp is easily torn apart, particularly at the ends where new growth occurs. Therefore, kelp releases large amounts of detritus, which is first modified by decomposers and only then consumed by animals. Second, the kelp canopy provides a hiding place and protection from predators, particularly seabirds. The fronds afford many escape routes that enable agile harbor seals and sea lions to evade their shark and killer whale predators, which are less agile and less able to maneuver through the kelp forest. In addition, kelp fronds and holdfasts provide substrate for encrusting organisms and their **grazers** or predators and many secluded places for small fishes and invertebrates to hide from predators.

The rocky seafloor and holdfasts of the kelp forest also sustain a surprisingly diverse community of nonhermatypic corals, **anemones**, crabs, shrimp, sea stars, **nudibranchs**, fishes, and other animals (**Fig. 15-9**). In many ways, kelp forest communities rival those of coral reefs in complexity and beauty.

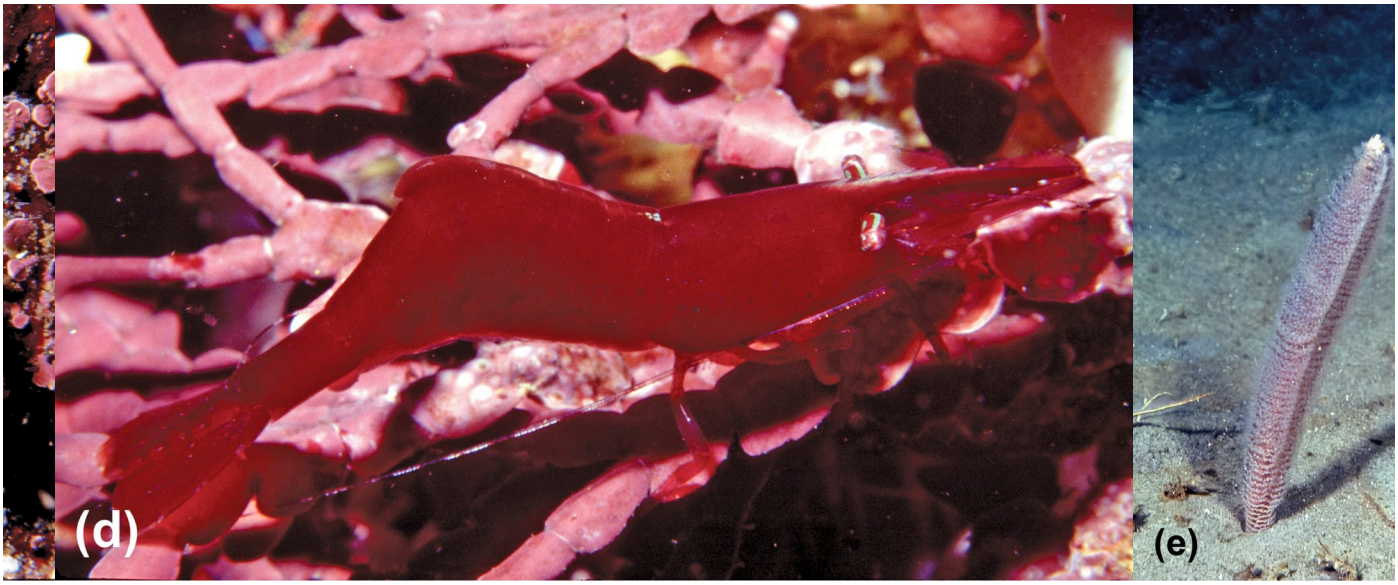


Figure 15-9. Members of the kelp holdfast community. (a) These small brown cup corals (*Paracyathus stearnsii*, California) cover many of the rocks at shallow depths. (b) Anemones, such as this rose anemone (*Tealia lineata*, California), are abundant on the kelp forest floor. (c) Many species of crabs, such as this northern kelp crab (*Pugettia producta*, California), feed on the abundant detritus and other animal life. (d) A kelp shrimp is camouflaged to look like the algae in which it lives and on which it feeds (Monterey, California). (e) Sea stars of many species live on the kelp forest floor, and sea pens are often found in the patchy areas of seafloor that are located in and around the forest (Monterey, California). (f) This horned nudibranch species (*Hermissenda crassicornis*, California) is common in the kelp community, together with many other nudibranch species. (g) Many rockfish species live and hunt in the kelp forest, such as these two: the quill-backed rockfish (*Sebastes maliger*) on the lower left, and the copper rockfish (*Sebastes caurinus*, California) swimming above, further to the right.

Kelp, Sea Otters, and Sea Urchins

Two of the most important residents of the kelp community are sea otters (**Fig. 15-10a**) and sea urchins (**Fig. 15-10b**). Sea urchins eat kelp, and sea otters eat sea urchins. Because of this relationship, healthy and abundant kelp forests depend to a large extent on a healthy population of sea otters. In areas where otters are abundant and many sea urchins are eaten, sea urchin populations are low and dominated by small, young individuals. Such populations are too small to affect the kelp significantly (**Fig. 15-10c**). However, if otters are scarce, the sea urchin population multiplies, and many large sea urchins are present to graze heavily on the kelp and reduce its abundance (**Fig. 15-10d**). Large sea urchins, in particular, feed preferentially on the kelp holdfasts and may destroy the forest by cutting the kelp loose from the seafloor, even if they do not consume it.

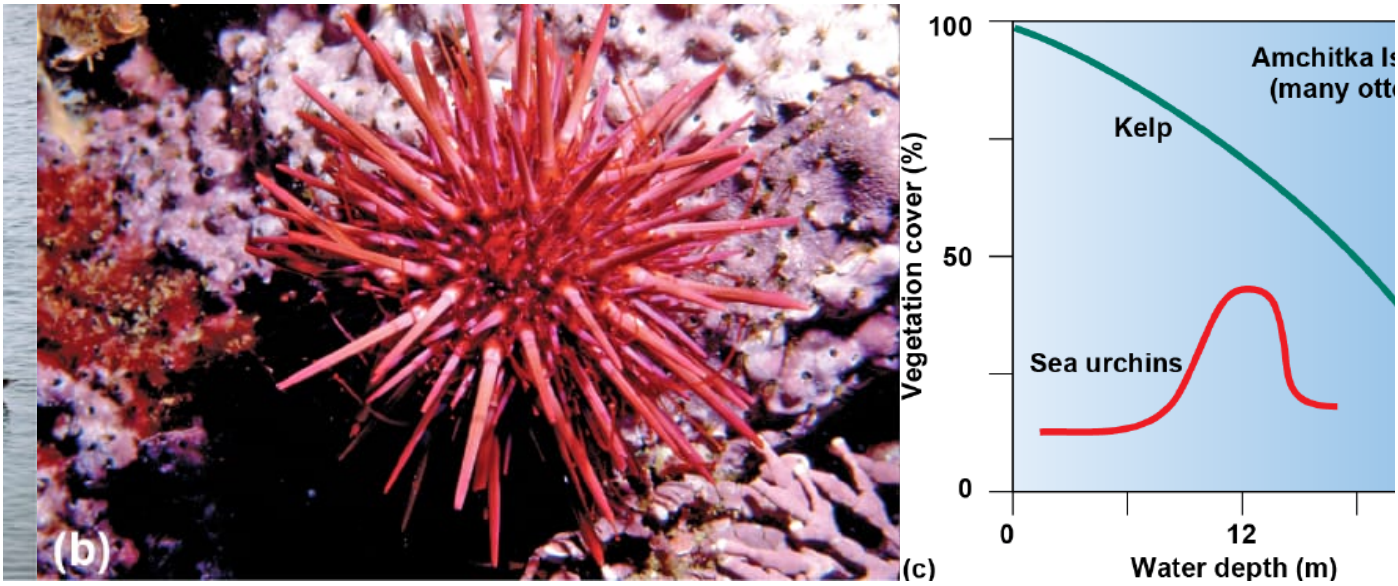


Figure 15-10. The sea otter–urchin–kelp relationship. (a) Sea otters (*Enhydra lutris*) were at one time abundant in the kelp forests of California, but their numbers plummeted during the nineteenth century because they were exploited for their pelts. Their populations are now recovering. (b) The sea otter's favorite and most important food is sea urchins, such as this red urchin (*Strongylocentrotus sp.*, California). (c) At Amchitka Island, Alaska, where sea otters are abundant, sea urchins are rare in shallow water because they are preyed upon by the otters. As a result, there is little browsing of the kelp forests by urchins and the kelp is abundant. (d) In contrast, at nearby Shemya Island, where otters have become scarce, urchins are abundant and kelp is very sparse as a result of browsing by the urchins.

The California coast was once almost completely fringed by kelp forests, but they were severely depleted by the early years of the twentieth century and have only recently begun to recover. The primary reason for the kelp forest decline is now understood to be hunting of sea otters during the eighteenth and nineteenth centuries, which drove the otters almost to extinction. Once the otters were removed, sea urchin populations increased and steadily overwhelmed and destroyed the kelp forests. Now sea otters are protected and their populations are slowly recovering. As a result, sea urchin populations are declining and kelp forests are gradually returning, although their future is threatened by climate change-driven warming of ocean waters. The sea otter is called a "keystone predator" because, without it to prey on urchins, the urchin population growth produces large and fundamental changes in the ecosystem. Keystone predators are found in many ecosystems.

This page titled 15.3: Kelp Forests is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.4: Rocky Intertidal Communities

The rocky intertidal community that lives between the **high-tide line** and the low-tide line on rocky coasts is unique in two important respects. First, this community can be easily observed and studied by anyone who visits a rocky coast at low tide. Second, the community is normally arranged in well-defined depth zones that run parallel to the shore at different heights above the low-tide line. Each zone is populated by different species. Species that live in each zone are determined by a combination of physical conditions of the environment and competition among species that have different tolerances to these conditions.

The most important characteristic of the rocky intertidal zone is the degree of exposure of the substrate to the atmosphere when the tide recedes. Unlike organisms that inhabit sandy **beaches** or mudflats, most rocky intertidal species cannot bury themselves in the rocky substrate to survive periods when they are exposed by the tide. Once exposed to the atmosphere, organisms are subjected to a number of stresses, such as:

- Temperature variations and extremes that far exceed those in seawater
- Variable water exposure (spray or rain) and **humidity**, and the consequent variable tendency to lose body fluids by evaporation
- Variable salinity because the remaining pockets or pools of water may be subject to evaporation and/or dilution with rainwater or snow
- Variable oxygen concentrations and **pH** because physical and biological processes quickly alter these conditions in the small volume of seawater that remains in contact with the organism, and this water is not renewed until the water level rises again
- Predation by birds and terrestrial animals

Zonation of communities on a rocky intertidal shore is critically dependent on the frequency and duration of their exposure to the atmosphere and, therefore, to these stresses (Fig. 15-11).

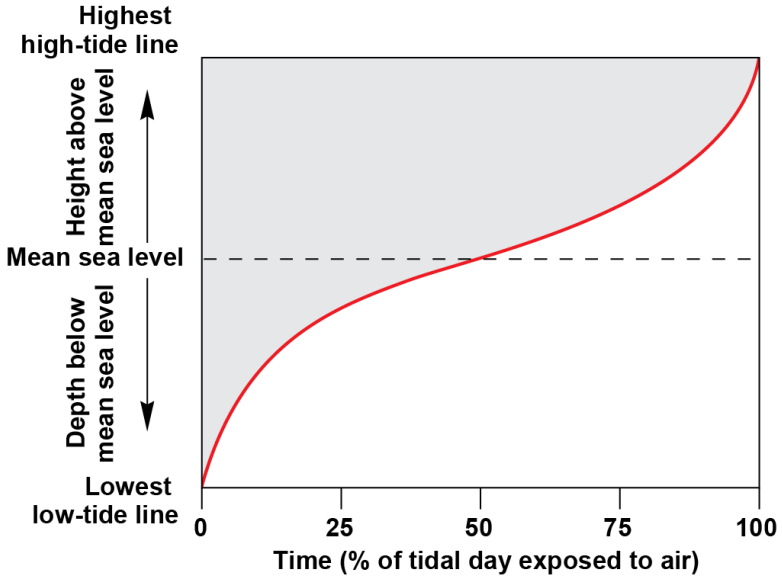


Figure 15-11. The length of time that a particular depth in the intertidal zone (and the organisms that live at that point) are exposed to the air during the tidal cycle varies with height above the low-tide line.

The shore can be divided into four zones on the basis of atmospheric exposure time (Fig. 15-12). The highest zone, the **supralittoral zone**, is above mean higher high water (Chap. 10) and thus essentially permanently out of the water, but it is frequently wet with seawater spray from breaking waves, at least during high tides. Below the supralittoral zone, the **high-tide zone** is covered with water for parts of the tidal cycle but remains exposed to the atmosphere most of the time. The **middle-tide zone** is usually covered by water, but it is exposed to the atmosphere during all or most low tides. Finally, the **low-tide zone** is water-covered almost permanently and exposed to the atmosphere only briefly during the lowest low tides. The boundaries between these zones are sometimes sharp, as revealed by abrupt changes in the biological communities that they support.

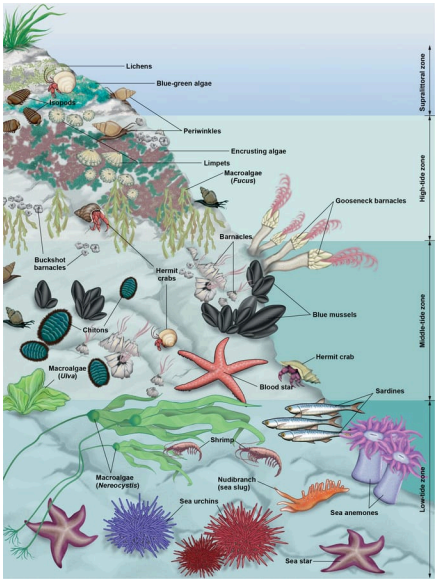


Figure 15-12. The rocky intertidal shoreline can be separated into four zones based on length of time exposed to the atmosphere. Many species have ecological niches that restrict them to only one of these zones, although closely related species may occupy different zones. The flora and fauna in this figure are typical of a mid-latitude, cold-water, rocky shoreline. Generally, many more species occur in each zone than are shown here. Competition between species often determines the vertical limits of a particular species' distribution within the zone that it occupies.

Even when covered by water, the zones of the rocky shore are subject to different physical conditions, particularly wave-induced **turbulence** and scour. Wave-induced turbulence decreases with depth but can still be substantial well below the low-tide zone.

Rocky coasts are present throughout the oceans, and each has its own unique communities with different species and species interactions. Consequently, only the general types of **flora** and **fauna** in rocky intertidal zones are described here. For most coastal areas, easy-to-read guides to shore and **tide pool** creatures can be found in local bookstores.

Supralittoral Zone

The supralittoral zone is either permanently exposed to the atmosphere or covered by water only during occasional extreme high tides or **storm surges**. Hence, this zone can be considered land and not part of the ocean ecosystem. Nevertheless, the continuous or frequent spray of seawater that reaches this zone provides moisture and nutrients that support the growth of **lichens**, encrusting **blue-green algae**, and small tufts of various green algae. These autotroph communities are sparse but provide food for a variety of animals, including species of periwinkles (marine snails), other marine snails, **limpets**, **isopods**, and crabs (**Fig. 15-12**).

Periwinkles, other snails (**Fig. 15-13a**), and limpets are grazers that feed on algae encrusting the rocks. The periwinkles are often species of the **genus** *Littorina*. Many *Littorina* species are well adapted to life in the supralittoral zone. Some "breathe" air and may even drown if fully immersed in water for long periods. Supralittoral species of *Littorina* are **viviparous**, giving birth to live young. In contrast, the marine species of this genus that live completely submerged deposit eggs on the rocks or release them to the water. Periwinkles can completely withdraw into their shell and seal off the opening with an operculum, a rigid disk of hornlike material. This seal is watertight and provides protection from predators and especially from dehydration during long periods of exposure to the atmosphere.

Figure 15-13. Animals and algae of the rocky intertidal zone. (a) Group of black turban snails (*Tegula funebralis*, Monterey, California). (b) A limpet (order Archaeogastropoda, California). (c) Barnacles in the high intertidal zone, often called buckshot barnacles. These are probably primarily *Chthamalus sp.*, with some interspersed *Balanus glandula*. (d) Lined chiton (probably *Tonicella lineata*, California). (e) Dense bed of California mussels (*Mytilus californianus*, Monterey, California). (f) These green algae are in the upper intertidal zone and have been out of water and in the sun for several hours. The algae survive this experience on a regular basis with no harm (south of San Francisco, California). (g) A hermit crab (*Pagurus sp.*, California) living in its borrowed gastropod shell.

Limpets (**Fig. 15-13b**) are not able to withdraw entirely into their shell. Instead, they cling to the rocks with their bodies and pull their shells tightly down on top of them. If the shell edge fits well with the surrounding rock, the animal is sealed inside, where it is protected from predators and desiccation. Some limpets rasp the rock away or release acidic secretions to create a perfectly fitted platform for their shell. When immersed or wet, limpets may move over the rocks to graze, but they return to their prepared location during periods of prolonged exposure to the atmosphere. The limpet's bond to the rock is so strong that it cannot easily be turned over, and thus its soft body is protected from predators such as birds. The limpet's defensive action of clinging tightly to the rock is easy to observe. Limpets can be surprised and easily removed from their rock location by a quick sideways stroke of a chisel or blade. However, if the limpet is warned, perhaps by a light tap from a finger, it clamps down and is impossible to remove without breaking the shell.

Isopods often live in the supralittoral zone in great numbers. These animals are **scavengers** that feed on organic debris. They are rarely seen by most visitors to the shore because they remain in hiding places within the rocks by day and emerge to feed only at night. Various species of crabs also inhabit the supralittoral zone. Many are scavengers, but some are grazing **herbivores** or predatory **carnivores**.

Each zone of the rocky intertidal ecosystem varies in width according to factors that include the slope of the rocks and the **tidal range**. The width of the supralittoral zone is also affected by factors that determine the degree of wetness of the rocks, including the intensity of wave action, the average air temperature and humidity, and the location, roughness, and orientation of the shore (which control the degree of shading from the sun). For example, the supralittoral zone is wide in areas such as central California and Maine, where cool, damp, and foggy days are common.

High-Tide Zone

In the high-tide zone, organisms are exposed to the atmosphere for long periods and, like inhabitants of the supralittoral zone, must be able to withstand extremes of temperature and salinity and must be protected from dehydration. In addition, these organisms must be able to withstand severe wave-induced turbulence. To offset these disadvantages, a reliable supply of nutrients, plankton, and suspended organic particles becomes available each time the rocks are covered by ocean water. Consequently, the high-tide zone can support species that cannot tolerate the limited nutrient supply and prolonged dry periods of the supralittoral zone.

Rocks of the high-tide zone support many species of encrusting algae, particularly in the upper parts of the zone. In contrast to the supralittoral zone, the high-tide zone has several species of macroalgae. All these species have thick cell walls, which give

them a leathery feel and protect them from excessive loss of water by evaporation when exposed to the atmosphere (**Fig. 15-13f**). They are firmly attached to the rocks by holdfasts, and their stipes (branches) are extremely flexible, enabling them to withstand wave turbulence. Although they are eaten by some species of snails, crabs, and other animals, these algae are tough and difficult to digest, and they contribute to the food web primarily in the form of detritus. The detritus is formed when parts or all of the algae are first broken loose from the rocks by waves or grazers and then broken down or modified by decomposers.

In many areas, the top of the high-tide zone is marked by a band dominated by small **barnacles** called "buckshot barnacles" (**Fig. 15-13c**). Several larger species of barnacles are present lower on the shore in the middle-tide zone. Because barnacles are suspension feeders, they cannot live above the highest high-tide line. However, they can survive in locations where they are able to feed for only a few hours on the few days of spring tides each month when they are immersed in water. When exposed to the atmosphere, the barnacle withdraws into its hard shell, which protects it from dehydration, predators, wave impacts, and wave-induced turbulence.

Because organisms of the high-tide zone, like species in the supralittoral zone, must withstand extended exposure to the atmosphere, the two zones sustain many similar species, including periwinkles and limpets. Toward the lower end of the high-tide zone, periwinkles and limpets become less abundant, whereas chitons and mussels become more abundant. This change marks the transition from the high-tide zone to the middle-tide zone. Chitons (**Fig. 15-13d**) feed and attach themselves to the rocks in the same way as limpets, but they have shells made up of eight separate connected plates. Mussels (**Fig. 15-13e**) are suspension feeders like barnacles, but they have a two-piece shell attached to the rocks by a network of strong threads called "byssal threads." These threads are formed by a liquid secreted from the mussel's foot that hardens in seawater. The threads are attached between the mussel shell and the rocks or, in dense mussel beds, between one shell and another. Periwinkles, barnacles, limpets, mussels, and other species living attached to rocks of the high- and middle-tide zones tend to have rounded shells, which can best withstand and dissipate the turbulent impacts of waves.

The lower limit of the zone inhabited by a rocky intertidal species is determined for many species by competition from other species. In contrast, the upper limit of the inhabited zone is normally determined by the tolerance limits of the species' fundamental niche.

Middle-Tide Zone

Macroalgae are generally less abundant or absent in the middle-tide zone because of competition by mussels and barnacles. In a middle-tide zone newly formed by vertical movements of the coast during earthquakes or by **lava** flows, or in a middle-tide zone partially denuded by extreme storms, macroalgae quickly establish themselves and cover the rocks. However, as mussel and barnacle **larvae** settle and grow into new colonies, the macroalgae are steadily overcome and eventually disappear from the zone.

Within the middle-tide zone, the lower limit of mussel beds is determined by competition from predatory sea stars, which can grip and slowly pull open a mussel with the many tube feet on their undersides. As you can see in **Figure 15-14**, mussel beds terminate abruptly at their lower limit on many shores, almost as though the mussels were incapable of growing below that depth. However, this limit is simply the depth at which recolonization by mussels is less effective than predation by sea stars (whose fundamental niche does not extend as high up the middle-tide zone as the fundamental niche of the mussels does). Although some sea stars do prey on mussels within the mussel bed zone, they must withstand stresses associated with being at the limits of their fundamental niche (e.g., atmosphere exposure and turbulence), or they must migrate up and down the shore's zones with the tides. Hence, mussels are able to outcompete the predatory sea stars in this upper zone.

Figure 15-14. On this shoreline at Rialto Beach in Washington State, you can see the zonation of species in the middle tide zone. There is a band of barnacles (the light greenish gray) above a band of mussels (black and shiny). The band of mussels extends down to a depth determined by sea star predation. Sea stars cannot survive the long air exposure between tides that occurs in the mussel zone.

The mussel beds on many rocky intertidal shores are ideal **habitat** for a variety of algae and animals that include **hydroids**, worms, snails, clams and other mollusks, and crabs and other **crustaceans**. Acorn barnacles are interspersed in the mussel beds. These barnacles, and to a lesser extent the mussels, are eaten by snails that drill through the barnacle plates or mussel shell or force the mussel shell open to get to their prey. The middle-tide zone is also populated by numerous species of **hermit crabs** (**Fig. 15-13g**) and a number of species of anemones that can withstand periodic exposure to the atmosphere.

Low-Tide Zone

The low-tide zone sustains a variety of macroalgae and encrusting algae nourished by nutrients brought to them with each tide. In this shallow zone, algae have ample light, even when turbidity is relatively high. In addition, because they are exposed to the atmosphere for only short periods at low tides, they do not need the protection against desiccation that algae in higher zones require. Although the low-tide zone, unlike zones higher on the shore, is dominated by algae, it sustains numerous species of animals, many of which use the abundant mats of algae for shelter. Animals in the low-tide zone are similar to those of the kelp community (**Fig. 15-9**) and include anemones, **sponges**, sea urchins, nudibranchs, shrimp, sea stars, crabs, sea cucumbers, and fishes. These organisms are infrequently exposed to the atmosphere, so they include delicate forms that would be dehydrated or thermally shocked if exposed for more extended periods.

The principal physical hazard in the low-tide zone is wave-induced turbulence. Some species withstand this turbulence by attaching themselves to the rocks. Other species are active swimmers or simply use macroalgae or cracks and holes in the rock as protection from wave action. Because the low-tide zone has abundant macroalgae and a continuous supply of suspended detritus and plankton, the low-tide community comprises many species of **filter feeders**, detritus eaters, scavengers, grazers, and carnivores.

Tide Pools

On many coasts, the rocky shore is convoluted or pitted sufficiently that seawater remains in many depressions, even when the tide recedes. These depressions are called “tide pools.” Depending on the size, location, and permanence of the tide pool, any of the species of any part of the rocky shore, from the high-tide zone to the low-tide zone, may be present.

Each tide pool is unique because evaporation, rainfall, solar heating, winter cooling, and other factors affect the physical properties (e.g., salinity, temperature, pH, oxygen concentration) of the water in each tide pool differently. Tide pools high on the shore are isolated for long periods between tides and are subject to the greatest changes. Deep tide pools have a greater volume of seawater per unit area than shallow ones and are less affected by evaporation, rainfall, heating, and cooling. Consequently, small tide pools tend to support only microscopic algae and highly tolerant **copepods** and other microscopic animals. In contrast,

large tide pools may contain many species of algae, sea urchins, anemones, crabs, shrimp, small fishes, and other animals that are tolerant to relatively small changes in salinity, temperature, and other physical characteristics of the tide pool water. The next time you visit a shore with tide pools, you can see which species are more tolerant by simply examining several different tide pools.

This page titled 15.4: Rocky Intertidal Communities is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.5: Sargasso Sea

The Sargasso Sea is the region of the North Atlantic Ocean surrounded by the North Atlantic **subtropical gyre**. The permanent thermocline in the Sargasso Sea is steep, persistent, and deep (**Chap. 8**). Hence, upwelling does not occur, and the surface layer is depleted of nutrients. Furthermore, there is little exchange between Sargasso Sea surface waters and adjacent surface waters from which nutrients could be resupplied. Consequently, primary productivity is very low, and we would expect the Sargasso Sea to be an ocean "desert." The phytoplankton-based **food chain** in the Sargasso Sea is indeed very limited. However, parts of the Sargasso Sea surface are covered by dense mats of floating brown macroalgae (**Fig. 15-15a,b**), within which lives a diverse community of animals. These mats often cover areas of many square kilometers.

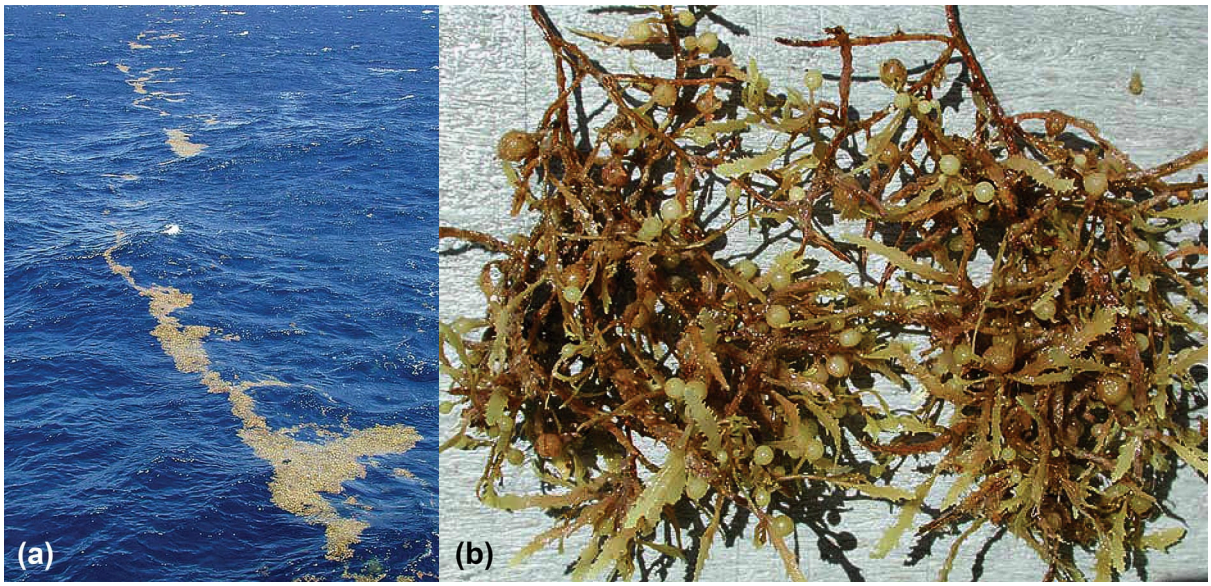


Figure 15-15. The *Sargassum* community. (a) *Sargassum* weed floats on the Sargasso Sea surface in rafts that can exceed hundreds of meters in diameter. (b) *Sargassum* weed. Note how thick the mat of algae fronds is. This makes *Sargassum* an ideal place for many small fishes and invertebrates to hide.

Why do these vast quantities of algae grow in what should be a desert? The answer illustrates the infinite adaptability of nature. The brown algae in the Sargasso Sea are species of the genus *Sargassum*. These species grow extremely slowly and are thought to have lifetimes of decades or even centuries. The very small amounts of nutrients available in Sargasso Sea surface waters are adequate to sustain the very slow growth of *Sargassum*. However, the critical factors that enable *Sargassum* to populate the Sargasso Sea so heavily are very weak currents and very limited wind mixing. Because the surface currents are weak and flow in directions that are oriented partially toward the center of the subtropical **gyre**, they tend to retain the floating algae within the Sargasso Sea, and the mats of the long-lived algae tend to stay together. The algal population therefore can develop a large biomass despite its very slow growth and reproduction rate. In addition, because wind mixing is very limited, nutrients released into the water column by the consumers and decomposers that live in the *Sargassum* are not rapidly diluted, and are available to be recycled back into new *Sargassum* growth.

Within the *Sargassum* mats lives a community of many different species of fishes, crabs, snails, and other animals, including a frogfish called the sargassumfish. Many in this community have a body that mimics the structure and color of the *Sargassum*, and they can cling to the branches of the algae unnoticed by predators or prey that swim by. In contrast to many other ocean ecosystems, the *Sargassum* community has relatively few grazers and a relatively small animal biomass in relation to the algal biomass. This composition reflects the evolution of the community to a stable condition. If there were large populations of grazers or other animals, the algal biomass would be rapidly reduced to the point where the animal populations could not be sustained. The animals that have survived in this community include many long-lived, slow-growing species with sedentary lifestyles. Each of these characteristics represents an adaptation to minimize food intake needs.

Although some species similar to those found in the Sargasso Sea are found within the interior of other subtropical gyres, for reasons as yet not fully understood, none of these other areas sustain the high biomass found in the Sargasso Sea.

This page titled 15.5: Sargasso Sea is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.6: Polar Regions

The marine ecosystems of the north and south polar regions differ from one another physically and biologically because of the configuration of the landmasses and because they are separated by warm tropical waters through which most cold-adapted marine species cannot transit. Nonetheless, the two ecosystems have a number of environmental similarities, including extreme seasonal variation in light availability, generally low surface water temperatures, and seasonally variable sea-ice cover.

In both polar regions, cooling of surface ocean waters and vigorous wind mixing caused by storms formed at the polar **front** (**Chap. 7**) prevent the formation of a permanent thermocline and promote vertical mixing. Therefore, nutrients are generally abundant, especially in surface waters of the upwelling region between the Antarctic **Convergence** and the near-coastal east-wind drift current that flows around Antarctica. Primary productivity in this region apparently is limited by the availability of **micronutrients**, particularly iron, even when light is ample, and nitrogen, phosphorus, and silica are abundant. Iron concentrations are low in this region because there is a very low supply of river inputs and atmospheric dust in this region.

Special Characteristics of Arctic Marine Environments

In the Northern Hemisphere, nutrients are abundant in surface waters of the **marginal seas** that surround the Arctic Ocean, particularly the Bering, Norwegian, and North Seas. Nutrients are less abundant in seasonally ice-free surface waters of the Arctic Ocean itself, because freshwater **runoff** and **ice exclusion** lower surface salinity and establish a strong **halocline** that inhibits the vertical mixing of higher-salinity, nutrient-rich deep water into the surface layer. In addition, the permanent ice cover of much of the Arctic Ocean and its location in an atmospheric **downwelling** zone (zone of weak winds) minimize wind-induced vertical mixing.

In coastal regions of the Arctic Ocean, nutrients are generally available in relatively high concentrations during summer because they are supplied to some extent in freshwater runoff. In addition, nutrients are returned to the water column through decomposition during the darkness of winter, when these regions are covered by ice. Hence, nutrients are readily available during at least the early part of the short spring–summer ice-free season.

Common Characteristics of Arctic and Antarctic Marine Environments

In both the Arctic and the Antarctic, substantial populations of microscopic ice algae live, or form resting phases, within liquid pockets in the ice or on the underside of the ice. During spring and summer, ice algae can grow in or under the ice as the light increases in intensity and begins to penetrate the ice. As the ice melts, the resting spores of many species of phytoplankton are released into the water column, where they grow rapidly in the nutrient-rich and now well-illuminated open water. Both the Arctic and the Antarctic have a zone of maximum productivity that coincides with the edge of the floating ice. Each spring, this zone moves poleward with the melting ice edge.

Several important characteristics of polar ecosystems determine the species that are able to live there. First, because nutrients are generally available, phytoplankton usually grow quickly and are abundant when light is also available. Second, the extreme seasonal variation of light intensity and duration, and in some areas the extent of sea-ice cover, limit the period of ideal conditions for phytoplankton growth to only a few weeks or months in summer. Third, upwelling and wind mixing, which supply nutrients to surface waters and at the same time affect the **residence time** of phytoplankton in the nutrient-rich photic zone, are particularly variable because of eddies, turbulence induced by seafloor topography, and especially changes in weather and **climate**.

As a result of the unique physical characteristics of their environment, many animals that live in polar regions must migrate seasonally or be able to obtain all the food they need for the year during the short productive summer period. In addition, they must be capable of ensuring the survival of their species during years when climatic extremes may result in a partial, or even total, loss of their food supply. The evolutionary response to these constraints is for the animals to be long-lived, to reach sexual maturity late, and to bear only one or a few offspring each year for several years. This response ensures that the species will survive if one or more successive years are poor years in which none of the offspring survive. In addition, the animals must be able to store the food energy needed to survive the winter when food is unavailable or to migrate to warmer regions for the winter. The food energy is stored as fat. Many species that live in polar ecosystems, including certain fishes, marine mammals, and birds, have these characteristics.

Many of the characteristics shared by polar animals make them more susceptible than the marine species of lower latitudes to the adverse effects of certain chemical compounds that can cause **contamination**. Because fat-**soluble** chemical compounds are concentrated in an animal's fatty tissues, they can accumulate for several years before polar animals reach reproductive age. This buildup maximizes the possibility of **carcinogenic**, **mutagenic**, and **teratogenic** effects in the species or its offspring (**CC18**). Fortunately, the Arctic and Antarctic marine ecosystems are substantially less contaminated than marine ecosystems at lower latitudes.

Antarctic Communities

The **pelagic** food web of the Southern Ocean is described in **Chapter 12** (**Fig. 12-12**). It consists of many species of whales, seals, and penguins that feed on the abundant **krill** and other zooplankton (many of which can be found at the ice edge feeding on phytoplankton populations). Most of the penguin and seal species live year-round in Antarctic waters and haul themselves out on the continent or one of the nearby islands to breed and bear their offspring (**Fig. 15-16**). Like many polar species, most of these mammal and bird species mature only after they are several years old, normally have one or two offspring per year for several years, and build up heavy layers of fat during summer, when food is abundant.

Figure 15-16. Some examples of seal species that inhabit the Antarctic ecosystem. (a) Crabeater seals (*Lobodon carcinophagus*, Antarctic Peninsula) have specialized teeth (not shown in the image) to filter out krill, their principal food source. They are somewhat misnamed since they do not eat crabs. (b) Antarctic fur seals (*Arctocephalus gazella*, South Georgia Island), heavily hunted for their fur during the 18th and 19th century, feed mostly on krill squid fish and penguins, (c) Leopard seal, (*Hydrurga leptonyx*, Antarctic Peninsula) a voracious predator that will eat krill, fish, squid, penguins and other seals.

Although fat layers act as insulation against the Antarctic cold, particularly when the animal is on land, their most important function is to provide energy during the period when the animal is not feeding. For example, the majority of whale species in the Antarctic are **baleen** whales that visit the region only during summer, when food supplies are abundant. The huge store of fat that they build up during this time is used to supply them with energy during the remainder of the year, when most of these species migrate to breeding grounds in the tropical or subtropical ocean. During the migration and breeding season, the whales feed little or not at all.

Human hunting has dramatically reduced the populations of many seal and whale species in the Antarctic (**CC16**). These species are now protected, and most populations are showing signs of slow recovery.

Although fishes of the Antarctic ecosystem are much less well studied than the marine mammals and birds, we know that a high proportion of these species are present only in Antarctica and appear to be adapted to the cold and seasonally variable food supply in much the same way that Antarctic marine mammals are. Some species have evolved a unique blood chemistry that enables them to live at temperatures below freezing. The natural antifreeze of these fishes is the subject of considerable research because it could have commercial and medical applications.

Arctic Communities

The biological populations of the Arctic region are similar in some ways to those of the Antarctic, particularly in their concentrations of marine mammals. However, species found in the Arctic are different from those found in the Antarctic. Many seal species live year-round in the Arctic and its adjacent seas and haul themselves out on land to breed, as other seal species do in the Antarctic. Most northern whale species migrate between their Arctic feeding grounds and tropical or subtropical breeding grounds. Penguins and leopard seals are not present in the Arctic. However, the Arctic is populated by polar bears and walrus (**Fig. 12-27d**), which are not present in Antarctica. Polar bears are voracious predators and superb swimmers, but they are land animals. However, because they range across the sea ice to hunt seals, they are primarily dependent on food from the marine environment, and so they are considered a part of the marine ecosystem.

Susceptibility to Climate Change

Global climate models all predict that climate changes will be amplified in the polar regions, and this conclusion is supported by historical data and by observations during the past several decades. The reasons for this special susceptibility are many and complex, but they include the positive feedback (**CC9**) due to reduced snow and ice cover. A covering of snow and ice is a good reflector of the sun's energy. Warming reduces snow and ice cover, which increases the amount of the sun's energy absorbed by land and ocean, causing further warming.

The polar climate has warmed during the past 30 years, resulting in substantial reduction in the area of permanent sea ice in the Arctic Ocean, and causing ice sheets to be reduced in size in the Antarctic, and glaciers to retreat in both polar regions. If sustained, this trend will not only cause sea level to rise worldwide, but will also have profound effects on polar biological communities. For example, polar bear populations are already adversely affected and are likely to be devastated because the bears hunt mostly on floating summer sea ice, which is steadily reducing in area and retreating farther from the coast. Similarly, many species of seals and penguins are already and likely will be much further affected by changing ice and snow cover in their traditional breeding grounds on the Antarctic coast. Furthermore, changes with unknown effects are likely to take place because a reduction in ice cover favors phytoplankton production over ice algae production in both polar regions. Profound changes in both the Arctic and Antarctic marine ecosystems already appear to be well underway.

This page titled 15.6: Polar Regions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.7: Beyond the Sun's Light

Below the surface layers of the oceans, organisms must be adapted to extremely low light levels, to uniformly low temperatures, to a detritus-based food supply that decreases steadily with depth, and to increasing pressure. These factors interact to give pelagic and benthic communities of the **bathyal zone** and **abyssal zone** characteristics that are very different from those of communities that live near the surface.

Relatively little is known about bathyal and abyssal communities because the vast volume of the ocean that they inhabit has been visited only for fleeting moments by research **submersibles**, which many denizens of the deep undoubtedly avoid. More is being learned from ROVs but even ROVs are expensive to operate at these depths, and likely also avoided by many species, so progress is slow. In addition, collecting samples from the deep oceans is very difficult and expensive.

The creatures that live in the deep oceans zones range from rather familiar forms that resemble fishes and invertebrates of the shallower ocean to incredibly bizarre-looking creatures that would be well suited to science fiction movies. **Figure 15-17** shows just a small selection of the fishes that inhabit the **aphotic zone**.

Figure 15-17. Mid-water fish species are often very strange-looking and have unusual adaptations, such as unhingeable jaws and hugely expandable stomachs, strange eyes, lures, and light-emitting photophores. Many of these adaptations aid these species to cope with the scarcity of food and difficulty of locating mates and prey.

Organisms living below the photic zone have four potential sources of food: particulate detritus that sinks slowly through the water column, carcasses of large animals that sink rapidly to the ocean floor because of their size, prey species that live in the aphotic zone, and prey species that live in the photic zone above. Of these, only prey species in the photic zone are abundant, so many deep-water species migrate to this zone to feed. All other deep-ocean biota must be adapted to a low and uncertain food supply.

Many species of crustaceans (such as shrimp, copepods, and **amphipods**), other invertebrates (such as squid), and numerous fish species live in the part of the water column immediately below the photic zone, between about 200 and 1000 m. Many of these species migrate vertically up into the photic zone at night to feed and then return to the depths during the day. Other species prey on species that live in their own depth zone. Vertical migrants include the unique *Nautilus* (**Fig. 12-24e,f**), which has a chambered shell to provide buoyancy. To avoid problems with pressure changes as the *Nautilus* migrates vertically, the rigid internal buoyancy chambers contain gas at a low pressure. To adjust its buoyancy it pumps small amounts of water into and out of its internal chambers. Since this increases or decreases its mass, and density is mass divided by volume, this allows the nautilus to control or change its depth. Oceanographers profiling floats achieve the same buoyancy control by applying the same principle but changing their volume rather than their mass by pumping small amounts of oil between their interior and an external expandable membrane.

There is very little or no light below 200 m, and no red light penetrates to this depth. Many organisms of this zone are red or

dark colored (**Figs. 15-17, 15-18**), so they do not reflect any of the ambient light, and this makes them difficult to see in the near darkness. However, many fishes and other species that live at these depths have greatly enlarged eyes to enable them to hunt prey in the dim light.

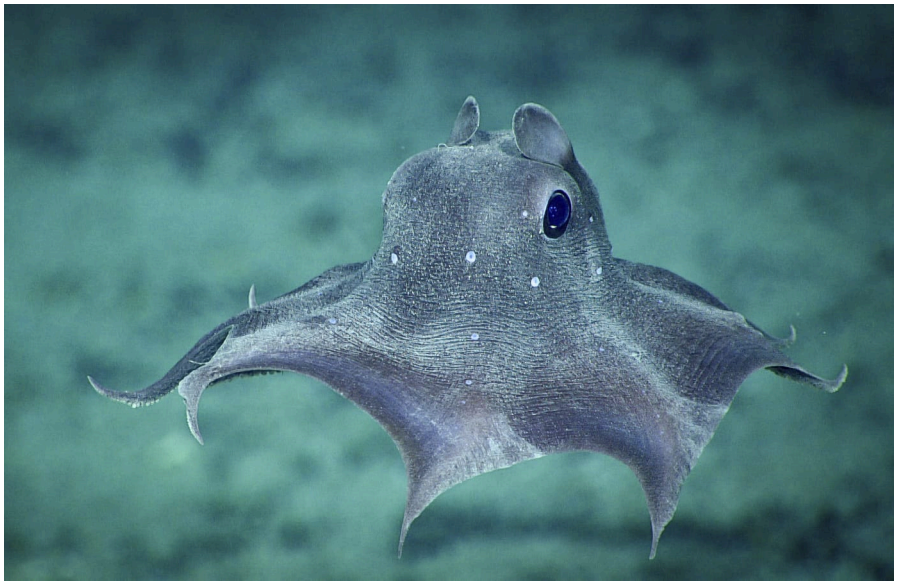


Figure 15-18. "Dumbo" octopuses named for their floppy ear-like fins that they use to move through the water, are members of a genus of octopi called *Grimpoteuthis*. The largest Dumbo octopus so far recorded was about 1.8 m in length and weighed about 5.9 kg. They live in the hadal and abyssal zones and have been observed as deep as 7000 m in the Java Trench. As food is rare they eat almost anything including worms, crustaceans, mollusks, and small fish. Unlike other octopuses, they can not swim by jet propulsion and do not produce ink to evade predators. These are likely adaptations that reduce metabolic energy expenditures needed for these functions in response to the relative absence of predators and light for them to hunt by, and the scarcity of food supplies for the octopuses themselves.

The easiest way to hunt visually for prey in waters below about 200 m where light still penetrates is to look upward into the very dim light filtering down from above and search for the dark silhouette of the prey species. Consequently, many fish species of this zone have eyes that look directly upward. To counter this hunting strategy, many prey species have a series of light-producing organs called "photophores" arrayed along the underside of their body. By illuminating their underside with these photophores, they can reduce the sharpness of their silhouette and blend better into the dimly lit background above. Many species also have photophores arrayed on other parts of their body, presumably as devices to identify members of their own species or to attract prey species.

With increasing depth below 1000 m, fewer and fewer species are present that migrate vertically. Many organisms are brightly colored, although, in the absence of light, color does not advertise their presence and so is irrelevant. Eyes become less prominent and absent in many species, but some species that live in the absolute darkness of the deep oceans do have eyes that, at least in some fishes, are adapted to detect bioluminescent light produced by many animal species and bacteria in this otherwise lightless zone, likely for recognition or mate attraction, and used by predators for prey location. Adaptations, found in some fish species suggest that they may be able to detect a single photon of bioluminescence and even to distinguish between wavelengths (see colors) which no other animals are known to be able to do at such low light levels.

Because of the low abundance and widespread dispersal of organisms, all fishes of mid and deep waters must be adapted to take advantage of any prey species they encounter. Therefore, many of these fishes have unhingeable jaws and soft expandable bodies, so they can swallow and digest prey species as large as, or larger than, themselves. Some also have highly sensitive chemosensory (chemical-sensing) systems that allow them to detect even slight changes in chemical traces coming from live or decaying prey.

On most of the deep-ocean floor, the food supply comes from above as a rain of detrital particles and as occasional carcasses of large animals. Much of the detrital material has already been subjected to substantial decomposition during its slow descent through the water column. Hence, what remains is primarily material that is difficult to digest and that decomposes only very slowly. This material has little immediate food value for animals. This detritus is normally consumed by **bacteria** in the surface sediment, and it is the bacteria that become food for larger animals. Because the suspended particulate food supply near the deep seafloor is at very low concentrations and is of low nutritional value, there are few suspension feeders. Most animals of the deep seafloor are deposit feeders, including many species of sea cucumbers and **brittle stars**. The particulate food supply in

some areas near the continents may be supplemented by detritus carried to the deep seafloor in **turbidity currents**.

Although we have little direct information about such events, it is known that bodies of large animals must sink rapidly to the seafloor with some frequency. When bait is lowered to the deep seafloor, a variety of fishes, sharks, crustaceans, and other invertebrate scavengers (**Fig. 15-17**) arrive to feed on the bait within as little as 30 min. Once the bait has been eaten, these animals disappear into the darkness. Some bait, such as a massive whale fall can create an ephemeral, yet long lasting ecosystem that can provide energy for decades. They can also act as "stepping stones" across the barren seafloor, providing energy for the gathering scavengers to disperse and colonize new areas. How these scavengers find the bait or their normal food is not known, but the speed at which they appear suggests that they must have extremely sensitive chemosensory organs for this purpose. It has been observed that many of these scavengers can sense, but cannot find, bait if it is suspended even a meter or less above the seafloor, which demonstrates that they are completely adapted to feeding on material that lies on the seafloor.

The speed with which food is found by deep-ocean scavengers after it reaches the seafloor indicates fierce competition for food, because food does not decompose rapidly at such depths. Bacterial decomposition is inhibited by high pressures, so food items that fall to the deep-ocean floor would decompose only very slowly if they were not immediately consumed. Inhibition of bacterial decomposition by high pressure was first discovered when the research submersible *Alvin* sank unexpectedly in more than 1500 m of water. A lunch box containing an apple and a bologna sandwich was inside the submersible when it sank. A year later, when *Alvin* was recovered, these food items were wet but almost undecomposed. They did decompose within weeks after their return to the surface, despite being refrigerated. Inhibition of bacterial decomposition by high pressures has been confirmed by a number of experiments conducted since the *Alvin* discovery.

This page titled 15.7: Beyond the Sun's Light is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.8: Hydrothermal Vents

Until 1977, it was thought that all areas of the abyssal oceans were biologically impoverished because of the limited availability of food that rains down from above. However, in 1977 the research submersible *Alvin* made a number of dives on the Galápagos Ridge that would change those ideas. The purpose of the dives was to study the geology and chemistry at this **oceanic ridge**. The researchers were looking for evidence that the high heat flow through the seafloor at the center of the ridge creates hydrothermal circulation. In hydrothermal circulation, seawater sinks through sediments or cracks in rocks of the seafloor and is heated, convected upward, and vented to the water column to be replaced by more seawater drawn through the rocks and sediments (**Fig. 15-19a**, **Chap. 6**). The heat comes from the upwelling **magma** and cooling volcanic rocks beneath the seafloor.

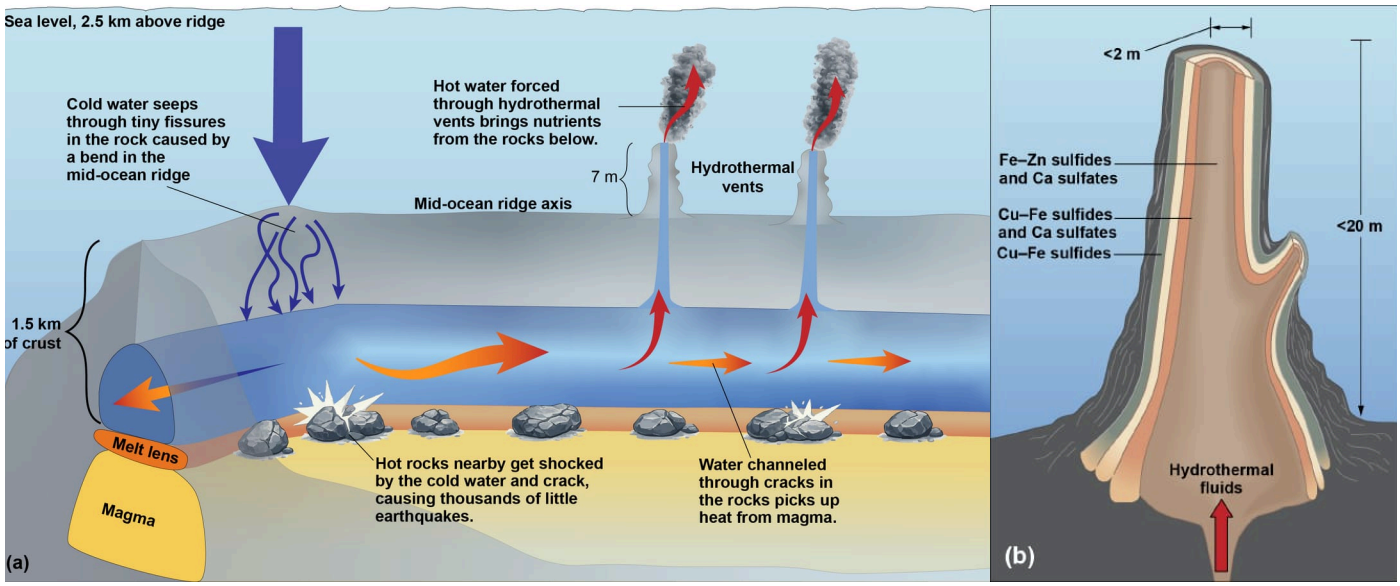


Figure 15-19. Black smoker formation and composition. (a) Hydrothermal vents on the oceanic ridges are believed to be formed because heat from magma below the oceanic ridge sets up a convection circulation. In this circulation, water from below the seafloor is heated and rises at the vents and is replaced by water drawn through the cracks in the rocks and sediment on the flanks or an adjacent section of the ridge. (b) At black smoker vents, the discharged water contains no oxygen but does contain high concentrations of iron, manganese, and other sulfides. As water is vented into cold ocean water, metal sulfides are deposited around the vent and often form a chimney-like structure.

The researchers found much more than they expected. In fact, their findings may represent one of the most surprising and profound scientific discoveries ever made. They found not only hydrothermal vents, but also dense communities of marine organisms surrounding those vents. The communities include **tube worms**, clams, mussels, and many other invertebrates, most of which belonged to previously unknown species and many of which were very large in comparison with similar known species. The biomass in these vent communities is hundreds of thousands of times greater than that in any other community at comparable depths in the ocean. However, even this finding of abundant oases of life in the "desert" of the abyss was to prove less surprising than the subsequent discovery that these communities do not depend on photosynthesis. In fact, they were found to be dependent on chemosynthesis for primary production of their food.

Hydrothermal Vent Environments

Initially, hydrothermal vents were thought to be rare and to occur only on fast-spreading oceanic ridges. We now know that hydrothermal vents and their associated biological communities are present in many locations, dispersed irregularly along the oceanic ridges in all oceans. Vents have even been found on the ultraslow-spreading Gakkel Ridge in the Arctic Ocean.

Vents have also been found on the submerged volcanoes at island arc **subduction zones** such as the Mariana Arc. These **back-arc** volcano vents are especially interesting to scientists because many are at much shallower depths than the oceanic ridge vents. The shallower depth makes them much easier to study, and the fluids they discharge are dispersed in the upper layers of the oceans, where they may have greater immediate effects on marine species that live in, or migrate periodically to, the photic zone. Some estimates now suggest that vents are abundant enough that a volume of ocean water equal to the entire volume

of the world oceans may be processed through high-temperature vents about every 10 million years, which is a relatively short period in geological time.

Each vent differs from the others in terms of the temperatures and chemical characteristics of the water it discharges. Two general types of vents are known to occur along the oceanic ridge axis. Vents of the first type, called black smokers, discharge hot water, usually about 360°C (but can be up to about 460°C) The second type of vent, called white smokers discharges somewhat cooler water (about 260 to 300°C), usually at flow rates lower than those at black smokers.

The water discharged by black smokers contains no oxygen or nitrate but has high concentrations of hydrogen sulfide and of certain metals, including iron (Fe) and manganese (Mn; **Fig. 15-19**). As the superhot water is discharged into cold, oxygenated seawater, it does not vaporize because of the high pressure. However, as it cools and mixes with oxygenated seawater, metal sulfides are precipitated to form a cloud of tiny black particles. This cloud gives the black smokers the appearance of a dirty smokestack and hence their name. As the black smoker continuously disgorges, precipitated metal sulfides are deposited in the area around the vent exit and may help to construct a chimney at the vent outlet that can be up to 20 m high and several meters wide (**Fig. 15-19b**). The deposits formed at these vents are rich in many metals such as copper, zinc, gold, silver and lead with the quantities of these metals varying with temperature and other characteristics of different vents. These deposits are commercially valuable because the sulfide deposits are rich in these metals and they are now often called seafloor massive sulfide (SMS) deposits (**Chap. 2**). Many of the metal sulfide ores mined on land today are believed to have originated in hydrothermal vent environments and then to have been scraped off and added to the continents at subduction zones (**Chap. 4**).

White smokers were not discovered until 2000. Unlike black smokers these vents are not located along the ridge axis. Instead they lie on the flank of the ridge, where the underlying rock is about 1.5 million years old. The fluids discharged at white smokers are cooler, have a higher pH, contain less sulfide but more silica, calcium, and magnesium than fluids discharged from black smokers. When these fluids are discharged and mix with seawater, calcium, magnesium, and other sulfates are precipitated out and deposited to form chimneys much like those at black smokers. However, some of these chimneys can be up to about 60 m high, much taller than black smoker chimneys. The populations of larger animals, such as crabs, are much less abundant at these vents than are often found at black smokers. However, the white smoker vents do support abundant microbial communities.

Biological Communities Associated with Hydrothermal Vents

Although individual species vary, biological communities that surround ridge axis black smoker hydrothermal vents in the Atlantic, Pacific, and Indian Oceans are composed of generally similar species, and similar communities form in distinct zones around the vent.

At many of the Pacific vents, numerous giant tube worms and clams (**Fig. 15-20**) live closest to the vent. Other invertebrates, such as one or more species of limpets, shrimp, and scale worms, are present but less abundant. Farther from the vent are other plume worms, crabs, amphipods, other shrimp species, and several species of snails. Still farther from the vent are a wide variety of hydroids, species of worms, shrimp, anemones, and snails that are different from those closer to the vent. Many of these species are filter feeders, and others are predators. The biomass decreases progressively and rapidly with distance from the vent, and the more normal sparse fauna of the deep seafloor are present a few tens of meters from the vent.

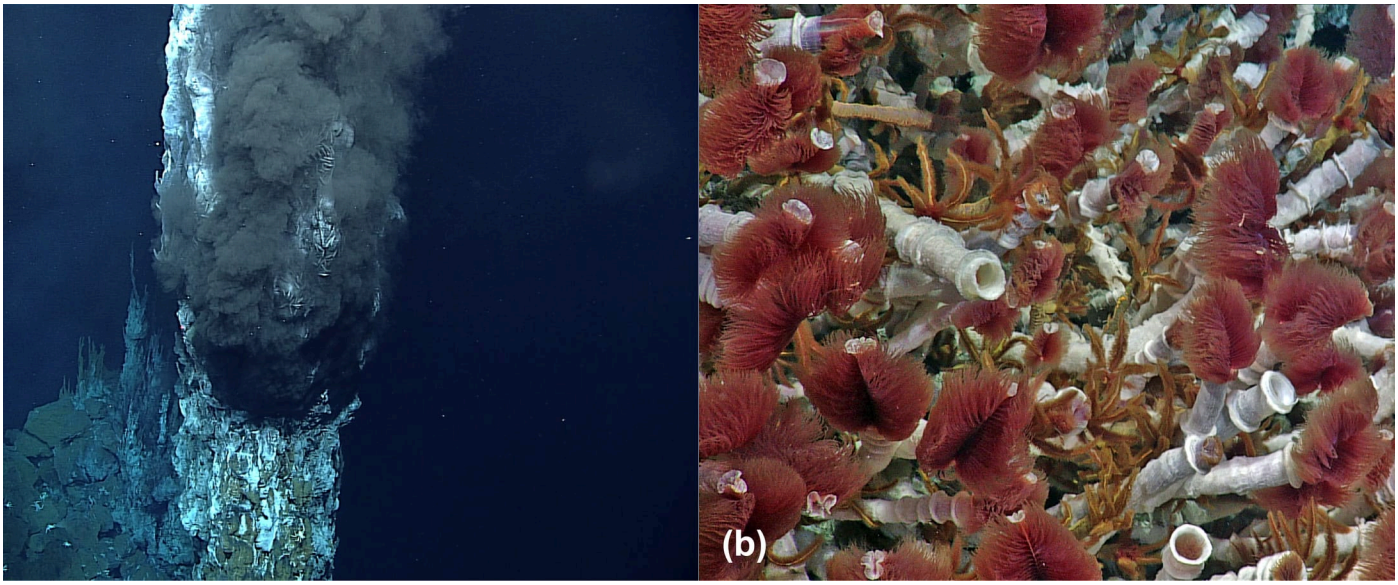


Figure 15-20. Hydrothermal vent communities support many species that are not found elsewhere. (a) A typical biological community near a black smoker, like the one seen here, includes tube worms, crabs, mussels, and other invertebrates. (b) One of the most important members of most hydrothermal communities is the tube worm *Riftia pachyptila*, which can be a meter or more in length and which sustains itself by feeding on symbiotic chemosynthetic bacteria.

More than 590 new species have been identified in the hydrothermal vent fauna, representing more than 20 new **families** and more than 100 new genera (plural of genus) and these numbers continue to climb steadily as new species are found at newly discovered vent areas. The discovery of this bewildering array of new species is unique in the history of biological science, rivaled only by the findings of the *Challenger* expedition in the 1870s (**Chap. 2**).

The biomass of many Pacific hydrothermal vent communities is dominated by the giant tube worm, *Riftia pachyptila*, a very strange creature (**Fig. 15-20b**). This species has no mouth and no digestive system, but it can grow to several centimeters in thickness and more than 1 m in length. Like the clams that live in the same region near the vents, it has red flesh and blood. Both species get their red color from hemoglobin in their blood, the same oxygen-binding molecule that is present in human blood.

The food source for the giant tube worm is apparently a population of chemosynthetic bacteria that it cultivates within its body in a symbiotic association similar to that between corals and zooxanthellae. The bacteria oxidize sulfide as an energy source to chemosynthesize organic matter (**Chap. 12**), and the tube worm assimilates either the waste products of this synthesis or the bacterial biomass itself, or both. This partnership is phenomenally successful, because the tube worms apparently grow very quickly. The tube worm also has a unique **enzyme** that is incorporated in its tissues, particularly in its surface tissues. The enzyme detoxifies hydrogen sulfide and protects the hemoglobin that carries oxygen needed for the worm to **respire**. However, this enzymatic protection is carefully adapted to allow a route by which the sulfide can be brought from outside the worm into the part of its body where the chemosynthetic bacteria reside.

Although it is certain that some hydrothermal vent species other than the tube worm have similar associations with chemosynthetic bacteria or **archaea** to provide a portion of their food, most species within the hydrothermal vent community are probably filter feeders. Hence, the bulk of their food must come from suspended particles. The source of these particles appears to be chemosynthetic bacteria and archaea that grow in profusion in the mouth of the vent and deep within the sediments and rocks of the seafloor. Clumps of the microbial biomass are broken loose periodically by the flow of water through the vent. The clumps fragment to form suspended particles that can be captured by the filter feeders. The concentration of these particles quickly declines with distance from the vent as the plume disperses and large particles are deposited.

Unanswered Questions about Hydrothermal Vents

Many questions about hydrothermal vents remain unanswered, not only about the abundance, geographic distribution, and physical/chemical characteristics of the vents, but also about the species that make up vent communities. For example, we do not know how the chemosynthetic bacteria or archaea can survive and grow at temperatures in excess of several hundred degrees. Also we do not fully understand how these species are able to survive the trip across many kilometers of abyssal ocean to colonize new hydrothermal vents. Most vent species are adapted to higher temperatures and different water chemistry than are present

in the abyssal ocean through which they would have to travel to colonize a new vent. Thus, vent species, or at least their eggs or larvae, must be able to survive a much greater range of environmental conditions than do most of the living organisms with which we are more familiar.

Most known vents are scattered along the ridges, some separated by substantial distances, and each may operate for a limited period, perhaps less than 20 years or so. However, observations have shown that new vents may be colonized within months or years. Bottom currents that could carry eggs and larvae on some ridges may tend to follow the ridge, but new vents may be upcurrent of the old ones. Thus, the physical mechanisms by which new vents are colonized, which also likely differ for individual species, may be complex and are likely aided by eddies formed as tidal currents flow over the elevated oceanic ridge topography. Also, recent findings have revealed that the types of species present at the vents are influenced greatly by the chemical and physical conditions at each vent so that, contrary to earlier expectations that new vents would be colonized by fauna from nearby vents, they may be colonized by completely different fauna.

Many vent species produce extraordinarily large numbers of larvae. One possible explanation for the very large size of some vent species in comparison with similar species elsewhere in the oceans may be a need to produce very large numbers of larval offspring. It has been suggested that, in some species, these larvae may have an arrested development phase, so they could be transported by ocean currents for many decades or even centuries before encountering a new vent to colonize. Eggs and larvae of vent species may rise into shallower layers of the ocean, perhaps entrained in megaplumes of water heated slightly above ambient temperature that are known to occur in hydrothermal vent areas. In the shallower layers, larvae may be widely distributed before settling to the seafloor for a chance encounter with a new vent. There is also evidence that the partially decomposed carcasses of whales and other large mammals that fall to the ocean floor, or other slowly decomposing organic matter such as wood that reaches the deep seafloor, may be ideal sulfide-containing environments to support vent species during a “stopover” while being dispersed across the deep oceans. Any, all, or none of these mechanisms may be involved in new vent colonization.

Genetic analysis of vent species has begun to reveal the rates and patterns of colonization of vents in different parts of the world ocean. Genetic and other studies of many more species from many more vents, exploration of the vast areas currently unsearched to catalog the distribution of hydrothermal vents worldwide, and studies to obtain much better knowledge of the current patterns, mixing, and dispersion in the deep oceans are all needed to enable us to unravel the mysteries of hydrothermal vent species life cycles. We are likely to have more surprises as these investigations proceed.

Other Chemosynthetic Communities

Since the discovery of hydrothermal vents on the oceanic ridges, several chemosynthetic communities have been discovered in other locations in the deep sea, as well as in shallow-water **anoxic** environments such as marshes. Chemosynthetic bacterial mats have been found at the base of the **continental slope** off the west coast of Florida. These bacteria use hydrogen sulfide in water that seeps out from the limestone underlying sediments of the continental slope. In and around these mats is a diverse community of animals that may obtain some or all of their food from the chemosynthetic bacteria.

Chemosynthetic communities have also been found at oil and gas seeps, also known as cold seeps, at a depth of 600 to 700 m in the Gulf of America (Golfo de México) south of Louisiana and elsewhere. These communities use either hydrogen sulfide and/or **hydrocarbons** as an energy source for their primary production. In the Juan de Fuca subduction zone and in other subduction zones, methane in **pore waters** squeezed out of the buried sediments provides the energy source for other chemosynthetic bacteria.

In addition to these ocean chemosynthetic communities, chemosynthetic communities that are generally dominated by archaea have been found to exist deep within the rocks of the Earth’s **crust**. Studies of the organisms able to live and reproduce in such extreme environments have assumed an important role in the search for the origins of life on the Earth and in the search for life or evidence of life in the past on other planets and moons of the solar system.

This page titled 15.8: Hydrothermal Vents is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.9: CHAPTER SUMMARY

Communities and Niches.

Species within an ecosystem are distributed nonuniformly, but they often cluster into communities with common characteristics. Each species has a survival niche that is defined by the ranges of environmental variables, such as salinity, temperature, turbidity, and nutrient concentrations, within which it can survive. Within its survival niche, each species has a fundamental niche within which it can survive and reproduce successfully. In some instances, a species may not occupy all of its fundamental niche, as a result of competition by other species.

Coral Reefs.

Reef-building corals grow only on the seafloor in the photic zone of tropical waters between about 30°N and 30°S. These corals house photosynthetic zooxanthellae within their tissues. Reef-building corals obtain some food from their zooxanthellae. Productivity is higher in coral reefs than in other, nutrient-poor tropical marine areas because nutrients are recycled between coral and zooxanthellae, the reefs cause some upwelling, and most organisms in the reef community are residents and do not export nutrients. High turbidity adversely affects corals by reducing photosynthesis by zooxanthellae and requiring the corals to expend energy to clear away deposited particles.

Primary production in coral reef communities is performed by zooxanthellae and by benthic microalgae and macroalgae, many of which are calcareous algae whose hard parts help to build the reef. A typical coral reef has a sheltered lagoon where coral growth is patchy because of variable salinity and turbidity and where many detritus, suspension, and deposit feeders are present. At the lagoon's seaward edge is a reef flat swept generally free of sediment by waves where coral growth is active and consists mainly of encrusting forms. Invertebrates live in holes and grooves cut in the reef flat.

Farther seaward, there is sometimes a low sandy island and usually an irregular ridge formed from broken coral thrown up periodically by waves and cemented by calcareous algae. Seaward of the ridge is the buttress zone, which may have a gradual to nearly vertical downward slope. It is scoured by waves to about the 20 m depth, characterized by massive robust corals, and often cut across by grooves in which there is little coral growth. Farther seaward, at depths below the reach of wave action, delicate forms of coral, including soft corals, grow in abundance.

Kelp Forests.

Kelp forests grow where the seafloor is stable, preferably rocky, and within the photic zone, and where the water is cold and rich in nutrients. Kelp fronds grow as much as half a meter a day, and kelp primary productivity is very high. Kelp reproduce both vegetatively and by releasing spores to the water column. Kelp forests provide shelter and habitat for many species of fishes and invertebrates. Only a few of these species eat kelp itself, but kelp releases large quantities of detritus that enter the food chain when it is consumed by detritus feeders.

Sea urchins eat kelp, and sea otters eat sea urchins. In the eighteenth and nineteenth centuries, hunting of sea otters off California reduced sea otter populations and thus their predation on sea urchins. The increased sea urchin populations ate more kelp and destroyed the forest in many areas. Sea otters are now protected, and kelp is slowly returning.

Rocky Intertidal Communities.

Species of the rocky intertidal zone are exposed to the atmosphere part of the time. They are subjected to variable conditions of temperature, water and air exposure, salinity, oxygen concentration, and pH, and they are vulnerable to birds and land predators. The rocky intertidal community is separated into four zones, distinguished by degree of exposure to air. The supralittoral zone is above high water and exposed permanently to air but is reached periodically by spray. It supports lichens, encrusting algae, grazers (including marine snails and limpets), and scavengers, primarily isopods. The high-tide zone is covered in water only during high tides and supports encrusting algae, tough attached macroalgae, filter-feeding barnacles, and periwinkle and limpet species different from those in the supralittoral zone. The middle-tide zone is covered and uncovered by water during most or all tidal cycles, has sparse macroalgae because of competition from mussels and barnacles, and supports a diverse community

of invertebrates. The low-tide zone is uncovered only during the lowest tides and supports macroalgae and many species of invertebrates and fishes. The upper limit of each zone is generally determined by tolerance of the species to air exposure and other environmental factors, whereas the lower limit is generally determined by competition with other species.

Tide pools undergo substantial changes in temperature, salinity, and other factors because they are small and isolated from mixing with ocean water for part of the tidal cycle. Because small tide pools tend to have greater changes than large tide pools, they support fewer species, and these species are more tolerant.

Sargasso Sea.

Extensive rafts of *Sargassum*, a macroalga, float on the surface of the Sargasso Sea, which is the interior of the North Atlantic Gyre. Nutrients are extremely limited in this region, and the large *Sargassum* biomass develops because it is very long-lived and currents tend to concentrate and retain it within the center of the gyre. A variety of small fish and invertebrate species live in the *Sargassum*, many of which are unique to this community. The animal biomass is very small in relation to the algal biomass, and there are few grazers because primary productivity is low and any food that is grazed is therefore replaced slowly.

Polar Regions.

Polar regions have extreme seasonal variation in light availability and ice cover, and generally cold surface waters. River runoff or ice exclusion creates strong haloclines in places, but in all other polar waters, vertical mixing due to storms is intense and nutrients are abundant. Microscopic ice algae are important in both polar regions. They grow rapidly during the ice-melting season. Phytoplankton bloom during only a few weeks of summer when sufficient light is available. Many animal species are adapted to the short primary production period and the large year-to-year variability by having a long life span, maturing late, bearing only a few offspring each year, and storing food energy as fat to survive the winter.

Beyond the Sun's Light.

Below the photic zone, food sources are limited to particulate detritus, carcasses that fall through the water column, and prey species. However, many species migrate vertically from the aphotic zone to the mixed layer, usually at night, to feed. Species that migrate vertically become less common with increasing depth. Many deep-sea animals are adapted to survive on very infrequent meals, and some are able to swallow prey larger than themselves. High pressures inhibit bacterial decomposition, which helps ensure that detrital food particles remain available.

Hydrothermal Vents.

Hydrothermal vents located along oceanic ridge axes and on submerged volcanoes of volcanic island arcs discharge seawater that has percolated through the seafloor and been heated by magma or cooling magmatic rock. The effluent from black smokers, which discharge the hottest water (270 to 380°C), has no oxygen and high concentrations of hydrogen sulfide and metal sulfides. Chemosynthetic bacteria use the sulfides to fuel primary production. Away from the ridge axis vents discharge cooler water and are called white smokers because they discharge much lower concentrations of metal sulfides, and instead contain abundant precipitated white particles of silica, and calcium and magnesium sulfates and carbonates.

Communities at hydrothermal vents are composed of many species that are unique to these environments. At many Pacific vents, the community closest to the vent is dominated by giant tube worms and clams that feed on chemosynthetic bacteria that they cultivate within their bodies. Various species of other invertebrates occupy zones at different distances from the vent. Most are filter feeders that live on clumps of bacteria and archaea grown at the vent or beneath the seafloor and sloughed off to become suspended particles. Food availability declines rapidly with distance from the vent. Many questions remain about these communities, including how they cross large distances of abyssal ocean to colonize new vents.

White smoker vents lie on the oceanic ridge flank, discharges fluids with little metal sulfide but instead particles rich in calcium and magnesium. Additionally, chemosynthetic communities that use methane or other hydrocarbons as their energy source are present in a few locations in the oceans.

15.9.1: Key Terms
15.9.2: Chapter Questions
15.9.3: References

This page titled 15.9: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.9.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- benthic
- biomass
- blue-green algae
- brittle stars
- buoyant
- calcareous
- carcinogenic
- carnivores
- climate
- communities
- contamination
- continental shelves
- continental slope
- convergence
- copepods
- coral reef
- corals
- crust
- crustaceans
- currents
- decomposers
- deposit feeders
- detritus
- diffuse
- dinoglagellate
- diversity
- downwelling
- ecologists
- ecosystem
- eddies
- encrusting microalgae
- environment
- enzyme
- families
- fauna
- filter feeders
- flora
- food chain
- food web
- fringing reef
- front
- fundamental niche
- genetic
- genus
- grain size
- grazers
- gyre

- habitat
- halocline

- hard parts
- herbivores
- hermatypic corals
- hermit crabs
- high-tide line
- high-tide zone
- humidity
- hydrocarbons
- hydroids
- ice exclusion
- invertebrate
- isopods
- kelp
- krill
- lagoon
- larvae
- latitudes
- lava
- leeward
- lichens
- limestone
- limpets
- low-tide line
- low-tide zone
- macroalgae
- magma
- mangrove
- marginal seas
- marine mammals
- micronutrients
- middle-tide zone
- mollusks
- mutagenic
- niche
- nudibrachs
- nutrient
- oceanic ridge
- pelagic
- pH
- photic zone
- photosynthesize
- phytoplankton
- pore waters
- primary productivity
- reef
- reef flat
- reef terrace
- residence time
- respire
- resuspended
- rocky intertidal zone
- runoff
- salinity
- scavengers
- scuba
- sea cucumbers
- sea fan

- sea stars
- sea urchin
- sedimentation rates
- sediments
- soft corals
- soluble
- species
- sponges
- spores
- spring tides
- storm surges
- subduction zones
- submersibles
- subtropical gyre
- supralittoral zone
- survival niche
- suspended sediment
- suspension feeders
- symbiotic
- teratogenic
- thermocline
- tidal range
- tide
- tide pool
- topographic
- tube worms
- turbulence
- turbidity
- turbidity currents
- upwelling
- viviparous
- western boundary currents
- windward
- zooplankton
- zooxanthellae

This page titled 15.9.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.9.2: Chapter Questions

Study Questions

1. Why don't species live in all locations where their fundamental-niche requirements are fulfilled?
2. Why do coral reefs sustain a very large variety of species, and why is the biomass of coral reef ecosystems greater than that of other ecosystems in tropical waters?
3. What factors determine the types of corals that live on different parts of a coral reef?
4. Describe the locations and environmental conditions that are appropriate for the development of kelp forests. Contrast these with the locations and environmental conditions appropriate for the development of coral reefs.
5. Why must rocky intertidal species be more stress-tolerant than many other marine species?
6. Why are rocky intertidal communities distributed in distinct bands of different species that follow depth contours?
7. Why does the Sargasso Sea sustain a community of photic-zone organisms that are different and distinct from those found elsewhere in the oceans?
8. What are the principal differences in physical characteristics between the Arctic Ocean and the ocean around Antarctica, and how do they affect primary production and species composition?
9. Why does the species composition of hydrothermal vent communities change rapidly with distance from the vents?

Critical Thinking Questions

1. Sea otters are increasing in numbers in California coastal waters, but one of their favorite foods, abalone, is heavily harvested, so they may rely more on urchins for their food supply. Describe what you think will happen to the kelp forests over the next several decades as a result of these changes, and how these changes will affect other species in the ecosystem.
2. Describe the feeding, hunting, defensive, and reproductive characteristics that are desirable for carnivorous species that live in the Sargassum community. Compare and contrast these characteristics with the desirable characteristics for a carnivorous benthic species that lives in the deep oceans. Explain each of the differences.
3. Hydrothermal vent communities throughout the oceans are separated, sometimes by substantial distances. However, some of the same species are found at many of the vents, regardless of their location. In addition, chemosynthetic bacteria species that are identical or very closely related to those found at the hydrothermal vents on oceanic ridges have been found in hydrothermal vent environments at the bottom of Crater Lake, Oregon, and Lake Baikal, Russia.
   (a) What are the possible reproductive strategies that could explain these observations?
   (b) Hypothesize ways that chemosynthetic organisms could be transported between their present locations and the locations of new vents or the locations of new lakes similar to Crater Lake and Lake Baikal.
   (c) What studies do you think would be needed to prove or disprove your hypotheses?
4. If the Earth's climate warms and both polar ice caps melt away completely, which ecosystem is likely to be changed more, the Southern Ocean ecosystem or the Arctic Ocean ecosystem? Explain the reasons for your answer.
5. The temperature and salinity of the abyssopelagic environment vary little with latitude or from ocean to ocean. This environment is also uniformly dark and the pressure uniformly high. Does this mean that, if we were able to sample every species in a cubic kilometer of deep-ocean water, we could be certain that we had sampled a large majority of all abyssopelagic species that exist in the oceans? Why or why not?
6. Coral reefs and kelp forests both have high species diversity compared to some other ocean ecosystems. What are the possible reasons for this high diversity?

This page titled 15.9.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

15.9.3: References

Figure 15-1 – Segar, D., and E. Stamman Segar, 1997: Deacon's Reef in Papua New Guinea.

Figure 15-3 – Adapted from Lerman, M., 1986: *Marine Biology: Environment, Diversity, and Ecology*. Addison-Wesley, 534 pp.
– Adapted from Chapmen, V. J., Ed., 1977: *Wet Coastal Systems*. Elsevier Science Publishing, 424 pp.

Figure 15-5a – Segar, D., and E. Stamman Segar, 1997: Sea cucumbers (family Synaptidae, Indonesia.

Figure 15-5b – Segar, D., and E. Stamman Segar, 1997: Rhinoceros, or horned, sea star (*Protoreaster nodosus*, Papua New Guinea).

Figure 15-5c – Segar, D., and E. Stamman Segar, 1997: Urchins (*Astropyga radiata*, Indonesia).

Figure 15-5d – Segar, D., and E. Stamman Segar, 1997: Sand dollars (*Clypeaster sp.*, Indonesia).

Figure 15-5e – Segar, D., and E. Stamman Segar, 1997: *Acropora sp.* (Papua New Guinea).

Figure 15-5f – NOAA, 2023: Elkhorn coral, accessed 23 March 2026, Source.

Figure 15-6 – Segar, D., and E. Stamman Segar, 1997: Palm trees and other plants on this island on a fringing reef in Fiji.

Figure 15-7 – Segar, D., and E. Stamman Segar, 1997: Black coral colony (probably *Cirrhipathes sp.*) with a longnose hawkfish (*Oxycirrhites typus*, Papua New Guinea).

Figure 15-8a,b – Adapted from Estes, J. A., and J. F., Palmisano, 1974: Sea otters: Their role in structuring nearshore communities. *Science*, **185**, 1058–1060, Source.

Figure 15-8c – Segar, D., and E. Stamman Segar, 1997: Canopy and kelp forest in Monterey, California.

Figure 15-9a – Segar, D., and E. Stamman Segar, 1997: Small brown cup corals (*Paracyathus stearnsii*, California).

Figure 15-9b – Segar, D., and E. Stamman Segar, 1997: Rose anemone (*Tealia lineata*, California).

Figure 15-9c – Segar, D., and E. Stamman Segar, 1997: Northern kelp crab (*Pugettia producta*, California).

Figure 15-9d – Segar, D., and E. Stamman Segar, 1997: Kelp shrimp (Monterey, California).

Figure 15-9e – Segar, D., and E. Stamman Segar, 1997: Sea stars of many species (Monterey, California).

Figure 15-9f – Segar, D., and E. Stamman Segar, 1997: Horned nudibranch (*Hermissenda crassicornis*, California).

Figure 15-9g – Segar, D., and E. Stamman Segar, 1997: Quill-backed rockfish (*Sebastes maliger*) and copper rockfish (*Sebastes caurinus*, California).

Figure 15-10a – Segar, D., and E. Stamman Segar, 1997: Sea otters (*Enhydra lutris*, California).

Figure 15-10b – Segar, D., and E. Stamman Segar, 1997: Red urchin (*Strongylocentrotus sp.*, California).

Figure 15-10c,d – Adapted from Estes, J. A., and J. F., Palmisano, 1974: Sea otters: Their role in structuring nearshore communities. *Science*, **185**, 1058–1060, Source.

Figure 15-13a – Segar, D., and E. Stamman Segar, 1997: Black turban snails (*Tegula funebralis*, Monterey, California).

Figure 15-13b – Segar, D., and E. Stamman Segar, 1997: Limpet (order Archaeogastropoda, California).

Figure 15-13c – Segar, D., and E. Stamman Segar, 1997: Buckshot barnacles, probably primarily *Chthamalus sp.*, with some interspersed *Balanus glandula*.

Figure 15-13d – Segar, D., and E. Stamman Segar, 1997: Lined chiton (probably *Tonicella lineata*, California).

Figure 15-13e – Segar, D., and E. Stamman Segar, 1997: Dense bed of California mussels (*Mytilus californianus*, Monterey, California).

Figure 15-13f – Segar, D., and E. Stamman Segar, 1997: Green algae south of San Francisco, California

Figure 15-13g – Segar, D., and E. Stamman Segar, 1997: Hermit crab (*Pagurus sp.*, California).

Figure 15-14 – Brownlee, E., 2024: Mussel beds in Rialto Beach, WA. With permission.

Figure 15-15a – NOAA OER, 2002: Lines of sargassum. Islands in the Sea 2002, accessed 23 March 2026, Source.

Figure 15-15b – NOAA OER, 2002: Small mass of sargassum weed. Islands in the Sea 2002, accessed 23 March 2026, Source.

Figure 15-16a – Segar, D., and E. Stamman Segar, 1997: Crabeater seals (*Lobodon carcinophagus*, Antarctic Peninsula).

Figure 15-16b – Segar, D., and E. Stamman Segar, 1997: Antarctic fur seals (*Arctocephalus gazella*, South Georgia Island).

Figure 15-16c – Segar, D., and E. Stamman Segar, 1997: Leopard seal (*Hydrurga leptonyx*, Antarctic Peninsula).

Figure 15-18 – NOAA OER: 2019: Dumbo octopus. 2019 Southeastern U.S. Deep-sea Exploration, accessed 23 March 2026, Source.

Figure 15-19a – Adapted from Detretskey, 2012: Hydrothermal vents. NSF, accessed 23 March 2026, Source.

Figure 15-19b – Adapted from Haymon, R. M., and Macdonald, K. C., 1985: The geology of deep-sea hot springs: Recently discovered hot springs burst from the seafloor in areas of active volcanism and deposit metal-rich minerals on the seabed. *Amer. Sci.*, **73**, 441–449, Source.

Figure 15-20a – NOAA OER, 2016: Black smoker, 2016 Deepwater Exploration of the Marianas. Accessed 23 March 2026, Source.

Figure 15-20b – NSF-OOI, 2015: Vibrant red tube worms, WHOI. Accessed 23 March 2026, Source.

This page titled 15.9.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar, Emily Brownlee, and Michelle Valkanas.

16: IMPACTS OF HUMANS ON THE OCEANS

Learning Objectives

- **CC8 Residence Time**: The residence time of seawater in a given segment of the oceans is the average length of time the water spends in that segment. The residence times of some coastal water masses are long, therefore some contaminants discharged to the coastal ocean can accumulate to higher levels in these long residence time regions than in areas with shorter residence times.
- **CC14 Phototrophy, Light, and Nutrients**: Phototrophy (which includes photosynthesis) is the primary process by which simple chemical compounds are made into the organic compounds of living organisms. Photosynthesis depends on the availability of carbon dioxide, light, and certain dissolved nutrient elements including nitrogen, phosphorus, and iron. Alterations of the availability of any of these elements by human activity can have adverse consequences in marine ecosystems.
- **CC17 Species Diversity and Biodiversity:** Biodiversity is an expression of the range of genetic diversity; species diversity; diversity in ecological niches and types of communities of organisms (ecosystem diversity); and diversity of feeding, reproduction, and predator avoidance strategies (physiological diversity), within the ecosystem of the specified region. Species diversity is a more precisely defined term and is a measure of the species richness (number of species) and species evenness (extent to which the community has balanced populations with no dominant species). High diversity and biodiversity are generally associated with ecosystems that are resistant to change. Changes in diversity are often used as an indicator of the impacts of human activities.
- **CC18 Toxicity**: Many dissolved constituents of seawater become toxic to marine life at levels above their natural concentrations in seawater. Some synthetic organic chemicals are especially significant because they are persistent and may be bioaccumulated or biomagnified.

Figure 16-1. Adverse effects on the oceans that constitute pollution come in numerous ways, many of which do not involve the dumping or discharge of wastes that most people associate with the term pollution, but one species is the ultimate origin of all such adverse effects.

By far the most important human pollution of the oceans is the release of carbon dioxide and methane to the atmosphere. The adverse effects on the oceans that are predicted to occur as a result of carbon dioxide releases since the industrial revolution far transcend all other human impacts on the oceans combined. Massive disruption of marine **ecosystems** and likely numerous marine species extinctions are anticipated to result from the changing temperatures, current patterns, sea level rise, and especially from the deoxygenation and increased acidity of the oceans that has already occurred, or will occur. These already unavoidable adverse effects will become more severe if carbon dioxide releases are not drastically reduced. This issue is so important that it was discussed in some detail in **Chapter 1** and referenced in many other chapters. This chapter explores the range of ways in which humans can, and do, adversely affect the oceans and examines the concept of pollution and of toxicity both of which are widely misunderstood. It then examines the wide range of human activities, such as sewage waste disposal,

industrial waste discharges, and oil spills that have been the focus of most public concern and debate with regard to ocean pollution. Finally, the chapter examines the human activities that have caused significant harm, or pose a much larger threat to ocean ecosystems, most of which receive little or no public recognition or debate. These include, fishing, the introduction of nonindigenous species, habitat alteration, and finally the most important and greatest threats of climate change, deoxygenation and acidification caused by **anthropogenic** releases of carbon dioxide releases to the atmosphere and anthropogenic nutrient inputs to the oceans.

Chapter 2 reviewed the historical importance of the ocean and the growing use of ocean resources for fisheries; transportation; trade; extraction of offshore oil, gas, and other minerals; pharmaceuticals; energy; and recreational and aesthetic opportunities. Each use potentially can or actually does have deleterious effects on ocean ecosystems and on the other uses of the ocean. Such deleterious effects are encompassed by the term **pollution**, which can be characterized as "the addition of substances to, or alteration of, the ocean ecosystem in a manner that is deleterious to the ocean ecosystem or its resources." This definition, or a similar one, is generally accepted by many national and international organizations that have responsibilities for managing and protecting the oceans and ocean resources. For example, the formal definition accepted by the premier world scientific body in this area, the United Nations Joint Group of Experts on the Scientific Aspects of Marine Environmental Protection (GESAMP), is as follows:

> Pollution means the introduction by humans, directly or indirectly, of substances or energy into the marine environment (including estuaries) resulting in such deleterious effects as harm to living resources, hazards to human health, hindrance to maritime activities including fishing, impairment of quality for use of sea water and reduction of amenities.

Notice that, according to the generally accepted official definition, the term pollution includes not only the discharge of harmful wastes, but also activities such as **overfishing**, construction of structures, and other human activities that adversely affect ocean ecosystems by changing **currents**, distributions of dissolved substances or heat, or **sediment** transport patterns. The definition also includes adverse impacts of one ocean use on another, such as the degradation of recreational and aesthetic value caused by trash on ocean **beaches** or offshore oil rigs "spoiling" the ocean view. Ocean pollution also includes harmful effects on ocean ecosystems of the anthropogenic release of carbon dioxide and other greenhouse gases to the atmosphere, as these releases indirectly add carbon dioxide, a "waste" material and heat energy to the oceans. Thus, ocean pollution is a very broad term that includes any anthropogenic activity that causes directly or indirectly harm to ocean ecosystems or resources. Notice also that there must be harmful ("deleterious") effects before any activity can be considered pollution. By contrast, the popular definition of pollution continues to be restricted to the addition of substances (chemicals, materials, or organisms) to the **environment**, and the term often is misused to include any such contamination, whether "deleterious" or not.

16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.4: Waste Disposal
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
16.8: Chapter Summary
16.8.1: Key Terms
16.8.2: Chapter Questions
16.8.3: References
16.9: Box 1 - Petroleum in the Marine Environment
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.11: Box 3 - Radioactive Wastes in the Marine Environment

This page titled 16: Impacts of Humans on the Oceans is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.1: Pollution versus Contamination

All too often, any human activity that releases wastes or introduces particulate or dissolved substances to the ocean, either by accident or incidental to other activities, is mistakenly reported as “pollution.” In many cases, however, such releases are benign or even beneficial to the ocean ecosystem and ocean resources. In such cases, the released material acts as a **contaminant** but no pollution has occurred. Only when the ocean ecosystem or ocean resources are damaged should the activity be called pollution. Hence, human activities may contaminate the oceans without polluting them.

One important note required here concerns the gross misuse of the term “pollutant.” This term is applied to many substances that can have, and may have had, adverse effects on some ecosystem in which the concentration of the substances has been elevated by human activity to a level that has caused adverse effects and, therefore, pollution. There has been a popular use of the term “pollutant” to describe such substances, and then a popular trend toward calling the specific substance (for example, arsenic or cyanide) a pollutant. This misuse of the term has led many to believe that these substances are toxic and unacceptable at any concentration. However, this is never true. There is no substance known to humans that is always toxic at any concentration in any ecosystem. For example, arsenic is a natural element that is present in humans and other living organisms almost always at concentrations that cause no adverse effect (arsenic is benign), was once used extensively for medicinal purposes, and is the source of energy (by oxidation of arsenate to arsenite) for some species of bacteria in certain ecosystems. There are many compounds called cyanides, but “cyanide” usually refers only to hydrogen cyanide, which is toxic to humans at low concentrations, but which is naturally occurring in many ecosystems, produced by certain types of bacteria, and can provide the only source of nitrogen for other types of bacteria in certain ecosystems. In summary, no chemical always causes adverse effects (pollution) regardless of concentration in all ecosystems so there is no substance that can, or should be, always called a pollutant.

For many years, the oceans were considered so vast that human populations could safely discharge all their wastes there and carelessly exploit ocean resources such as fisheries and mineral deposits without causing adverse effects. In recent decades, the recognition that the oceans can be harmed by human exploitation and waste disposal has led to another viewpoint: that ocean ecosystems are so fragile that they must be protected from any human influence that may change them, and that no inputs of any contaminants can be permitted.

Just as the historical view of the oceans as limitless was wrong, this new view, although idealistic, is also incorrect. In some cases, the new view may have led to political decisions that, although they may possibly have reduced ocean pollution, also increased human health risks and terrestrial pollution. The oceans are naturally changeable, and humans have caused many changes in the oceans, just as we have in the terrestrial environment. If human civilization is to continue, further changes in both the terrestrial and ocean ecosystems are inevitable and necessary. We must learn to view the planet as a whole and recognize that we must use its resources wisely. We must accept contamination where necessary or desirable, and avoid pollution wherever possible, not only in the oceans but also on land and in the atmosphere.

Assimilative Capacity

The oceans receive millions of tonnes per year of many dissolved elements and organic substances from rivers, dust, and rain (**Chaps. 5, 6**) and have done so since long before humans appeared on the Earth. These substances include many elements, such as copper, zinc, arsenic, and mercury, that are toxic to humans and other **species** if present at high enough concentration. The oceans also receive organic matter from soils, plants, and animal wastes, the composition of which is substantially the same as that of human fecal and urinary wastes.

The amounts of these substances introduced naturally must have varied substantially as **plate tectonics** modified and moved the continents and as the Earth’s climate changed. Hence, ocean ecosystems can and have changed to accommodate a range of input rates of these substances. The maximum rate at which the oceans can accommodate such inputs without adverse effects is called the **assimilative capacity**. **Chapters 5 and 6** discussed some of the chemical and biological processes that remove substances from ocean waters to balance inputs and prevent concentrations from rising continuously. The assimilative capacity is exceeded if the input rate increases so rapidly or by so much that removal processes are overwhelmed, and the concentration rises to a level that is toxic or has other adverse effects on the ecosystem.

Besides naturally occurring compounds, the oceans have an assimilative capacity, albeit sometimes small, for synthetic organic chemicals produced by human civilization. Although these compounds are new to the oceans, all (including plastics) are broken down by **decomposers** and chemical reactions into other compounds and eventually inorganic compounds. Some are broken

down quickly, whereas others, including **DDT** (dichloro-diphenyltrichloroethane), **PCBs** (polychlorinated biphenyls), and many plastics are broken down very slowly.

Although assimilative capacity is a useful concept, it is very difficult to apply. Each element or substance has its own unique **residence time**, natural concentration, concentration at which it becomes toxic, variable toxicity to different species, and chemical and biological decomposition and removal processes. Each of these factors must be understood before the ocean's assimilative capacity for a single substance can be estimated. The oceans are not instantly and uniformly mixed. Consequently, the assimilative capacity can be exceeded for part of the oceans if inputs to a specific region exceed the rate at which they can be removed by chemical and biological processes and by mixing with the rest of the oceans. Hence, assimilative capacity and residence time are linked.

CC8 describes how residence times can be determined for individual substances. In a geographically distinct region, an increased input rate of a contaminating substance will cause the substance's concentration in the water (and thus, generally, in the sediments and biota of the region) to increase. The increase in concentration is greater if the residence time is longer. Thus, for example, organic material in the large volumes of sewage of a major city can far exceed the assimilative capacity of a river or bay if the water body has a relatively long residence time. The discharged organic material can be decomposed and deplete dissolved oxygen faster than the oxygen can be replaced by mixing with oxygenated ocean or river water or resupplied from the atmosphere. As a result, the water becomes **anoxic**, and toxic hydrogen sulfide is generated by bacterial action (**Chap. 12**). Parts of San Francisco Bay (**Fig. 16-2**), the New York Harbor, and many other **estuaries** developed chronic sewage-induced anoxia in the 1950s and 1960s.

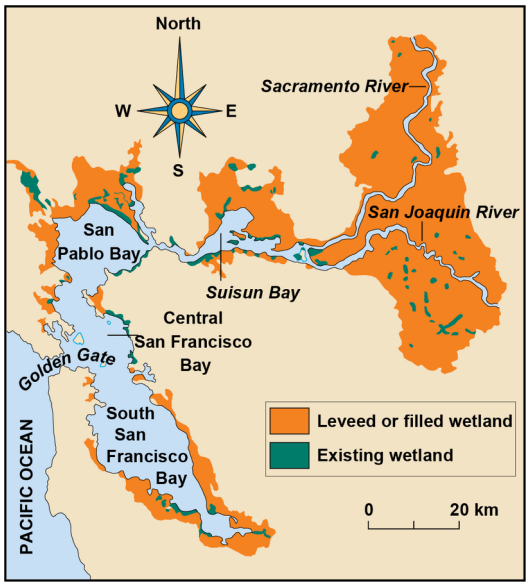


Figure 16-2. This map of San Francisco Bay and its delta shows the areas that were formerly wetland or part of the Bay but that have been filled or confined by levees in the past. The area of the Bay is now approximately one-half of what it was before 1850, and the area of remaining wetland is only a very small fraction of what it was then. Small areas that were previously filled have now been restored.

The anoxia that affected many estuaries in the U.S. and elsewhere only several decades ago has now been alleviated in most instances. Today, all sewage is treated to remove most organic matter before being discharged as **effluent**. In addition, some, but not all, treated sewage has been diverted to **outfalls** discharging directly to coastal oceans, where the residence time is much shorter and the assimilative capacity much greater. Unfortunately, even the reduced-contaminant inputs exceed the assimilative capacity in some estuaries. Much of the oxygen demanding organic matter is now removed from all sewage in most industrialized nations, and this has led to the revival of many estuaries that formerly were severely degraded. However, treated sewage along with agricultural runoff are still largely responsible for areas of anoxia or **hypoxia**, commonly called dead zones, in a number of estuaries such as Chesapeake Bay. In many locations globally, treatment of sewage effluents has revived an estuary where relatively short residence times and restricted light penetration prevent rapid algal bloom formation only to create expanding seasonal or permanent anoxia and dead zones in the adjacent coastal ocean areas where residence time is much longer than in the estuary and turbidity is lower so algae can bloom freely as long as the nutrients (principally nitrogen), most of which remains in the treated sewage effluent during treatment, are available to sustain the growth. **Coastal zone** anoxia was discussed in more detail in **Chapter 13**.

A critical lesson here is that a discharge (nutrients in treated sewage effluents) in one location (the estuary) can produce

contamination but no pollution, whereas a discharge of identical volume and composition in a region with a longer residence time (the adjacent coastal ocean) can produce a serious pollution problem. Hence, residence time and assimilative capacity are important parameters that must be considered when evaluating any activity that releases or may release contaminants. Because these parameters are location-specific, a release of a given type or amount of contaminant that has caused pollution problems in one location may not cause the same problems at other locations. Also, a release of a given type or amount of contaminant that has not caused pollution problems at one location may cause problems at another location.

This page titled 16.1: Pollution versus Contamination is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.2: Adverse Effects of Human Activities

Contamination and other human activities cause a variety of adverse effects, some of which are described here. In many instances, one activity can lead to more than one of the impacts discussed.

Interference with Photosynthesis and Primary Production

Many contaminant inputs to the oceans contain large quantities of **suspended sediment** and light-absorbing organic matter. If retained in the water column, high concentrations of these materials block the penetration of sunlight and reduce the amount of light available for **photosynthesis** (**CC14**).

Human-caused discharges of suspended particles to the oceans primarily affect coastal regions (**Fig. 16-3**), where much of the ocean's **primary production** occurs. **Benthic algae**, both **microalgae** and **macroalgae**, are important primary producers in many coastal and estuarine areas. Excess suspended sediment reduces light penetration and, thus, primary production by benthic algae.

Figure 16-3. There are many sources of suspended sediment contaminant discharges to the oceans including some you might not expect. For example, sugar cane fields are often burned after harvest to clear the land for new planting. This practice increases erosion, and large amounts of suspended sediment are discharged in runoff. These fields are on the north shore of Hawaii. You can see the smoke from one of the fires and the plume of suspended sediment that spreads in the nearshore zone where there are reef-building coral formations that can be destroyed by excess sedimentation.

Adverse effects on primary production can be caused by discharges of large quantities of **nutrients**. Alterations in relative concentrations of nutrients can cause changes in **phytoplankton** species composition (**Chap. 13**). The result can have adverse effects on the food web if the newly advantaged phytoplankton species are unsuitable food for the zooplankton population.

Excessive nutrient discharges can increase primary productivity so dramatically that it exceeds the **grazing** capacity of consumers. In such cases, phytoplankton reproduce and grow rapidly in a **bloom** until the **limiting nutrient** is depleted. The bloom then collapses, and dead phytoplankton cells sink and decompose rapidly, removing some or all of the oxygen from the water. Reduction of oxygen concentration can adversely affect the **respiration** of fishes and **invertebrates**.

Oxygen depletion due to the presence of excessive nutrients or organic matter is called **eutrophication** and has long been a menace in lakes, rivers, and estuaries where nutrient inputs are large and residence times long. In many such areas, the nutrient and organic matter inputs are from sewage discharges. Sewage treatment has dramatically reduced eutrophication in U.S. lakes, rivers and estuaries since the early 1970s. However, nutrients remaining in the effluent after sewage treatment, when combined with other nutrient inputs, are now causing increased eutrophication problems (hypoxia and anoxia) in coastal oceans near estuaries. Hypoxia and anoxia that result from excessive nutrient inputs, and the increasing severity of these problems, were labeled in a 2000 report by the National Research Council of the National Academy of Sciences as the most pervasive and

troubling pollution problem facing U.S. coastal waters.

Oxygen concentrations can also drop if photosynthesis is reduced. For example, increased concentrations of suspended sediment, copper, or other substances, such as herbicides, can inhibit photosynthesis and/or primary production without significantly affecting respiration in a marine ecosystem.

Human activities adversely affect oxygen concentrations in at least two special situations. First, when rivers are constrained within levees, the surface area is diminished but the river depth and currents increase. Often the **photic-zone** depth is naturally shallow and may be further reduced by higher **turbidity** caused by faster river currents that **resuspend** particulate matter. Because both the surface area and the depth of the photic zone are reduced, the total volume of the photic zone is smaller. Furthermore, the reduced surface area lowers the rate of oxygen resupply from the atmosphere.

In the second special case, oxygen concentrations are reduced because power plant, or other industrial, effluents have a higher-than-ambient temperature. Because oxygen has a lower **saturation solubility** at higher temperature, oxygen concentrations in the heated effluents are generally lower than ambient concentrations. This disparity is seldom an immediate problem, because the saturated oxygen concentration is sufficient to support respiration at all but the highest ocean water temperatures (above about 30°C). However, the low concentration of dissolved oxygen reduces the assimilative capacity for **oxygen-demanding** substances. Because tropical waters have high ambient temperatures and low ambient oxygen concentrations, oxygen depletion and sulfide production can be caused by smaller inputs of nutrients or organic matter in the tropics than in other **latitudes**. Hence, water bodies in tropical regions have a lower assimilative capacity for organic wastes than do water bodies with similar residence times at higher latitudes.

Habitat Alteration

Each species that lives in the oceans has its own requirements for the physical and chemical characteristics that constitute its **habitat** (**Chaps. 14, 15**). Benthic species are particularly dependent on the nature of sediments in or on which they live. **Pelagic** species require, or prefer, certain ranges of temperature, **salinity**, turbidity, and current or wave regime. Many pelagic species depend on the benthic environment for food or shelter during part of their life cycle, whereas many others depend on the shallow-water environments of **mangrove** swamps, coastal **wetlands**, rivers, and estuaries during their juvenile phases.

Human activities have caused the destruction of vast areas of coastal wetlands and have caused adverse habitat alteration in other areas. For example, rivers have been dammed, preventing **anadromous** fishes, such as salmon, and **catadromous** fishes, such as eels, from migrating to the upper reaches of the rivers on which they depend. Levees, marinas, and ports have substantially altered current patterns, and thus suspended sediment transport routes, in many estuaries. Vast quantities of freshwater have been removed from many rivers, thus changing the salinity and other chemical characteristics of estuaries. Coastal structures have interfered with sand transport along many coasts. Large volumes of wastes have been discharged directly to the oceans or through rivers, and many parts of the seafloor have been damaged by dredging and dynamite blasting, and by anchors and fishing gear. All of these activities cause changes in **sedimentation rate**, sediment **grain size**, and sediment chemistry on the seafloor of affected regions. Sandy seafloor can be turned into mud and vice versa, and stable rocky or **reef** bottoms can be covered with sediment or broken up and **eroded** away.

Whenever marine habitat is altered, the species composition changes. Some species are disadvantaged and others benefit. Most often, the alteration causes at least a temporary reduction in species **diversity** (**CC17**) and dominance by species that are less sensitive to changing habitat. Opportunistic species that are not normally a major part of the biota commonly move into and dominate an environment when it has been altered by human activity, especially if the disturbance is ongoing, such as waste disposal or dredging. Although in some instances opportunistic species can enhance the **biomass** of the natural food web, more often they are less desirable or worthless in the natural food web.

Community Structure Alteration

The many species that make up a marine **community** depend on each other for food and in many other ways (**Chaps. 14, 15**). The balance among species is determined by centuries or millennia of competitive and cooperative interaction that has enabled the community to reach a relatively stable state. If the balance is disturbed, the community structure can become unstable and change unpredictably.

The greatest direct human disturbances of community structure are caused by preferential exploitation of one or more species in a fishery or by the introduction, accidentally or otherwise, of **nonindigenous species** (discussed later in this chapter). Human

activities may also introduce nonindigenous disease-causing organisms that affect some species important to the ecosystem. Other human influences, such as substrate alteration and introduction of potentially toxic substances, can also advantage or disadvantage certain species and cause community structure to be altered.

Marine communities are periodically subjected to habitat disturbances due to natural events, such as earthquakes and climate changes. Some additional human disturbance can be tolerated and accommodated by the ocean ecosystem. However, in some cases, anthropogenic disturbance may be more rapid or drastic than natural disturbances. Furthermore, human disturbances are often continuous or increase progressively, and their scale may be unprecedented in some coastal and estuarine areas.

Contamination of Food Resources

Aquatic species can obtain elements and compounds from food and directly from solution. Therefore, substances that are toxic to humans introduced in dissolved form or associated with organic particles can be assimilated by most marine organisms and passed through the food web. Many marine species are tolerant of relatively high concentrations of potentially toxic elements or compounds in their environment or food, probably because their body surface and respiratory tissue are continuously exposed to seawater. Rather than building defenses against the absorption of potentially toxic substances from their environment, many marine species simply take up such compounds and store them in some organ where they cannot interfere with essential **biochemical** processes. This method of detoxification has limits, but it enables many marine species to tolerate high concentrations of some substances.

Fish and **shellfish** with high concentrations of stored potentially toxic substances may suffer no adverse effects but may still pose a significant risk to human health. Shellfish, such as oysters, are particularly adept at concentrating trace metals, and many synthetic organic compounds, such as DDT and PCBs which are both concentrated in fatty tissues of most marine animals.

High concentrations of metals and synthetic organic compounds have been found in the biota from many locations where human activities release such contaminants. In these locations, the fishery or shellfishery is closed, or people are advised to eat only limited amounts of seafood or a specific seafood species. Thus, the value of the fishery resource is diminished or lost. Fortunately, in only one recorded instance have people died from ingesting seafood contaminated with industrial toxic substances discharged into the oceans. That incident, in Minamata, Japan, is discussed later in this chapter.

Contamination of fish and especially shellfish with microorganisms that are human **pathogens** is a serious problem. Because some seafood is eaten raw or only lightly cooked, any microorganisms present will be passed on to the consumer. Seafood may be contaminated during handling and processing, although refrigeration and hygienic food-handling techniques have greatly reduced this problem in most developed countries. Most microbiological contamination of seafood now comes from the harvested waters. The problem areas are generally contaminated by discharges of raw or treated sewage or by animal feces carried in street and land runoff. The contamination is concentrated in estuaries and the coastal zone because human pathogens are progressively diluted and destroyed by marine microbes as they are transported to the open oceans. Unfortunately, shellfish beds are located mostly in coastal and estuarine zones.

Some pathogen-contaminated shellfish can be collected and cleansed by being kept in pathogen-free, constantly running and renewed seawater, but this procedure is expensive and requires a source of reliably pathogen-free seawater. Consequently, the only practicable way to prevent the spread of disease by contaminated shellfish is to prohibit harvesting in contaminated areas. At present, many potentially valuable shellfishing areas of the coastal oceans and estuaries in the contiguous U.S. are closed to shellfishing for at least part of the year. The total area closed is increasing, despite immense expenditures for sewage treatment and control of other contaminant sources since 1972, when the U.S. passed the Clean Water Act requiring secondary treatment of almost all sewage.

Toxin-producing phytoplankton blooms are also a growing contamination problem. These blooms, the toxins they produce, and their effect on marine ecosystems and seafood values are discussed in **Chapter 13**.

Beach Closures and Aesthetic Losses

Many people look to the coastal oceans as places of aesthetic beauty where they can renew their contact with the natural world through various recreational activities. As a result, one of the largest industries on the planet has developed around coastal recreation. However, the value of the coast for such activities is compromised and diminished in many locations by the presence of human structures that mar the natural beauty (**Fig. 16-4**) and interfere with natural processes.

Figure 16-4. Coastal industries tend to reduce the aesthetic value of the shoreline. (a) The Moss Landing Power Plant, a very prominent structure located on the coast in the middle of Monterey Bay, California. (b) An oil refinery at Anacortes, Washington. (c) A marina at Antioch, California, constructed on a former wetland in the San Francisco Bay Delta.

Many coasts that otherwise would be areas of high recreational value are sites of human industrial and residential developments that bar the public from reaching the **shore**. Such developments often cause major changes in the coastal form and function through direct modifications of the **shoreline**, such as **bulkheads**, piers, and groins, and through alteration of the coastal current, wave, and sediment transport patterns (**Chapter 11**). In addition, vast areas of biologically important coastal wetlands have been drained and filled to accommodate development (**Fig. 16-2**). Today there is a growing realization that many of these developments were ill-planned. Public opinion now favors protection of the coast from unreasonable degradation by development, although the legal system lags behind this imperative.

Outfalls discharge treated sewage, industrial waste, and storm-water runoff to rivers, estuaries, and oceans. Rivers also receive various materials that have been carelessly or deliberately dumped by humans. Vessels often discharge wastes directly into the ocean. Many beaches are periodically closed to swimming because the water is contaminated with pathogens from improperly treated sewage. In addition, floating debris mars beaches and, in extreme cases, leads to the temporary closure of beaches for recreational purposes. For example, medical wastes, including syringes, apparently dumped in nearby rivers or from vessels, have periodically washed up on New York and New Jersey beaches, prompting closure of the beaches for fear that the materials could carry pathogens, including the HIV **virus**.

This page titled 16.2: Adverse Effects of Human Activities is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.3: Toxicity

Toxic substances are substances that have adverse effects on organisms, including humans. The term toxic chemical is now among the most emotion-evoking in our society. In fact, there is a widespread belief, fostered by the media and many environmental interest groups, that the release of any quantity of toxic chemicals to the environment, particularly the oceans, is harmful and must be stopped. This belief is unfounded (**CC18**). Many chemicals that are toxic at high concentrations are naturally occurring elements or compounds that have been present in the oceans throughout geological time. Some additional quantities, in many cases substantial quantities, of these substances and even synthetic chemicals that do not occur naturally can be safely accommodated by ocean ecosystems. However, if they are to cause no harm, wastes must be disposed of in ways that ensure that the ocean's assimilative capacity is not exceeded either globally or locally. Marine pollution problems occur when chemical discharges exceed the assimilative capacity and cause adverse effects, including lethal or non-lethal toxicity to species within the ocean ecosystem. Therefore, determining this capacity for each local region into which toxic substances are discharged is an important scientific task.

A second widespread belief is that the ultimate solution to contamination by toxic chemicals is to recycle everything. Recycling is a highly desirable and rapidly growing practice that continues to reduce quantities of potentially toxic chemicals released to the environment. However, recycling can never be 100% effective. Chemicals that can be toxic at high concentrations are present in small quantities in almost every natural and synthetic product in our society, including human waste excretions. For many of society's waste products, the energy costs and associated adverse environmental effects of recycling and removal of traces of potentially toxic chemicals may far outweigh the environmental costs of properly managed disposal of the waste. In addition, storing all toxic chemical–containing wastes in "secure" landfills is not feasible. In most instances, using valuable land to dispose of large-volume wastes that have low concentrations of toxic chemicals is inappropriate. Perhaps more importantly, no landfill can ever be secure on a geological timescale. Wastes containing high levels of long-lived potentially toxic substances that are buried in landfills will eventually release these substances to freshwater and **groundwater** and from there to the oceans. The Earth's freshwater resources are severely limited, and any contamination of the freshwater poses a threat to human health and terrestrial ecosystems. Consequently, many environmental scientists believe that properly managed ocean disposal of wastes may be the environmentally preferable management option for wastes that cannot be efficiently recycled and that have low concentrations of potentially toxic chemicals.

The fate and effects of chemicals in the oceans are determined by how they are transported by currents, how they are removed and incorporated in sediments, and how they are taken up by the marine biota. Several chapters in this text describe the movements of ocean water and suspended sediment and the processes of marine **sedimentation** that determine the distribution and ultimate fate of dissolved and particulate constituents in the oceans.

Effects of Toxic Substances on Marine Organisms

CC18 describes the general principles that determine the toxicity of substances to living organisms. The essential point is that most substances are toxic to a specific organism only when their concentration in food or in solution in the surrounding water exceeds a certain level, above which, the concentration of the chemical within the organism is high enough to interfere with one or more biochemical processes critical to the organism's life cycle. For all potentially toxic chemicals, there is a concentration below which the chemical has no adverse effects.

The only exception to this rule may be for compounds that are **carcinogens** (cancer-causing), **mutagens** (causing **genetic** changes in the offspring by altering the parents' **DNA**), or **teratogens** (causing abnormal development of the **embryo**). There are conflicting views about whether a threshold concentration exists for such compounds. If there is no threshold concentration, any concentration of the compound will increase the incidence of disease. However, below a certain concentration, the number of individuals affected by a carcinogenic, mutagenic, or teratogenic chemical will be far smaller than the number similarly affected by naturally occurring compounds and natural **radioactivity**, which also have such effects. Hence, even for these compounds, there is a concentration below which they have no significant adverse effects.

Evaluating Toxicity

Assessing the toxicity of substances to marine organisms is very difficult. Two general approaches are used. First, the distribution of species in an ocean region affected by inputs of that substance can be compared with the distribution in similar but

unaffected regions or in the affected region before the anthropogenic inputs occurred. Second, **bioassays** can be performed in the laboratory to test the responses of organisms to various concentrations of the chemical (**CC18**). The laboratory results must be extrapolated to what might occur in the environment. Both approaches are difficult and prone to errors and uncertainties.

The field approach to toxicity evaluation requires a very detailed survey of the affected ecosystem and of control sites. Researchers must characterize the population levels and health of many species to be certain that the most susceptible species are included. Moreover, changes observed in the ecosystem may be caused by natural factors, such as climate variations. Hence, even if a significant difference in the ecosystem is observed between test and control sites, researchers usually cannot eliminate the possibility that it is natural and unrelated to contaminant toxicity. In addition, most contaminated locations receive inputs that contain a variety of toxic chemicals, all of which vary in concentration with time. Assessing which contaminant might be responsible for an observed change is difficult.

At present, we do not have a detailed understanding of the effects of toxic substances in the oceans, and we cannot predict with certainty the consequences of specific concentrations. Management of ocean uses that involve the release of toxic chemicals to the marine environment is therefore difficult and controversial. However, our understanding of the effects of toxic chemicals in terrestrial, freshwater, and groundwater ecosystems, and on human health, is not significantly better than our understanding of their effects in the marine environment. There is general agreement that we must exercise caution and, where practical, limit the release of potentially toxic chemicals to the environment. Nevertheless, wastes containing low concentrations of potentially toxic chemicals will always require disposal. If properly managed, ocean disposal of these wastes may be both safe and environmentally preferable to other disposal methods.

Bioaccumulation and Biomagnification

Two processes cause concentrations of substances that are potentially toxic to humans in marine organisms to become elevated in relation to concentrations in their food or environment (**CC18**): **bioaccumulation** and **biomagnification**.

Uptake and excretion of toxic substances by marine organisms are typically complex processes that involve transfers of the toxic substance among several different tissues of the organism, as well as to and from the organism's food and the surrounding water. Hence, some toxic substances may be taken up quickly when the environmental concentration increases, but released much more slowly when the environmental concentration decreases. The complexity of these bioaccumulation equilibria generally increases as the organism becomes more complex.

Higher order animals tend to store toxic substances in tissues where they are least harmful. Toxic substances are excreted only very slowly from these tissues after the exposure is reduced. In some cases, the loss is so slow that the concentration of the toxic substance tends to increase progressively during the individual organism's lifetime. The cumulative buildup of metals, such as lead, mercury, and arsenic, in human beings during their lifetime is an example. In such situations, short-term laboratory bioassay tests cannot accurately reflect the effects of long-term exposure to toxic substances or wastes.

Biomagnification is a special situation in which substances that are taken up by an organism from food or water are stored in tissues and there is no route by which these stored substances can be released. Biomagnified substances not only accumulate throughout an organism's lifetime but are transferred up the food chain so that the substance accumulates to higher concentrations with each step in the chain.

Trace metals are not biomagnified in marine food webs, except in a few instances when the metal is organically combined, such as in the methylated form of mercury. Biomagnification in ocean food webs appears to be limited to compounds that are highly soluble in fatty tissues and have relatively low solubility in water. These mainly include synthetic organic compounds, of which DDT and PCBs, and perhaps a few synthetic or naturally produced metal-organic compounds, such as methylmercury and tributyltin, are the primary ones.

Biomagnification of DDT and its decomposition products was responsible for the decline of pelican, sea lion, and elephant seal populations off California and throughout the eastern Pacific Ocean during the 1950s and 1960s. Almost all uses of DDT were banned in the U.S. in 1971. Since then, each of the affected species has recovered steadily in the eastern North Pacific Ocean. However, DDT continues to be used elsewhere in the world, particularly in tropical regions. Both DDT and its toxic but longer-lasting decomposition products are still found at high concentrations in marine species at higher **trophic levels** in all parts of the oceans.

Synthetic and Naturally Occurring Toxins

Two distinct classes of toxic substances are released to the marine environment by human activities: naturally occurring substances, including trace metals (e.g., copper, lead, cadmium) and petroleum hydrocarbons; and synthetic chemicals produced only by human industries (e.g., DDT and PCBs). These two classes should be viewed differently.

Marine organisms are adapted to generally low but variable concentrations of natural toxic substances in their environment. For example, lead and other toxic metals are either safely stored within marine organisms' tissues where they do not affect critical biochemical processes or safely excreted back to the environment. Petroleum hydrocarbons are metabolized to harmless substances by many fish species and used as food by many marine decomposers. Adverse effects occur only when these natural detoxification processes are overwhelmed by high concentrations of such compounds. This is fortunate, because total elimination of the releases of such compounds by all human activities is impossible.

Synthetic chemicals, such as DDT and PCBs, are not present naturally, and marine organisms may not have mechanisms to detoxify them. Consequently, these compounds have more unpredictable fates and effects in the marine environment. If they are persistent and not easily broken down by the chemical processes or metabolic processes of marine organisms, they tend to accumulate, persist, and cause adverse effects. Unlike contaminants that occur naturally, a synthetic chemical can be eliminated completely from our society and therefore from any further introduction to the environment. Alternate synthetic chemicals have now been designed and developed that adequately fulfill the purposes for which DDT and PCBs were intended but that, in contrast, are readily and rapidly decomposed in the environment. Unfortunately, developing, testing, and manufacturing new, effective, and rapidly biodegradable chemicals is very costly. The widespread adoption of such expensive replacement products is impeded primarily by economic factors, especially in poorer nations.

This page titled 16.3: Toxicity is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.4: Waste Disposal

For centuries, the oceans were viewed as a limitless sink for waste disposal. Consequently, during the industrial era, the oceans were used to dispose of sewage, dredged materials, construction dirt and debris, trash and garbage, chemical wastes, radioactive wastes, fish-processing wastes, used machinery and boats, and almost anything else that people needed to discard. Wastes have been dumped from vessels or discharged through outfalls into rivers, estuaries, and the ocean. Since the 1970s, indiscriminate use of the oceans for waste disposal has been recognized to cause adverse environmental impacts, some of them severe. As a result, ocean waste disposal is now viewed in the U.S. and elsewhere as being universally unacceptable. With the exception of dredged material, ocean dumping from vessels is now illegal in the U.S. and most other developed nations. In addition, stringent rules are in place that require sewage and industrial wastes to be treated before they are discharged through outfalls.

Contrary to popular belief, many marine scientists believe that properly managed disposal of certain wastes in the oceans is not only acceptable, but may be environmentally preferable to any possible alternative disposal or recycling technology for these wastes. Properly managed disposal of certain organic and nutrient-containing wastes, such as sewage and fish-processing wastes, might even have beneficial effects if done appropriately. However, because all ocean disposal is considered unacceptable by the media and the public, little or no research is being done to define how, where, and what types of wastes can and should be safely disposed of in the oceans. This situation may eventually prove detrimental to the global environment. Many wastes, particularly wastes containing toxic trace metals, might be safely dispersed in and assimilated by the oceans and/or removed by natural processes to ocean sediments (where they reenter the global biogeochemical cycle at a point where they are removed from contact with the biosphere for millennia), but instead they are buried in “secure” landfills from which they will eventually be released into the ground, groundwater, and then rivers and oceans.

This page titled 16.4: Waste Disposal is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.5: Ocean Pollution Issues

Petroleum

If asked to identify the sources of serious pollution in the oceans, most people would name oil spills at, or near, the top of their list. This is not surprising, since oil spills, unlike any other form of ocean pollution, provide spectacular images and video of leaking tankers (**Fig. 16-5a**), oiled marine animals and birds being rescued or dead on a beach (**Fig. 16-5b**), and towering flames and smoke from burning oil rigs (**Fig. 16-6**). As a result, media coverage of oil spills is always extensive, and since bad news sells better than good news, the media tends to emphasize and often exaggerate the extent of the environmental harm. In contrast, most other ocean pollution is almost impossible to photograph or video so gets little or no media attention. The net result of this is that the public perception is that oil spills are the most serious form of ocean pollution and that huge sums of money must be spent to monitor the impacts of each spill and to provide absolute assurance that spills "never occur again." Many marine oil spills have been extensively studied over more than the past half century and the findings of such studies support a different conclusion. While spill prevention is important and oil spills do have adverse impacts on the area in which they occur, the adverse effects are geographically limited and the affected marine ecosystems recover to a natural state (not to the same state as before since ecosystems are naturally always changing and a previous state can never be restored exactly) within a few years.

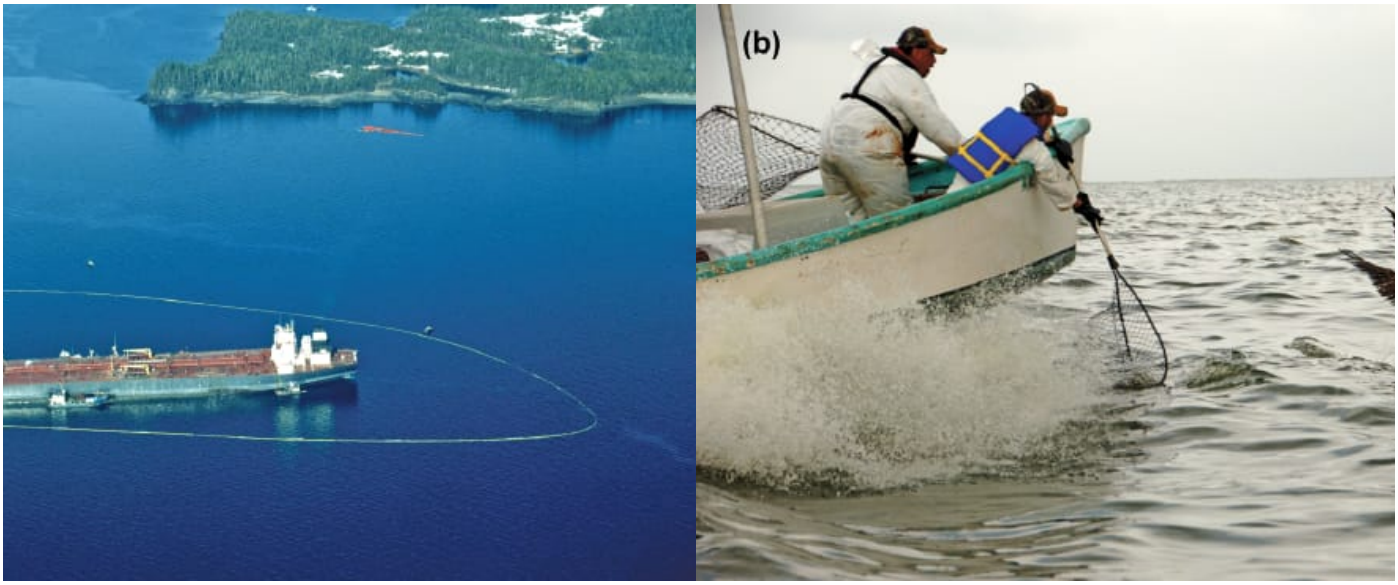


Figure 16-5. Oil spills provide spectacular images that ensure media coverage. (a) The *Exxon Valdez* tanker 48 h after it ran aground on Bligh Reef in Prince William Sound, Alaska, in 1989. In this photograph, the tanker is seen still leaking some oil, and it is surrounded by a boom that has been placed on the water surface in an attempt to contain this oil so that it can be collected. (b) A rescuer tries to capture an oiled Pelican after the 2010 *Deepwater Horizon* spill in the Gulf of Mexico. Rarely do heavily oiled birds survive even when cleaned up.

Figure 16-6. Major oil spills from the offshore oil industry are rare but dramatic events such as the *Deepwater Horizon* rig which is shown burning before it sank in the Gulf of Mexico, April 2010.

Oil spills are discussed in more detail in **Online Box 16B1**, but some of the more important findings are summarized here.

1. Oil is a natural substance that is consumed by microbial decomposers.
2. Oil spills have short-term adverse effects in the spill area including, deaths of birds and some marine animals due to oiled fur or feathers, oiled beaches, and disruption of plankton communities. Long-term effects are uncertain but may include some reduction in reproductive success of some marine species.
3. Ecosystems subject to spilled oil recover to a natural state within years in most areas to a decade or more in cold regions.
4. The use of dispersants or aggressive cleanup techniques such as hot water washing of rocky coastline or beaches generally leads to greater harm to the ecosystem than simply skimming off as much oil as can be done easily and allowing the microbial community to remove the remainder (adding nutrients to aid these microbial organisms may be appropriate in some cases).
5. Efforts to rescue and clean oiled marine animals and birds are often very expensive but result in very low survival rates for the cleaned animals.
6. Measures taken over the past several decades to minimize oil spills from tanker accidents and other accidental spills have been largely successful in reducing the frequency of large spills.
7. Chronic inputs of petroleum products from dispersed sources such as runoff from highways are still large and may have chronic adverse effects in some locations, especially rivers and estuaries where such inputs are large.

Sewage

Some human waste has been disposed in oceans, estuaries, and rivers for millennia. However, once sewers systems were introduced, they acted as a public health tool to ensure that human wastes were carried easily and efficiently for disposal in nearby bodies of water. Sewers are historically so important that human wastes are now called sewage. Sewage is a combination of liquid and solid natural materials that was disposed of in rivers, estuaries and oceans for centuries after sewers were introduced, generally without causing adverse effects on the discharge ecosystem. Most cities were, and still are, located next to or near rivers, estuaries or a coastline for reasons that included the availability of these bodies of water into which sewage could be discharged. As the human population grew and cities became ever larger, the quantities of sewage discharged increased as did the concentrations of sewage waste in bodies of water near these cities. By the mid 20th century, many rivers and estuaries worldwide were anoxic. This anoxia was caused by microbial decomposition of sewage organic matter that consumed oxygen faster than it could be replaced from the atmosphere. The ecosystems of many rivers, estuaries, and even some coastal ocean areas were severely damaged and waterside locations in many areas became undesirable places to live or work because of the smell of sulfide and partially decomposed organic matter. This was one of the primary driving reasons for the environmental revolution of the 1960s and 1970s.

The problems caused by sewage waste discharges prior to the 1970s were primarily due to the high organic matter loading of untreated sewage. Sewage treatment was developed and designed to address this problem by collecting sewage wastes in

treatment plants where the larger particle solids could be filtered or precipitated out (primary treatment) and then treated with bacteria to decompose the organic matter that remains after primary treatment (secondary treatment). Both steps generate sludge (mostly particulate solids with enough water to allow them to flow as a thick slurry). Secondary treatment of sewage is now required in the U.S. and in most nations and large cities, and has proved to be effective in alleviating the anoxia problems. Most river, estuary and ocean ecosystems that were damaged by overloading of organic matter from sewage have now recovered.

Despite the success of sewage treatment, several issues still remain because secondary sewage treatment is not designed to remove trace metals, many soluble organic compounds, human pathogens, or dissolved inorganic nitrogen and phosphorus, although it does partially remove most of these. In most areas, sewage entering sewage treatment plants is a mixture of: human waste water, water used for washing and cleaning that often contains potentially toxic chemicals, wastewaters from some industries, and in some areas storm drain runoff. These are all mixed in one waste stream. In the U.S. and many other nations, industrial pretreatment programs have been highly effective in drastically reducing inputs to sewage of trace metals and potentially toxic organic compounds from industry before they reach the treatment plant. Street cleaning and other measures have somewhat reduced storm runoff of metals and potentially toxic organics. Efforts to improve the removal of remaining non industrial continue to make slow progress.

Most pathogens in sewage are killed quickly in estuaries and ocean water but extensive monitoring programs are still required to assure that pathogen concentrations on swimming beaches are safe. Beach closures due to sewage discharges still occur infrequently when treatment plants have technical problems or when those sewage treatment plants that must treat storm water runoff combined with the sewage waste stream are overwhelmed by large rainfalls.

Sewage treatment has been perhaps the greatest success story in environmental management since the environmental movement began. Nevertheless, some issues remain, and a new one is increasingly becoming apparent. This issue is the nutrient loading due to treated sewage wastes. Recall, that secondary treatment was not designed to reduce the concentrations of nutrients like nitrogen and phosphorus and does not do so effectively. Recall also from earlier chapters that anoxia and dead zones driven by discharges of nutrients from rivers are a major and growing problem worldwide (**Chaps. 1, 12, 13**). Treated sewage discharges to rivers are a major contributor to the nutrient inputs that cause coastal dead zones and likely also cause an increase in total primary production in the surface oceans that contributes to the problem of deoxygenation of the deep oceans discussed in **Chapter 1** and later in this chapter. Tertiary treatment technologies do exist to reduce nutrient concentrations in sewage but they are not widely used. These tertiary treatment technologies are costly to perform and require large capital expenditures to upgrade and expand existing sewage treatment plants.

A more detailed discussion of sewage and sewage treatment can be found in **Online Box 16B2**.

Urban and Agricultural Runoff

Many chemicals are used in urban and agricultural communities. Fertilizers and pesticides are applied in large quantities to fields and gardens. Oil and rubber dust (containing toxic contaminants such as cadmium) are left on road surfaces by vehicles. Hydrocarbons and other chemicals are released from road-paving and other building materials. Particulates injected to the atmosphere from the burning of fossil fuels settle everywhere. Paints, solvents, acids, and other chemicals are spilled or released in many different ways, and toxic chemicals are still deliberately dumped to avoid costs of proper disposal. There are also many other ways for toxic contaminants and nutrients to be deposited on the land.

When it rains, contaminants are either absorbed into the soil or washed off into storm drains, streams, rivers, and eventually estuaries and the ocean. Many contaminants are carried by runoff in dissolved form, but most are carried on small organic-rich particles. These particles are carried by streams and rivers to estuaries and coastal ocean, where they can become trapped in the estuarine circulation and sediments (**Chap. 13**). Contaminants from urban and agricultural runoff reach estuaries through numerous drainage channels spread throughout the watershed. Consequently, their sources are often referred to as nonpoint sources. Nonpoint source inputs are highly variable. Such inputs are greatest during the first few hours of a rainfall, especially if it follows a prolonged dry period. Thus, contaminants that are carried into the estuarine and marine environment generally are diluted by large quantities of freshwater. These factors make the identification and control of nonpoint sources of contaminants extremely difficult and make the cost of possible treatment in most instances prohibitively high.

Urban and agricultural runoff contributes a major proportion of the nutrients and contaminants that enter many estuaries and coastal embayments. Substantial efforts have been made to control nonpoint sources. However, success has been limited, and nonpoint sources are still much greater than industrial and sewage inputs in most estuaries. Consequently, where such substances cause pollution problems, further controls on industrial and sewage inputs will have only a minor effect. The difficult task of controlling nonpoint source inputs must be addressed successfully if estuarine and coastal ecosystems are to be fully

protected and restored.

Pollution problems caused by urban and agricultural runoff are many and varied. Two examples illustrate the complexity of such problems. First, in Chesapeake Bay, nutrient-induced eutrophication has led to widespread, persistent anoxia. Although sewage discharges contribute some nutrients, the predominant sources are fertilizer and animal waste in runoff from surrounding agricultural land, as well as atmospheric inputs. Extensive efforts have been made to reduce these nonpoint source inputs—for example, by reducing fertilizer use and avoiding its application near streams. These efforts have resulted in some improvement, but the problem remains a difficult one.

The second example is the "dead zone," of the nearshore **continental shelf** of the Gulf of Mexico in which the bottom waters are seasonally hypoxic (**Chap 13**). The Gulf dead zone is located in the middle of one of the most important commercial and recreational fisheries in the U.S. The Mississippi River basin drains about 41% of the Lower 48 states of the U.S.—a total of over 3.2 million km^2. The drainage basin encompasses all or part of 30 states, a population of about 70 million people, and extensive agriculture. The Mississippi River discharge contains a high nutrient level, some of which is natural but much of which comes from runoff of fertilizers applied to farmlands and from treated sewage. In the last four decades of the twentieth century, the discharge of nitrogen by the Mississippi River basin tripled. It is this excess nutrient load that is believed to have created the dead zone, which was first documented in 1972. Efforts to reduce the nutrient loads of the Mississippi River are under way, but, as in the Chesapeake Bay region, these efforts are difficult and will take many years if the dead zone is to be eliminated.

Industrial Effluents

Most industries produce solid or liquid wastes that differ widely in composition among industries and even among factories within an industry. Most of these wastes contain various potentially toxic chemicals. A few decades ago, most industrial wastes were discharged to sewers or a local waterway or dumped in a nearby landfill.

The U.S. and most other nations have had laws for several decades to control contaminant discharges to aquatic environments. Enforcement of most of these laws has focused on industrial discharges and sewage treatment. Industries have been required to reduce drastically the concentrations of contaminants in their liquid effluents before discharge. Most industries have changed technologies to reduce the amount of wastes they generate and avoid or reduce the expense of treating effluents.

As a result of decades-long efforts to reduce toxic substances, nutrients, and other contaminants in industrial effluents, industry is now only a minor contributor to the contamination of estuaries and coastal oceans in most areas. Problems remain with some older industrial plants, with industrial discharges that are poorly located where residence time is long, and with the enforcement of discharge permits at a few plants whose unscrupulous owners or operators find ways to discharge wastes illegally. Nevertheless, the general public mistakenly still believes that industry is the principal polluter of the marine environment. There are several reasons for this misunderstanding. The most important is probably that it is easier for the media, politicians, and the public to blame industry for pollution problems than to face the difficult problems of cleaning up individual actions, homes, public utilities, and farms. Another reason is that the only demonstrated and confirmed incident in which the discharge of toxic chemicals to the oceans caused human deaths involved industrial effluents.

For many years, Chiso Chemical Corporation in Minamata, Japan, generated an organic form of mercury, methylmercury, as a by-product of its manufacture of acetaldehyde using mercury as a catalyst. A small portion of the methylmercury was discharged continuously to Minamata Bay in wastewater. Mercury, a toxic metal, accumulated in water, sediments, and fishes. Because methylmercury is more fat-soluble than elemental mercury and is biomagnified (**CC18**) in marine food webs, its concentration in fishes rose to very high levels.

The discharge of methylmercury to Minamata Bay started in 1952. By 1953, many cats were becoming ill, behaving erratically, and dying. Dead fishes were periodically found floating throughout the region (**Fig. 16-7a**). At the same time, some of the human population contracted a puzzling "disease" that produced numbness, disturbances in vision and hearing, and loss of the control of motor functions, similar to drunkenness. The "disease" quickly assumed epidemic proportions, and between 1953 and 1962 more than 40 people died and as many as 2000 other victims suffered what has later proven to be a persistent disability (**Fig. 16-7b**). In 1957, it was found that "Minamata disease" could be induced in cats by feeding them fishes taken from near the Chiso plant outfall. Although fishing was quickly banned, the discharge of methylmercury was not stopped until 1968, because Chiso denied that methylmercury caused the disease until it was overwhelmed by the weight of scientific evidence.

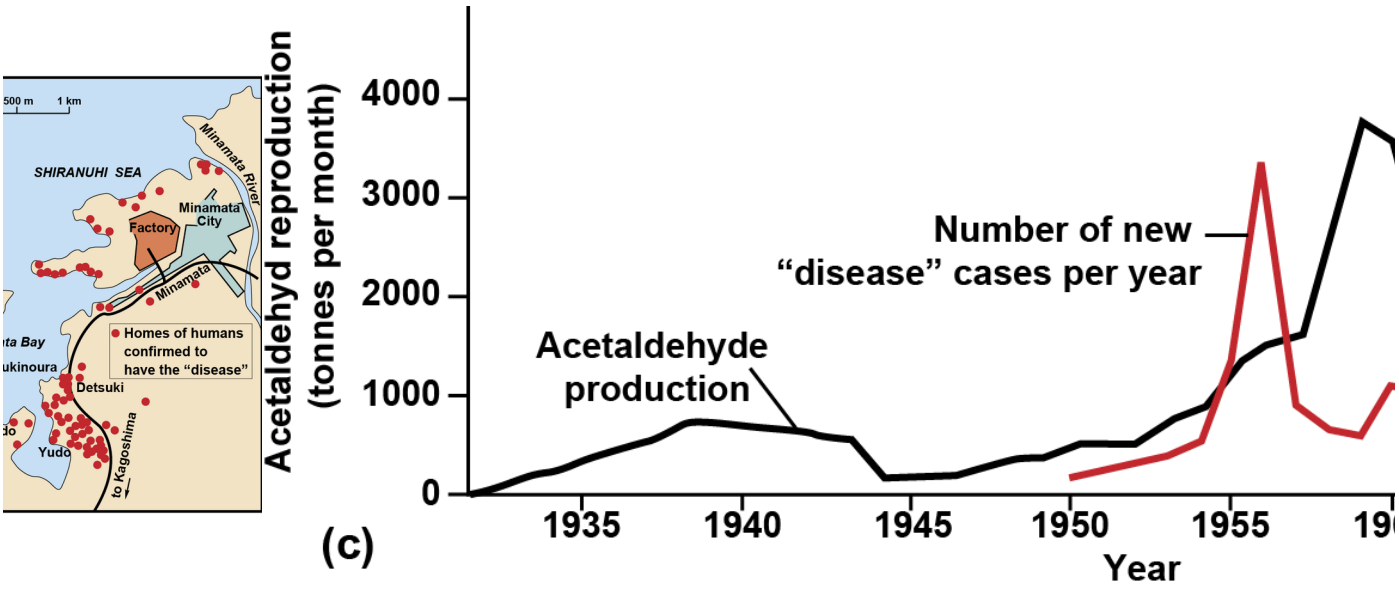


Figure 16-7. Methylmercury, discharged by a chemical factory at Minamata, Japan, and concentrated in fish and shellfish tissues, caused a major pollution problem in the 1950s. (a) Map of the locations around Minamata where diseased cats and dead fishes were reported many kilometers away from the discharge point. (b) Detailed map of the Minamata city area identifying the locations where humans lived who were confirmed to have contracted the "disease." (c) Forty-six people died of mercury poisoning, and many more suffered lasting effects. Even after the early poisonings, the Chiso Chemical Corporation continued to increase its production of acetaldehyde and its waste byproduct, methylmercury. Poisonings continued for years after the release of mercury from the plant was eventually curtailed.

Minamata is apparently the only documented case of human deaths caused by industrial inputs to the marine environment, but a number of cases of deaths of fishes and other marine organisms caused by industrial effluents have been documented. For example, in 1970, the Montrose Chemical Corporation in Los Angeles produced about two-thirds of the world's DDT. Starting in 1953, the plant discharged effluent containing DDT through a sewer to Santa Monica Bay. By 1970, this and other sources of DDT, which is fat-soluble and biomagnified, had caused the total collapse of many species at the top of the food web. It particularly affected pelicans, whose eggshells were so thinned by the DDT that they broke before hatching, and sea lions, which produced large numbers of stillborn offspring. DDT discharge was stopped, and the use of DDT was banned in the U.S. and Europe early in the 1970s.

Pelican, sea lion, and other affected populations have slowly recovered in the decades since the DDT ban. However, DDT and its persistent and toxic decomposition products can still be found in elevated concentrations in sediments and the biota of the southern California coastal zone. Unfortunately, despite its ban in the U.S. and many other countries, estimates suggest that more DDT is produced and used today than in 1970. Most of it is used for mosquito and malaria control in developing tropical countries that cannot afford more costly and less effective alternatives to DDT, especially since the resurgence of malaria since the 1990s.

Dredged Material

With the exception of a few deep-water ports, most ports and harbors do not have sufficient natural depth for vessels to navigate safely and berth. Consequently, most ports and harbors must be dredged to increase navigable depth. Once sediments have been removed to deepen a channel or basin, suspended sediments tend to settle in the new **topographic** depression. Therefore, all dredged channels and harbors must be re-dredged periodically to remove newly accumulated sediment. In some high-energy regimes where the new channels are regularly swept by currents, or areas where very little suspended sediment is transported to refill the channel, dredging may be necessary only once every several years. However, other channels must be dredged more often, and some require almost continuous dredging.

Dredging is done either by scooping up buckets of sediment or by sucking sediment from the seafloor through a pipe lowered from a surface vessel. In most cases, the dredged material is loaded onto barges and transported to another location before being dumped into an estuary or the ocean.

Dredging destroys the benthos, causes increased turbidity, and, if the dredged material is contaminated, releases toxic substances into the water and suspended sediment at both dredging and dumping sites. These effects may be serious at some dredging sites, but they are generally more damaging at the disposal site. The U.S. has more than 150 aquatic disposal sites for

dredged material. Most are in bays or coastal waters very close to the mouth of the estuary from which the material is dredged.

When dredged material is dumped, it descends quickly until it strikes the seafloor. Upon impact, it spreads rapidly across the seafloor as a suspended sediment cloud. Coarse-grained materials settle out quickly. Some smaller particles are partially buried with the coarse-grained material, and others are carried off by currents and deposited elsewhere where current speed is slower. Most or all benthos in the area buried by the dredged material are killed. The area is recolonized over the next few weeks or months except where the dredged material is badly contaminated with toxic substances or if dumping is frequent or continuous.

Because most dredged material consists of sediment taken from channels close to industrial and urban areas, it is usually contaminated with toxic substances from urban runoff, industrial discharges, sewage, and spills. Dumping at most dredged-material dump sites is continuous or is done several times each year. Consequently, the bottom at almost all such dump sites is covered with contaminated sediment, and the benthos is severely damaged on a continuing basis.

A proportion of the fine-grained contaminated dredged material is released as suspended sediment during dumping. Because the contaminants associated with these particles will be at least partially **bioavailable**, dredged material can contribute substantial quantities of toxic substances to the marine environment. For example, until the 1990s the amount of several toxic contaminants dumped annually at the dredged-material dump site near Alcatraz Island in San Francisco Bay (**Fig. 16-8**) exceeded the combined total annual input of these toxic substances to San Francisco Bay from more than 100 industrial and treated sewage discharges. Some of the dredged material previously dumped at the Alcatraz site has now been diverted to disposal on land or at a deep-ocean dump site.

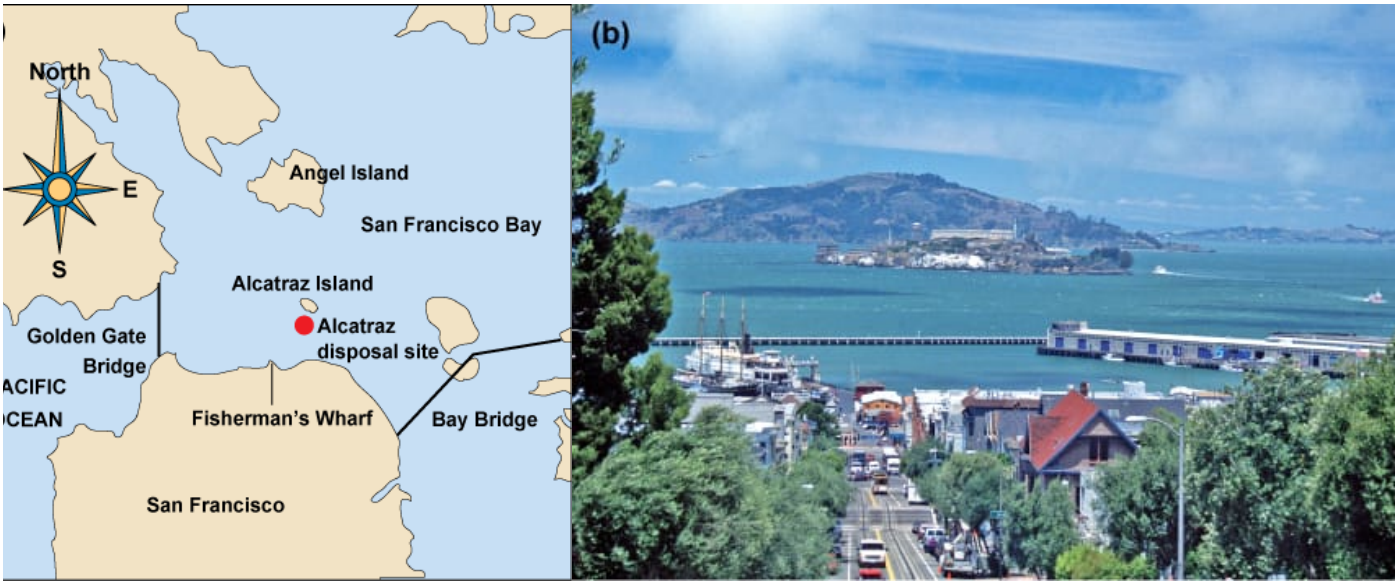


Figure 16-8. Dumping of dredged material in San Francisco Bay. (a) The Alcatraz dredged-material dump site in San Francisco Bay is located close to the estuary mouth and to the city of San Francisco. (b) The dump site, which is located between Alcatraz Island and the shoreline in this photograph, is only a few hundred meters from one of the famed tourist attractions of San Francisco, Fisherman's Wharf, and it is within easy view from much of the city.

At some dump sites, very heavily contaminated dredged material is covered within a few days by a cap of clean, sandy dredged material several tens of centimeters deep. Once under this cap, contaminants are effectively removed from the biosphere, but it is not clear what fraction of the contaminants, especially those associated with fine-grained materials, escape before being capped or whether erosion caused by waves and currents eventually breaches the cap.

Most dredged-material dump sites are located in low-current areas to ensure that as much of the dumped material as possible is retained at the site. However, at some sites, such as the Alcatraz dump site in San Francisco Bay (**Fig. 16-8**), with its fast tidal currents, all but the largest dredged particles are resuspended and swept away soon after their disposal. Alcatraz and other similar dump sites were originally selected near estuary mouths because it was asserted that the dumped material would be quickly swept to sea by the river flow. Unfortunately, the dredged material dumped at many such sites, including the Alcatraz dump site, is instead transported back into the estuary by estuarine circulation (**Chap. 13**).

The practice of dredging and dumping the dredged material within estuaries has two important effects. First, the material is partially transported back to, and deposited at, dredging sites within the estuary, thereby increasing the frequency and cost of

dredging. Approximately 40% to 50% of the dredged material dumped at the Alcatraz site returns to dredging sites and must be re-dredged. Second, contaminants that were discharged to estuarine sediments decades ago, when waste discharges were largely uncontrolled, are continuously re-dredged, dumped, and resuspended within the estuary, where they come into contact with the biota. The Alcatraz site has been used for dumping dredged material since the 1890s. Since that time, the once extremely valuable commercial fish and shellfish populations of San Francisco Bay have been largely destroyed. Dredged material dumping has undoubtedly contributed to their decline.

Even if treatment techniques were available, treating estuarine sediments to remove historical contamination would not be possible because the sediment volumes are too large. In addition, land disposal of the dredged material is practical only in a few locations, and then only if the material is not contaminated. However, if dredged material disposal sites were moved offshore, beyond the biologically critical coastal zone, the contaminated material, especially if properly capped, would have less environmental impact on the limited benthic communities of the outer continental shelf and continental slope. In addition, if dredged material were dumped in the ocean, and if contaminant source control continued in the estuaries, contaminant concentrations in the estuaries would decrease, and dredged material eventually would be clean enough to be used for beneficial purposes on land.

Plastics and Trash

Although plastics are now known to be decomposed by bacteria in the oceans, they decompose very slowly in the environment. However, some plastic debris first breaks down into smaller particles that can be mistaken for food and eaten by marine organisms. Many species can neither digest the particles nor pass them through the gut, and plastic particles can accumulate in the gut until it is blocked, leading to starvation and death. Larger fragments of plastics are ingested by large animals, especially turtles, which mistake plastic bags and other plastic debris for their favorite food, **jellies**. Plastic fishing line, six-pack rings, plastic fishing nets, and all kinds of other plastic debris act as drifting traps that entangle and kill turtles, birds, and marine mammals (**Fig. 16-9a**).

Figure 16-9. Plastics can cause more than just aesthetic problems in the marine environment. (a) This northern fur seal pup is entangled by a section of lost or discarded gill net and would not have survived long without human intervention. (b) The remains of a dead Laysan Albatross chick (*Phoebastria immutabilis*) show the large variety and quantity of plastics that it had ingested. The ingested plastics were likely at least a contributing cause to the chick's death.

It has been found that microscopically small plastic fragments are ubiquitous in recent marine sediments and seawater. These fragments have been found in preserved plankton samples from the 1960s, and examination of more recent samples shows that their concentration has increased significantly since then. Adverse effects of plastic ingestion have been observed in larger animals that ingest relatively large particles of plastic (**Fig. 16-9b**), and there is substantial evidence that the ubiquitous microscopic fragments have adverse effects on zooplankton and benthic animals.

International law now prohibits the disposal of plastics in the oceans, but enforcement of this prohibition is virtually impossible, and violations are common. Plastic also reaches the oceans from many sources other than direct disposal, including storm-drain

runoff and helium-filled balloons released at sporting events and other festive occasions.

In the 1980s, it was discovered that there was a large area at the center of the North Pacific subtropical gyre where floating plastic debris collects, carried into the center of the gyre by the Ekman transport processes described in **Chapter 8**. Subsequently, similar concentrations of plastic debris have been found in the centers of each of the five subtropical gyres (**Fig. 16-10a**). The area of greatest plastic debris concentration has been referred to as the North Pacific garbage patch, and the media have portrayed this area as being a huge visible debris field. However, this is a significant mischaracterization since the debris in most of the affected region (and affected regions in the four other subtropical gyres) primarily consists of particles that are generally invisible to the naked eye, with usually widely scattered larger plastic pieces or items. Nevertheless, the accumulation of plastic debris in these mid-ocean regions poses a threat to at least some marine organisms and birds. Since 1980, substantial efforts have been made in most countries to reduce the amount of plastics that enter the oceans, but these have only had modest success. There are also ongoing studies investigating how plastic debris might be collected and removed from the centers of the gyres. However, the amount of plastics is very large, and most of it is widely dispersed, so any removal efforts will be extremely costly and difficult.

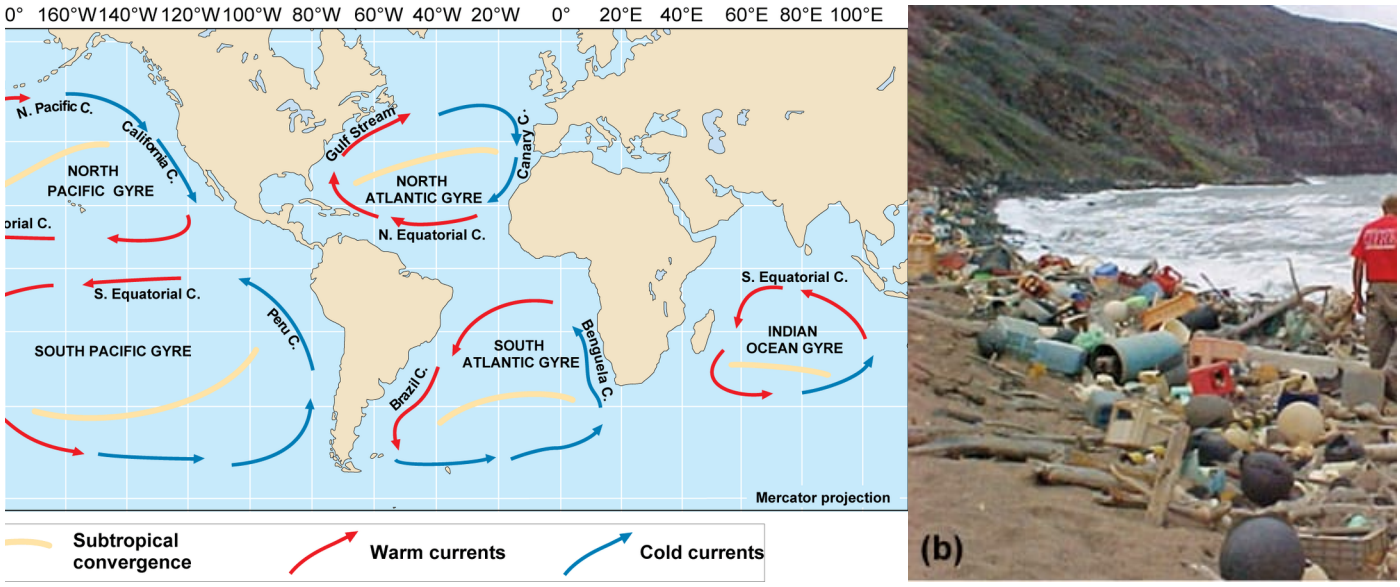


Figure 16-10. (a) Plastic debris collects in the subtropical convergence zones at the center of each of the subtropical gyres. The largest concentration is apparently in the center of the North Pacific gyre and has been dubbed the "Pacific Garbage Patch." (b) Plastic debris also washes up on the beaches of islands, such as the Hawaiian Islands, that are located within the subtropical gyres.

Metal, paper, and other trash items cause fewer problems in the ocean than plastics because they eventually corrode or decompose. Nevertheless, these discarded materials, including plastics, cause aesthetic problems when they wash up on beaches and shores or litter the seafloor that divers visit (**Fig. 16-10b**). Included among the floatable wastes are medical wastes, such as syringes, illegally discarded from vessels and into rivers and storm drains. These wastes have led to precautionary beach closures when they wash ashore. Storm drains are considered to be the primary source of most types of floatable wastes present in the oceans.

Antifouling Paints

Unless protected, almost any surface introduced into the ocean is quickly covered, or fouled, with a wide variety of marine animals and algae. Barnacles have an especially strong natural adhesive that enables them to attach and hold on to surfaces even in swift currents that can sweep off other fouling organisms. Hence, barnacles dominate the fouling community on vessel hulls and on structures in strong currents.

When a normally smooth vessel hull or other moving surface is fouled, **friction** between the hull and water increases, causing the vessel to either lose speed or use more fuel to maintain speed. Consequently, vessel hulls, turbines of tidal power plants, and water intakes and drainpipes must be protected from fouling.

The predominant and most effective means of reducing fouling is to coat the substrate with paint containing a toxic substance that will kill any organism that settles. The toxic substance must be bioavailable to perform its function, so the paint must release its toxic substance slowly into solution. Consequently, vessels and other structures release these toxic substances to the marine

environment, causing an especially serious problem in harbors where boats are concentrated and water residence times are long.

For many years, most antifouling paints contained copper as their active toxic ingredient. Copper is highly toxic at moderate concentrations in its ionic form. However, at lower concentrations, it is nontoxic and readily **complexes** with organic matter, reducing its toxicity. Thus, as copper is released from an antifouling paint, it has the desired antifouling action, but it is quickly diluted and complexed with natural organic matter as it **diffuses** away from the treated surface. Copper does accumulate in the water of harbors and marinas, but adverse impacts on biota other than those fouling the painted surface have been extremely rare.

Although copper-based antifouling paints are effective, their effectiveness diminishes as the copper progressively dissolves. To reduce the frequency with which ships' hulls must be cleaned and repainted, a more effective antifouling paint was developed in which tributyltin replaced copper. Tributyltin is much more toxic than copper and is released more slowly from antifouling paints. Hence, each application of tributyltin-based paint was effective for a longer period than an application of copper-based paint.

Unfortunately, once released into the water, tributyltin is degraded only slowly into a less toxic chemical form. Consequently, it retains its toxicity and tends to concentrate in the water in areas where many boats have such paints. Tributyltin has caused serious marine pollution problems in some areas. Many ports and marinas where tributyltin paint was used extensively were essentially denuded of all animal life. Because tributyltin is particularly toxic to **mollusks**, it destroyed populations of oysters and other species over a wide area far from its source in some estuaries. There has been a worldwide ban on the use of tributyltin-based paints since 2008. Fortunately, unlike DDT, tributyltin does not persist in the environment for decades. Rapid recolonization and recovery have occurred in many areas previously affected by tributyltin.

Radionuclides

Among the most feared words in the English language are nuclear, radioactive, and radionuclide. This is, of course, because most people associate these terms with nuclear bombs. As a result, during the nuclear testing era from 1945 to about 1990, there was considerable attention, both by the public and among environmental researchers, to the introduction of radioactive materials to the environment, including the oceans, of radioactive fallout from nuclear tests. Dumping of radioactive wastes and discharges or accidental releases from nuclear industrial facilities. Some of the radionuclides from these sources do not occur naturally on Earth. Public fears of radioactive materials in the environment were heightened by a nuclear power plant accident in Chernobyl, Ukraine in 1986 and further exacerbated by the severe damage of several nuclear reactors near Fukushima Japan in 2011 following a massive earthquake and tsunami.

There has been extensive research on the effects of radioactive material from nuclear bomb test fallout, and radioactive materials discharged or dumped, either deliberately or as a result of accidents. The results of this research are summarized in **Online Box 16B3**, and the principal findings are summarized here.

1. In the marine environment, radionuclides enter **biogeochemical cycles** and behave in the same way as the stable **isotopes** of their elements (if stable forms exist).
2. Each element and its **radioisotope**(s) have their own distinct behavior in the oceans. For example:
    1. Tritium (hydrogen-3), released in nuclear explosions, quickly combines with oxygen to form water molecules that join the ocean circulation system, where they behave almost exactly as other water molecules do.
    2. An iodine radioisotope, iodine-131, enters solution in ocean water and is concentrated by marine biota, particularly certain species of algae.
    3. Plutonium does not occur naturally but is rapidly attached to particles in the ocean environment. Most plutonium is carried to sediments by these particles, where it remains until it decays to other elements.
3. Radioisotope concentrations are easily measured even in environmental samples that contain only a few atoms of the isotope. For this reason, radioactive isotopes are used as **tracers** to study geochemical, biological, and physical processes in the oceans (**Chap. 8**). They are also commonly used as tracers in biological systems, including human medical scans.
4. Radioactivity can have adverse effects on marine ecosystems and on human health, primarily because some anthropogenic radioisotopes concentrate in seafood and increase the risk of cancer in consumers.
5. Except for very limited areas surrounding some nuclear bomb testing sites, concentrations of anthropogenic radioisotopes in seafood are well below background levels of naturally occurring radioactive elements.
6. The accepted theory of carcinogenicity is that any increase in radioisotope intake will increase the risk of cancer. The oceans' assimilative capacity for anthropogenic radioactive materials may therefore be zero. In reality, a very small increase in radioactivity above natural background levels will not produce a measurable or significant increase in cancers.
7. Because the ocean's assimilative capacity for radioactive materials is so small, international agreements require that anthropogenic inputs be eliminated. Almost all nations adhere to these agreements.

8. A very small amount of liquid radioactive waste continues to leak from the Sellafield nuclear industrial complex in Sellafield, Northern England.
9. The former Soviet Union routinely dumped and discharged large quantities of solid and liquid radioactive waste into the Arctic Ocean, the Sea of Japan, and rivers that empty into the Arctic Ocean, until 2005, when Russia agreed to the global treaty that bans these practices.
10. In 2011, the Fukushima Daiichi nuclear power reactors in Japan were damaged by an earthquake of magnitude 9.0 (a once in a century occurrence) and the more than 40 m-high tsunami it caused (a once in a thousand years occurrence). Large quantities of water containing radioactive elements were eventually released into the oceans.
    1. Studies showed that radioactivity levels rapidly reduced with distance from the reactors.
    2. Within five years, levels were too low for any adverse effects to be found in any marine organisms studied (ranging from microalgae to mollusks and fishes), even in ocean areas close to the accident site.
    3. The major impact on the oceans appears to have been the precautionary closure of the area to fisheries.

Fears of radioactivity releases fueled by media accounts of the Fukushima and Chernobyl, Ukraine (1986) power plant accidents, and a 1979 accident at Three Mile Island in Pennsylvania (that released small amounts of radioactivity to the atmosphere) have caused public perception to conclude that nuclear power plants are unsafe and nuclear power should not be used in the future. While these fears are understandable, they are not based on scientific facts.

1. At present, there are about 420 nuclear power plants operating worldwide, and these provide about 9% of the world's electrical power generation capacity and needs.
2. Nuclear power is the only electricity-generating technology that is
    1. Free from carbon dioxide emissions to the atmosphere,
    2. Operates 24 h per day and can be operated at variable levels to meet hourly demand fluctuations, and
    3. Can provide city-scale amounts of power from a single plant with a small land-use footprint.
    4. Has a long history (since the 1950's) of safe operations. The Chernobyl accident occurred at a poorly designed Soviet reactor with no safety containment vessel. The Fukushima accident was caused by one of the most powerful earthquakes, plus one of the largest tsunamis in human history. No other major accidents have occurred that have caused deaths or significant radioactivity exposure beyond a small number of plant employees.
3. Waste disposal is a significant issue with nuclear power. However, there are breeder reactors that reduce the radioactivity of waste from other nuclear power plants, or wastes could be placed in safe storage and disposal sites. Disposal sites must be located where wastes will remain undisturbed until radioactivity is reduced to background levels. The best such geologically stable long-term storage locations may be beneath the sediments in the center of oceanic tectonic plates. Technologies to emplace radioactive waste deep within ocean sediments were extensively researched and developed in the U.S. in the 1970s. This research was abandoned after the London Dumping Convention adopted a ban on all disposal of wastes in the oceans in 1975.

Noise

The ocean environment has always been noisy, with sounds generated by many marine species that use sound for communication or other purposes (**Chap. 14**) and by rain, waves, and earthquakes. However, the oceans are becoming progressively noisier due to sound from ships, offshore resource extraction, including offshore oil and gas development, from military and civilian use of sonar, and from seismic surveying. Very little is known about the individual sources of sound in the oceans or about how anthropogenic noise may affect marine species. However, there is concern that anthropogenic noise may adversely affect some species, especially marine mammals that communicate and navigate by using sound.

In March 2000, 14 beaked whales and two minke whales became stranded on beaches in the Bahamas after they were exposed to sound emitted by the U.S. Navy in testing of a new high-intensity, mid-frequency sonar. Six of the beaked whales died and were found to have internal bleeding that apparently resulted from close-range exposure to the high-intensity, mid-frequency sonar. As a result, use of this sonar, which uses higher intensity than other anthropogenic sound sources and a frequency not used in other sonar applications, has now been severely restricted. Precautions are now taken to ensure that no marine mammals are within a wide exclusion zone whenever this sonar, which is believed to be essential for national security, is tested. Also, as a result of the finding that excessive sound can harm marine mammals, ongoing research aims to study the effects of all anthropogenic sounds in the oceans.

Nonindigenous Species

Marine species are transported around the world in ships' ballast water, attached to ships' hulls, and in other unintended ways.

Species are also deliberately transported from one location to another to be used in aquariums or for aquaculture, and many of these are inadvertently or deliberately released to marine ecosystems in which they do not occur naturally. Once in their new environment, these species may die if they are unable to tolerate the new conditions or avoid their new predators, or they may survive and reproduce with little effect on the rest of the ecosystem. However, some nonindigenous species not only survive but also reproduce, spread rapidly, and out-compete native species. Estimates suggest that approximately 15% of introduced species cause severe harm to their new environment.

Many examples can be given of nonindigenous species that have severely damaged terrestrial and freshwater ecosystems. Fewer examples of damage are known in the marine environment, probably for several reasons. First, marine biota are not as easily observed as freshwater and land biota; thus, adverse changes due to nonindigenous species may not be noticed. In addition, because the ecology of marine ecosystems is poorly known, declines of important species and other ecological changes may be attributed to other factors, even if caused by nonindigenous species. Open-ocean ecosystems are generally well connected with each other and more uniform than terrestrial or freshwater ecosystems. However, coastal marine ecosystems, especially estuaries, are more isolated from one another and therefore more vulnerable to the introduction of nonindigenous species.

San Francisco Bay is an excellent example of an estuary that has been severely damaged by nonindigenous species. It is the only major estuary on a long stretch of coast, and as a result, it is isolated from other estuaries. Therefore, the Bay had many unique species when European settlers arrived. The settlers deliberately introduced nonindigenous species, such as the eastern oyster (*Crassostrea virginica*), in an attempt to populate the bay with valuable food species. The oyster did not survive, but a variety of other species brought in with the oyster, either attached to its shell or in the seawater in which it was transported, did become established.

Eastern oysters were the first of many introductions. Among these introductions, one notable species was striped bass, which became a major anadromous sport fish. However, the striped bass competes for food and habitat with indigenous species, such as salmon and sturgeon. These and other introductions of nonindigenous species into San Francisco Bay have caused dramatic changes. Most original species in the Bay are now gone, and the majority of species present today were introduced. For example, almost half of the fish species in the critical wetland habitat are nonindigenous. Nearly all invertebrates that now inhabit shallow, nearshore parts of the Bay are also non-indigenous. Indigenous clams and other edible mollusks have been largely replaced by a small, economically valueless species of Asian clam.

Habitat Alteration

Alteration or destruction of habitat is among the most damaging forms of pollution in estuaries and the coastal zone. The most serious habitat destruction is the filling of wetlands to create dry land. An estimated 450,000 km^2 (almost 50%) of the historical wetlands of the U.S., excluding those in Alaska, have been lost since European settlers arrived. In some areas, such as San Francisco Bay, less than 10% of the historical wetland area remains (**Fig. 16-2**). Many tens of thousands of acres of wetland habitat for aquatic birds, juvenile fishes, and other species have been lost. Other forms of adverse habitat alteration or loss include beach erosion or loss due to coastal structures (**Chap. 11**), dredging of channels through mudflats, burial of seafloor by dredged material, and damage to coral reefs by anchors.

Less obvious habitat alteration is caused by human activities that lead to changes in salinity, temperature, and turbidity. Soil erosion has increased in many watersheds because trees and other vegetation have been removed. Increased suspended sediment loads have altered sediment grain size in many rivers, making the sediments unsuitable as **spawning** habitat for anadromous fishes. Excess turbidity has reduced light penetration and thus rendered both benthic and pelagic habitats less suitable, or unsuitable, for algae growth, with resultant effects throughout the food web. For example, more than 90% of the historical **sea grass** cover has been lost in Galveston Bay, more than 75% in Mississippi Sound, and more than 50% in Tampa Bay.

Reduction of freshwater flow into estuaries is a particularly serious form of habitat alteration. In many estuaries, species zonation is based on the salinity distribution. For example, plants rooted in freshwater are replaced by saltwater species in the lower parts of the estuary, where salinity is higher. If freshwater inputs are drastically reduced, as when rivers are dammed, salinity ranges can be shifted many kilometers up an estuary, dramatically altering the balance between freshwater-dominated and seawater-dominated habitats.

In some estuaries, such as San Francisco Bay, freshwater input reductions have caused the critical **brackish** water zone, where much estuarine productivity occurs, to migrate from a wide wetland-fringed part of the estuary into a deeper, narrower, faster-flowing section. Primary productivity has been lowered because of the smaller area of suitably brackish water and the high turbidity. Such loss of primary productivity may have contributed to dramatic declines in anadromous fish species in this

and other estuaries. Reduction of freshwater inflows also tends to increase residence times within the estuary. Consequently, contaminants have longer residence times and reach higher concentrations.

Fishing

Far from being the traditional environmentally friendly activity that it is often portrayed to be by the media and thought to be by the general public, fishing is acknowledged by most marine scientists to be the human activity that has caused more adverse impacts on the oceans than anything other than anthropogenic releases of carbon dioxide and nutrients. Fisheries and overfishing, their principal impact, are discussed in **Chapter 2**.

Fishing, particularly commercial fishing and subsistence fishing in some nations, contributes to ocean contamination or pollution in many ways beyond overfishing. The most significant impact of fishing on ocean environments is the habitat disturbances and destruction caused by trawl fishing for fish and shellfish that involve dragging nets across the seafloor. Trawling generally destroys or collects all epibenthos, including delicate and extremely slow-growing deep sea corals, which take centuries to regrow even if they can recolonize the altered seafloor sediments.

Large quantities of fishing gear, including fishing line, fishing nets, and Styrofoam floats, much of which are made from plastics, are lost and can have adverse effects on marine organisms, as discussed earlier in this chapter. This lost or discarded fishing gear is so widespread and problematic that the media often refers to it as “ghost gear.” In addition, vessels intentionally or accidentally discharge oil and diesel fuel, oily bilge waters, sewage, and food and packaging wastes. Such discharges are regulated or banned by many nations, including the U.S. However, as is true of fishing regulations, anti-discharge laws are difficult to enforce on the high seas, and especially difficult in those areas that are not part of any nation’s exclusive economic zone (EEZ), where many of these laws do not even apply. Fishing vessels also have hull paints that contain toxic chemicals to combat fouling by organisms such as barnacles, and these toxic chemicals can damage fishery resources.

This page titled 16.5: Ocean Pollution Issues is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.6: Climate Change, Acidification, and Deoxygenation

The global radiation budget is not currently in equilibrium. The imbalance between heat energy entering and leaving Earth's atmosphere is difficult to measure because it varies with the sun's energy output, which varies in short term to decadal and longer cycles time scales. However, it has been estimated that in 2025 Earth radiated to space about 1.2 W/m^2 less energy than it receives from the Sun, creating an imbalance of 0.35% of the total incoming solar radiation at the top of the atmosphere. This imbalance is estimated to be growing at a rate of about 0.045 W/m^2 per year. In response to this radiation imbalance, additional energy is accumulating in Earth's climate system and Earth is warming. Satellite sensors and instrumented buoys and floats at sea show an accelerating increase in heat content of the upper 2 km of the ocean since the 1970s.

Because Earth's climate system has considerable thermal inertia, which is mostly due to the ocean's high heat capacity, a considerable length of time is required for Earth's climate system to achieve equilibrium under new environmental conditions. Unless there are unforeseen changes (such as a major increase in global volcanic activity) global climate models predict that rising concentrations of atmospheric carbon dioxide and other greenhouse gases will cause global warming to persist through this century and beyond. Even if greenhouse gas emissions were to stabilize at present levels, global warming would continue well beyond the 21st century. The magnitude of warming will depend on the rate of continued greenhouse gas emissions and related feedbacks.

Climate change is geographically non-uniform in both magnitude and sign so the rise in global mean annual temperature predicted by global climate models does not represent what might happen everywhere. Climate models predict polar temperature amplification, meaning that warming will be greater at higher latitudes.

The residence time of carbon dioxide in the atmosphere is between about 300 and 1000 years. The slow rate of removal of carbon dioxide from the atmosphere means that, if the rate of carbon dioxide emissions were to stabilize at the current level, the atmospheric concentration of carbon dioxide would still continue to increase. If anthropogenic emissions can be substantially reduced but not eliminated entirely, carbon dioxide would still enter the atmosphere faster than it is cycled out, and the atmospheric carbon dioxide concentration would continue to increase, although at a slower rate. Only the complete elimination of anthropogenic carbon dioxide emissions would stabilize and then, eventually, reduce the concentration of atmospheric carbon dioxide from current levels.

Approximately 40% of the estimated anthropogenic carbon dioxide released into the atmosphere since the beginning of the Industrial Revolution has been absorbed by the oceans. Currently, about 15% to 30% of anthropogenic carbon dioxide gas released to the atmosphere annually is removed from the atmosphere to the oceans. Some of the anthropogenic carbon dioxide absorbed by the oceans is transported to the deep oceans below the permanent thermocline by sinking of cold surface water in the global thermohaline circulation. Some carbon dioxide is also transported to the deep waters by sinking of organic matter produced by photosynthesis in the surface layer. The residence time of ocean deep waters is long, so this carbon dioxide is sequestered for as much as a thousand years before returning to the atmosphere-ocean interface.

Climate change, driven by rising concentrations of atmospheric carbon dioxide and other greenhouse gases, causes ocean surface waters to warm. Warming increases ocean density stratification and slows thermohaline circulation. Slowing the thermohaline circulation causes deep ocean water to spend longer in the deep oceans, where its oxygen concentration declines steadily due to the transport of organic matter from above and the respiration of decomposers. The amount of organic matter transported to the lower layers of the oceans is increasing due to enhanced photosynthetic primary production in the surface waters, which is due to higher concentrations of carbon dioxide and nutrients introduced by humans. The increased primary production contributes to declining oxygen concentration or deoxygenation in deep oceans. Absorption of anthropogenic carbon dioxide in ocean water also causes the water to acidify.

Thus, the oceans are subjected to acidification, deoxygenation, and circulation changes all caused by the same anthropogenic activity - carbon dioxide release to the atmosphere. Each of these can cause adverse effects on ocean ecosystems, and the synergistic effects of their acting together can be severe. Potentially so severe that this combination of rapid change in atmospheric greenhouse gases, ocean acidification, and deoxygenation of the deep oceans has been associated in the historical record with most, if not all, of the five major mass species extinctions that have occurred in Earth's 4 billion year history. The possibility that our anthropogenic releases of carbon dioxide could be driving the ocean ecosystem toward another mass extinction is both real and serious. The consequences of a 6th mass species extinction to humans and Earth's ecology would be catastrophic beyond anything that could possibly be caused by any of the other ocean pollution issues discussed in this chapter. Consequently, we discuss these three interrelated issues together in this section.

The Oceans and Climate Change

The ocean is the largest heat sink in Earth's climate system and has been found to contain more than 90% of the additional heat energy captured by Earth due to anthropogenic greenhouse gases. About one-third of this heat energy is now in the deep ocean (below 700 m). Since 2005, the rate of total anthropogenic heat accumulation in the deep ocean has exceeded the rate in the upper layer (0 to 700 m).

Models predict that the ocean will continue to warm during the 21st century, with the strongest ocean warming projected to occur in the surface layer in tropical and Northern Hemisphere subtropical regions. At greater depth, the warming will be most pronounced in the Southern Ocean.

Temperature and Salinity Changes in the Oceans

Atmospheric global warming has accelerated in recent years, with the top 10 warmest years occurring since 2015. Since 1979, the rate of warming has been about 0.20±0.02°C (0.36±0.04°F), resulting in approximately 0.9°C (1.6°F) of warming over the last 45 years. The atmospheric warming has been accompanied by the warming of ocean waters

A warming atmosphere and ocean surface lead to a redistribution of rainfall and ocean salinity patterns. Salinity increases in regions where there is increased evaporation of ocean water. Conversely, increased inputs of fresh water from rivers, precipitation, and melting ice decrease salinity. Observations show that mid-latitude surface ocean water has become saltier, indicating the evaporation rate has increased in this zone. Surface waters in tropical and polar regions have become less salty. In other words, the moisture surpluses from the mid-latitudes condensed into liquid that precipitated in nearly equal amounts in these tropical and polar regions. This pattern is expected to lead to weakening of the meridional overturning circulation, impacting the climate of northern latitudes, contributing to increasing deoxygenation of the deep water layers of the ocean, as discussed in more detail later in this section.

Polar Regions

Polar temperature amplification means that the current global warming trend is greater at higher latitudes. Amplification of warming at higher latitudes threatens the ice sheets of Antarctica and Greenland. About 90% of Earth's glacial ice blankets Antarctica, and melting this glacial ice would cause a catastrophic rise in sea level.

Two ice sheets cover most of Antarctica, separated by the Transantarctic Mountains. The larger of the two, the East Antarctic Ice Sheet (EAIS), is situated on a continent about the size of Australia, averages about 2 km in thickness, and accounts for two-thirds of Antarctic ice. Complete melting of the EAIS would raise mean sea level by about 60 m, although such large-scale melting is highly unlikely to occur under the climate predictions that are deemed likely during the next century or so, which depend on assumptions of future global carbon dioxide emissions. Geological evidence suggests that the EAIS has been stable for the past 30 million years and remains fairly stable today. However, it has undergone episodes of rapid disintegration and may have completely melted at least once in the past 600,000 years. The West Antarctic Ice Sheet (WAIS) sits on a series of islands and the floor of the Southern Ocean, with parts of the ice sheet more than 1.7 km below mean sea level. Complete melting of the WAIS would raise mean sea level by an estimated 5.8 m.

Researchers have concluded that the annual temperature of West Antarctica increased by about 0.20±0.18°C (0.35±0.32°F) per decade between 1957 and 2024: one of the fastest-warming regions on the globe. This warming has been accompanied by higher SSTs around Antarctica and the breakup of ice shelves along the Antarctic Peninsula coast. The majority of glaciers of the Antarctic Peninsula are retreating at an accelerating rate. In fact, the melting rate for the WAIS has tripled over the last 10 to 15 years. Scientists find that the section of the WAIS with accelerated melting appears to be in persistent decline, which will lead the glaciers in this area to melt into the ocean and cause global sea level rise. The question is whether that occurs sooner or later (that is, centuries or millennia).

Unlike the Antarctic ice sheets, which are polar (cold), the Greenland Ice Sheet is temperate. While the Antarctic ice sheets are well below the pressure-melting point (temperature at which ice melts at a given pressure), the temperature of much of the ice on Greenland is near the pressure-melting point. The melting point of ice decreases with increasing confining pressure (or depth within the glacier). Temperate ice sheets more readily generate meltwater and, with less frictional resistance, tend to flow downhill faster than polar ice sheets. The Greenland Ice Sheet has exhibited accelerated melting, and indications are that this melting rate might accelerate even further. Complete melting of the Greenland Ice Sheet would raise mean sea level by an estimated 7.3 m. The consequences of sea level rise are one of the most important adverse effects of climate change on human civilization.

Another major concern associated with the current global warming trend is shrinkage of Arctic sea-ice cover. Arctic sea-ice cover is shrinking at an accelerated rate so that the Arctic Ocean may be free of summer ice within a few years or decades. Although melting of floating sea ice does not raise sea level, it can alter climate significantly. Less sea-ice cover on the Arctic Ocean is causing the humidity of the overlying air to increase, leading to more cloudiness. Clouds cause both cooling (by reflecting sunlight to space) and warming (by absorbing outgoing infrared radiation and emitting infrared radiation in all directions, including some directed towards Earth's surface). During the long, dark polar winter, additional cloud cover is expected to have a net warming effect. In summer, the impact of greater cloud cover depends on the altitude of the clouds. Cooling would be expected with an increase in low-altitude cloud cover, whereas warming would likely accompany an increase in high cloud cover. The mean annual air temperatures in the Arctic region have increased at a rate of 0.67±0.08°C (1.2±0.15°F) per decade, more than triple the global temperature increase rate.

In addition to shrinking sea ice, Northern Hemisphere snow cover has decreased, mountain glaciers are shrinking, permafrost is beginning to thaw, and fresh water runoff into the ocean has increased. Input of more fresh water from rivers and melting glaciers could impact the ocean thermohaline circulation by reducing the salinity and thus density of surface water in the area where dense, cold, salty water sinks at high latitudes of the Atlantic Ocean to become a critically important part of the MOC. Reductions in the strength of MOC could impact Northern Hemisphere weather patterns and would increase ocean deoxygenation.

Global Mean Sea Level

During the last glacial maximum, about 20,000 to 19,000 years ago, mean sea level was estimated to be 113 to 135 m lower than it is today. Sea level is currently close to the highest level it has been during the past 100,000 years, but it has been substantially higher, as much as several hundred meters higher, during warmer periods prior to that. Sea level falls during ice ages when global temperatures cool, and much of the world's water is stored in glacial ice on land. In contrast, sea level rises when the climate and ocean water warms and glacial ice melts.

The mean rate of globally-averaged sea-level rise between 1901 and 2010 was about 1.7 mm• yr^{-1}, which increased to 3 mm•yr^{-1} between 1993 and 2020. Considering the inertia in the Earth's climate system, it is a virtual certainty that global mean sea-level rise will continue beyond 2100 and that sea-level rise due to thermal expansion will continue for many centuries. Due to the rapid melt now occurring in Greenland and Antarctica and the instability of parts of the Antarctic ice sheet, sea-level rise by the end of this century is expected to range from 0.3 to 1 m, depending on future anthropogenic release rates of greenhouse gases.

The impacts of sea level rise are extensive. Higher mean sea level is expected to accelerate coastal erosion by wave action, allow ocean water to inundate wetlands, estuaries, and islands, and make low-lying coastal plains more vulnerable to storm surges. Rising sea level is also expected to disrupt coastal ecosystems, ruin agricultural lands, threaten historical, cultural, and recreational resources, and displace coastal populations. Ports, airports, and other coastal infrastructure are likely also to be disrupted by rising sea levels, causing large economic losses and extensive costs to either protect them from flooding or relocate them. In some coastal areas, higher sea level is also likely to exacerbate saltwater intrusion into groundwater.

According to a report by the U.S. Office of Science and Technology, a 0.5 m rise in sea level, close to the lower end of the range expected before the end of the century, would result in a substantial loss of coastal land, especially along the U.S. Gulf and southern Atlantic coasts. Particularly vulnerable is South Florida, where the elevation of one-third of the Everglades is less than 0.3 m above sea level. For people currently living on low-lying islands (such as the Maldives and Tuvalu), abandonment appears to be their only option as sea level rises. Globally, 147 to 216 million people live on land that will be below sea level or regular flood levels by the end of this century.

Marine methane hydrates (**Chapter 2**) are not stable at surface temperatures and pressures, but form in pore spaces within ocean floor sediments at depths greater than 400 m, where temperatures are sufficiently low and/or pressures are sufficiently high. When methane hydrates are brought to the surface, the reduction in pressure and rise in temperature cause the hydrates to become unstable and decompose, releasing methane. Warming of ocean water may release methane from methane hydrates and, if large quantities of methane are released and not absorbed by marine organisms before reaching the atmosphere, may contribute to global climate change. Atmospheric methane is a greenhouse gas, more potent than carbon dioxide, although with a shorter lifetime in the atmosphere. If the released methane is consumed by marine bacteria that use it as a food source, this will consume dissolved oxygen and contribute to deoxygenation. Methane concentrations in the atmosphere have been monitored since about 1985 and have indeed been rising, with a faster rate since 2007 than in previous decades.

Marine Ecosystems

Marine organisms are vulnerable to a changing climate. Materials and energy flow from one organism to another via food webs within and between ecosystems. Climate change alters the physical and chemical conditions in the ocean, possibly exceeding the tolerance limits of many organisms. If these organisms are unable to avoid or adapt to these changed conditions, they may individually perish or, as a species, become extinct.

One important example of the vulnerability of marine organisms is the phenomenon of coral bleaching, which is not limited to corals. In addition to hard corals (**Figs. 13-12**) and sea fans (**Fig. 14-8e**), many mollusks such as giant clams (**Fig. 14-35d**), some flatworms, many anemones, and some protists all host symbiotic zooxanthellae. In response to elevated sea surface water temperatures, zooxanthellae become toxic, and their coral polyps expel them. The specific temperature susceptibility varies with the particular strain of the zooxanthellae hosted by each species. Without the colorful zooxanthellae, the corals appear nearly transparent against the white exoskeleton, so the condition is known as coral bleaching. The same "bleaching" process also occurs in anemones, clams, and other zooxanthellae hosting species.

For reef-building corals, a rise in SST of 1 to 2°C is sufficient to stress the corals, causing temporary bleaching. While corals can recover from bleaching, persistent or severe episodes cause coral polyps and the reef to die. Furthermore, higher ocean temperatures and bleaching make corals more vulnerable to infection and likely acidification, adding to reef mortality. In many tropical coral reef ecosystems, sea surface temperature has been high enough to cause coral bleaching during some El Niño events, but temperatures have been low enough between events to allow the corals to recover. However, as sea surface temperatures have continued to rise, temperatures between El Niño events are becoming progressively higher, and bleaching has occurred over greater areas every year. As a result, coral bleaching events are becoming more frequent and are now occurring annually. Because the corals no longer have sufficient time to recover between bleaching events, these recurring events often result in the collapse and death of much of the reef. Many reefs have already been observed to collapse and die from coral bleaching.

A key consideration in the potential impacts of climate change on marine organisms is the rate of change. Marine ecosystems can more readily adjust to gradual rather than abrupt changes in the ocean environment brought on by global-scale climate fluctuations. The current rate of climate change due to the enhanced greenhouse effect is approximately 10 times faster than at any time during the past 10,000 years.

Acidification

Recall from **Chapter 1**, that as a result of anthropogenically released carbon dioxide, the ocean is becoming more acidic. Evidence shows that at times the ancient ocean was more acidic than today, but modern acidification is happening faster than in the past. Currently, the ocean is still weakly alkaline, but the average pH value has dropped from 8.2 to 8.1 during the past 200 years. This corresponds to a 30% increase in acidity. By 2100, the ocean is expected to be 100 to 150% more acidic than today. This rate of acidification is expected to be too fast to allow many marine species to adapt or evolve. As a result, marine organisms and ecosystems may be adversely affected and many marine species may become extinct.

A more acidic ocean makes calcium carbonate, the material of many hard parts (shells and skeletons) of marine organisms, more soluble and therefore more difficult for marine organisms to grow and maintain. The adverse effects of ocean acidification on marine ecosystems are already significant. In those areas of the ocean where acidity levels are highest, bivalves, marine snails, and other mollusks show decreases in the growth rate of their shells, especially in the early larval stages. Echinoderms, most notably sea urchins, have shown a decrease in fertilization success, developmental rates, and larval size. For vertebrates like fish, the direct impacts are less understood, but recent studies indicate that fish, like invertebrates, are most susceptible at the egg and larval life stages. Research has documented changes in fish behavior and issues with the development of their otoliths, a calcareous structure of the inner ear linked to balance, orientation, and sound detection. These effects, especially when taken together, can disturb food webs and the entire marine ecosystem.

The observed adverse effects of acidification on oysters and pteropods in both the Bering Sea and the California Current off the west coast of the U.S. and Canada are described in **Chapter 1**.

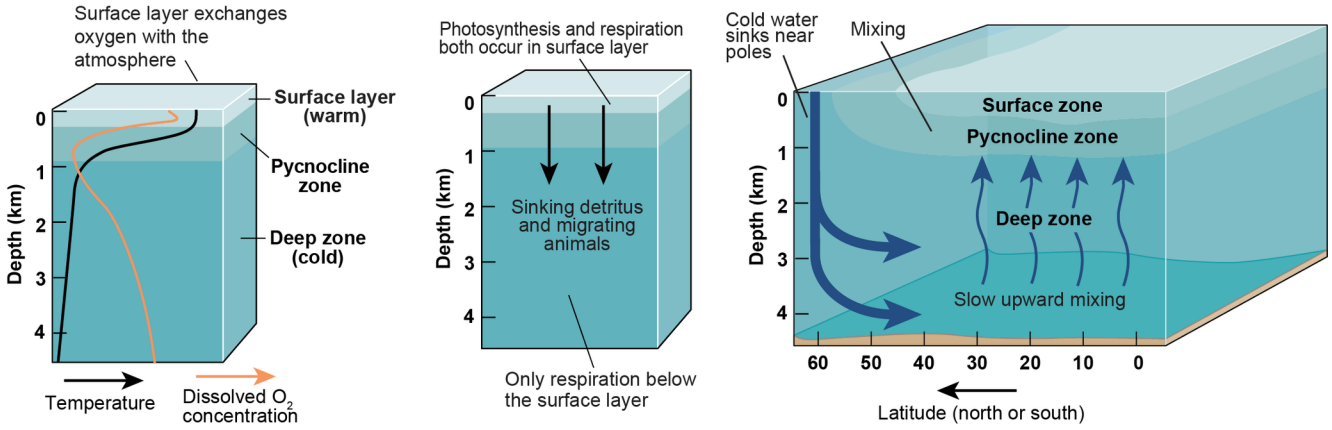
Deoxygenation

Ocean deoxygenation is the loss of dissolved oxygen from the ocean. Multiple sources of observations show that during the past century, oxygen concentration in the ocean has declined. According to research published in 2017, scientists detected a decline of

more than 2% in global ocean oxygen content between 1960 and 2010, while some areas experienced a larger drop. For example, the largest volume of oxygen loss occurred in the North Pacific, but the largest percent of decline occurred in the Arctic. A 2% drop may not seem significant, but, similar to the global mean annual temperature and ocean water acidity, small changes have large implications. Decreasing oxygen levels in the ocean pose a huge risk to marine ecosystems and, consequently, to humans.

The solubility of gases in water decreases as temperature increases, so warmer water holds less dissolved gas, including oxygen. Approximately 15% of the observed decline in global ocean oxygen content is attributed to warmer ocean temperatures. In addition, as surface ocean water warms, it becomes less dense and less likely to sink and mix with colder, less dense deep water. The overall result is a reduction in physical mixing, which translates to a longer residence time and a decrease in oxygen content of the deep ocean.

Recall from **Chapter 1** (**Figure 1-3**) that the photic zone is the top layer of the ocean where sunlight penetrates. Here, oxygen is produced much faster by photosynthesis than it is consumed by animal respiration and the concentration of dissolved oxygen in the upper few meters of ocean water is saturated for that temperature. Photosynthesis declines rapidly with depth, but respiration continues throughout the ocean, so below the photic zone, the concentration of oxygen decreases continuously as the ocean water flows through the subsurface oceans. Just below the pycnocline is where the water mass has been out of contact with the atmosphere for the longest period of time and where the rate of respiration is high due to the rapid consumption of detritus falling from the photic zone. At this depth, there is an oxygen concentration minimum coinciding with the nutrient maximum.



(a) Layered structure of the oceans. Pycnocline zone is a transition zone where water temperature decreases and density increases rapidly with depth and vertical mixing is slow.

(b) Oxygen is used up by respiration in the pycnocline and deep zones. Rate of respiration is higher just below the pycnocline zone than at deeper depths.

(c) Ocean circulation carries oxygen rich surface water into deep layers. This oxygen rich water is slowly mixed back toward the surface layer while its oxygen concentration is progressively decreased by respiration.

Figure 1-3. Oxygen concentration in ocean water is controlled by both physical and biological processes. The oceans consist of two main layers, a warm surface layer and a cold deep layer, with a transition zone, called the pycnocline zone, between them. (a) Surface layer water is well mixed and in contact with the atmosphere, and its oxygen concentration is primarily controlled by gas exchange with the atmosphere. (b) Photosynthesis by marine life in this layer converts carbon dioxide to organic matter and releases oxygen, but this is offset by respiration, which converts oxygen to carbon dioxide. Respiration also occurs in the deep layer, fueled by animals that migrate vertically or live in the deep and decomposition of organic material (detritus), but there is no photosynthesis and no contact with the atmosphere, so respiration continuously consumes oxygen in this deep layer, and, therefore, the longer the water remains in the deep layers, the less oxygen it contains. (c) The deep layer water is continuously replenished by sinking cold water in some near-polar regions and then circulates through the oceans while being slowly mixed upward, eventually into the surface layer again. The pycnocline provides a barrier that inhibits mixing, so the upward mixing process is slow.

Currently, large areas of the tropical and subtropical oceans exhibit a strong oxygen minimum zone below the pycnocline (**Figure 16-11**). In these areas, the oxygen minimum zone water has oxygen concentrations below 70 $\mu mol \bullet kg^{-1}$, below which marine species (other than those microbial species that are anaerobic) are subject to oxygen deficit stress, so the water is considered to be hypoxic. In smaller areas, where the concentration is essentially zero (anoxic), aerobic marine species can not survive. The volume of both hypoxic and anoxic water has been observed to have expanded during the past five decades. The hypoxic zone has expanded by an area that is more than half the area of the continental U.S., while the area of the anoxic zone has quadrupled. If this trend continues, it will become increasingly more likely that the low oxygen zone will mix upward to affect the upper water layers where there are higher concentrations of marine animal habitats. Recall that for the past few years, hypoxic (concentration of dissolved oxygen low enough to be detrimental to organisms), and at times anoxic (totally depleted of dissolved oxygen), water from this water mass has seasonally moved or been mixed onto the Oregon continental shelf, forming a dead zone (**Chapter 13**).

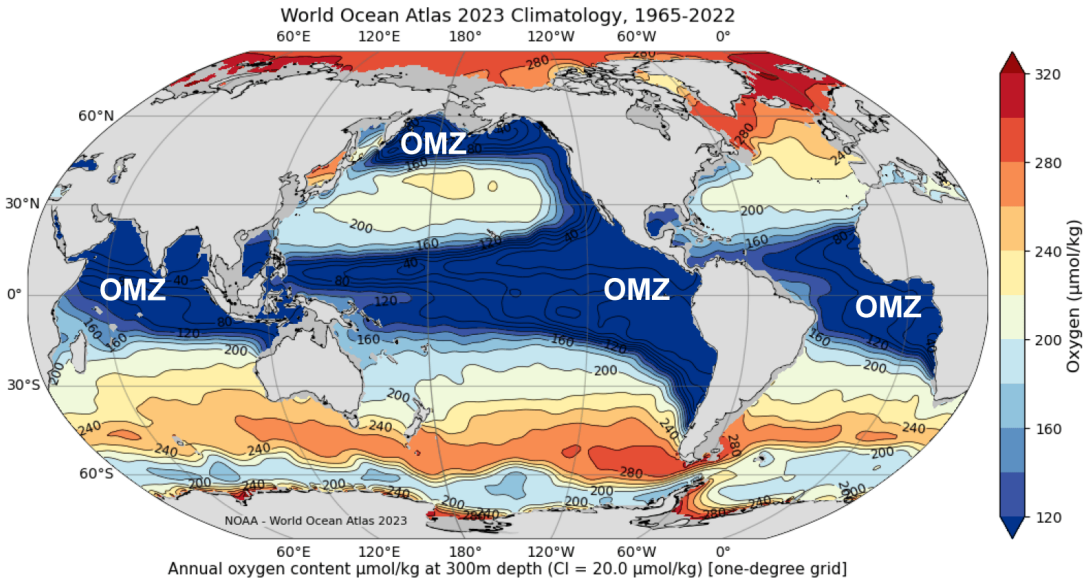


Figure 16-11. The concentration of dissolved oxygen at 300 m depth. This depth is generally below the pycnocline and at approximately the depth where an oxygen minimum occurs. Concentrations in the figure are expressed as µmol•kg^{-1}. The water is considered to be hypoxic at concentrations below 70 µmol•kg^{-1}.

While the cause of the expansion of oxygen minimum zones is not yet fully known, it has been estimated that about 15% of the total oxygen loss in the oceans is due to the warming-induced reduction in the solubility of oxygen in ocean water. However, warming is responsible for about 50% of the loss in the upper 100 m of the oceans. The remaining 85% of total oxygen loss in the ocean deep waters is likely due to increased stratification, which slows the sinking of oxygen-rich surface water, combined with faster transport of carbon-rich particles to the deep water as primary production increases as a result of nutrient enrichment and warming of the ocean surface layers. It is expected that oxygen in the deeper layers will decline further during the next several hundred years as the deep water is replaced by the meridional overturning circulation, and that this oxygen loss will occur even if there is no continued increase in stratification or nutrient enhancement of primary production. This means that the areas of oxygen minimum zones, in which the oxygen concentration is below the concentration needed to support aerobic marine species, will continue to grow for centuries, even if humans were to reduce anthropogenic carbon dioxide and nutrient emissions to zero or near zero immediately.

Recent data and modeling suggest that the post-industrial loss of oxygen concentration now observed in deep ocean waters represents only about 25% of the loss that will occur over the next several hundred years, even if anthropogenic emissions were to be completely stopped immediately. Unless anthropogenic emissions are reduced sufficiently and quickly enough to stop additional warming and further nutrient enrichment of the oceans, the rate at which the oxygen minimum zones will expand will be higher, likely substantially higher, than the now unavoidable rate of increase that will result from the existing warming and solubility loss in surface waters. We are already committed to much expanded deep ocean hypoxic and anoxic zones in the deep oceans with consequent but largely unknown effects on ocean biology and long-term climate. These already committed changes will not be reversed for as much as a thousand years or more even if we immediately stop all anthropogenic greenhouse gas emissions and nutrient releases.

Hypoxia and anoxia have also grown in coastal and estuarine regions during the past 50 years. During this time period, over 500 locations in coastal waters worldwide have been known to experience hypoxia. Fewer than 10% of these ocean areas are known to have experienced any hypoxia before the mid-20th century.

Hypoxic and anoxic water is currently found in only relatively limited areas of the ocean, mostly in the oxygen minimum zones below the pycnocline in the open ocean, and in an increasing number of coastal regions and estuaries where the hypoxia or anoxia is most often caused by inputs of nutrients from sewage and agricultural runoff. However, geologic evidence shows that periods of widespread anoxia within the ocean have occurred throughout Earth's history. Uncertainties remain about the causes of these anoxic events, but they were usually associated with warmer climate conditions, rises in sea level, and occasionally with mass extinctions. For example, global ocean anoxia has been proposed as the main cause for the mass extinction event at the close of the Permian (245 ma).

Climate Change Summary

In summary, while climate change will cause sea-level rise and changes in nutrient supply and ocean circulation that may adversely affect some ocean ecosystems, the more serious global threat to marine life from climate change comes from the combined effects of rising ocean temperature, increased acidity, and decreased oxygen concentration. Given the anticipated range of changes to each of these parameters, many marine species will be stressed or killed, and at least some species will become extinct within the next several decades. Indeed, climate-driven changes in each of these parameters have already been observed to cause harm to marine ecosystems in some parts of the ocean. For example, as discussed above, warming ocean water has caused a rise in the frequency of coral bleaching and damaged coral reef ecosystems. Increased acidity has caused mortality of oyster larvae and reduced reproductive success of pteropods in some parts of the northern ocean. Decreased oxygen concentration has caused episodic dead zones in many parts of the coastal and estuarine ocean and in the deep layers of some ocean areas.

As stated above, deoxygenation has been implicated as a cause of at least one of five mass extinctions in Earth's past and probably at least contributed to the other four mass extinctions. Warming and acidification are thought to have also played a part in these five mass extinctions. Elevated carbon dioxide concentrations in the atmosphere have also been associated with most extinctions. Some scientists have suggested that increased carbon dioxide in the atmosphere drives warming, acidification of the ocean, and deoxygenation of the ocean, and it is the combined effects of these stresses that were responsible for most of Earth's mass extinctions in the past. Moreover, the rate of species extinction has increased in the past few centuries, leading some scientists to hypothesize that Earth may already be experiencing a sixth mass extinction of species.

Much more detailed reliable scientific information on the likely future impacts of climate change, acidification and deoxygenation can be obtained from reports of the Intergovernmental Panel on Climate Change (IPCC).

This page titled 16.6: Climate Change, Acidification, and Deoxygenation is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.7: Risk Assessment and Management

None of the issues discussed in this chapter can be totally resolved or avoided without causing environmental and economic impacts. Measures that we might take to reduce the risk of impacts almost always come with unavoidable risks of harm to some other environment or other side effects, and no environmental issue can be solved without economic and social costs. In a world where human civilization exists, it is scientifically impossible to avoid all adverse impacts on all ecosystems. Thus, good stewardship of the Earth requires that we seek ways to minimize the root causes of degradation of global natural ecosystems rather than try to eliminate impacts or risk from any one particular cause among the many discussed in this chapter. The scientific way to minimize the total impacts of humans on the environment is through risk assessment and management, focused on prioritizing the reduction of the highest risk impacts.

Risk-based Earth stewardship is not simple nor easy, since risk itself is composed of several elements, and each element is very difficult to adequately assess and assign a value to. The two primary elements of risk are, first, the probability that an undesirable outcome will occur without corrective action, and, second, the severity of the outcome if corrective action is not taken before the outcome occurs. Further complications exist because the public perception of risk cannot be ignored and is often not based on either of the factors mentioned in the previous sentence. For example, the risk of a significant oil spill from an offshore oil rig is extremely small, many orders of magnitude less than the chance that numerous humans will be killed in auto accidents in a single day. Also, the effects of even the largest spill from any offshore oil rig, while dramatic and certainly undesirable, are limited to the region local to the spill, and are temporary (adverse effects disappear within a few years or so, a blink of the eye in geological time). Nevertheless, the public perception is that this risk should be avoided at any cost.

Although quantitative risk assessment is difficult, qualitative assessment of risks using the two primary risk elements can be useful. For example, all of the risks discussed in this chapter are considered highly likely to occur. However, with the exception of climate change/acidification/deoxygenation, all the issues discussed in the chapter would impose adverse impacts only in limited areas of the oceans and for geologically short time periods. Climate change, acidification, and deoxygenation are also likely to occur, but we have less certainty that their worst-feared effects will occur. However, the potential global consequences of climate change/acidification/deoxygenation set this issue aside from all others discussed in this chapter, since the consequences of not taking action could result in global mass extinctions, and severe damage to human food supply and the infrastructure that supports human civilization. Human civilization, and even the survival of the human species, are also within the scope of possible risk consequences. These adverse effects might take centuries to fully develop, but they would be geologically long-lasting events.

A reasonable assessment of ocean pollution priorities, based on what we know about the effects and risks associated with each area of concern discussed in this chapter, is that climate change/acidification/deoxygenation is by far the dominant ocean pollution issue and that carbon dioxide and nutrients, principally nitrogen, are the most harmful ocean contaminants.

This page titled 16.7: Risk Assessment and Management is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.8: CHAPTER SUMMARY

Assimilative Capacity.

The oceans can safely assimilate wastes if the rate at which the waste is introduced does not exceed the assimilative capacity, which is the point at which toxic or other adverse effects occur. Assimilative capacity varies by waste and location because residence time determines the concentration of contaminants, and each contaminant has a different concentration–effect relationship. Release of a specific type and amount of contaminant that causes pollution in one location may not do so elsewhere.

Adverse Effects of Human Activities.

Increased turbidity can reduce primary productivity. Excessive nutrient inputs can alter phytoplankton species composition and can cause blooms, eutrophication, and hypoxic or anoxic bottom waters.

Habitat can be altered by development and by discharges of wastes. Many areas of mangrove swamps and coastal wetlands have been filled and developed. Damming of rivers has prevented or inhibited anadromous and catadromous fish migrations and has altered salinity regimes and circulation in estuaries. Dredging, anchoring, fishing, and other activities disturb and alter seafloor habitats.

Community structures within ocean ecosystems can be altered when human activities affect habitat or affect some species but not others. Selective harvesting of fish, invertebrates, and marine mammal species has caused major changes in communities in some areas.

Toxic substances released to the oceans can be concentrated in seafood and present a human health risk or result in economic loss if the resource is unfit to eat. Contamination of seafood, particularly shellfish, has caused many shellfisheries to be closed and sometimes has caused human disease outbreaks.

Coasts have been marred and altered in many locations by structures. Many beaches are periodically closed because of contamination with improperly treated sewage or medical wastes.

Toxicity.

With the possible exception of carcinogens, teratogens, and mutagens, toxic substances are toxic only above a critical concentration that varies by substance, species, and even life stage within a species. The toxicity of a waste can be evaluated by comparing contaminated sites with control sites or by performing laboratory bioassays. Both approaches are difficult, and their results must be carefully interpreted. Many toxic substances are bioaccumulated by most species. Certain toxic substances are biomagnified, so their concentration increases at each trophic level.

Petroleum.

Natural inputs from seeps are the largest source of most oil released to the oceans. Stormwater runoff and other discharges from the everyday uses of petroleum are also a very large source. Oil tanker accidents are a far smaller source, and releases from offshore platforms are even smaller.

When oil is spilled, a slick forms and the volatile compounds evaporate to the atmosphere or dissolve, leaving lumps of tar or, if seas are rough, a gummy suspension called “chocolate mousse.” If the slick reaches shore, oil coats any substrate but is eventually decomposed by bacteria.

Oil spill events damage the ecology of shores reached by the oil and kill seabirds and marine mammals in the spill area. Some damaged areas recover naturally within a few years. Wetlands and other low-energy coastal habitats recover more slowly than high-energy rocky shores or beaches. Some spill cleanup efforts may have little or no benefit and may even retard natural recovery. Long-term effects may persist in some spill areas, but they are difficult to detect and document.

Sewage.

Sewage is discharged to the marine environment to protect public health. Sewage is a natural material, composed mostly of water and particulate and dissolved organic matter that is contaminated with chemicals from household wastes and industry. Sewage also contains pathogens that can be classified as pollution because the pathogens can infect humans through water contact or seafood consumption. Organic matter and nutrients, especially nitrogen, can cause algal blooms and anoxia, and toxic compounds can accumulate in discharge-area ecosystems. Sewage treatment removes floatables, particulates, and some dissolved organic matter, pathogens, and some potentially toxic chemicals. Treatment has substantially reduced the incidence of anoxia in rivers and estuaries.

Particulate organic matter from sewage outfalls can accumulate in sediments around the outfall and cause alteration of benthic infaunal communities. If outfalls discharge large amounts of sewage, the benthic infauna is degraded in a zone around the outfall, but the effects decrease with distance. If the same amount of sewage were discharged from several small outfalls instead of a single large one, no areas would be degraded, and the benthic infauna around each outfall would be enhanced.

Urban and Agricultural Runoff.

Runoff from streets and agricultural land contains toxic chemicals, nutrients, particles from paints, solvents, oil, and other substances deposited by vehicles, combustion products, pesticides, fertilizers, and many other sources. In estuaries, these nonpoint source inputs are often much larger than sewage or industrial inputs.

Industrial Effluents.

Industrial discharges of toxic substances and other contaminants have been considerably reduced in recent decades. However, violations of regulations and illegal discharges to sewers and streams are common. The industrial effluents of a chemical company in Minamata, Japan, in the 1950s caused the only documented case of human deaths due to toxic substances (methylmercury) discharged to the marine environment. DDT discharges have caused ecological damage in a number of coastal ocean areas.

Dredged Material.

Navigation channels that are dredged to maintain navigation depths are often sites of accumulation of past and present contamination. Dredging damages the benthos at the dredging site and at the dump site, which is usually nearby in the estuary or coastal ocean. Toxic contaminants in dredged material are preferentially released to the suspended sediment during dumping and may be bioavailable. Most of the fine-grained, contaminant-rich dredged material dumped in estuaries or just outside estuary mouths is transported back into the estuary by the estuarine circulation and may accumulate in channels to be dredged again.

Plastics and Trash.

Trash, particularly floating trash, is dumped into rivers and the ocean and can cause aesthetic problems when washed up on beaches or accumulated in areas visited by divers. Plastics are a problem because they degrade very slowly, and they may strangle or choke birds and marine organisms or accumulate in the gut until the animals starve to death.

Antifouling Paints.

To be effective, antifouling paints must be toxic and must be slowly released to solution. Tributyltin, used in some paints, is very persistent and highly toxic. It has accumulated in some estuaries to toxic levels, drastically reducing populations of commercially valuable shellfish and other invertebrates.

Radionuclides.

The oceans have been contaminated by radionuclides from testing of nuclear weapons, liquid waste discharges, and dumping of radioactive wastes. Other than at a few nuclear bomb test sites, radionuclide concentrations in the oceans do not present

a significant human health or ecological risk. However, the former Soviet Union disposed of nuclear submarine reactors and large quantities of radioactive wastes in the Arctic Ocean and the Sea of Japan, and there is concern that these materials may eventually release radionuclides to the food web.

Noise.

The oceans are filled with numerous sources of natural sound. However, additional noise is contributed by ships, resource extraction activities, and military and civilian use of sonar. Intense sounds at certain frequencies have been found to be harmful to marine mammals. As a result, there is considerable concern that the increased anthropogenic sound in the oceans may cause harm to marine species.

Nonindigenous Species.

Species are carried in ships, in ballast water, and attached to hulls and are inadvertently introduced to ecosystems where they do not occur naturally. Species are also deliberately introduced. Many introduced species do not thrive, but some outcompete important species in their new habitat and may totally disrupt natural food webs. Estuaries have been particularly severely damaged by nonindigenous species.

Habitat Alteration.

Alteration or destruction of habitat, especially filling of wetlands, is among the most damaging forms of pollution, especially in estuaries and the coastal zone. Habitat can also be altered in many other ways, such as by dredging, by the construction of structures that affect circulation and erosion, and by the reduction of river flow rates that results from the withdrawal of freshwater for human uses.

Fishing.

Many ocean fisheries, perhaps most, are overfished. Many fish stocks have collapsed due to overfishing, and some do not recover even if the fishing pressure is relieved. Fishing also contributes to ocean pollution in many other ways, especially through loss of fishing equipment, much of which is now made of synthetic materials that are highly resistant to decomposition, and bottom trawling that destroys benthic habitat.

Climate Change.

Global climate change has caused the oceans to warm. The warming has caused changes in a number of ocean processes that have resulted in adverse impacts on marine ecosystems and humans, some of which are likely to become greater and many of which will continue to drive change in the oceans for centuries to come. Impacts include changes in ocean circulation and evaporation/precipitation patterns; increases in wave and storm energy affecting coastlines; melting of sea ice and glaciers; rising sea level; and impacts on marine biological systems, including coral bleaching due to warmer water. The rate of change of ocean temperature, acidity, oxygen levels, and other ocean conditions due to climate change is faster than has occurred at any time in Earth's recent history, and these changes are expected to result in species extinctions and continuing disruption of marine species life cycles and distribution.

Acidification.

Anthropogenic releases of carbon dioxide cause carbon dioxide to be taken up by seawater, where a series of chemical reactions causes the acidity of ocean water to rise. Adverse effects of elevated acidity have already been observed in several areas of the world's oceans involving a variety of marine species. Most at risk are coral reefs and marine species that use calcium carbonate to construct their skeletal material, such as oysters and pteropods.

Deoxygenation.

Elevated ocean temperatures due to climate change have reduced the solubility of oxygen and, therefore, the concentration of dissolved oxygen in ocean waters. Lower oxygen concentrations in water sinking to form deep ocean waters, together with changes to stratification and primary production caused by anthropogenic releases of carbon dioxide and nutrients, especially nitrogen, have expanded the area of open ocean oxygen minimum zones in which hypoxia and anoxia are present. Upwelling has caused some of this water to be transported onto the continental shelf off Oregon, killing marine species and causing what is known as a dead zone. Similar processes have also led to the development of hypoxia and anoxia in more than 500 coastal areas that have resulted in dead zones and the deaths of marine organisms that are unable to avoid these zones.

Climate Change Summary.

The release of anthropogenic carbon dioxide, together with anthropogenic releases of nutrients, especially nitrogen, causes multiple impacts on the oceans. These impacts include sea level rise, loss of sea ice, coral bleaching, ocean acidification, and deoxygenation due to the warming of ocean water. These anthropogenic releases also cause changes in ocean stratification and primary production that also cause deoxygenation. The combination of elevated carbon dioxide in the atmosphere, warming and acidification of ocean water, and deoxygenation of the oceans has been associated with at least some of Earth's five mass extinctions.

Risk Assessment and Management.

Scientific risk assessment must consider both the probability that a feared impact might occur and the severity of the impacts if they do occur. The consequences of climate change/acidification/deoxygenation could include mass extinctions and other global and long-lasting environmental and economic impacts. These potential consequences make this the most important ocean pollution issue.

16.8.1: Key Terms

16.8.2: Chapter Questions

16.8.3: References

This page titled 16.8: Chapter Summary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.8.1: Key Terms

You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- anadromous
- anoxic
- anthropogenic
- assimilative capacity
- beaches
- benthic algae
- bioaccumulation
- bioassays
- bioavailable
- biochemical
- biogeochemical cycles
- biomagnification
- biomass
- bloom
- brackish
- bulkheads
- carcinogens
- catadromous
- coastal zone
- community
- complexes
- contaminant
- continental shelf
- currents
- DDT
- decomposers
- diffuses
- diversity
- DNA
- ecosystems
- effluent
- embryo
- environment
- eroded
- estuaries
- eutrophication
- friction
- genetic
- grain size
- grazing
- groundwater
- habitat
- hypoxia
- invertebrates
- isotopes
- jellies

- latitudes
- limiting nutrient
- macroalgae

- mangrove
- microalgae
- mollusks
- mutagens
- nonindigenous species
- nutrients
- outfalls
- overfishing
- oxygen-demanding
- pathogens
- PCBs
- pelagic
- photic-zone
- photosynthesis
- phytoplankton
- plate tectonics
- pollution
- primary production
- radioactivity
- radioisotopes
- reef
- residence time
- respiration
- resuspend
- salinity
- saturation solubility
- sea grass
- sediment
- sedimentation
- sedimentation rate
- shellfish
- shore
- shoreline
- spawning
- species
- suspended sediment
- teratogens
- topographic
- tracers
- trophic levels
- turbidity
- virus
- wetlands

This page titled 16.8.1: Key Terms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.8.2: Chapter Questions

Study Question

1. What is the difference between contamination and pollution?
2. Explain assimilative capacity. How are assimilative capacity and residence time linked?
3. How can human activities alter primary production?
4. Which human activities cause alterations in marine habitats? How many can you list? In what way do each of these activities alter habitat?
5. How can we evaluate the toxicity of a substance to marine species? Why is it so difficult to decide what concentration is safe in the environment?
6. What is the difference between bioaccumulation and biomagnification? Why should we be more concerned about toxic substances that are biomagnified?
7. In what ways can fishing damage the marine ecosystem?
8. List the principal sources of contamination in the oceans. Summarize the characteristics of each type of waste or other source
9. What are antifouling paints? What would be the desirable characteristics of a future antifouling paint?

Critical Thinking Questions

1. What is the difference between naturally occurring toxic substances and synthetic toxic substances that suggests that they might be considered differently by policy makers and managers concerned with marine pollution?
   (a) Do you think discharges of these two types of substances should be subject to different rules? Why or why not?
   (b) Do you think the production of all synthetic toxic substances (without any exception) should be banned?
   (c) If only some such substances are to be banned from being produced while others are allowed to be produced but their discharges regulated, what factors would you consider in deciding which of these two policies should be applied to a specific synthetic toxic substance?
2. In this chapter, it is speculated that the use of oceans for sewage waste disposal was one of the most effective advancements in human health protection in history. Do you think this statement is correct? Why or why not?
3. DDT has been banned from production and use in many countries, primarily in mid-latitudes, where it was previously used on crops for insect control. However, DDT is still produced and used in large amounts in some nations, particularly developing nations in tropical regions where malaria is rampant. Although effective substitutes for DDT exist, they are not used in these countries, because they cost too much. What should be done about this situation, and how?
4. On an annual average, petroleum hydrocarbons from street runoff contribute more than twice the volume of oil spilled in the oceans from tanker accidents.
   (a) What are the sources of petroleum hydrocarbons in street runoff?
   (b) What would you suggest politicians do to reduce the amount of petroleum hydrocarbons released to bays and estuaries by this route?
   (c) Should politicians require all storm water runoff to be treated to remove the hydrocarbons?
   (d) Should politicians attempt to control the release of hydrocarbons to the environment so that they do not get into the storm water? How could this be done?
   (e) What can you personally do to reduce the amount of hydrocarbons that you release to the environment?
5. Describe the principal sources of petroleum contamination in the oceans. Discuss what would probably cause more pollution of the oceans: production of oil from drilling platforms in U.S. coastal waters and transport of the oil ashore in seafloor pipelines, or purchase of oil from foreign suppliers and transport to the United States in oil tankers.
6. Because of environmental concerns, offshore oil exploration is currently banned in large areas of the Pacific and Atlantic continental shelves of the United States, despite the successful safety record of the oil industry operating under U.S. laws and regulations. Some oil companies are drilling instead in other areas, such as the ice-filled waters near the Russian island of Sakhalin, where the safety and environmental rules are weaker and sometimes totally unenforced, and from where the oil must be transported long distances to the locations where it is used.
   (a) Discuss the implications that drilling operations like the one conducted off Sakhalin Island might have for the ocean environment as a whole.
   (b) Should the drilling bans in U.S. waters be continued? Describe all the reasons for your answer.
   (c) If the ultimate political decision were to increase the drilling and oil activity in U.S. waters, what studies should be done

or actions taken before such drilling took place?
(d) Have such studies been done already?

7. Describe the principal adverse effects of sewage discharged to the oceans and how they can be reduced or eliminated. What actions have been taken to reduce or eliminate sewage pollution of the oceans in the United States? Have these actions been effective?
8. Billions of dollars have been spent on the construction of secondary sewage treatment plants in all cities and towns regardless of the locations of their discharges.
(a) Was this uniform approach justified scientifically? If so, why? If not, what other approaches to sewage treatment and disposal might have been taken?
(b) What are the characteristics of sewage that you would wish to modify by alternative treatment approaches before discharging it to the oceans?
(c) What are the factors that you would consider in deciding where you might not require secondary treatment or where you might allow only primary treatment?
(d) How would you determine the environmental benefits or detriments of alternative treatments?
9. Numerous sources release trace metals, toxic organic substances, nutrients, and pathogenic microorganisms to the environment. These releases are eventually discharged to the oceans in sewage, urban storm-drain runoff, and runoff from agricultural land.
(a) List as many of these sources as possible.
(b) What substances do the various sources contain
10. After you have answered Critical Thinking Question 9 above, take a look at the labels on all of the containers of cleaning fluids and powders, cosmetics, shampoos, paints, and other materials around your house.
(a) What additional sources can you now list?
(b) Where do these products and their chemical constituents go after you have used them?
(c) Make two lists: one of the products whose containers tell you enough about their composition for you to decide what contaminants they might release to the environment, and another of those that do not. Do you know what is in the products that do not tell you on their label what they contain?
(d) Why do some products have information on their labels that divulge their composition, while other products do not? Should this situation be changed? If so, how?
11. Discuss why it is more difficult to control contamination of the oceans from urban and agricultural inputs than from industrial inputs.

This page titled 16.8.2: Chapter Questions is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.8.3: References

Figure 16-1 – Segar, D., and E. Stamman Segar, 1997: People at the beach.

Figure 16-2 – Adapted from Nichols, F. H., J. E. Cloern, S. N. Luoma,and D. H. Peterson, 1986: The modification of an estuary. *Science*, **231**, 567–573, Source.

Figure 16-3 – Segar, D., and E. Stamman Segar, 1997: Sugar cane fields burned after harvest to clear the land for new planting in Hawaii.

Figure 16-4a – Segar, D., and E. Stamman Segar, 1997: The Moss Landing Power Plant in the middle of Monterey Bay, California.

Figure 16-4b – Siegmund, W., 2008: Anacortes Refinery (Tesoro Corporation). Accessed 11 July 2025, Source, CC BY-SA 3.0, CC BY 2.5.

Figure 16-4c – Segar, D., and E. Stamman Segar, 1997: A marina at Antioch, California, constructed on a former wetland in the San Francisco Bay delta.

Figure 16-5a – Segar, D., and E. Stamman Segar, 1997: The *Exxon Valdez* tanker 48 h after it ran aground on Bligh Reef in Prince William Sound, Alaska, in 1989.

Figure 16-5b – Miller, J., 2010: Saving wildlife in Louisiana. FEMA, accessed 11 July 2025, Source.

Figure 16-6 – US Coast Guard, 2010: *Deepwater Horizon* offshore drilling unit on fire. Accessed 11 July 2025, Source.

Figure 16-7a – Adapted from Smith, W. E., and A. M. Smith, 1975: *Minamata: The story of the poisoning of a city, and of the people who choose to carry the burden of courage*. Holt, Rinehart, and Winston, 192 pp.

Figure 16-7b,c – Adapted from Gerlach, S. A., 1981: *Marine pollution: Diagnosis and therapy*. Springer-Verlag, 218 pp.

Figure 16-8b – Segar, D., and E. Stamman Segar, 1997: The Alcatraz dredged-material dump site in San Francisco Bay.

Figure 16-9a – NFWS, 1985: Northern fur seal entangled in netting. NOAA.

Figure 16-9b – Wright, D., 2006: Albatross chick plastic. USFWS, accessed 11 July 2025, Source.

Figure 16-10b – NOAA, 2008: Marine debris on Hawaiian coast. Accessed 11 July 2025, Source.

Figure 16-11 – Garcia, H.E. and Coauthors, 2023: The World Ocean Atlas Volume 3: Dissolved Oxygen, Apparent Oxygen Utilization, and Oxygen Saturation. A. Mishonov, Tech. Ed. NOAA Atlas NESDIS 91, accessed 11 July 2025, Source.

This page titled 16.8.3: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.9: Box 1 - Petroleum in the Marine Environment

Almost every year, somewhere in the world, one or more oil tankers accidentally spill some or all of their cargo into the oceans. Tanker accidents (**Fig. 16B1-1a,b**) and oil well blowouts (**Figs. 16B1-1c,d,e, 16B1-2**) generate enormous public and media attention. These dramatic events provide gripping visual images of stranded and damaged ships; of thick black oil floating on the water and washing up on beaches; of oiled and dying or dead seabirds (**Fig. 16B1-1c**), otters, and other animals; and of frantic efforts to clean up the spilled oil (**Fig. 16B1-1d,e**). Consequently, tanker spills and oil platform blowouts are widely believed to be the major source of oil contamination and the most damaging form of ocean pollution. Neither of these beliefs is correct.

Figure 16B1-1. Oil spills provide spectacular images that ensure media coverage. (a) The *Exxon Valdez* tanker 48 h after it ran aground on Bligh Reef in Prince William Sound, Alaska, in 1989. The tanker is still leaking some oil, and it is surrounded by a boom that has been placed on the water's surface in an attempt to contain this oil so that it can be collected. (b) The *Amoco Cadiz*, Brittany, France, 1978. (c) A rescuer tries to capture an oiled Pelican after the 2010 *Deepwater Horizon* spill in the Gulf of Mexico. Rarely do heavily oiled birds survive even when cleaned up. (d) A boom skimmer towed by two small boats corrals surface oil so it can be burned. Controlled burns require only low-cost equipment and are among the most effective spill cleanup techniques. (e) Cleaning up Gulf of Mexico beaches after the *Deepwater Horizon* spill. Aggressive beach cleanup often causes more environmental damage than would occur if the beach were left to clear naturally.

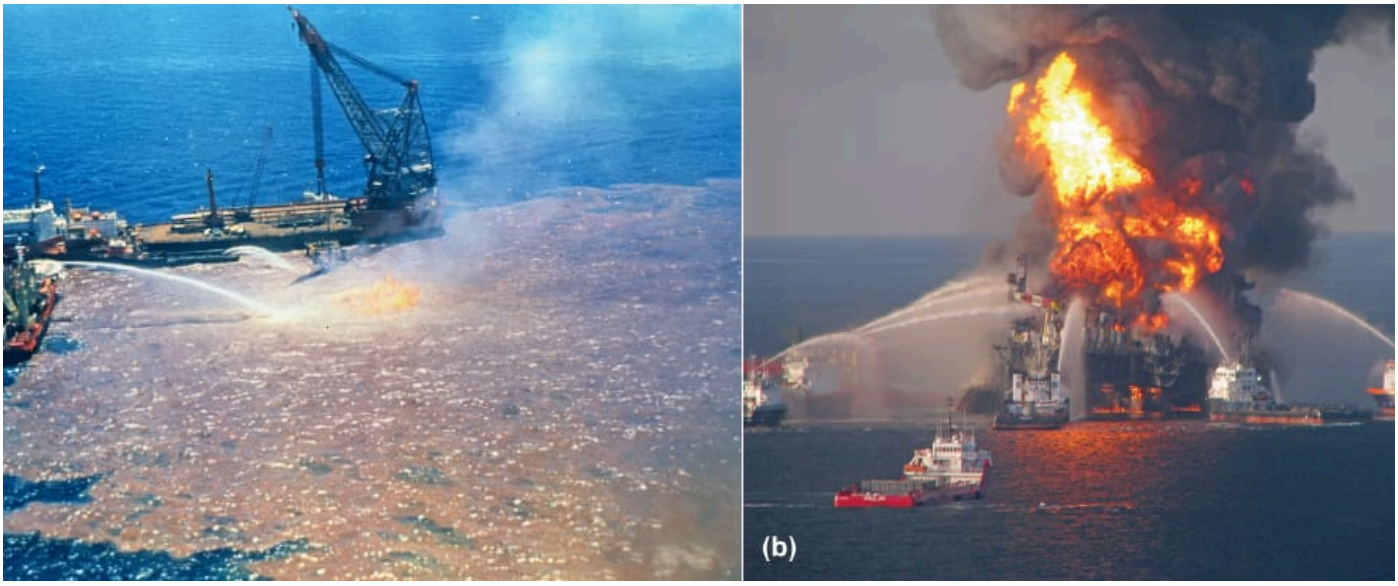


Figure 16B1-2. Major oil spills from the offshore oil industry are rare but dramatic events. (a) This spectacular blowout and fire at the Ixtoc I oil rig in the Gulf of Mexico in 1979 resulted in a continuous release of oil that lasted for several months before the well was finally closed (b) The *Deepwater Horizon* rig burned before it sank in the Gulf of Mexico in April of 2010.

Sources of Petroleum Contamination

Sources of oil in the oceans include natural seeps from the seafloor and from terrestrial watersheds that drain to the oceans, large and small spills, and incidental releases by vessels and offshore platforms. In a 2022 report, the National Academy of Sciences reported that natural seeps are the major source of petroleum to the oceans both nationally and worldwide. According to the report, oil from individual cars and boats, lawn mowers, jet skis, marine vessels, and airplanes contributes the most anthropogenic oil contamination to the ocean both in the U.S. and globally. This oil enters the oceans through land runoff from oil slicks on urban roads and through atmospheric deposition of hydrocarbons. The report estimates that use-related oil pollution accounts for about 87 percent of the oil from human activity in North American waters, an amount that dwarfs the inputs from oil and gas production activities. From 1990 to 1999, the report estimated that worldwide natural releases of petroleum were 15 times greater than releases from offshore oil platforms, and that oil transportation accidents (mostly spills from tankers) contributed about 4 times more oil to the oceans than oil platforms (**Fig. 16B1-3**).

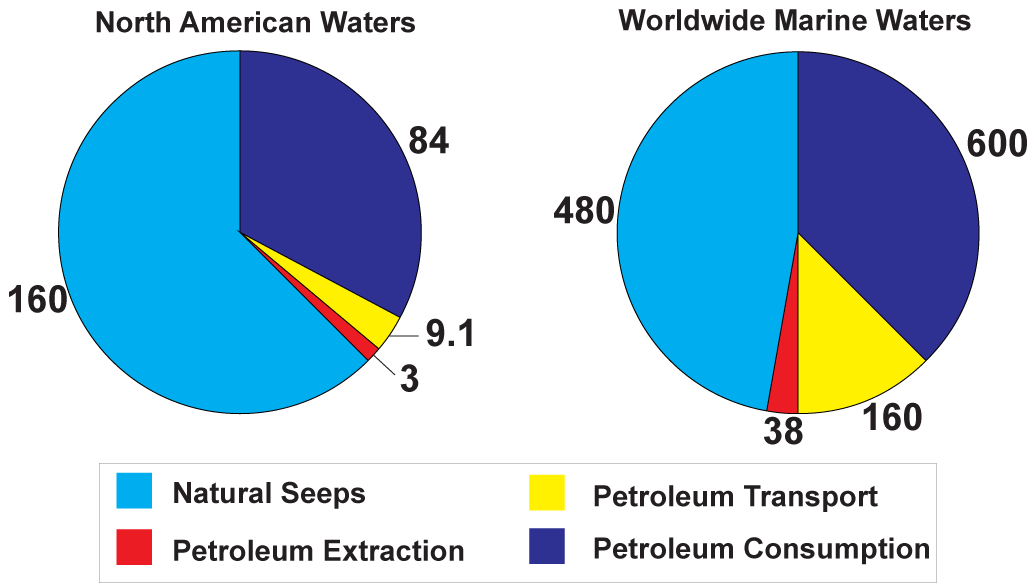


Figure 16B1-3. There are many sources of oil contamination of the marine environment. According to the National Academy of Sciences report, Oil in the Sea III, while incidental releases (in tonnes) during the consumption stage, for example, from automobiles, far exceeds the inputs from spills and other releases during the production and transportation of petroleum, especially in the U.S.

Since the *Exxon Valdez* oil spill in Alaska in 1989, extensive efforts to reduce accidents and incidental oil releases from vessels and offshore oil platforms have substantially reduced the input from these sources, particularly in the U.S. For example, no large spills from tankers have occurred since 2002 (**Table 16B1-1**). In contrast, there has been relatively little success in reducing the inputs of oil through urban and river runoff or in incidental discharges from recreational and other vessels. Major spills such as the 2010 *Deepwater Horizon* explosion in the Gulf of Mexico still occur, but they are now drastically less frequent than they were a decade or two ago.

Table 16B1-1. *Selected Oil Tanker Spills since the 1960s*

Tanker	Date	Location[a]	Oil Spilled (tonnes)[b]
Atlantic Empress	1979	Off Tobago	287,000
ATB Summer	1991	700 n.m. off Angola	260,000
Castillo de Bellver	1983	70 n.m. off Cape Town, South Africa	252,000
Amoco Cadiz	1978	Brittany, France	223,000
Haven	1991	Genoa, Italy	144,000
Odyssey	1988	700 n.m. off Nova Scotia, Canada	132,000
Torrey Canyon	1967	Scilly Isles, United Kingdom	119,000
Sea Star	1972	Gulf of Oman	115,000
Irenes Serenade	1980	Navarino Bay, Greece	100,000
Urquiola	1976	La Coruña, Spain	100,000
Hawaiian Patriot	1977	320 n.m. west of Hawaii	95,000
Independenta	1979	Istanbul, Turkey	95,000
Jakob Maersk	1975	Leixões, Portugal	88,000

Tanker	Date	Location[a]	Oil Spilled (tonnes)[b]
Braer	1993	Shetland Isles, United Kingdom	85,000
Aegean Sea	1992	La Coruña, Spain	74,000
Sea Empress	1996	Milford Haven, United Kingdom	72,000
Khark 5	1989	120 n.m. off Morocco, Atlantic	70,000
Nova	1985	75 n.m. off Kh¯ark Island, Persian Gulf	70,000
Katina P.	1992	Off Maputo, Mozambique	72,000
Prestige	2002	Off the Spanish coast	63,000
Sinclair Petrolore	1960	Off Brazil	60,000
Epic Colocontris	1975	60 n.m. northwest of Puerto Rico	60,000
Corinthos	1975	Marcus Hook, Philadelphia, USA	53,000
Assimi	1983	60 n.m. off Masqat, Oman	52,000
Metula	1974	Strait of Magellan, Chile	50,000
Andros Patria	1978	Off Cape Finisterre, Spain	50,000
World Glory	1968	90 n.m. off Durban, South Africa	48,000
Pericles GC	1983	200 n.m. off Doha, Qatar	46,000
British Ambassador	1975	Pacific Ocean	44,000

Tanker	Date	Location[a]	Oil Spilled (tonnes)[b]
Ennerdale	1970	Off Port Victoria, Seychelles	41,000
Mendoil II	1968	340 n.m. off Washington State, USA	40,000
Wafra	1971	Off Cape Agulhas, South Africa	40,000
Juan A. Lavalleja	1980	Arzew, Algeria	40,000
Trader	1972	Off southwestern coast of Greece	37,000
Exxon Valdez	1989	Prince William Sound, Alaska, USA	37,000
Thanassis A.	1994	200 n.m. off Manila, Philippines	37,000
Burmah Agate	1979	Galveston, Texas, USA	36,000

Note: For incidents in which some of the spilled oil burned, not all of the amount listed here was released into the ocean.

[a]n.m. = nautical miles

[b]All values are estimates. Reports of the amounts of oil spilled have varied for some spills

Fate of Petroleum in the Oceans

When oil is spilled into the ocean, it spreads on the water's surface to form a slick. The oil's fate then depends on several factors, including the oil composition, air and sea temperatures, concentrations of suspended sediment, presence of breaking waves, and whether the oil reaches a shore (**Fig. 16B1-4**).

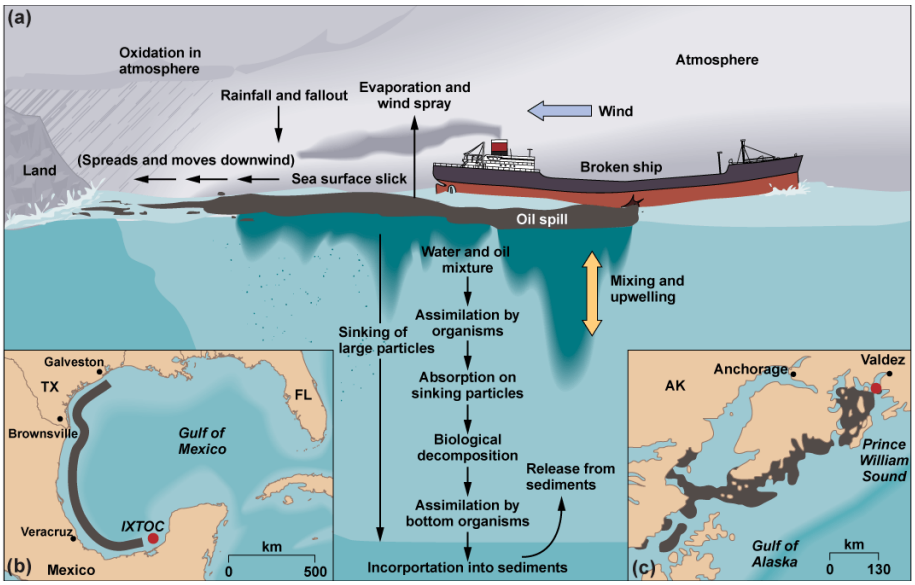


Figure 16B1-4. The fate of oil spilled in the ocean depends on many factors, including the type of oil, the type and proximity of the coastline, and the extent of wave energy. (a) Some oil evaporates quickly, but most coagulates to form tar balls or is adsorbed by particles and sinks to the sediments. The oil is eventually decomposed, primarily by bacteria. Oil spills generally do not cause lasting environmental damage, unless the oil reaches an ecologically sensitive shoreline, especially in high latitudes where oil degrades slowly because of the low temperature. (b) Areas affected by oil after the 1979 blowout of the Ixtoc I well, which took several months to cap. (c) Areas affected by oil after the 1989 *Exxon Valdez* accident.

Oil contains many different chemical compounds called hydrocarbons. Individual types of hydrocarbons differ widely in volatility, solubility, toxicity, and chemical properties. Crude oil composition depends on its source, and refined oil products are very different in composition from crude oil. Generally, refined oil products, such as gasoline and diesel fuel, contain a greater proportion of low-molecular-weight, more volatile hydrocarbons and are therefore more toxic than crude oil.

After a spill, the volatile components of the oil evaporate into the atmosphere or dissolve in the water. These volatile compounds are largely evaporated from the slick within a day or two. As volatile components are removed, the oil becomes more viscous and, unless the seas are rough, begins to aggregate into lumps. In rough seas, oil may be mixed with air and water to form a gummy suspension that resembles and is often called “chocolate mousse.” The oil dissolves or forms lumps, called “tar balls,” composed primarily of high-molecular-weight, less-volatile hydrocarbons, or the oil is attached to sediment particles and deposited. If the slick does not encounter a shore, it eventually completely dissipates. If the slick does reach a shore, oil clings to any substrate as an oily film. Once ashore, the oil film persists until it is washed off, buried in sediments by continuous strong wave action, or slowly decomposed by bacteria.

Because oil is a naturally occurring material, many decomposers can use hydrocarbons as food. Consequently, oil spills are eventually “cleaned up” naturally. The severe damage to birds and mammals and to intertidal epifauna and infauna that provides gripping television coverage after a spill is generally limited to relatively small stretches of shore. Even the most severely damaged shores normally recover and are almost indistinguishable from their original condition within a few years to a decade after the spill.

Recovery and recolonization are generally faster on rocky or other high-energy shores where physical processes limit the extent of oil accumulation during the spill and maximize its removal and dispersion into the open ocean in the post-spill period. Low-energy shores, particularly wetlands, recover more slowly because oil can accumulate more easily, and sediments into which it is mixed often have low concentrations of oxygen, an element that some bacteria need in order to decompose the hydrocarbons. Because both bacterial decomposition and evaporation are reduced at low temperatures, oil spills may persist longer, and recovery may be slower in high-latitude environments than in warmer regions.

Effects of Major Spills

One of the largest oil spills from a tanker accident occurred in Brittany, France, in 1978 (**Table 16B1-1**). The tanker *Amoco Cadiz* (**Fig. 16B1-1b**) spilled its entire cargo of more than 200,000 tonnes of oil over several days after it hit a rock 13 km offshore and broke up. This was more than six times the amount of oil spilled in the *Exxon Valdez* accident in Alaska in 1989. Strong currents rapidly spread the oil slick along the Brittany coast, and strong wave action prevented at-sea cleanup efforts from recapturing

much of the oil. Within a few days, 300 km of shore was affected by oil. Oil entered a number of low-wave-energy and low-current-energy estuaries and other embayments, where it accumulated in large quantities.

The biological impacts of the *Amoco Cadiz* spill were immediate, dramatic, and severe. More than 7000 seabirds, mostly diving birds such as cormorants, were oiled and died. The plankton biomass was substantially reduced for at least two months after the accident, and mortalities of benthic organisms, including sea urchins, clams, and amphipods, were massive. The major commercial species of the area, oysters, survived but were heavily contaminated and thus unfit for human consumption for many months. The rooted vegetation in coastal salt marshes and the fauna of intertidal mud-flats were severely damaged. In contrast, only a small number of fishes were reported killed within the immediate vicinity of the wreck (about 10 km). Commercial flatfishes, including plaice and sole, showed no significant changes in population, although their average size in the spill year was somewhat below average, presumably because of the reduction in biomass of the juveniles' plankton food.

In the weeks and months after the *Amoco Cadiz* accident, there were extensive cleanup and oil removal efforts, especially in the low-energy estuarine environments. Visible signs of oil persisted in the water for as long as six months and in sediments for more than three years. However, all but very limited areas, such as the mudflats, were repopulated with most or all of their original species within several years. Some mudflats were subject to increased erosion caused by the loss of their vegetation, and oil contamination still could be detected in the mudflat sediments more than a decade after the spill. Nevertheless, this massive spill, which occurred in a particularly vulnerable area with extensive and important tidal wetlands and estuaries, caused severe ecosystem disruption in only a limited area, and the system recovered almost totally within a decade. This experience has been repeated in many other spills, including the *Exxon Valdez* accident (**Fig. 16B1-1a**), the huge Ixtoc I platform spill (**Fig. 16B1-2b**), and the massive deliberate spill by Iraq in the Persian Gulf during the Gulf War in 1991.

Both the Ixtoc I spill and the Gulf War spill were estimated to be about the same size as the *Deepwater Horizon* spill in the Gulf of Mexico in 2010. Even just one year after the beginning of the *Deepwater Horizon* spill, the accumulating scientific data showed that the Gulf of Mexico was already well on its way to recovering from the spill and further studies suggest that any the long term damage will be minimal, except for degradation of the benthos in a limited area around the spill site, which was most likely caused primarily by dispersant chemicals used to minimize the amount of floating oil. For example, the data show that Gulf fisheries catch rates were higher in 2011 than they were before the spill. This has been attributed by some scientists to the fact that stocks were allowed to recover when fishing was banned in the oil-affected area of the Gulf for several months. An equally valid hypothesis has been proposed that the improved fishing is due to the large pulse of additional food supplied by the microbial biomass that decomposed the spilled oil in the open waters of the Gulf so effectively that oil-related compounds were back to normal (the Gulf has many ongoing natural seeps, so oil-related compounds are never zero) background concentrations within months. Nevertheless, public pressure has demanded and required intensive future studies to identify the long-term impacts of the *Deepwater Horizon* spill. As in all past spills that have been studied, it is likely that changes will be observed in the Gulf ecosystem in future years that could possibly be caused by the spill or could alternatively be due to natural variation. In the meantime, the public and the media will almost certainly continue to almost totally ignore the annual occurrence of a human pollution caused annual "dead zone" in the Gulf of Mexico (**Chapter 13**) which is almost certainly a far greater cause of adverse ecosystem impact on the Gulf than the 2010 oil spill, especially since the dead zone is an ongoing and growing pollution problem that remains largely unaddressed.

Cleanup activities after a spill may be essential to remove as much oil as possible and speed natural recovery. However, if cleanup is too aggressive and protracted, the environmental benefit of the additional actions quickly diminishes, and extending the "cleanup" beyond a certain point can cause more damage than would otherwise occur. Generally, it is beneficial to skim up and remove as much floating oil as possible and to mop up oil from the shore that can be easily removed without disturbing the sediment. In addition, oil can be washed back into the water to be skimmed and removed, but only from high-energy beaches and rocky areas where the high-pressure water jets used for cleaning essentially simulate extended strong wave action.

Other cleanup efforts, including aggressive removal of oiled sediment from low-energy environments and extended efforts to remove oil from below the surface of coarse sand or gravel beaches, are costly, provide little or no environmental benefit, and in some instances, cause additional ecological damage and retard recovery. In isolated wild areas, such as Alaska's Prince William Sound, the extended presence of people and their cleanup activities on beaches can have adverse effects on shorebirds, terrestrial wildlife, and marine mammals. In addition, the use of chemicals to disperse and dissolve oil can be damaging because such chemicals can be more toxic and persistent than the oil. Burning the oil as soon as possible after a spill or as oil reaches the surface from an ongoing spill (such as *Deepwater Horizon*) and then skimming off what does not burn may well produce better overall protection for the environment, but that option is never considered to be viable due to the inevitable public and media reaction that would be generated by "sensational" video of the burning oil. About 7500 tonnes of dispersants were used in the 2010 *Deepwater Horizon* spill. It is likely that future research will show that the dispersants, while minimizing floating oil and perhaps oiling of beaches, caused significant, perhaps more serious damage than would have occurred without their use. For example, without the dispersants, oil (and dispersant) concentrations near the seafloor would have been much lower except

immediately surrounding the well. The reported damage to deep water corals at substantial distances from the well might not have occurred without the dispersants.

Unfortunately, public pressure often requires that everything possible be done to clean up all the oil after a spill. As a result, spill cleanups often continue beyond the point at which many technical specialists believe they should be ended. Large sums of money are wasted in cleanup that yields no environmental gain or is even detrimental. Nature is very efficient at cleaning up oil spills within a few years. Our role should be to remove as much oil as possible quickly, and then let nature take its course. However, in some instances, naturally occurring hydrocarbon-degrading bacteria, or nutrients that encourage the growth of such bacteria, might be beneficially added to the oiled ecosystem to help nature heal itself.

Whether we should attempt to rescue, clean, and rehabilitate oiled birds and marine mammals in an oil spill area is debatable. For example, efforts to clean and rehabilitate sea otters after the *Exxon Valdez* spill are estimated to have cost more than $80,000 for each animal that was captured and eventually returned to the ocean (a total of only 197 animals). It is estimated that only about one-half of those animals survived a year after reintroduction, and many that did survive would probably have survived without cleaning because they were so lightly oiled. Mounting an animal cleanup program after a spill is emotionally satisfying, provides good material for the media, and provides those responsible for the spill with the ability to claim that they are making every effort possible to minimize the damage. However, almost all such efforts have proven ineffective.

Environmental scientists generally accept that even major oil spills do not cause lasting and widespread destruction of ocean ecosystems. However, concerns remain that major oil spills may have long-term adverse effects on some fish and other species because of the chronic toxicity of some of the more persistent hydrocarbons, particularly polyaromatic hydrocarbons (PAHs). For example, the unusually low numbers of salmon, herring, and other species returning to Prince William Sound in some (but not all) years after the *Exxon Valdez* spill raised suspicions that the low returns were in some way a chronic effect of the spill. These returns are highly variable from year to year because of natural factors, such as climatic variations and disease outbreaks. Consequently, determining whether variations, such as those in Prince William Sound since the 1989 spill, are natural or related to the spill or to other anthropogenic factors is exceedingly difficult.

The difficulty in determining the cause of these low returns is partly related to the lack of studies of year-to-year variability before the spill. After many years of intensive research following the *Exxon Valdez* spill, there is a consensus in the scientific community that, although some lingering long-term effects could be identified, many of these were caused by the aggressive cleanup efforts after the spill and that the sum of the effects that still remained a decade or more after the spill was far less significant than natural changes in the ecosystem (possibly related to the climate change associated with the Pacific Decadal Oscillation; **Chap. 7**).

A related issue is the widespread belief that if scientists are able to still find any trace of the spilled oil in the environment of a former spill location, the area is still polluted. Scientists can detect incredibly low concentrations of hydrocarbons, far lower than the levels found naturally in many areas, and well below any known sublethal toxicity limit. Measuring these small amounts means that these environments may be considered contaminated despite there being no known lingering effects, but they should not be considered polluted.

Chronic Inputs

Many hydrocarbons are toxic to some species, even in small concentrations. PAHs especially have a range of toxic effects, including teratogenicity and carcinogenicity (**CC18**). Products like motor oil and lubricants have elevated levels of PAHs, so any chronic long-term impacts of oil contamination from sources other than spills are likely to be more serious and widespread than those of the much more dramatic spills. Nonspill contamination sources are usually most highly concentrated in estuaries and the coastal zone, particularly near ports, harbors, and major cities. Hence, many scientists believe that greater research emphasis should be placed on the possible chronic effects of these other sources of oil in environmentally important coastal areas than on the massive research programs that follow major spills.

Lessons Learned

Several lessons can be learned from a careful review of oil pollution studies. Most importantly, the extraordinary public attention to and concern about major spills is misplaced. Recovery after such spills is relatively rapid, and their long-term effects are almost certainly less than the effects of the much more widespread chronic oil contamination from other sources.

Second, although cleanup of oil from the area of a spill helps to speed recovery, cleanup must be limited and done carefully, because overly aggressive cleanup can cause more damage than the spill itself.

Third, public opposition to offshore oil drilling in U.S. waters may result in greater oil pollution because offshore production has been historically safer than tanker transport. Although the U.S. must ultimately reduce its use of oil, it cannot do so immediately. If we do not produce oil in the continental U.S., we must continue to produce it from offshore oil rigs or import it in tankers and incur a greater risk of oil spills. If the imported oil is produced from offshore oil platforms in other countries, the global environmental cost of using oil will include any oil spills from these platforms, plus the transport risks.

Finally, ecosystems are naturally variable, so after an oil spill or other major disturbance, appropriate actions may restore the ecosystem to a natural state, but the balance of species in this new state will be different from the preexisting natural state because ecosystems are in a constant process of change.

References

Figure 16B1-1a – Segar, D., and E. Stamman Segar, 1997: *Exxon Valdez* tanker on Bligh Reef in Prince William Sound, Alaska.

Figure 16B1-1b – NOAA, 2007: *Amoco Cadiz* grounding. Accessed 18 February 2026, Source.

Figure 16B1-1c – Miller, J., 2010: Saving wildlife in Louisiana. FEMA, accessed 18 February 2026, Source.

Figure 16B1-1d – Williams, J.T., 2010: Controlled Burn. USCG, accessed 18 February 2026, Source.

Figure 16B1-1e – Kelley, P., 2010: Workers contracted by BP clean up oil on the beach in Port Fourchon, LA. USCG, accessed 18 February 2026, Source.

Figure 16B1-2a – USCG, 2010: Early response to the *Deepwater Horizon* oil rig explosion. Accessed 18 February 2026, Source.

Figure 16B1-2b – USCG, 2010: *Deepwater Horizon* offshore drilling unit on fire. Accessed 18 February 2026, Source.

16B1-3 – Data from The National Academy of Science, 2003: Oil in the Sea III: Inputs, Fates, and Effects. Accessed 18 February 2026, Source.

Table 16B-1 – Data from NOAA, 1992: Oil Spill Case Histories 1967–1991, accessed 18 February 2026, Source.

– Data from ITOPF, 2025: Oil tanker spill statistics, accessed 18 February 2026, Source.

16B1-4a – Adapted from NOAA, 2002: Trajectory Analysis Handbook, accessed 18 February 2026, Source.

16B1-4b – Adapted from Sun, S., C. Hu, and Tunnell Jr. J.W., 2015: Surface oil footprint and trajectory of the Ixtoc-I oil spill determined from Landsat/MSS and CZCS observations. *Mar. Pollut. Bull.*, **101**, 632–641, Source.

16B1-4b – Adapted from Piper, E., 1993: The *Exxon Valdez* Oil Spill Final Report, State of Alaska Response. ADEC, accessed 18 February 2026, Source.

This page titled 16.9: Box 1 - Petroleum in the Marine Environment is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.10: Box 2 - Sewage Wastes in the Marine Environment

The most fundamental and unavoidable wastes produced by human populations are human feces and urine. Almost all cities, towns, and villages in the U.S. and most of the world's major cities have extensive sewer systems to collect these materials and transport the sewage away from our living environment. The development of sewers was probably the single most effective advancement for human health protection in history. In medieval Europe, sewage was simply thrown into the streets in many cities. Apart from the obvious aesthetic problems caused by human excrement in the streets, this practice almost certainly fostered the devastating plagues of disease that decimated European urban populations during this period. Hence, for many decades, the practice of collecting sewage and discharging it into a river or ocean to be diluted and washed away was viewed as essential and beneficial. This view has changed almost totally since the 1970s. Sewage discharge into rivers and oceans is now seen by many as a pollution problem that must be completely stopped.

Figure 16B2-1. Sewage treatment plants are often located on the water's edge. This large secondary-treatment, the Hyperion Water Reclamation Plant in Los Angeles, is located on a prime stretch of Pacific Ocean beach just west of the Los Angeles International Airport.

Nature of Sewage Wastes

Sewage is a natural material that consists primarily of the water used to carry solid fecal material through sewers. The water contains small concentrations of dissolved nutrients, trace metals, and organic compounds that originate either in the water supply or in human excretions. In addition, sewage contains solid organic matter that has a wide range of particle sizes. This solid matter, which is composed partly of paper fibers, also includes low concentrations of nutrients and trace metals and a variety of bacteria, parasites, and viruses. Many of these organisms are benign to humans and aquatic species, but others are pathogenic to humans, and some are thought to be pathogenic to some aquatic species.

Although aquatic and ocean ecosystems can assimilate sewage without harmful effects, sewage discharges have caused pollution in many lakes, rivers, estuaries, and coastal waters. Pollution has resulted from one or more of three problems. First, pathogenic organisms in sewage can reinfect humans if sewage is discharged into waters where people swim or otherwise come into contact with the water, or if it is discharged into waters where harvested shellfish concentrate the pathogens. Second, organic matter and nutrients in sewage can cause blooms, hypoxia, and anoxia in discharge waters if the assimilative capacity of receiving waters is exceeded. Finally, sewage is often contaminated with potentially toxic compounds that enter the sewers from industrial discharges, street runoff, or household chemicals.

Only a few decades ago, many rivers, estuaries, and coastal areas near major cities were severely impacted by sewage pollution. Many waterways were simply overwhelmed by large quantities of organic matter. As a result, they became anoxic, causing biota, including fishes, to die, and creating significant aesthetic problems (odors and floating material). As environmental consciousness arose and city populations began to look toward the water for recreation, two approaches to solving the sewage pollution problem

were adopted. First, pipelines called “outfalls” were built to carry sewage offshore into the ocean, where dilution would be greater and where pathogens would die before reaching beaches. Second, sewage treatment plants were designed and developed to reduce oxygen demand and floatable material in the sewage before discharge.

Sewage Treatment

The many different technological approaches to sewage treatment can all be characterized by one of three treatment levels (**Fig. 16B2-2**):

- Primary treatment usually involves removing floating material, grinding up solids, and removing some particulate material by allowing it to settle out to form sewage sludge, which is not discharged but still must be disposed of. Some, but not all, pathogens die or are removed to the sewage sludge.
- Secondary treatment usually includes a primary treatment step, followed by a secondary step in which bacteria that use sewage organic matter for food are encouraged to grow in the sewage (**Fig. 16B2-2**). Most of these bacteria are removed in the secondary sewage sludge. Secondary treatment generally removes more than 90% of the BOD (which stands for “biochemical oxygen demand” and is a measure of oxygen-demanding organic matter) and more than 90% of the bacteria found in raw sewage. Some nutrients, trace metals, and toxic organic compounds are also removed to the sludge, but secondary treatment is not designed for this purpose, and a high proportion of each of many of these compounds remains in the treated liquid effluent that is discharged.
- Tertiary treatment processes are generally designed to remove nutrients, such as nitrogen and phosphate, from the effluent that remains after secondary treatment.

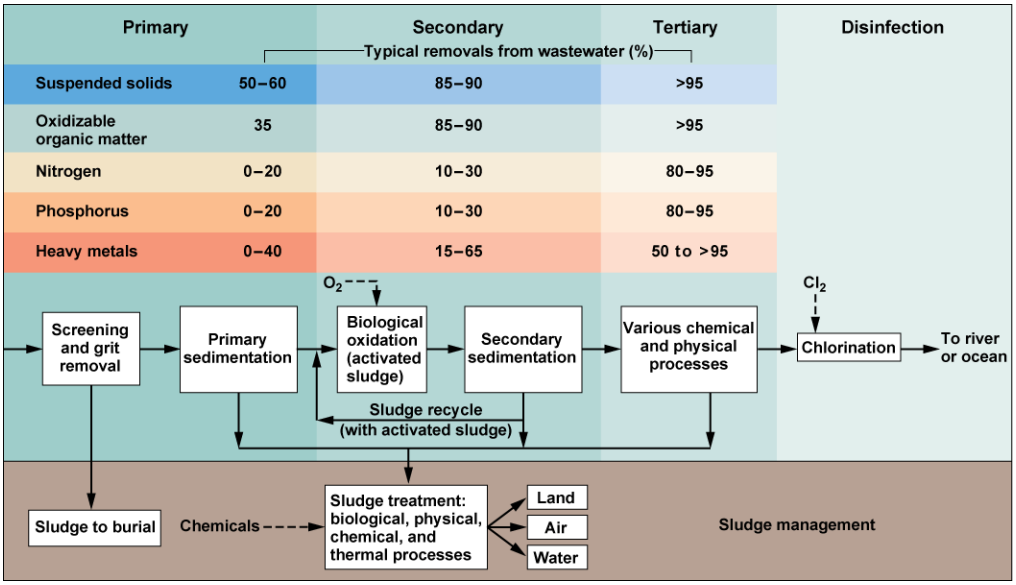


Figure 16B2-2. Sewage treatment is designed to remove suspended solids and organic matter from the waste stream before it is discharged to rivers, estuaries, or the ocean. Most areas of the U.S. are served by secondary-treatment plants that remove most of the suspended solids and organic matter, but are not very effective at removing nitrogen, phosphorus, and toxic trace metals. Tertiary treatment, which removes nutrients, and disinfection, which kills most remaining pathogens, are employed in very few locations because they are technologically complex and costly. All sewage treatment processes produce a semisolid sludge that must be disposed of on land or by incineration. In some locations, this sludge was disposed of for many years by being dumped into the ocean, but in the U.S. this practice was stopped in 1992.

In the U.S., all communities are required to perform secondary treatment, although in a very few exceptional cases this requirement is not met or is legally waived. Only a few communities are required to use tertiary treatment, which is very costly and often technologically unreliable.

Sewage treatment and diversion of treated sewage to the open ocean have dramatically improved the water quality in many rivers and estuaries. For example, salmon have returned to the Thames River in England after it had been anoxic and little more than an open sewer for many decades before treatment was started. Because some parts of San Francisco Bay were anoxic in the 1960s, the shore was covered in rotting debris and slime, and odors drove people away from the Bay. San Francisco Bay is now dramatically improved and no longer anoxic. Many other bays and estuaries have undergone similar improvements, but the story is not all good.

The United States has spent tens or hundreds of billions of dollars to build and operate secondary-treatment plants in almost all communities. In many locations, expenditures on sewage treatment have brought major environmental improvements. In others, where sewage inputs are small or where residence time at the discharge site is short, secondary treatment has brought little or no environmental benefit because assimilative capacity was not exceeded before treatment was started.

Many pathogens and toxic compounds are partially deposited in the sewage sludge, which is disposed of separately from the effluent discharge. However, secondary treatment was not designed to remove pathogens and toxic compounds, so a large proportion of many of these contaminants remains in the treated effluent and is discharged. Consequently, pollution by pathogens or toxic compounds in treated sewage effluents still remains a problem, particularly where residence times of the receiving waters are long and where commercially valuable shellfish beds are near discharges. For example, even though the Bay is no longer anoxic, shellfish beds remain closed, and several metals at times exceed water quality criteria in South San Francisco Bay as a result of the inputs of treated sewage.

Sewage sludge must be incinerated, deposited in landfills, or spread on agricultural land. Toxic substances and pathogens in the sewage sludge cause environmental impacts with each of these options.

Ocean disposal of treated sewage still causes significant environmental problems in some areas. Problems include nutrient-induced eutrophication of the coastal ocean, contamination by toxics, contamination of beaches and shellfish by pathogens, and damage to benthic ecosystems as a result of high organic loadings.

Eutrophication

Eutrophication of the coastal ocean and dead zones are discussed in **Chapter 13**. The primary cause of this eutrophication is believed to be nitrogen compounds discharged in agricultural runoff and treated sewage. Tertiary sewage treatment is successful in reducing phosphate concentrations in effluents, but potential technologies for nitrogen removal are more complex and costly. Consequently, the growing incidence and widespread recurrence of blooms and anoxia caused by eutrophication of the coastal oceans are growing pollution problems related to sewage effluent, which, at present, sewage treatment practices are inadequate to address.

Toxic Substances

Although some toxic compounds are removed in sewage sludge, a considerable proportion of many toxic substances remains in the treated effluent. Once in the environment, toxic substances are taken up by organisms and sediment particles (especially fine-grained particles). In discharge locations where receiving waters have long residence times or low-energy sedimentation regimes, the continuous introduction of toxic substances can cause a buildup of concentrations in sediment or waters. In most locations where toxic substances accumulate from sewage discharges, other sources (e.g., runoff and industrial discharges) also contribute, making it difficult to assess the relative importance of sewage versus these other sources.

Although toxic substances cannot be completely eliminated from sewage, they are unlikely to cause environmental problems if their concentrations in sewage reflect only the background levels of these compounds in human food and water supplies. Consequently, major efforts are underway to prevent toxic substances from entering sewer systems. In the United States, a vigorous and extensive pretreatment program requires removal of toxic substances from industrial wastes before they are discharged to sewers. This program has drastically reduced the inputs of toxic substances to sewage plants from industrial sources. However, only limited efforts have been made to reduce the inputs of toxic substances from cleaning products, other household sources, and illegal disposal of chemicals. The principal source of toxic substances in most major cities' sewage is therefore the home. In some areas, sewers are connected to storm drains through which urban runoff may contribute substantial quantities of toxic substances. Frequently, toxic industrial wastes are dumped illegally into storm drains and sewers to save treatment costs or the costs of disposal in secure waste disposal landfills.

Drugs

A number of drugs used by humans to combat medical problems are either excreted from the body in wastes or enter the sewage stream when people dispose of unused drugs by flushing them down a toilet or washing them down a sink drain. Some of these drugs survive the sewage treatment process and are discharged with the liquid effluents. Consequently, low concentrations of a number of these drugs have been detected in estuaries and the oceans. Because many of these chemicals have powerful pharmaceutical effects on humans, it is likely that some also affect at least some marine species. Little is yet known about these possible effects. There is particular concern that estrogen and other fertility control drugs may interfere with the reproductive

cycles of some marine species.

Pathogens

Sewage treatment is not designed to destroy or remove pathogens, but it does substantially reduce concentrations of some bacteria. Studies of the fate and effects of pathogens in sewage effluents discharged to the oceans are extremely difficult because of the low concentrations of pathogens in the environment and difficulties inherent in their isolation and measurement. Consequently, little is known about the fate of most pathogens in the oceans. Beach water quality and the safety of shellfish from pathogens are monitored by measuring concentrations of indicator bacteria, usually *Escherichia coli* (*E. coli*) or, more recently, *Enterococcus*. These organisms are present in all mammalian feces, not just human sewage, but they pose no significant human health risk themselves. Indicators are not perfect because they are not specific to human feces and because pathogenic organisms are transported in different ways and die at different rates in the environment.

Bacterial monitoring for indicator organisms has reduced the incidence of diseases carried by shellfish or contracted through water contact. However, bacterial pollution still necessitates the periodic closing of beaches, and vast areas of coastal and estuarine zones in the United States and elsewhere in the world are closed to shellfish harvesting. Many pathogenic bacteria and viruses have been found to survive in seawater for much longer than was believed to be the case until near the end of the past century, and more human health problems may be due to sewage in the marine environment than was previously thought. In particular, the incidence of temporarily debilitating intestinal upsets caused by sewage-carried bacteria may be much higher than has been recognized. Such relatively mild upsets are poorly monitored and reported by health authorities.

Effects on the Benthos

The benthic ecosystem surrounding some ocean sewage outfalls is substantially altered by the discharge. The alterations are related to and probably caused primarily by the organic sewage particles deposited in the sediment. If current energy is low, these particles accumulate in sediments surrounding the outfall, and the effects on the benthos depend on the rate at which organic matter accumulates. At rates only slightly above the background rate, the benthos are enhanced because the greater food supply increases the biomass, and species diversity remains high (**Fig. 16B2-3a**). At higher rates, some species are disadvantaged by the higher suspended particulate loads, and others are advantaged. Hence, biomass increases or remains high, but species diversity declines. As loading rates increase further, the benthos become dominated by a few tolerant species, and biomass eventually declines (**Fig. 16B2-3a**). Because loading rate usually decreases progressively with distance from the outfall, the effects also decrease as this distance increases.

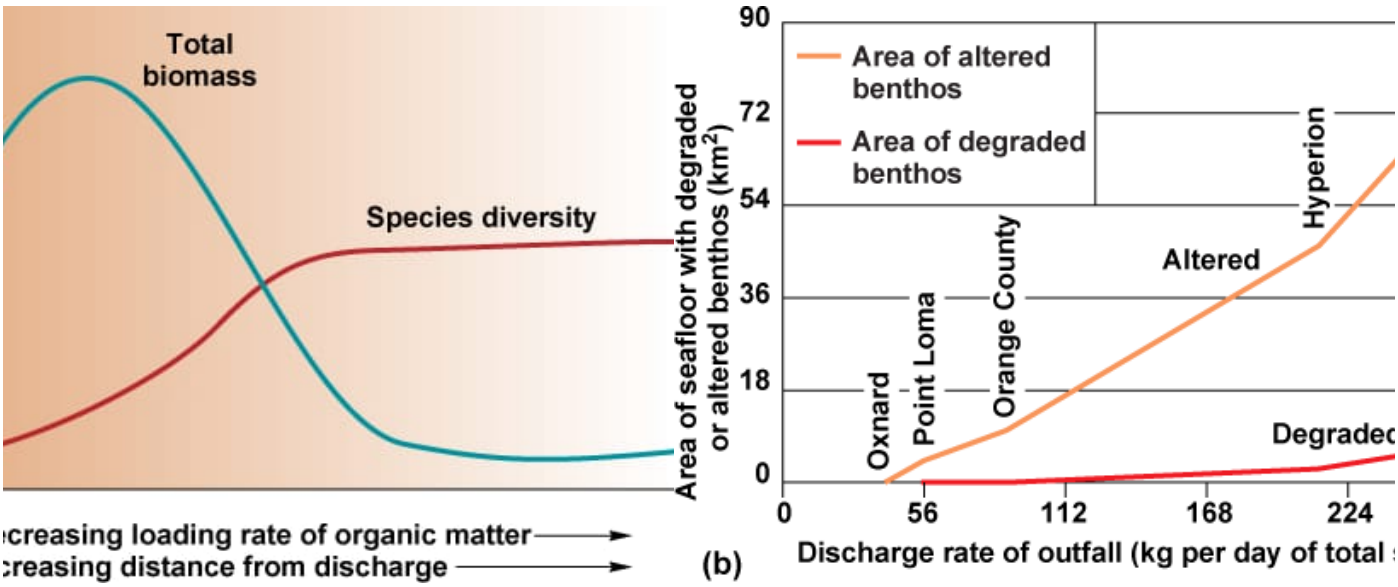


Figure 16B2-3. Degradation of the benthic environment in areas affected by sewage discharges to the marine environment. (a) Species diversity and biomass vary across a gradient of increasing sewage pollution. In the area on the right side where the loading rate of sewage is low, species diversity is not decreased, and the total biomass is increased. Increasing biomass without decreasing the species diversity could be considered an enhancement of the environment. (b) The area of sediments with altered and degraded benthic infauna increases as the rate of sewage discharge from outfalls increases in similar hydrographic regimes. The large outfalls off the California coast—JWPCP (the Joint Water Pollution Control Plant in Los Angeles County) and Hyperion Water Reclamation Plant—adversely affect very large areas, whereas the smallest outfalls (Oxnard and Point Loma) do not cause any area to be degraded.

Although it is somewhat arbitrary, a point can be identified in the gradient of effects around an outfall where the more severely impacted sediments are considered degraded, and the less severely impacted sediments are not. The area of degraded sediments surrounding an outfall varies with the rate of discharge of organic particulate matter (**Fig. 16B2-3b**). For outfalls in similar current regimes, a small volume of discharge may cause no degraded sediments, whereas a higher volume of discharge may cause a significant area of degraded sediments. Hence, if sewage is discharged through many small outfalls that are designed to maximize the dispersion of organic matter, the benthos will be moderately enhanced. However, if the same treated sewage is discharged through a few large outfalls, significant areas of the ocean floor may have degraded benthos.

These effects are analogous to those of fertilizer applied to a lawn. If the fertilizer is spread, the lawn will flourish. If it is dumped in one pile, it will burn and kill the grass close to and under it. Unfortunately, in the U.S. and elsewhere, environmental regulators have pursued a strategy of permitting only a small number of large outfalls in the belief that this restriction would limit any adverse impacts to only a few areas of the ocean.

For many years, large volumes of sewage sludge were dumped through an ocean outfall offshore of Los Angeles and from barges into the oceans offshore of New York, Philadelphia, and the United Kingdom. This practice was highly controversial. The media and public incorrectly blamed it for a wide variety of pollution problems, including the 1976 nutrient-induced anoxia off the New Jersey coast (**Chap. 13**) and the washup of floating medical wastes, including syringes, on New York beaches. Sewage sludge dumping contributed to these problems, at most, in a very minor way.

Lessons Learned

Public fears have caused sewage sludge dumping in the ocean to be prohibited. However, many scientists believe that the high cost and environmental and human health impacts associated with land-based disposal alternatives far outweigh any environmental gains realized by terminating ocean dumping.

Sewage sludge dumping has caused or contributed to some pollution problems. These problems were related to toxic chemicals and pathogenic organisms in the sewage sludge and organic overloading of the seafloor, particularly at the Los Angeles outfall. If toxic substances were prevented from entering sewers and pathogenic organisms in sewage sludge were killed by modified treatment processes, sewage sludge could be disposed of safely in the ocean and might even have beneficial effects because of its food value. However, contrary to the historical practice of restricting dumping to small sites, disposal would have to be spread over a sufficiently large area to prevent overloading of sediments or the water column with oxygen-demanding organic matter.

Sewage treatment, although expensive, has dramatically reduced the worst impacts of historical sewage disposal in the marine environment. The remaining impacts can be reduced only by the adoption of new and different approaches to sewage treatment technologies that more effectively remove toxic substances and nutrients, especially nitrogen, and kill all pathogenic organisms. Removal of all organic matter from sewage effluent is not necessary for disposal in the oceans, because, if adequately dispersed, such material will be assimilated without adverse effects and may in some instances even be beneficial.

References

Figure 16B2-1 – Crawford, K., 2006: The Hyperion Wastewater Treatment Plant and South Bay from the air, accessed 18 February 2026, Source.

This page titled 16.10: Box 2 - Sewage Wastes in the Marine Environment is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

16.11: Box 3 - Radioactive Wastes in the Marine Environment

During the nuclear testing era from 1945 to about 1990, the oceans were contaminated with many radionuclides. They were introduced in fallout from atmospheric nuclear bomb tests, liquid waste discharges from nuclear facilities, and deliberate dumping of solid and liquid radioactive wastes at ocean dump sites. Some of these radionuclides from these sources do not occur naturally on Earth.

Once in the marine environment, radionuclides become involved in biogeochemical cycles and behave in the same way as the stable isotopes of their elements (if stable forms exist). Thus, each radioisotope has its own distinct behavior in the oceans according to which element it is an isotope of. For example, tritium (hydrogen-3), a radioactive form of hydrogen released in nuclear explosions, quickly combines with oxygen to form water. The radioactive water molecules then join the hydrologic cycle and ocean circulation system, where they behave almost indistinguishably from other water molecules. An iodine radioisotope, iodine-131, enters solution in ocean water and is concentrated by marine biota, particularly certain species of algae. Plutonium does not occur naturally but is rapidly attached to particles when it enters the ocean environment. Most plutonium is carried with these particles to the sediment, where it remains until it has decayed to other elements.

Concentrations of radioisotopes are easily measured, even at exceedingly small concentrations, and sometimes even in an environmental sample that contains only a few atoms of the isotope. For this reason, radioactive isotopes have proven invaluable as tracers to study geochemical, biological, and physical processes in the oceans (**Chap. 8**). They are also commonly used as tracers in biological systems, including human medical scans. However, radioactivity can also have adverse effects on marine ecosystems and on human health. The primary concern is that some anthropogenic radioisotopes are concentrated in seafood and that people who eat the seafood will have an increased risk of cancer.

With the exception of very limited areas surrounding a few former nuclear bomb testing sites, the concentrations of anthropogenic radioisotopes in seafood are well below background levels of naturally occurring radioactive elements. Nevertheless, the prevalent, although not universally accepted, theory of carcinogenicity is that any increase in radioisotope intake will increase the risk of cancer. The assimilative capacity of the oceans for anthropogenic radioactive materials may therefore be considered essentially zero. In reality, a very small increase in radioactivity above natural background levels will not produce a measurable or significant increase in cancers. Hence, the oceans do have an assimilative capacity for radioactive materials, but it is small and not easily defined, because there is no generally accepted level of incremental risk.

Because the ocean's assimilative capacity for radioactive materials is so small, international agreements require that the release of such materials to the oceans be eliminated entirely. Almost all nations adhere to these agreements. The United Kingdom does continue to have a leak of some liquid radioactive wastes from its nuclear industrial complex at Sellafield, but the amount of radioactivity now released in discharges is very small, the radioactive materials release is apparently now largely contained within soil at the site, and the area is continuously monitored. In addition, after the fall of the Soviet Union, it was revealed that the Soviet Union had routinely dumped and discharged very large quantities of solid and liquid radioactive wastes directly into the Arctic Ocean and the Sea of Japan (**Fig. 16B3-1**), and into rivers that empty into the Arctic Ocean. This ocean dumping continued until 2005, when Russia finally agreed to a global treaty that bans such practices.

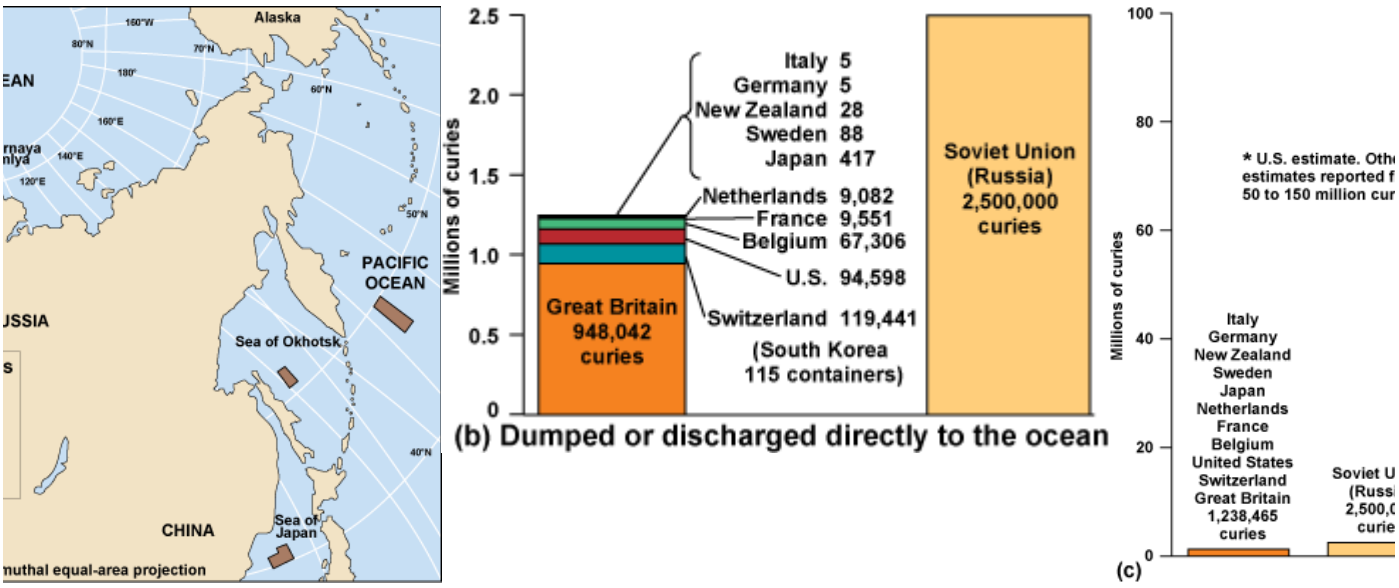


Figure 16B3-1. During the latter half of the 20th century, the former Soviet Union dumped large quantities of nuclear wastes in the oceans. (a) Damaged or decommissioned nuclear submarines were dumped and abandoned primarily in the Arctic and Pacific Oceans. (b) The quantities of radioactive wastes dumped in the oceans by the former Soviet Union far exceeded the total amounts dumped by all other nations. (c) Even though considerable amounts of nuclear wastes have been dumped in the oceans, these amounts are small compared to the amount released to the environment as a result of the 1986 Chernobyl, Ukraine, nuclear power plant accident. The massive earthquake and huge tsunami that damaged the Fukushima Daiichi, Japan nuclear plant in 2011, released about one-quarter as much radioactive material as Chernobyl. The Three Mile Island, Pennsylvania, nuclear power plant accident in 1979 released only a few curies of radioactivity, an amount far too small to be visible in these figures.

The quantities of radioactive wastes dumped in the oceans by the former Soviet Union far exceeded the total amounts dumped by all other nations (**Fig. 16B3-1b**). The material dumped by the former Soviet Union material includes a number of nuclear submarine reactors still containing their nuclear fuel (**Fig. 16B3-1a**). These fuel rods contain very large amounts of potentially dangerous radionuclides, including strontium-90 and cesium-137, which could bioaccumulate and threaten human health if released to solution and dispersed. At present, very little radioactive material appears to have been released from sunken submarines and solid wastes. However, there is still some concern that these wastes, and radioactive wastes dumped on land and into rivers, may eventually be released to contaminate seafood, especially in the Arctic Ocean and the Greenland and Bering Seas, which are among the world's most important fisheries. Fortunately, it is known that many elements for which radioactive isotopes are used in, or produced by, nuclear power plants are strongly absorbed by particles, so the probability of any release and transport of significant quantities of radioactive material from sediments at dumpsites is very low.

The massive earthquake and huge tsunami that damaged the Fukushima Daiichi nuclear power reactors in Japan in 2011 led to a sequence of events that caused large quantities of water containing radioactive elements to be released into the oceans. The total amount released to the oceans was only about one quarter of the amount released to the environment by the Chernobyl, Ukraine nuclear power disaster in 1986 (**Fig. 16B3-1c**), but considerably larger than the total estimated amount of radioactive waste dumped in the oceans prior to 2005. Despite the large quantities of radioactivity discharged to the oceans at Fukushima, studies showed that radioactivity levels in the environment were rapidly reduced with distance from the plant site and within five years were too low for any adverse effects to be found in any marine organisms studied, ranging from microalgae to mollusks and fishes even in ocean areas close to the accident site. The only major impact on the oceans appears to have been precautionary fishery closures of a substantial area of the oceans around Fukushima.

Fears of radioactivity releases fueled by media accounts of the Fukushima and Chernobyl, Ukraine (1986), power plant accidents, and a 1979 accident at Three Mile Island in Pennsylvania (that released small amounts of radioactivity to the atmosphere) have caused public perception to conclude that nuclear power plants are unsafe and nuclear power should not be used in the future. While these fears are understandable, they are not based on scientific facts. At present, there are about 4280 nuclear power plants operating worldwide, and these provide about 9.11% of the world's electrical power generation capacity and needs. Among all the technologically available sources of electrical power, nuclear power is the only technology that is free from carbon dioxide emissions to the atmosphere, can operate 24 h per day and can be operated at variable levels to meet hourly demand fluctuations, and that can provide city scale amounts of power from a single plants with a small land-use footprint (i.e., does not require large areas of agricultural or natural landscape to be altered). Other than the Chernobyl accident that occurred at a poorly designed reactor with no safety containment vessel, and the Fukushima accident that happened as a response to one of the most powerful

earthquakes, plus one of the largest tsunamis ever experienced in human history, no other major accidents have occurred that have caused deaths or significant radioactivity exposure beyond a small number of plant employees. Nuclear power's generation is a mature technology that can be ramped up rapidly worldwide using reactors that have benefited from decades of safety improvements both in design and site location, many scientists believe it deserves further consideration as it offers one of the only readily available technology that can reduce carbon dioxide emissions rapidly and, thus, enable humans to reduce the risk of incurring climate change consequences that could extend to planet-wide mass extinctions.

Waste disposal is a significant issue faced by the nuclear power industry. However, this can be addressed by means such as the use of breeder reactors that reduce the radioactivity of waste from other nuclear power plants or by safe storage and disposal of the wastes in a location where they will remain in place, undisturbed, for long enough for radioactivity to decay to background levels. The best such geologically stable long-term storage locations are likely to be beneath the sediments in the center of oceanic tectonic plates. It is not well known that technologies to emplace radioactive waste deep within ocean sediments were extensively researched and developed in the U.S. in the 1970s and 1980s. However, this research was abandoned after the London Dumping Convention adopted a ban on all disposal of radioactive wastes in the oceans. This ban, established in 1993 despite substantial scientific disagreement, effectively removed two-thirds of the Earth's surface from any consideration as a suitable site for geological storage of radioactive wastes.

References

Figure 16B3-1a – Data from Yablovkov, Y., and Coauthors, 1993: *Facts and Problems Related to Radioactive Waste Disposal in Seas Adjacent to the Territory of the Russian Federation*. Office of the President of the Russian Federation, 72 pp.

Figure 16B3-1b – Data from Broad, W.J., 1993: Russians Describe Extensive Dumping Of Nuclear Waste. The New York Times, accessed 31 March 2026, Source.

– Data from IAEA, 1999: Inventory of radioactive waste disposals at sea. Accessed 31 March 2026, Source.

Figure 16B3-1c – Data from Steinhauser, G., A. Brandl, and T.E. Johnson, 2014: Comparison of the Chernobyl and Fukushima nuclear accidents: a review of the environmental impacts. *Sci. Total Environ.*, **470–471**, 800–817, Source.

This page titled 16.11: Box 3 - Radioactive Wastes in the Marine Environment is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17: CRITICAL CONCEPTS

The **Critical Concepts** are the heart of the science that you must learn in order to understand the basic processes of the Earth and its oceans. They are also concepts you will need to know before you can make sense of many issues that you will encounter in your future work or that will affect your life.

Figure CC-1. Image of the Earth from space showing land, ocean, and clouds.

Some of these concepts are quite difficult to fully understand, especially without the use of mathematics at a level that is not appropriate for an introductory text like this one. Therefore, the Critical Concepts that follow are structured to allow easy learning of the essential nature of each concept. This essential nature is encapsulated in a section entitled "**Essential to Know**." In this section, the essential aspects of each Critical Concept are listed as brief statements of all the key facts and characteristics of the concept. The information in these "**Essential to Know**" lists can be learned quickly and then used for easy reference while you study the areas of ocean sciences in which they are applied. This information is all you need to know to fully understand and enjoy the broad introduction to ocean sciences that this textbook provides. However, a fuller understanding of these Critical Concepts might substantially enhance your introductory learning experience and will be necessary if you choose to study any aspect of ocean sciences at a more advanced level. Therefore, for each Critical Concept, the "**Essential to Know**" section is followed by a section entitled "Understanding the Concept," which provides a more detailed explanation.

In the "Understanding the Concept" sections, the essential nature of the concept is explored in more depth. This section also describes some aspects of the scientific basis for, and general applications of, the concept in ocean sciences. To fully understand the "Understanding the Concept" section, you will need to read carefully and make constant reference to any figures noted. You will likely need to reread some of this material, perhaps several times, to fully understand it. However, you should remember that, as long as you learn the "**Essential to Know**" material, you will be able to understand the rest of this textbook and apply these concepts appropriately in class discussions, written projects, and exams.

17.1: CC1 Density and Layering in Fluids

17.2: CC2 Isostasy, Eustasy, and Sea Level

17.3: CC3 Convection and Convection Cells

17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension

17.5: CC5 Transfer and Storage of Heat by Water

17.6: CC6 Salinity, Temperature, Pressure, and Water Density

17.7: CC7 Radioactivity and Age Dating

17.8: CC8 Residence Time
17.9: CC9 The Global Greenhouse Effect
17.10: CC10 Modeling
17.11: CC11 Chaos
17.12: CC12 The Coriolis Effect
17.13: CC13 Geostrophic Flow
17.14: CC14 Phototrophy, Light, and Nutrients
17.15: CC15 Food Chain Efficiency
17.16: CC16 Maximum Sustainable Yield
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.19: CC19 Metagenomics
17.20: References

This page titled 17: Critical Concepts is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.1: CC1 Density and Layering in Fluids

Essential to Know

- **Fluids are separated by gravity into layers, each layer having a lower density than the one below it.**
- **Less dense fluids rise through more dense fluids.**
- **More dense fluids sink through less dense fluids.**
- **A fluid is neutrally buoyant when the fluid above has a lower density, and the fluid below has a higher density.**
- **A neutrally buoyant fluid will spread horizontally to form a layer.**
- **A fluid in which each higher layer is of lower density than the layer below it is stably stratified. There are no density-driven vertical motions within such a stably stratified fluid.**
- **The depth range in a fluid within which there is a marked vertical gradient of density is called a "pycnocline." A pycnocline inhibits vertical mixing between fluid above and below the pycnocline layer.**

Understanding the Concept

All substances exist in one of three physical states: solid, liquid, or gas (**Chap. 5**). Both liquids and gases are fluids. Seawater is a liquid, and the atmosphere is a gas, so both are fluids that flow in response to forces such as **gravity**. Some parts of the Earth's interior are liquid. Other parts, including the **asthenosphere** and **mantle** (**Chap. 4**), although they are solid, can flow extremely slowly at the high temperatures and pressures present within the Earth. These solids behave like fluids over the million-year timescales in which geological processes occur.

When two fluids are brought into contact, gravity causes them to separate vertically according to their **densities**. The density of a fluid is the mass of the fluid divided by its volume. Pure liquid water has an absolute density (specific gravity) of 1000 kg·m–3 at 4°C and 1 atmosphere (atm) of pressure. Generally, density is stated as a relative density (r), which is the ratio of the fluid's specific gravity to the specific gravity of pure water at 4°C and 1 atm pressure. Thus, pure water at 4°C and 1 atm has a relative density of 1.

The effects of gravity and other forces on fluids of different densities account for many of the processes studied in this text, including the circulation of ocean water and atmosphere, and the formation of ocean basins and continents.

Many fluids do not mix with one another. Examples include oil and water, and water and air. When two such fluids are placed in a container, the denser fluid sinks to the bottom, leaving the less dense fluid above. If the less dense fluid is a gas, such as air, it expands to fill the upper part of the container not occupied by the denser fluid. If the less-dense fluid is a liquid such as oil, it forms a separate layer on top of the denser fluid (**Fig. CC1-1a**). Even if we vigorously shake such mixtures, they quickly separate again into layers after we stop shaking.

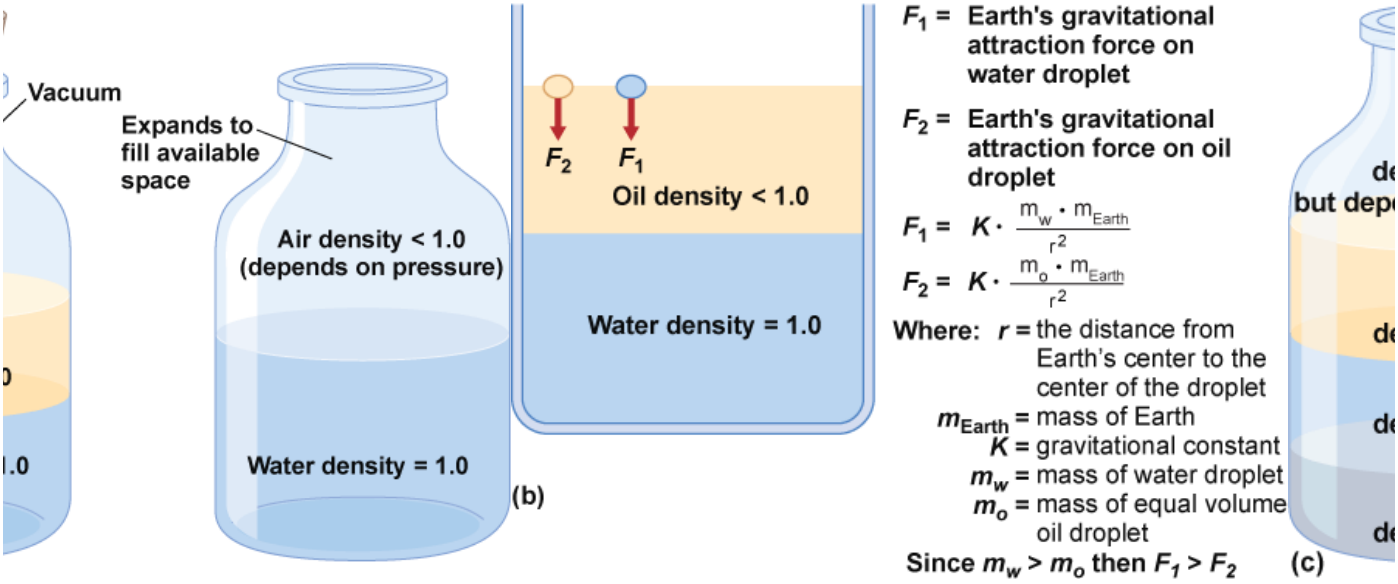


Figure CC1-1. Fluids (liquids and gases) that do not mix with each other are separated into layers in such a way that the density of each layer is greater than that of the layer above. (a) Oil has a lower density than water and forms a separate layer that floats on the water. If this layer is very thin, it is called a "slick." Air, which has a much lower density than either water or oil, forms a layer (the atmosphere) above the water. (b) Fluids are separated according to density because a more dense fluid has a greater mass and is subject to a greater gravitational force than the same volume of a less dense fluid. (c) Air, oil, water, and mercury separate into four layers, with density increasing toward the Earth's center (that is, with depth).

Gravity is the principal force that separates fluids of different densities. Consider a container full of water with an oil layer on top, into which we introduce a small drop of water (**Fig. CC1-1b**). Because water is denser than oil, the drop of water has a greater mass than the same volume of oil. The gravitational force between any object on the Earth's surface and the Earth increases in proportion to the object's mass. Therefore, the gravitational attraction force between the Earth and the water drop is greater than that between the Earth and the oil, and the water droplet is pulled down by the Earth's gravity more than the oil is. Because the oil can flow, the water droplet sinks, and the oil flows around it. When the water droplet reaches the top of the water layer, it is surrounded by a fluid of the same density, and the gravitational attraction on the water in the drop is now equal to the gravitational attraction on all the other water at its level. Therefore, the droplet sinks no farther. However, it does mix with the surrounding water by diffusion.

Gravitational sorting of fluids into layers according to density is called **stratification**. This process works equally well with more than two fluids. For example, if we mix mercury, water, oil, and air, the mixture will separate with water on top of the mercury, oil on top of the water, and air above the oil (**Fig. CC1-1c**). This process does not work quite the same way with solids because they cannot flow.

Density stratification can occur not only between fluids that do not mix, but also between fluids that do mix. For example, if we carefully add cream to a cup of cold coffee, it remains as a floating layer on top of the coffee until we stir the coffee and thoroughly mix the two layers. After two liquids are thoroughly mixed, the mixture has uniform density, and the two layers do not re-form.

Density stratification can also occur in a single fluid, such as pure water or air. Just as cream forms a layer on coffee, a less dense layer of any fluid will remain on top of a denser layer of the same fluid until they are mixed (**Fig. CC1-2**). In the natural environment, we often find fluids with multiple layers, each of slightly higher density from the shallowest to the deepest layer (**Fig. CC1-2**). When the density of the fluid increases progressively with depth, no layer has a tendency to sink or rise, and the fluid is said to exhibit stable stratification. Ocean waters are arranged in layers of different density in this way (**Chap. 8**) because the density of water varies with temperature and salinity (**CC6**). The atmosphere is also vertically stratified (**Chap. 7**), and the components of the Earth's interior are arranged similarly (**Chap. 4**).

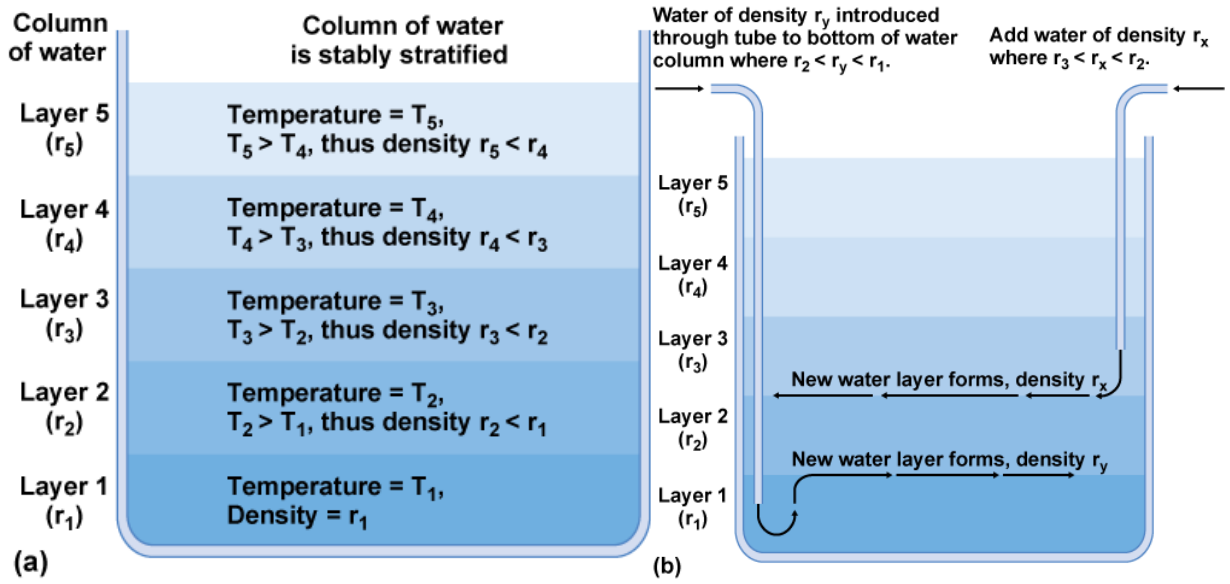


Figure CC1-2. Water may form distinct layers, each of which has a different density because it has a different temperature or salinity. (a) A water column is stably stratified if each successive layer has a higher density than the one above it and, thus, no water sinks or rises. (b) If water is introduced to a stratified water column, it will sink or rise until it reaches an equilibrium level at which the density of the water above it is lower and the density of the water below it is higher. It will then spread out to form a new layer at this depth. Some mixing of the introduced water with other water layers will occur as it sinks through layers of lower-density water or rises through layers of higher-density water. However, when the vertically moving volumes of water are very large (as they are in the oceans), such mixing is limited.

If we add a fluid to a column of other fluids, each with a different density and none of which mix, the added fluid will sink or rise until it reaches a depth where the fluid immediately above it is of a lower density and the fluid immediately below it is of a higher density. At this equilibrium level, the introduced fluid is neutrally buoyant, so it will neither sink nor rise, but will spread out to form its own layer (**Fig. CC1-2**). For example, if we introduce oil underwater to an air-water layered system, the oil will rise to the surface of the water and spread out. This is exactly what happens when sunken ships release their oil.

When a fluid of slightly different density is added to a density-stratified column of the same fluid or of a fluid with which it mixes, some mixing will occur as the added fluid sinks or rises to its equilibrium level. We can see this effect if we pour cream from a height into a cup of coffee. Both the cream and the coffee are primarily water, but different chemicals are dissolved or suspended in each, so they have different densities. If we pour the cream in carefully, it will be partially mixed as its momentum carries it down into the coffee and as it subsequently rises because of its lower density. The cream, mixed with some coffee, will rise to form a layer on top of the coffee and will then mix only very slowly (by molecular diffusion) unless we stir the mixture (and create turbulent diffusion). When the volumes of fluid in each layer are very large and vertical motions are slow, as they are in the oceans, mixing is very limited. The oceans are vertically stratified (**Chap. 8**), and density-driven vertical motions of water masses move large volumes of water vertically to find their equilibrium level with little mixing between the ascending or descending water masses and the layers of water through which they rise or sink.

Our cream and coffee experiment can illustrate another important concept. After we carefully place a layer of cream on the coffee, there is a distinct cream layer of lower density and a distinct lower coffee layer of higher density. Between these layers is a thin region where coffee and cream are mixed in varying amounts. The vertical distribution of density in this experiment is shown in **Figure CC1-3**. Density is uniform in the surface layer of cream, becomes progressively higher (this is referred to as a density gradient) in the transition layer of mixed cream, and is uniform in the lower layer of coffee. The transition region in which there is a vertical density gradient is called a **pycnocline**. More energy would be needed to move a molecule of water vertically through this pycnocline than would be needed to move a molecule the same vertical distance in either the cream or the coffee. Thus, pycnoclines act as barriers to vertical mixing.

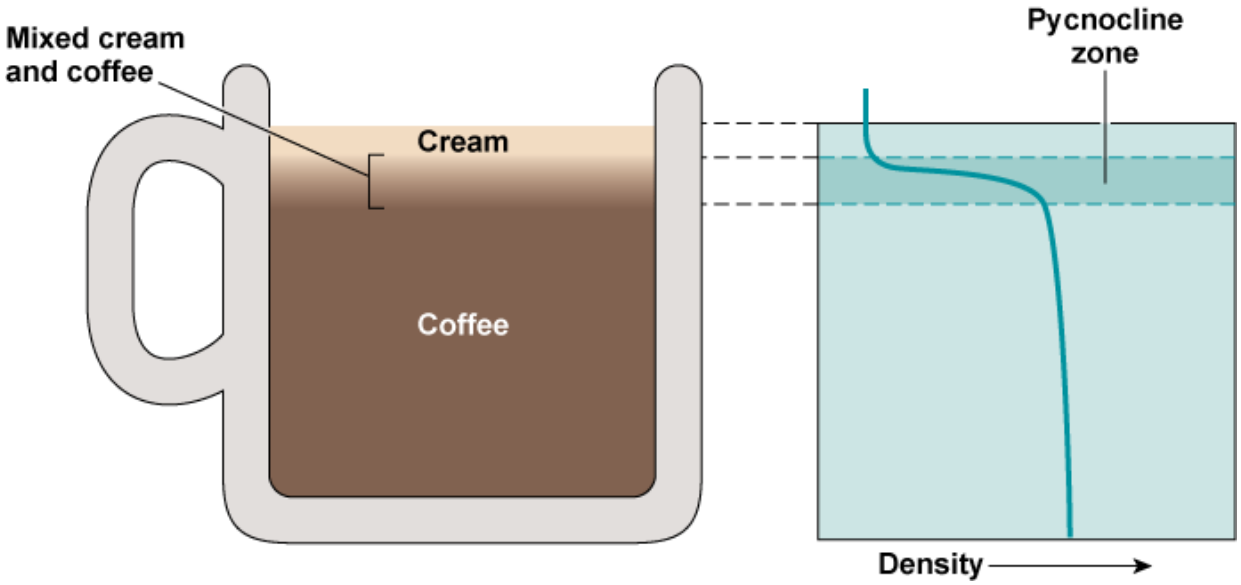


Figure CC1-3. If cream (which has a lower density than coffee) is introduced at the surface of a cup of cold coffee, the cream forms a surface layer overlying the coffee. Between the cream and coffee layers, there is a zone within which the cream and coffee are mixed in proportions that change progressively with depth from pure cream to pure coffee. In this zone between the coffee and cream layers, there is a rapid change in their respective proportions, which results in a sharp increase in density with depth. This zone, in which density changes rapidly with depth, is called a pycnocline zone. Vertical mixing between the two layers takes place only slowly across a pycnocline because the lower-density upper layer (cream) has no tendency to sink, and the higher-density layer (coffee) has no tendency to rise.

The strength of the density gradient in a pycnocline determines how strong a barrier to vertical mixing it is. Try the experiment again, but use milk instead of cream. Milk has a slightly higher density than cream (because it contains less fat). It is almost impossible to pour the milk into the coffee without the two mixing, even though milk does have a lower density than water. In this case, the density gradient, and therefore the pycnocline between milk and coffee, is very weak.

Pycnoclines form in many parts of the oceans, particularly where warm surface water overlies colder deep water, and they are important because they inhibit vertical mixing. The density of a fluid is altered as heat is gained from the sun at its surface and as heating or cooling occurs by conduction and radiation at its interfaces with its container or other fluids surrounding it. Density is also altered by changes in pressure and by changes in fluid composition, such as the addition of salts to water (**Chap. 5**) and of water vapor to air (**Chap. 7**). The factors that affect the density of air are discussed in **CC5** and **Chapter 7**, and the factors that affect the density of water are examined in **CC6** and **Chapter 5**. **CC3** explains more about the vertical motions created by such density changes.

This page titled 17.1: CC1 Density and Layering in Fluids is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.2: CC2 Isostasy, Eustasy, and Sea Level

Essential to Know

- **The level at which a solid floats in a liquid is determined by the relative densities of the solid and the liquid.**
- **Solids of equal density float in a liquid with the same percentage of their volume submerged.**
- **Two solids of different densities will float in a liquid in such a way that the solid with the greater density has a greater fraction of its volume submerged than the less dense solid.**
- **Lithospheric plates float on the asthenosphere. Because oceanic crust is denser and thinner than continental crust, the surface of the oceanic crust is always lower than the surface of the continental crust.**
- **Changes occur in the density and thickness of lithospheric plate sections as a result of various processes. The affected plate section rises or falls in response to such changes in a process called "isostatic leveling."**
- **Isostatic leveling is slow. Equilibrium may not be achieved for millions of years after an event, such as the start or end of an ice age, that changes the distribution of glaciers.**
- **Sea level changes on a section of coast as the section of the plate rises or falls isostatically. Isostatic sea-level changes occur at different rates on different coasts.**
- **Eustatic sea-level changes are caused by changes in the volume of ocean water or in the volume of the ocean basins that occur as a result of a variety of plate tectonic and climatic change processes.**
- **Eustatic changes of sea level take place simultaneously and uniformly throughout the world. Eustatic equilibrium is easily and quickly attained.**
- **Changes of sea level on different sections of the world's coasts occur at different rates, and even in different directions, because of the interaction of eustatic and isostatic changes.**

Understanding the Concepts

Solid objects will sink through a fluid in which they are placed if the **density** of the solid is higher than that of the fluid. For example, a rock thrown into a pond sinks to the bottom. Solid objects whose density is less than that of the fluid will float. For example, wood or Styrofoam will float on water, and helium balloons will float on air.

When a solid object floats on a liquid, the depth to which the object is immersed is determined by its density and the density of the liquid. A simple principle describes this relationship. Called Archimedes' principle, it states that the floating solid will be immersed in the liquid deep enough that the water it displaces has exactly the same total mass as the solid.

We can understand Archimedes' principle by considering a rectangular block of wood floating on water (**Fig. CC2-1**). If the wood has a density of 0.5, or half the density of water, the wood floats with exactly half of its volume underwater and half exposed. If the wood has a density of only 0.1, or one-tenth the density of water, the wood floats with only one-tenth of its volume underwater and nine-tenths exposed (**Fig. CC2-1a**). If we place a weight on top of the less dense piece of wood, the wood is forced to sink into the water. It sinks until the mass of the total volume of water displaced is equal to the total mass of the wood plus the weight (**Fig. CC2-1b**).

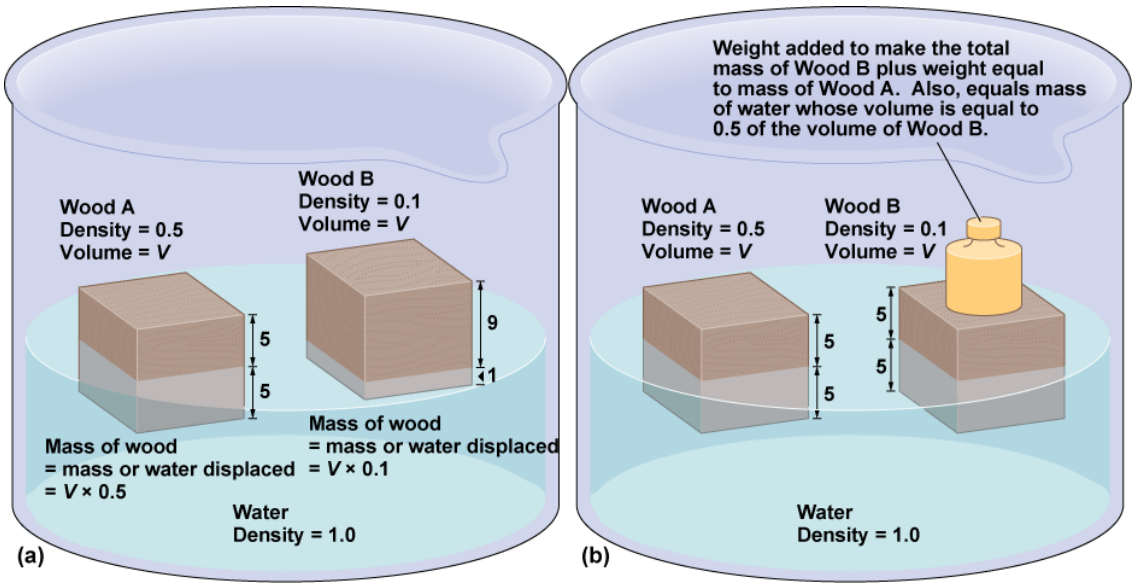


Figure CC2-1. Archimedes' principle determines how much of a floating solid's volume is below the surface. The solid floats at a depth such that the weight of the volume of water or other fluid that it displaces (that is, the volume of the part of the solid that is below the water or fluid surface level) equals the total weight of the solid. (a) Lower-density materials will float higher than higher-density materials. (b) Adding weight to a floating, low-density solid increases its mass and causes it to float lower and thus displace more water. The volume of water displaced will have a weight that is equal to the total weight of the floating solid plus the added weight that it supports.

If two pieces of wood have the same density but different thicknesses, the thicker piece will float with its upper surface higher above the water surface than the thinner piece, and it will extend deeper into the water (**Fig. CC2-2**). The two pieces have the same percentage of their individual volumes submerged. If we have two pieces of wood of identical dimensions but different densities, the wood with higher density will float lower in the water (**Fig. CC2-1a**). This effect can be observed in a river, lake, or ocean. For example, most logs or driftwood pieces float with much of their volume immersed below the waterline. In contrast, Styrofoam and balsa wood seem almost to sit on the water, because their low density means that only a small percentage of their volume is immersed.

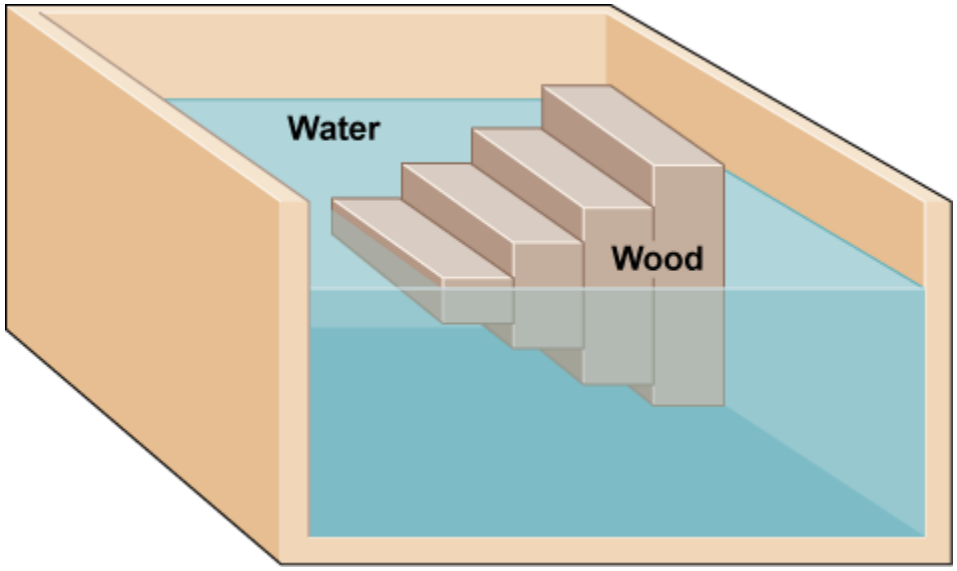


Figure CC2-2. Solids of the same density but of different thicknesses always float with the same proportion of their volume below the surface. Thicker blocks float with their upper surface higher above the surface of the fluid on which they float, and their lower surface deeper below.

Lithospheric plates float on the fluid surface of the **asthenosphere** in the same way that wood floats on water. Therefore, a **buoyancy** equilibrium is established in which the plates float freely on the material beneath them at a level that is determined by their density and thickness. This level is called isostasy, or "isostatic equilibrium."

Continental crust has a lower density than oceanic crust (**Chap. 4**). Therefore, lithospheric plates covered with continental crust float with a smaller percentage of their volume "submerged" than plates covered with oceanic crust. If the plates with continental crust and oceanic crust were the same thickness (**Fig. CC2-3**), the upper surface of the continental crust plate (the land surface) would be higher than the surface of the oceanic crust plate (the seafloor). In addition, the lithosphere-asthenosphere boundary would be shallower (farther from the Earth's center) under the continents than under the seafloor. However, lithospheric plates with oceanic crust are thinner than those with continental crust. Because the two types of crust are of different thicknesses, the lithosphere-asthenosphere boundary is shallower below ocean crust than it is below continental crust (**Fig. CC2-3**), and the continent surface is more elevated above the oceanic crust surface than it would be if ocean and continental crust were the same thickness.

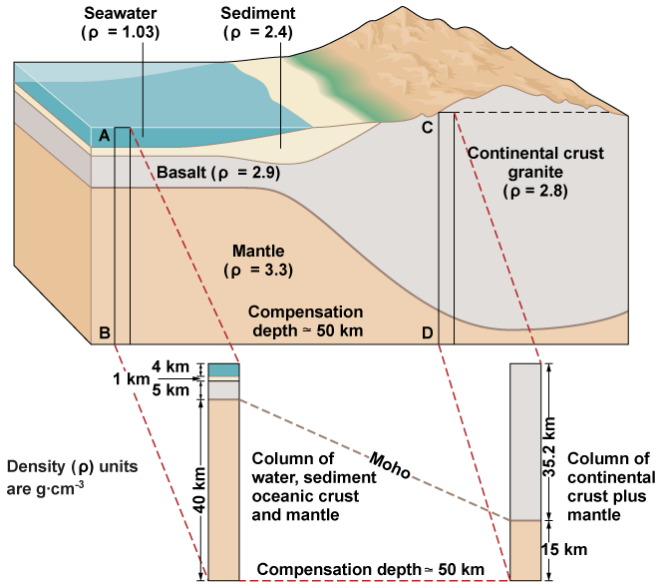


Figure CC2-3. Isostasy is the condition in which blocks of lithosphere float on the asthenosphere at the equilibrium level determined by Archimedes' principle. Above the compensation level (the depth at which the asthenosphere behaves as a fluid such that it can be displaced by floating lithospheric plates), the total weight of a column of continental crust plus mantle at isostasy will equal the total weight of a column of water, sediment, oceanic crust, and mantle. They do not quite do so in this figure, because the numbers are rounded.

If the density and thickness of the lithospheric plates were invariable, the continents and seafloor would always remain at a fixed equilibrium height above the asthenosphere. However, both the density and the thickness of lithospheric plates are altered by a variety of processes. First, the crust can be heated, which reduces its density; or cooled, which increases its density. Second, the thickness of the crust can be increased by the formation of mountains or volcanoes or by the **deposition** of large amounts of **sediment**, or it can be thinned by the **erosion** of mountains or stretching of the lithospheric plate. Third, the thickness of the lithospheric plate can be increased by the cooling and solidification of upper mantle material. This material is added to the bottom of the plate. Conversely, the plate can be thinned by the heating and melting of mantle material on the underside of the plate. Finally, the volume and mass of the continental crust can be increased by the development of **glaciers**, or decreased if glaciers melt, or altered by several other means, such as the growth of **coral reefs**.

If the density of a section of lithospheric plate is increased, it sinks until it reaches its new equilibrium level in a process known as isostatic leveling (**Fig. CC2-4**). Similarly, crust whose density has decreased will rise isostatically. Changes in thickness of the crust will also cause isostatic leveling (**Fig. CC2-4**).

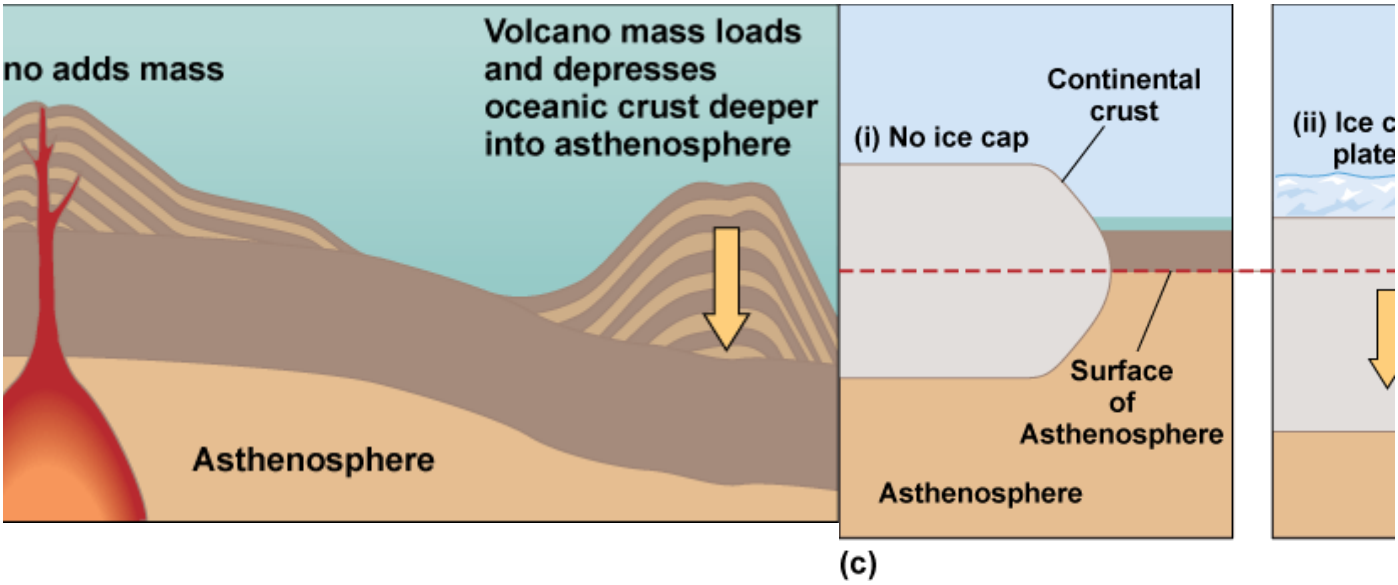


Figure CC2-4. Isostatic leveling takes place in response to crustal density changes, plate thickening, and plate loading processes. (a) As oceanic crust moves away from an oceanic ridge, it cools, its density increases, and it sinks lower in the asthenosphere. The plate also thickens as mantle material solidifies and is accreted to the cooling underside of the plate. (b) Volcanoes formed at hot spots add mass to the crust. The increased mass causes the plate to bend and sink lower under the volcano, especially after the old volcano moves away from a hot spot and cools, thus increasing its density. (c) A section of continental crust on which glaciers are formed will sink to a new isostatic equilibrium because of the additional weight of the glacier (like the wood in **Figure CC2-1b**). If the ice melts, the continent will rise in a process called "isostatic rebound" until it reaches its new equilibrium level. Isostatic changes take place very slowly compared to the climate changes that can cause large variations in the area of the continents that is covered by glaciers.

Isostatic leveling is very slow because the asthenosphere is very **viscous** and flows extremely slowly to accommodate the rising or sinking lithospheric plate. The processes that cause changes in isostatic level are often localized to certain sections of a lithospheric plate and its overlying crust. Therefore, one section of a plate may be rising while another is sinking. Lithospheric plates can bend to accommodate this process. When the section of a lithospheric plate that supports the edge of a continent changes its isostatic level, sea level rises or falls along this **coast** but does not necessarily change on other coasts.

Sea level can be altered not only by isostatic leveling of the continents, but also by processes that increase or decrease either the volume of ocean water or the volume of the ocean basins. Changes in water volume or in ocean basin volume cause changes in sea level that occur simultaneously and almost uniformly along all the world's coasts. Simultaneous worldwide changes in sea level are called **eustatic** changes (**Fig. CC2-5b**).

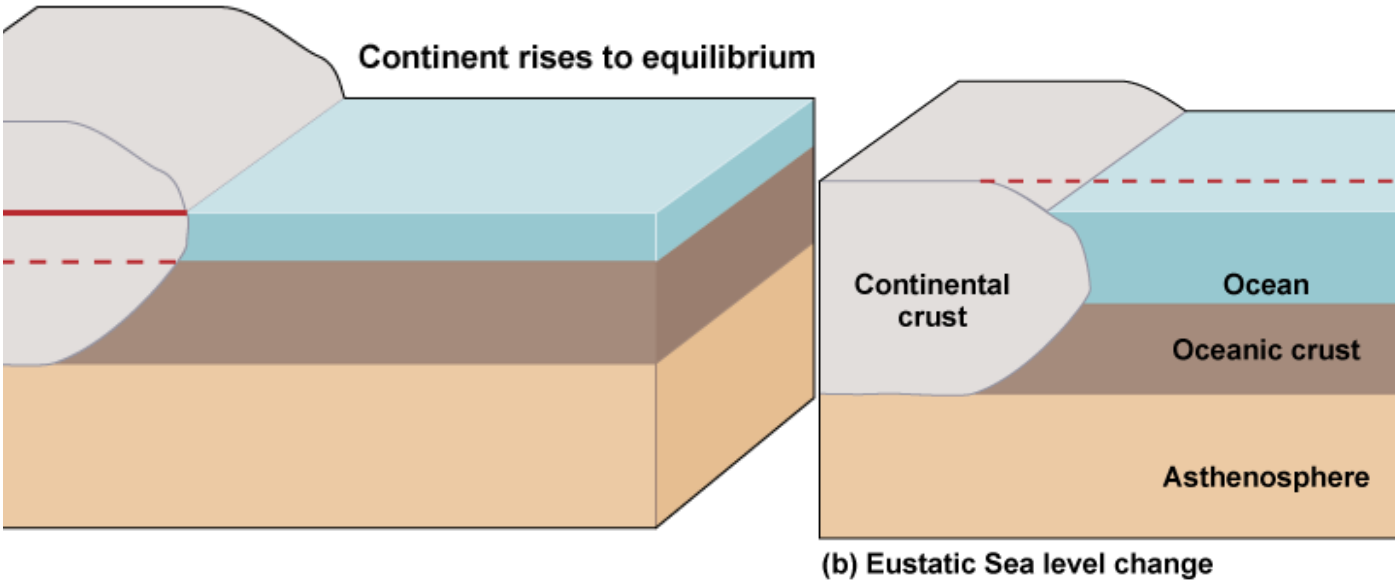


Figure CC2-5. (a) Isostatic sea-level changes take place when the land rises or falls while the seafloor and ocean depth remain the same. In this case, a continent that was previously depressed by the weight of an ice sheet rises slowly, exposing more of the continent to the atmosphere. (b) Eustatic sea-level changes take place when the volume of water in the oceans changes. In this case, the ocean has cooled and the water has been transferred to an ice cap on the land, thus lowering the sea level. Notice that the continent has not sunk lower in the asthenosphere in this diagram. In the situation depicted, the continent would eventually sink isostatically, but isostatic changes are much slower than eustatic changes.

Eustatic sea-level changes can be caused by several processes. For example, the volume of water in the oceans can be increased by the addition of water from melting glaciers or by elevation of the average temperature of ocean water that causes the water to expand. Correspondingly, the volume of ocean water can be decreased by increased **glaciation** or by a decline in average ocean water temperatures.

The volume of the ocean basins (actually the volume available for seawater to fill below a fixed level, independent of vertical movements of the continents) can be decreased if a larger percentage of the ocean basin floor is occupied by **oceanic ridges**. During the parts of **spreading cycles** when continents are being broken up, such as at present, there are many continents moving apart and many oceanic ridges between them. Hence, the percentage of the ocean floor covered by oceanic ridge is larger than it is during periods of reassembly of the continents. As a result, the ocean basin volume tends to be diminished, and sea level is correspondingly high.

Changes in the rate of **seafloor spreading** within a spreading cycle also affect sea level. When seafloor spreading is fast, the young, warm oceanic crust extends far from each oceanic ridge. Therefore, the volume of the ocean basin is reduced, sea level is higher, and more continental crust is covered by oceans. When seafloor spreading is slow, a smaller amount of young, warm oceanic crust is produced, the volume of the ocean basins is increased, and sea level is lower. During periods when the Earth's continents are assembled in supercontinents such as Pangaea, the number of oceanic ridges and the rate of seafloor spreading are reduced. Therefore, young, warm oceanic crust covers a smaller percentage of the ocean floor, the ocean basin volume is increased, and sea level is lower. The processes, just described, that affect the volume of the ocean basins are complicated and interact with each other as the various **plate tectonic** processes simultaneously change the thickness, relative abundance, and isostatic level of the continental and oceanic crusts. At the same time, **climate** changes, which themselves may be linked to tectonic processes, also change the volume of water in the oceans.

Processes that alter sea level occur on a variety of different timescales, from centuries to millennia. Eustatic equilibrium is reached quickly in response to such changes, whereas isostatic equilibrium is attained very slowly. Sea level may be rising, static, or falling on any individual section of the world's coasts as various eustatic and isostatic changes work in concert or in opposition in different regions.

At present, there is concern that human enhancement of the **greenhouse effect** (**Chaps. 1**, **7**, **CC9**) will lead to substantial additional global warming. One fear is that if sufficient warming occurs, it will increase the volume of ocean waters as the average temperature increases and the water expands, and as partial melting of the polar ice sheets contributes additional water. The result is likely to be an acceleration in the eustatic rise in sea level, which would have severe consequences because many coastal regions and cities would be inundated by the sea. Indeed, the acceleration of the rate of sea level rise has been observed in

currently available data. Therefore, measuring the current rate of eustatic sea-level change is considered to be critically important.

Measurement of the rate of eustatic change of sea level is complicated by the interaction of isostatic and eustatic changes that occur simultaneously on any given section of coast. For example, the sea-level change measured on the northeast coast of the United States will be the net result of several processes: rising sea level due to any greenhouse warming effect, rising sea level due to progressive cooling and isostatic sinking of the continental crust, and falling sea level due to isostatic rise caused by the geologically recent melting of the **ice age** glaciers that once lay on this crust. As a further complication, human activities such as **groundwater** and oil withdrawal cause the subsidence of certain sections of the coast. Because of these complicated interactions, we are able to measure any greenhouse-induced sea-level rise only by studying sea level change on many different coasts or by extremely precise and difficult satellite measurements.

This page titled 17.2: CC2 Isostasy, Eustasy, and Sea Level is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.3: CC3 Convection and Convection Cells

Essential to Know

- **Vertical motions in a fluid can be caused by temperature or compositional changes that alter the density of parts of the fluid.**
- **When density continuously decreases within one layer of a fluid (often at its lower boundary with another fluid or solid) and continuously increases in a higher layer (often the fluid surface), a convection cell is established.**
- **In a convection cell, a plume of lower-density fluid rises within the fluid until it reaches an equilibrium level where the surrounding fluid is of equal density. If the rising fluid has a lower density than all fluid layers above it, it rises to the surface. At its equilibrium level (or the surface), the fluid spreads out. If it is cooled or otherwise altered so that its density increases, the fluid eventually becomes dense enough to sink back to its original level, where it replaces rising fluid and reenters the cycle.**
- **Some convection cells have toroidal (doughnut-shaped) circulation, in which a rising column of fluid spreads out laterally in all directions at the top of the convection cell and returns by sinking in a ring-shaped band surrounding the rising plume. This pattern may be reversed with a column of sinking fluid surrounded by a ring of rising fluid.**
- **In other convection cells, both the rising fluid and the sinking fluid form an elongated curtain. The simplest example of such a convection cell is cylindrical.**
- **The rising plume is called an "upwelling plume," and the sinking plume is called a "downwelling plume."**
- **At the top or bottom of a convection cell, the fluid flows horizontally away from a location at a divergence. The fluid flows toward a location at a convergence.**
- **At the top of a convection cell, divergences are regions of upwelling, and convergences are regions of downwelling. At the bottom of the convection cell, divergences are regions of downwelling, and convergences are regions of upwelling.**
- **Several convection cells may be formed side by side within a fluid. Convergences and divergences always alternate across the top or bottom of the cells.**

Understanding the Concept

CC1 explains how density differences cause stratification in fluids, in which successive layers have higher density with depth. If the density of the layers in a vertically stratified fluid does not change, the system is stable and vertical motions do not occur.

Three stratified fluids are important to oceanographers: the Earth's interior layers, the oceans, and the atmosphere. Heat is introduced into the Earth's layers by energy released during **radioactive** decay in the core and **mantle** (Chap. 4). In the oceans, a small amount of heat is introduced by conduction from the mantle through the seafloor, and a much larger amount by solar heating of surface water. Heat is also exchanged (gained in some areas and lost in others) between oceans and atmosphere by **conduction**, radiation, evaporation, and precipitation (Chaps. 5, 7). The atmosphere exchanges heat with the oceans and land, gains heat from solar radiation, and loses heat by radiation to space (Chap. 7).

The heat transfer processes between sun, atmosphere, ocean, and land all vary with time and with location on the Earth. Variable heat transfers produce density changes in the mantle, ocean, and atmosphere and cause the stratification to become unstable. Changes in composition can also alter density. For example, changes in the **salinity** of ocean water (Chap. 5) or in the water vapor pressure of air (Chap. 7) alter the fluid density. When stratification becomes unstable, vertical motions called **convection** occur.

We can best understand convection using a simple kitchen analogy: the motion of water in a saucepan heated on the stove (Fig. CC3-1). Initially, when we place the saucepan on the burner, all the water within the pan is at room temperature. No vertical motion occurs because the water is uniform in density. When we begin heating the base of the saucepan, the water at the bottom is heated, while the water above remains at room temperature. As bottom water is heated, its density decreases, and it rises through the overlying water to the surface. If you watch a saucepan of water carefully as you start to heat it, you can see the vertical water movements as swirls of motion.

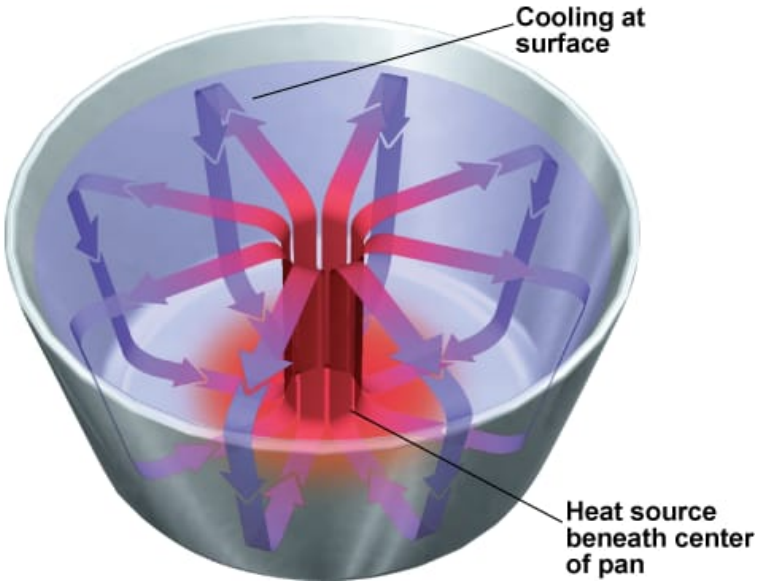


Figure CC3-1. In a saucepan heated at the center of its bottom, water will establish a circulation in which heated water rises to the surface, spreads toward the pan's sides, cools, and then sinks back to the bottom. This toroid-shaped convection cell is established because the water's density decreases when it is heated and increases when it is cooled. The idealized toroidal convection cell depicted would, in practice, normally be distorted because saucepans are usually heated across their entire base, not just at the center, and by turbulence in the flow patterns caused by variations in the rate of heating and cooling.

When the water is heated above room temperature, it begins to lose heat through the saucepan's surface and sides (**Fig. CC3-1**). Heated water rises to the surface, spreads out, and cools. As more warmed water rises to the surface, it forces the cooler surface water toward the sides of the saucepan. Because heat is lost through the water surface, the surface water that is displaced to the edges of the pan is cooled slightly and has a slightly higher density than the newly warmed water rising to the surface at the center of the pan. Accordingly, warmer water continuously flows toward the edge of the pan, cooling as it flows. As we continue to heat the saucepan, a continuous recycling motion is established. Water heated at the bottom of the water column rises at the center of the pan and is replaced by cooler water flowing in from the sides. The heated water rises to the surface, then spreads out, cools, and sinks down the sides of the pan to be heated again (**Fig. CC3-1**). The closed cycle of fluid circulation, driven by warming at depth that causes the fluid to rise, cooling at a lower-density level, and sinking back to its original depth to rejoin the cycle, is called a **convection cell**.

Convection is not as uniform as depicted in **Figure CC3-1**. The heating of the saucepan bottom and the heat loss through the water surface and the pan sides vary. Therefore, the location of the rising plume of the convection cell fluctuates, and more than one rising plume is often present. In addition, the water flow is **turbulent** rather than smooth. Turbulence can be seen in the internal swirling, churning motions in fast-running streams.

If we set our stove on low heat, the water in the saucepan does not boil, and convection cell motion continues indefinitely. Heat is added continuously and lost at the same rate that it is added. In nature, most convection cells operate in this way.

Convection cells transport heat vertically within a fluid. Convection also establishes areas on the surface and at the bottom of the cell where warmer fluid (in the center of the saucepan) and cooler fluid (at the edges of the saucepan) are concentrated. These features of convection cells are important in many ocean processes, including **plate tectonics** (**Chap. 4**), ocean water circulation (**Chap. 8**), and atmospheric circulation (**Chap. 7**). The upper and lower boundaries of convection cells are often the interfaces between two fluids (e.g. water and air) or between a fluid and a solid (e.g., the ocean surface and the seafloor). However, convection cells may develop between any two density surfaces within a vertically stratified fluid.

The convection cell just described is toroidal (doughnut-shaped), with a concentrated rising plume of heated water at its center and a ring of sinking water surrounding this plume (**Fig. CC3-1**). In another configuration of convection cells, both the rising plume and the sinking plume are extended laterally. In its simplest form, this type of convection cell is cylindrical in shape (**Fig. CC3-2**). Because fluid cannot both rise and sink at the same location, convection cells must be arranged so that the rising and sinking plumes alternate (**Fig. CC3-2**). The areas where fluid rises are upwelling zones, and areas where it sinks are downwelling zones. Upwelling and downwelling areas must alternate across the top or bottom of the fluid. We cannot go from one upwelling zone to another upwelling zone without passing through a downwelling zone. The locations of upwelling and downwelling in the atmosphere, ocean, and Earth's mantle are important to many aspects of Earth's geology, chemistry, climate, and biology.

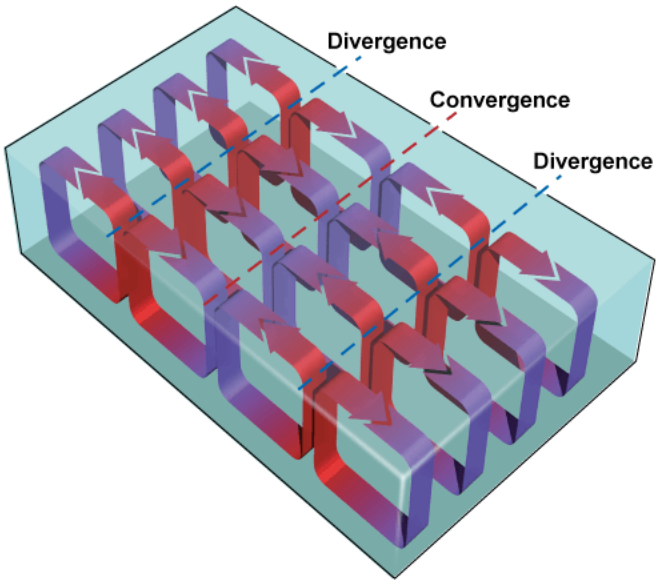


Figure CC3-2. Cylindrical convection cells. Note that convergences and divergences alternate between adjacent cells.

At the top of a convection cell, fluid flows horizontally away from upwelling areas as it is displaced by upwelled fluid. Such areas are called divergences (**Fig. CC3-3**). Similarly, fluid flows into downwelling areas to replace downwelled fluid. Such areas are called convergences (**Fig. CC3-3**).

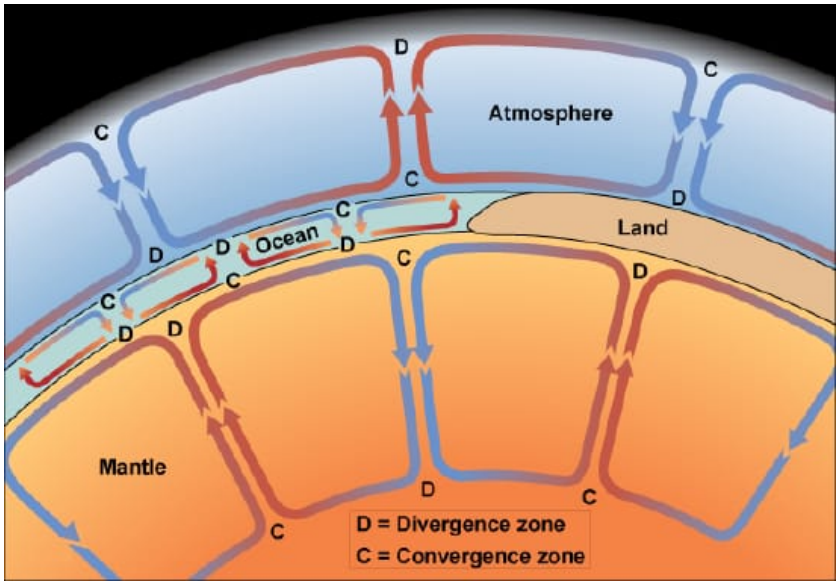


Figure CC3-3. The mantle, oceans, and atmosphere each have convection cells with alternating divergences and convergences across their upper and lower boundaries. In the atmosphere, we are generally concerned with the lower (sea-level) boundary of the convection processes, where divergences are downwelling zones, and convergences are upwelling zones. In the mantle and oceans, we are generally concerned with the upper (sea surface and upper surface of the asthenosphere) boundaries of the convection processes, where divergences are upwelling zones, and convergences are downwelling zones.

This text primarily considers processes that occur at the tops of convection cells (e.g., in the upper mantle and ocean surface waters), where upwelling zones are divergent, and downwelling zones are convergent (**Fig. CC3-3**). When we consider atmospheric circulation, we are concerned primarily with the bottom of the convection cell at the Earth's surface, where upwelling zones are convergences and downwelling zones are divergences (**Fig. CC3-3**).

This page titled 17.3: CC3 Convection and Convection Cells is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension

Essential to Know

- **The sinking rate of suspended particles in the ocean is determined primarily by particle size. Large particles sink quickly, and small particles sink more slowly.**
- **Large particles are deposited near where they enter the oceans (unless transported by turbidity currents). Very fine particles may be transported long distances by currents before they eventually settle on the seafloor.**
- **Particles can be resuspended after they reach the seafloor if the current speed is sufficiently high.**
- **Because fine-grained particles are cohesive, a high current speed is necessary to resuspend them from some fine-grained sediments.**
- **Particles may be alternately deposited and resuspended many times, where current speeds are variable.**
- **The particle size of grains within deposited sediment is determined by the range of current speeds and the size range of particles transported into the area. Areas with high maximum current speed generally have coarse-grained sediments. Fine-grained sediments are present only where the minimum current speed is low, and the maximum current speed is not extremely high.**

Understanding the Concept

Solid particles, or **suspended sediments**, are transported through the water by **currents** in the same way that dust particles are carried through the air by winds. Just as dust particles settle when the air is calm, waterborne particles (suspended sediments) sink to the seafloor when currents are slow or absent.

Particles sink through the water in response to **gravity**. However, particles do not all sink at the same rate, because of **friction** between them and the water molecules they must push aside. Higher-**viscosity** fluids have higher frictional resistance to sinking particles. The viscosity of water is low compared with, for example, molasses or motor oil.

Large, dense particles are not slowed significantly by friction as they fall through the water column. However, viscosity becomes more important as particle size decreases. To understand why smaller particles are more affected by viscosity, we must consider three factors. First, the gravitational attraction on a particle falling through the water column is directly proportional to its mass (for particles of the same **density**, it would be directly proportional to the volume). Second, because viscous friction occurs at the particle surface, particles with larger surface area are subject to greater friction. Third, the ratio of the surface area of a particle to its volume generally increases as particle size is reduced. The ratio of viscous friction to gravitational force, therefore, increases as particle size is reduced. Consequently, for small particles, such as those in suspended sediment, the particle's settling rate decreases progressively (the particle sinks more slowly) as the particle size decreases.

There are some exceptions to this rule. First, the density of the particle is important. Less dense particles sink more slowly than denser particles. This fact explains how some marine organisms avoid sinking (**Chap. 14**). Second, the shape of the particle is important. A parachute or feather falls more slowly through the atmosphere than a rock or rice grain because the surface area of a parachute or feather is large in relation to its volume and weight. Thus, air resistance or viscous friction is enhanced. Similarly, some marine organisms have body designs that maximize their surface area to prevent them from sinking (**Chap. 14**).

Most suspended particles (mineral grains) in the oceans, other than organic **detritus**, are of similar density. Hence, they sink at rates that are determined primarily by their diameter. **Table CC4-1** reports typical sinking or settling rates for particles of varying sizes. Sand-sized particles sink rapidly, and smaller particles sink much more slowly. Organic detritus particles are generally of lower density than mineral grains and sink more slowly.

Table CC4-1. *Settling Velocity of Particles of Average Density and Approximately Spherical Shape When No Current Is Present*

Property	Sand	Silt	Clay
Particle diameter	0.1 mm	0.01 mm	0.001 mm
Settling velocity	2.5 $cm \cdot s^{-1}$	0.025 $cm \cdot s^{-1}$	0.00025 $cm \cdot s^{-1}$
Time to settle 4 km (average ocean depth)	1.8 days	6 months (185 days)	50 years

Particle-sinking rates are modified by the presence of currents. As current speed increases, **turbulence** increases and particle sinking slows. This effect can be seen in a glass of orange juice. If the glass sits undisturbed for a while, the particles of orange pulp settle to the bottom of the glass. If the orange juice in the glass is gently stirred to keep the liquid in motion, the pulp will not settle. The effect of currents on particle-sinking velocities depends on both current speed and particle size. For a given particle size, sinking rate is reduced as current speed increases until the turbulence is sufficient to prevent the particle from sinking at all. The current speed at which particles no longer sink varies with particle size. This relationship is shown in **Figure CC4-1a**. At high current speeds, only large particles settle to the seafloor, and smaller particles remain in suspension. At lower current speeds, smaller particles also settle to the seafloor.

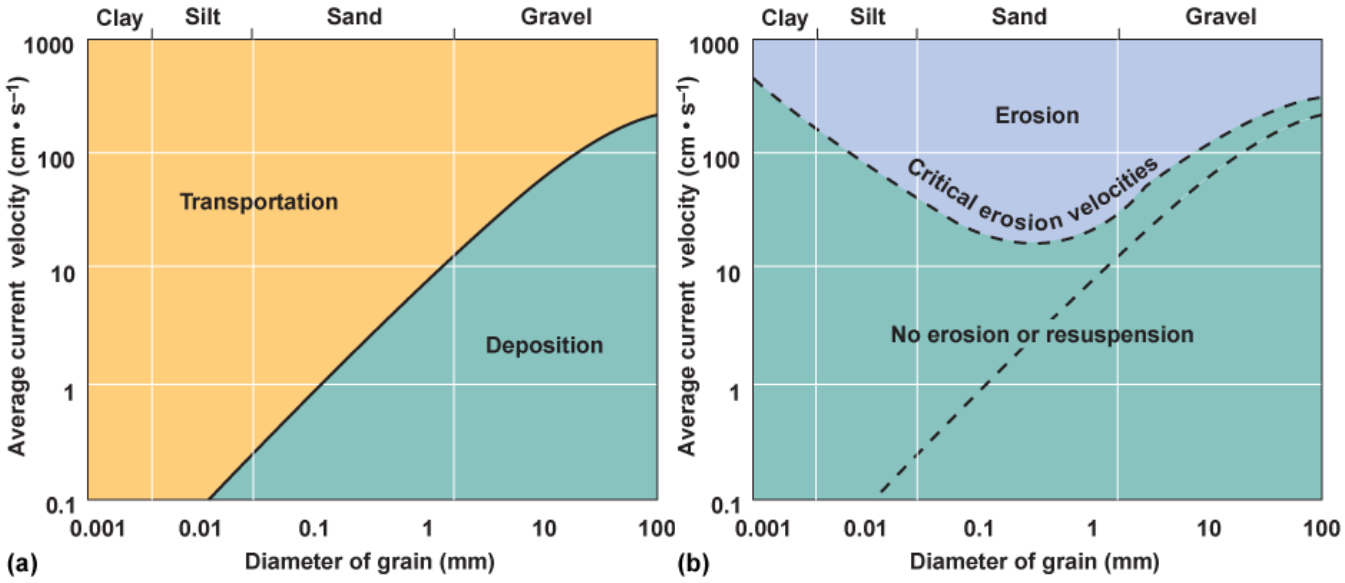


Figure CC4-1. Particle settling, resuspension, and current velocity. (a) Settling rate changes with current velocity. (b) Particle size and current velocity affect particle resuspension.

Once a particle has settled to the seafloor, it can be **resuspended** if the current speed at the seafloor is sufficiently high. **Figure CC4-1b** shows the current speeds needed to resuspend sediment particles of different sizes from sediments that consist primarily of grains of that size range. For all particle sizes, the speed needed to resuspend them is higher than the speed needed to prevent them from sinking (**Fig. CC4-1**). High current speeds are necessary to resuspend large particles, and somewhat lower speeds are needed to prevent them from sinking once resuspended.

As particle size decreases, both the resuspension speed and the speed required to prevent sinking decrease at approximately the same rate until silt- or clay-size particles are reached. The current speed needed to resuspend these smaller particles increases with decreasing particle size, so the speed needed to resuspend the finest sediment particles is higher than that required to

resuspend sand-sized sediments. The reason is that fine sediment particles are irregularly shaped and have a very large surface area relative to their volume (size). The grains are so close to each other that they are held by **electrostatic** attraction between the individual particle surfaces. In addition, they tend to lock together because of their irregular shapes. Organic matter may also help the particles stick to each other. Because they cling together so tightly, fine-grained sediments are considered **cohesive**.

Figure CC4-2 shows how suspended sediments of various grain sizes behave at different current speeds. The settling and resuspension threshold lines on this chart are the same as those in **Figure CC4-1**. Three areas are shown in **Figure CC4-2**: one where conditions are such that a particle would sink to the seafloor; another where conditions are such that particles would remain suspended; and a third where conditions are such that particles would not only remain suspended, but would also be resuspended if they were deposited on the seafloor. Referring to **Figure CC4-2**, consider an ocean area with no local sources of particles, where the current speed ranges from A to B. No sediment of grain size larger than d_2 is brought into this area by currents. Particles of grain diameters between d_3 and d_4 would be deposited, and sediments would consist of these grain sizes. Particles smaller than d_1 would be transported through this area and would not settle.

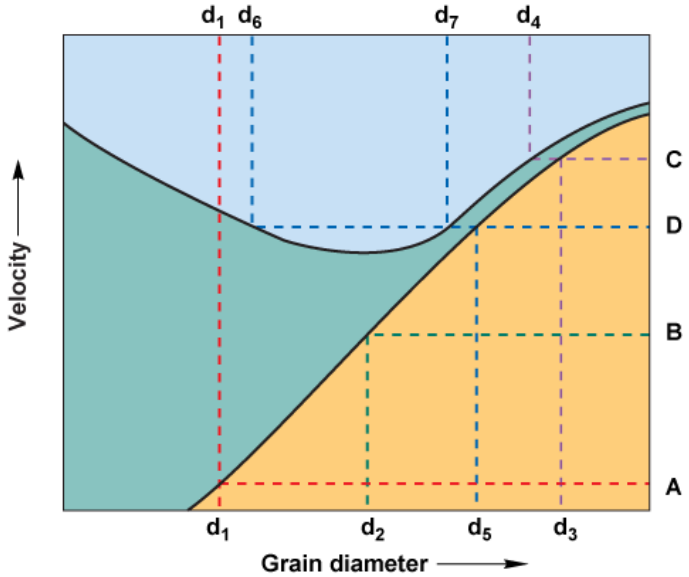


Figure CC4-2. Depiction of why most sediments are sorted by grain size, as explained in the text. Note how grain size sorting is determined by the range of wave and/or current speeds at a particular point on the seafloor.

If the current speed in **Figure CC4-2** ranged from A to D, particles between sizes d_1 and d_5 could enter the area and be deposited, but finer-grained particles would not be deposited. Particles between sizes d_6 and d_7 could be resuspended and removed, but fine-grained sediments between sizes d_6 and d_1, once deposited, could not be resuspended. Particles of size range d_4 to d_7 also would accumulate. However, in most locations, fine-grained particles far outnumber larger particles in the suspended sediment. Therefore, at this site the sediments would be fine-grained muds consisting primarily of grains between d_1 and d_6 in size.

From **Figure CC4-2**, we can conclude that large-grained particles cannot be carried far from areas where they are introduced to the oceans unless maximum current speeds are very high (but they could be transported by **turbidity currents**; **Chap. 6**). In contrast, fine-grained particles tend to be carried long distances before settling in an area where the minimum current speed is low, and they can not accumulate in any area where the minimum current speed is high. The diagram further demonstrates that the sediment particles that accumulate in a given location are restricted to a range of sizes determined by the current regime.

This page titled 17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.5: CC5 Transfer and Storage of Heat by Water

Essential to Know

- **Water has a high heat capacity. More heat is needed to raise the temperature of water by 1°C than is required to do the same for almost any other substance.**
- **Large amounts of heat can be absorbed and stored in the waters of the mixed layer (approximately the upper 100 m) of the oceans without causing major water temperature changes.**
- **Heat can be transported with water masses and released at another location by radiation, conduction, and evaporation. This is a mechanism by which heat is transported from the tropics to higher latitudes.**
- **Water has a high latent heat of fusion. More heat is needed to melt ice than is needed to melt almost any other substance.**
- **Ice in the polar regions acts as a thermostat because of water's high latent heat of fusion. Large amounts of heat can be lost or gained seasonally by the conversion of ice to water, or vice versa, without changing the temperature of the ocean surface water.**
- **Water has a high latent heat of vaporization. More heat is needed to vaporize or evaporate water than is needed to vaporize any other substance.**
- **Water evaporated from the ocean surface contains large quantities of heat as latent heat of vaporization. This heat energy is transported with the air mass and then released to the atmosphere when the water vapor condenses. This is another mechanism by which heat is transported from the tropics to higher latitudes.**

Understanding the Concept

Water has a very high **heat capacity** (**Chap. 5**, **Fig. 5-10**). This means that it takes much more heat energy to increase the temperature of water by 1°C than it does to cause the same temperature change in other substances, such as rocks or air. Consequently, in regions where solar radiation is intense, the surface layer of the oceans can absorb large quantities of heat while the water temperature undergoes little change.

Most of the solar heat absorbed by the oceans is initially absorbed in the upper few meters of water because solar energy is absorbed rapidly and does not penetrate far into the water column. The heat energy is distributed rapidly by **turbulence** throughout the ocean's upper **mixed layer**, stirred primarily by winds (**Chap. 8**). Thus, solar heat is distributed and stored in large volumes of water. In contrast, solar heat reaching the land is transferred downward only slowly by **conduction** (a very inefficient heat transfer mechanism) and is mostly radiated back into the atmosphere and space at longer **wavelengths**.

Because large quantities of heat can be stored in the surface layers of the oceans, heat can be transported with ocean **currents**. As the heated water travels, the heat energy can be released to the atmosphere. Thus, the high heat capacity of water facilitates the transfer of heat from tropical regions to colder regions near the poles (**Chap. 7**). The mild **climate** of parts of Europe in comparison with the climate of eastern North America at similar **latitudes** is partially the result of this mechanism. Heat is transported from the tropics into European seas by the warm Gulf Stream water and then released to the atmosphere (**Chap. 8**).

Water also has a high **latent heat of fusion** and a high **latent heat of vaporization**. This means that a large amount of heat energy is needed to convert ice to water, or water to water vapor (**Chap. 5**, **Fig. 5-10**). Heat energy added to ice to form water, or to water to form water vapor, is released when water freezes, or water vapor condenses. Thus, heat can be stored and transferred from one location to another if the conversions between ice and water, or between water and water vapor, occur at different times or places. For example, heat from the ocean surface can be used to vaporize or evaporate water. Water vapor can be transported through the atmosphere for thousands of kilometers until it recondenses and releases its heat energy to the atmosphere. Also, icebergs can be transported to lower latitudes by currents, where they melt as they absorb heat from the sun or surrounding water.

The high latent heat of fusion is important in polar regions. If water had a lower latent heat of fusion, the extent of the Arctic Ocean ice and the Antarctic ice sheet would vary more between summer and winter than they do at present. The presence of ice moderates the climate in these regions. As heat is lost in winter and gained in summer, most of the loss or gain goes to the freezing or melting of ice, and the ocean water temperature remains at the freezing point. Thus, the heat stored in the water by the melting of ice in summer is returned to the atmosphere during the winter as the water refreezes. In this way, the polar ice acts virtually as a thermostat, preventing air temperatures from becoming much higher in summer and much lower in winter.

The high latent heat of vaporization is important globally because it allows large amounts of heat energy to be transferred from the oceans to the atmosphere. Once in the atmosphere, the heat is redistributed by atmospheric circulation. Water need not boil for water molecules to be transferred from the liquid to the vapor phase (**Chap. 5**). At any temperature, water molecules are transferred from ocean to atmosphere by evaporation at the sea surface, but the rate of evaporation generally increases with increasing water temperature. When a water molecule evaporates from the oceans, it carries with it its latent heat of vaporization, which is more than 1000 times the amount of heat needed to raise the temperature of the same amount of water vapor by 1°C. Once evaporated, the water carries its latent heat through the atmosphere until the air cools, causing water vapor to condense to liquid water (clouds and rain).

When water molecules condense, they release their latent heat of vaporization to the surrounding air molecules. Because the heat capacity of all atmospheric gases is less than that of water, one water molecule can release enough heat to significantly increase the temperature of the air around it. Thus, evaporation and condensation of small amounts of water can transfer large amounts of heat energy from the oceans to the atmosphere. This mechanism is responsible for transporting heat from the equator toward the poles, thus moderating climate differences that otherwise would be extreme with increasing latitude (**Chap. 7**). In addition, this mechanism is responsible for the more moderate climate of coastal areas than of inland areas of the same latitude. It is also the driving energy of air movements in the atmosphere, including storms and **hurricanes** (**Chap. 7**).

This page titled 17.5: CC5 Transfer and Storage of Heat by Water is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.6: CC6 Salinity, Temperature, Pressure, and Water Density

Essential to Know

- **The density of water increases as the salinity increases.**
- **The density of seawater (salinity greater than 24.7) increases as temperature decreases at all temperatures above 4°C**
- **The density of seawater increases with increasing pressure. Density changes about 2% because of the pressure difference between the surface and the deep seafloor. The effect of pressure on density usually can be ignored because most applications require density comparisons between water masses at the same depth.**
- **Water with a salinity less than 24.7 has an anomalous density maximum. Pure water has its maximum density at about 4°C, but the maximum density of water occurs at lower temperatures as salinity increases.**
- **Between 4°C and the freezing point, the density of pure water decreases as temperature decreases.**
- **The relative importance of changes in temperature and salinity in determining seawater density varies with water temperature. Temperature variations are more important in warm ocean waters, whereas salinity variations are more important in cold ocean waters.**

Understanding the Concept

Below the ocean surface layers, movements of water masses are caused predominantly by differences in density between water masses (**CC1, CC3**). For this reason, water density, which depends on temperature, the concentration of dissolved salts, and pressure, is among the most important properties of seawater.

Increasing pressure compresses liquids, squeezing more mass into a smaller volume. Thus, density increases with increasing pressure. However, the molecules of water can be forced together only slightly, even by large pressure increases. Water, unlike air, is therefore almost incompressible, and the effect of pressure on water density is small. The density of seawater is only about 2% greater at the deepest depths of the oceans than it is at the surface. Consequently, the effects of pressure on density are generally ignored because oceanographers are usually interested in density differences between water masses at or near the same depth.

The density of water increases as **salinity** increases. This is because most **ions** have a higher density than water molecules do, and dissolved substances reduce the clustering of water molecules.

Decreasing temperature generally causes liquids to contract. Therefore, decreasing temperature generally increases density. However, the behavior of pure water is an exception to this rule. Between its boiling point and 4°C, water behaves normally: density increases as temperature decreases. However, between about 4°C and its freezing point, pure water decreases slightly in density as the temperature decreases. In other words, water density has a maximum at about 4°C (**Chap. 5**).

As salinity increases, the temperature at which the density maximum occurs decreases. At salinities of 24.7 and higher, the water density maximum is at the freezing point. The salinity of open-ocean waters is generally above 24.7. Therefore, in contrast to pure water and water with salinity less than 24.7, open-ocean seawater increases in density as temperature decreases at all temperatures above its freezing point.

The relationships among seawater density and pressure, temperature, and salinity are complex and are discussed in more detail in **Chapter 5** and **Figure CC6-1**. However, they are illustrated by the following summary:

- Pure water at 4°C and 1 atmosphere (atm) pressure has, by definition, a relative density of exactly 1.
- Density varies almost linearly with salinity. Density increases by about 0.00080 (±0.00004) for each unit of salinity increase, where salinity is measured in practical salinity units, abbreviated PSU. The density of "average" seawater is about 1.028 at salinity 35, 0°C, and 1 atm pressure, which is about 3% more than that of pure water at 4°C.
- Density varies almost linearly with pressure. Density increases by about 0.00045 for every 100 m of depth in the oceans.
- The rate of change of density with temperature is a function of both temperature and salinity. Density increases as temperature decreases, as shown in **Table CC6-1**.

Table CC6-1. *Density increase with salinity and temperature*

Salinity	Temperature (°C)	Density Increase per 1°C Reduction
0	1	-0.00007[a]
0	10	0.00009
0	20	0.00021
0	30	0.0003
20	1	0.00001
20	10	0.00014
20	20	0.00024
20	30	0.00033
40	1	0.00017
40	10	0.00018
40	20	0.00027
40	30	0.00034

[a]The density of pure water decreases with decreasing temperature below 4°C.

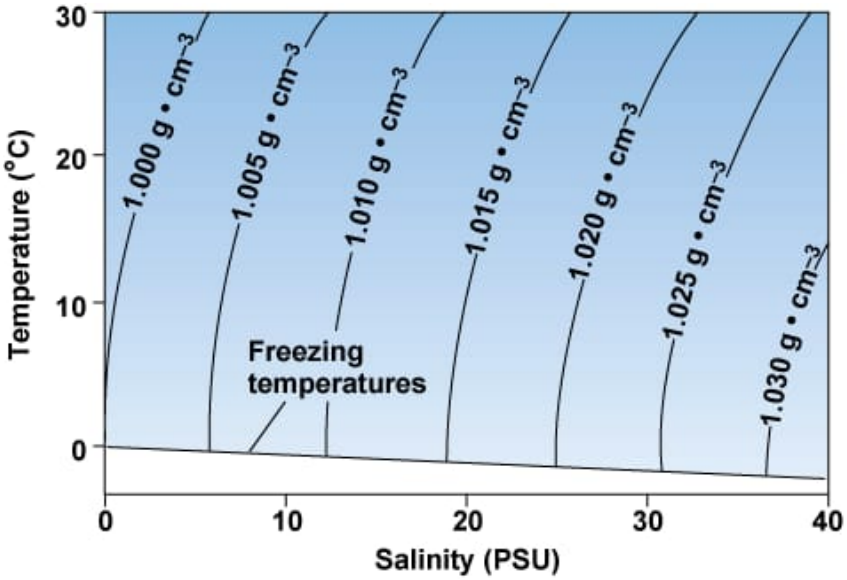


Figure CC6-1. The relationships among salinity, temperature, and density are complex. Generally, density increases as temperature decreases or as salinity increases. This figure is reproduced with additional information in **Chapter 5** (**Fig. 5-13a**).

The relative importance of temperature and salinity in determining density varies with water temperature. In cold waters, variations in salinity are more important than variations in temperature. In warm waters, variations in temperature are more important.

The density of seawater in the surface layer is altered continuously by solar heating, which changes temperature, and in turn evaporation, precipitation, and ice formation and melting, which alter salinity. These processes take place at the ocean surface. Vertical motions of water masses in the oceans are caused primarily by the sinking of high-density surface water formed by cooling or evaporation (increased salinity).

The vertical motions resulting from changes in water density are discussed in **CC3** and in **Chapter 8**.

This page titled 17.6: CC6 Salinity, Temperature, Pressure, and Water Density is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.7: CC7 Radioactivity and Age Dating

Essential to Know

- **Elements usually have several isotopes that have a different number of neutrons and differ in atomic weight but are essentially identical in chemical properties.**
- **Some individual isotopes of many elements are radioactive. Others are stable.**
- **Radioisotopes called "parent isotopes" decay by losing part of the atom, becoming a different isotope, called a "daughter isotope." Often, the daughter is an isotope of a different element.**
- **Radioactive decay releases heat and radioactive particles, including alpha particles, beta particles, and gamma rays.**
- **Each radioisotope decays at a certain fixed rate, expressed by the isotope's half-life.**
- **One half-life is the time it takes for exactly half of the atoms of a particular radioisotope in a sample to decay.**
- **The age of a rock, archaeological artifact, or skeletal or undecomposed remains of an organism can often be determined by measuring the concentrations of both the parent and daughter isotopes in the sample.**
- **For such age dating to be accurate, no daughter isotope must have been present when the sample was formed. In addition, no parent or daughter isotope atoms must have been gained or lost by the sample after it was formed. These conditions are often not met.**
- **Different parent-daughter isotope pairs must be used to date samples of different ages. Parent isotopes with long half-lives are useful only for dating older samples.**
- **Radioisotope dating is often used to calibrate other, less expensive dating techniques, such as fossil and stratigraphic dating, that reveal only the relative dates of samples from within a group of samples under study.**

Understanding to Concept

Atoms of any element may occur in several different forms called isotopes. Isotopes differ from each other in atomic weight, but they are virtually identical in their chemical properties. Some isotopes are stable, and others are radioactive. **Radioisotopes** (but not stable isotopes) called "parent isotopes" are converted, or decay, by losing part of the atom to become a different isotope (often of a different element), called a "daughter isotope."

During **radioactive** decay, the radioisotope releases heat and one or more types of radioactive particles, including alpha particles, beta particles, and gamma rays. Radioactive decay takes place at a fixed rate, different for each radioisotope. This rate is expressed by the radioisotope **half-life**. One-half of the atoms of a specific radioisotope in a sample will decay during one half-life for that isotope, regardless of how many atoms were present initially. Thus, after one half-life, one-half of the atoms originally present have decayed. After two half-lives, one-half of these remaining atoms have also decayed, and one-quarter of the original number of atoms of the isotope remain. After three half-lives, one-eighth of the original atoms of the isotope remain; after four half-lives, one-sixteenth of them remain; and so on.

The age of various materials, including rocks, archaeological artifacts, and the remains of living organisms, can be determined by radioisotope dating. To use this technique, the fraction of the atoms of certain radioisotopes that remain undecayed from the time the rock or other sample was formed must be determined. We can do this by measuring the concentrations of both the parent and daughter isotopes in the sample. If the sample is one half-life old, the parent and daughter will be present in equal concentrations. After two half-lives, there will be three times as many daughter atoms as parent atoms. After three half-lives, there will be seven times as many daughter atoms as parent atoms, and so on for other ages.

Radioisotope dating depends on several critical assumptions. We must usually assume that no atoms of the daughter isotope were present when the sample was formed and that neither parent nor daughter has been added or removed, except by radioactive decay, since then. These assumptions are often incorrect and are always difficult to test because samples undergo physical and chemical changes over time that can add or remove elements. Therefore, to be certain of the measured age, we must often date a sample by two or more different methods.

A particular radioisotope is not useful for age dating if only a very small fraction of the original parent atoms have decayed or, conversely, if almost all the original parent atoms have decayed. Hence, the ideal radioisotope for dating depends on the sample's age. Several of the most frequently used radioisotopes and their daughters and half-lives are listed in **Table CC7-1**. We

can see that carbon-14 is the most useful radioisotope for dating geologically recent samples (if they contain carbon), whereas uranium-235 is more suitable for older materials. The other radioisotopes listed are suitable only for dating ancient samples.

Table CC7-1. *Some radioisotopes used for age dating and their half-lives*

Parent Isotope	Stable Daughter Isotope	Half-Life* (years)
Carbon-14	Nitrogen-14	5560
Uranium-235	Lead-207	700 million
Potassium-40	Argon-40	1.3 billion
Thorium-232	Lead-208	1.4 billion
Uranium-238	Lead-206	4.5 billion

*Dating methods are most accurate when used on samples with ages between 0.5 and 3.5 times the half-life

The very sensitive, accurate measurements required for radioisotope dating, along with the equipment, can be expensive. In addition, the determined dates must be verified because the necessary assumptions for this technique are not always valid. For these reasons, radioisotope dating is often supplemented by relative dating techniques based on magnetic properties or **fossils** (**Chap. 6**). In some studies, radioisotope dating is performed on only a few representative samples, and the dates of these samples are used to calculate absolute dates for other samples dated by relative dating methods.

This page titled 17.7: CC7 Radioactivity and Age Dating is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.8: CC8 Residence Time

Essential to Know

- **Substances, including water, move through different parts of the oceans, atmosphere, and biosphere at different rates.**
- **The concentration of a substance in a part of the ocean, atmosphere, or biosphere is determined by how fast, on average, a molecule of the substance moves from input to output within that particular segment of the environment. This rate is measured as a residence time.**
- **Residence time is calculated by dividing the total quantity of a substance within an environmental segment (or "box") by the rate of either its input or its removal from the segment.**
- **Residence time can be used to estimate the concentration increases that will result from increased discharges of a substance into a segment of the ocean. It is particularly useful for pollution assessments in estuaries and coastal waters. Long residence time and small water volume lead to high concentrations of a contaminating substance.**
- **In certain situations, we can use residence times to estimate the relative magnitudes of inputs to, or outputs from, a particular ocean segment without measuring them directly.**

Understanding the Concept

Marine and atmospheric sciences involve the study of the movement of elements, water, and other substances within the oceans and atmosphere, and between the oceans, atmosphere, biosphere, and Earth's mantle and crust. These movements are components of complex **biogeochemical cycles** (Chap. 5). The complexity of chemical movements within these cycles makes it very difficult, if not impossible, to study a cycle as a whole. Accordingly, the **environment** must be divided into segments for most studies.

A segment of the environment is chosen by a characteristic that distinguishes it from adjacent segments in some way. A segment, which, for convenience, is usually called a "box" or "compartment," can be defined as almost any part of the Earth, ocean, or atmosphere. For example, a box could be defined as the entire world ocean, a single basin such as the Mediterranean Sea, or a simple square section of open ocean that is identified only by boundaries defined by an oceanographer. A box could also be the entire ocean **biomass** or the members of a particular fish species within a specified geographic area. Boxes are often defined to investigate the **residence times** of substances.

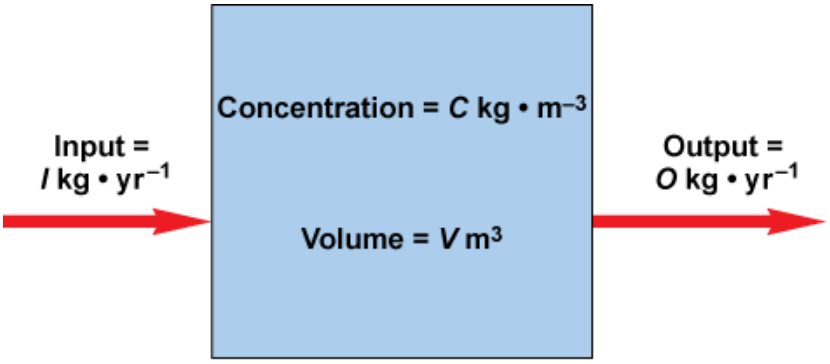
The substance of interest enters a box through certain identifiable pathways (inputs) and leaves through other identifiable pathways (outputs). Substances passing through the box can move quickly from input to output, or remain within the box for a long period before exiting. The average time that an atom or molecule of a relevant substance remains in the box is an important characteristic of the system and is expressed as its residence time.

To calculate the residence time in a defined box, one must measure the total quantity of the substance in the box, and either its rate of input to or its rate of output from the box. For example, if we want to determine the residence time of sodium in the Mediterranean Sea, we estimate the total quantity of sodium by measuring the sodium ion concentration at different places, averaging the result, and multiplying by the Mediterranean Sea's volume. The rate at which sodium is entering or leaving the Mediterranean must also be measured. All of the sea's inputs or outputs must be considered, including river **runoff**, atmospheric fallout and rain, and exchange of water between the Mediterranean and the Atlantic Ocean through the Straits of Gibraltar, between the Mediterranean and the Black Sea through the Bosporus, and between the Mediterranean and the Red Sea through the Suez Canal. We could determine the residence time of the water itself by measuring the total quantity of water and its inputs and outputs from the Mediterranean in the same way, but in this case, we would have to consider an additional output: evaporation.

Once we know the total amount of the substance of interest in the box and the rate of its input or output, we can calculate a residence time if we assume that the system we are studying is in a **steady state** (in other words, if we assume that the amount of substance in the box remains stable over time). Because the Earth's biogeochemical cycles have had billions of years to reach equilibrium, we can usually accept that assumption when we are studying large boxes, such as an entire ocean basin. However, the smaller the box, the less likely the assumption is to be true. Although the Earth's biogeochemical cycles are in approximate steady state when averaged globally, these processes can vary substantially at any one place and time because of such changes

as variations in the year-to-year rate of river flow.

If the total amount of substance in the box remains stable over time, the rate of input must equal the rate of output (**Fig. CC8-1**). The residence time is calculated as the total quantity of the substance in the box divided by either the output or input rate. Measured in years or another time unit, residence time is the time it would take for the outputs to remove all the substance from the box if all inputs were stopped, and vice versa.



$$\text{Mass of substance in box} = M = C \cdot V$$
$$\text{If } M \text{ is constant}$$
$$I = O$$
$$\text{Residence time (yr)} = M/I = M/O$$
$$\text{or } M \text{ (in kg) divided by } I \text{ or } O \text{ in kg} \cdot \text{yr}^{-1}$$

Figure CC8-1. This simple model shows how we calculate residence time by dividing the mass of the substance in the environmental segment ("box" or "compartment") by either the rate of input or the rate of removal (output) of the substance from the box.

The concept of residence time has many uses. First, it provides a method of estimating the rates of some processes that are difficult to measure directly. For example, if we can measure the concentration of a particular element in Mediterranean seawater and the rate of input of this element from rivers and adjacent seas, we can calculate the rate at which this element is being removed to the **sediment**. We did not include **sedimentation** (or evaporation and salt precipitation) in our preceding examination of sodium cycling in the Mediterranean because, for sodium (as opposed to other elements), the rate of removal to the sediment is extremely slow at present, so it is negligible in relation to other inputs and outputs.

A second use of residence time is to determine the probable fate of contaminants released to the oceans. If we release **contaminants** with long residence times, they will remain in the oceans for a long time and may cause **pollution**. If contaminants have short residence times, human additions to the oceans will be quickly removed (usually to the ocean sediment). Residence times of the elements in the world oceans are discussed in **Chapter 5**.

One important characteristic of biogeochemical cycles is that mechanisms for the removal of a substance from a box are generally related to the concentration of the substance in the box. If the concentration increases, the output rate will also increase. For example, if the concentration of a substance in the Mediterranean is increased, the rate of loss of the substance from the Mediterranean will increase as seawater flows out of this sea to an adjacent sea. Similarly, higher concentrations lead to faster uptake by **biota** and faster removal to sediment, although there is often not a simple linear relationship in these instances.

An increase in the input rate to a box generally raises the concentration in the box until the output rate matches the elevated input rate. If the expected increase in input is known, the resulting increase in concentration can sometimes be calculated and predicted. This method is extremely important in determining the potential of rivers, bays, or segments of the coastal ocean to be affected by contaminant inputs. A hypothetical example illustrates how this assessment can be made. **Figure CC8-2** shows the equations we could use for an **estuary** into which is discharged river water that carries a natural concentration (C) of a substance. An industrial complex is proposed that will discharge additional amounts of this substance to the estuary. For simplicity, we will assume that the substance of interest is not removed to sediment within the estuary and that it can leave the estuary only dissolved in seawater. From the calculations in **Figure CC8-2**, we see that the change in concentration (δC) that can be expected to occur in the estuary after the additional input (δI) begins is easily estimated if the residence time, volume of the estuary, and magnitude of δI are known. The important observation to be made about this result is that the concentration increase caused by an increased input of a contaminant substance is greater if the residence time (T) in the discharge water body is long in relation to its volume (V).

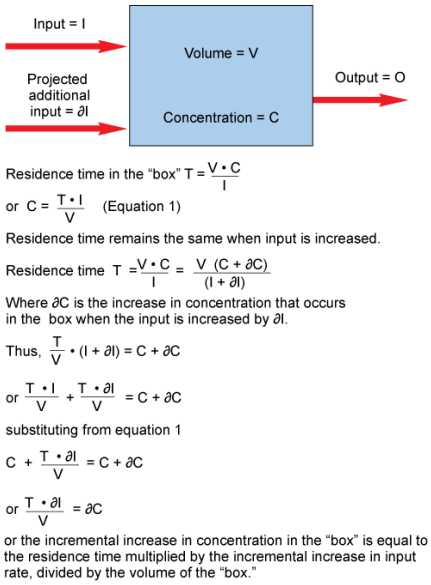


Figure CC8-2. A calculation, which is based on the residence time of a substance in an estuary or similar water mass, can be used to predict the effects of a new source of contamination that increases the input rate of the substance to a defined segment of the environment, or "box."

This finding is important because some water bodies, particularly estuaries, may have long residence times and small volumes. Such locations are not good choices for waste discharge. San Francisco Bay is a particularly good example. It has two segments: South San Francisco Bay, which has a water residence time of weeks to months; and North Bay, which has a similar volume but a water residence time of only a day or two. Inputs of treated municipal wastewater to South San Francisco Bay have caused serious water quality problems, whereas much larger inputs of treated municipal wastewater and industrial wastewater to North Bay have caused comparatively minor problems. The difference is the longer residence time in South San Francisco Bay.

Residence time calculations can be more complicated when there are multiple, separated inputs or outputs in the area of interest. For example, the calculation in **Figure CC8-2** is more complicated if a fraction of the contaminant is removed to sediments within the estuary. Nevertheless, residence time observations and calculations can be useful tools in such situations. If we can measure only the inputs and the contaminant concentrations in the estuary shown in **Figure CC8-2**, the calculated residence time will not tell us how much of the substance is flushed out to the ocean or how much is removed to the estuary sediment. We need not perform the relevant calculations here, but if we measure the inputs of the substance, its concentrations, and the residence time of water (not the residence time of the substance) in the estuary, we can calculate how much of the substance is removed with the water to the ocean and how much is retained in the estuary sediments and/or biota.

This page titled 17.8: CC8 Residence Time is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.9: CC9 The Global Greenhouse Effect

Essential to Know

- **The average temperature of the Earth's atmosphere is determined by the balance between solar energy that penetrates the Earth's atmosphere and is absorbed by gases, liquids, and solids, and energy that is lost to space by the Earth's radiation.**
- **The sun's radiated energy is concentrated in the visible portion of the electromagnetic spectrum, whereas the Earth's radiated energy is concentrated at longer wavelengths in the infrared portion of the spectrum.**
- **The Earth's atmosphere absorbs energy in the infrared and ultraviolet portions of the spectrum more effectively than it does in the visible portion of the spectrum.**
- **Because the Earth's atmosphere absorbs incoming solar radiation less effectively than it absorbs energy reradiated to space by the Earth, the Earth's climate is warmer than it would otherwise be. This is the greenhouse effect.**
- **The efficiency of absorption of infrared energy by the atmosphere is increased by higher concentrations of certain gases, including carbon dioxide, methane, chlorofluorocarbons, nitrogen oxides, and ozone, which are called "greenhouse gases."**
- **Concentrations of greenhouse gases in the atmosphere have increased and continue to increase as a result of human activities. Carbon dioxide concentrations have risen more than 50% since 1800.**
- **Radiation of heat from the Earth to space has probably been reduced and will continue to be reduced as a result of the absorption of radiated infrared energy by the increased greenhouse gas concentrations. In contrast, the amount of solar radiation passing through the Earth's atmosphere has not been reduced by such absorption. If this were the only change occurring, the average temperature of the Earth's atmosphere would increase. This is the global climate change that occurs as a result of the enhancement of the greenhouse effect by greenhouse gas releases.**
- **The Earth's average atmospheric temperature has increased by between 1.1 and 1.3°C since 1880 (the industrial era).**
- **The increase in greenhouse gas concentrations in the atmosphere and the consequent reduction in the Earth's radiative energy loss to space have many complex effects on the atmosphere, oceans, and biological systems. Many of these changes also have consequences that, on their own, would lead to warming or cooling of the Earth's atmosphere. These positive and negative feedbacks would, respectively, enhance or offset predicted global warming.**
- **Most analyses and models of this system predict that the net effect of greenhouse gas emissions will be a global warming of an additional 2°C to 6°C in the next century, but the magnitude of such warming is not certain. Indeed, the ultimate net climate change may be global or regional cooling. The only certainty is that the increases in greenhouse gases will cause global climate changes of currently unknown magnitude.**
- **The consequences of global warming would be damaging to human civilization. Possible changes include flooding of coastal cities due to rising sea level and drastic changes in rainfall patterns that would lead to the loss of agricultural land.**

Understanding the Concept

We experience temporal and spatial fluctuations of temperature in the Earth's atmosphere as weather and seasonal change. However, the average temperature of the Earth's atmosphere remains almost constant from year to year. Changes in average temperature of several degrees have occurred in the geological past (**Fig. CC9-1**). These changes, although generally small, profoundly affected the Earth's **climate**. For example, they caused **ice ages** to begin and end.

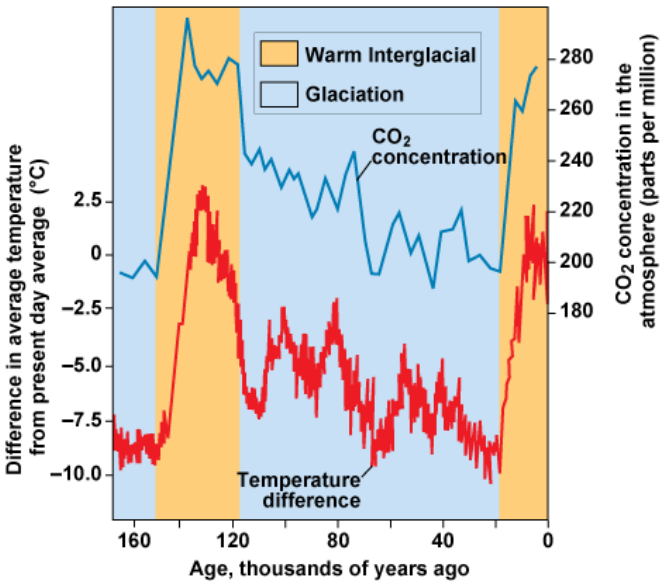


Figure CC9-1. Diagram showing the close relationship between variations in average air temperature and atmospheric carbon dioxide concentration during the past 160,000 years. These data were obtained from an ice core drilled into the Antarctic ice sheet. Notice that the Earth's average temperature at present (in geological time), is approximately at its highest level since more than 120,000 years ago. The diagram would show the post-industrial temperature change of about 1°C as only a thin line at the edge of the plot, and if the 50% CO2 concentration increase were shown, it would plot above this figure, off the top of this page.

Changes in the chemistry of the Earth's atmosphere due to human activity have already caused about a 1°C increase in the Earth's average temperature and may cause it to increase by as much as 1°C to 3°C more within the next decade or so. This predicted change is usually called the greenhouse effect or "global warming." The fear is that the Earth's average temperature could change faster than it has ever changed during human history, and that the extent of the predicted change could be great enough to have major effects on the Earth's climate. Rapid climate changes would have numerous damaging effects on nature and on humans. Consequently, there is considerable interest in determining how much of the post-industrial atmospheric temperature increase is caused by anthropogenic greenhouse gas emissions, if the increase is likely to continue and how fast, and whether the problem can be reduced by altering human activities.

The only significant source of heat energy to the atmosphere is solar radiation. Several other sources, including heat **conducted** from the Earth's interior, tidal friction, and heat released by burning **fossil fuels**, are all negligibly small in comparison with solar input. Part of the solar radiation reaching the Earth is reflected directly back to space (primarily by clouds or snow). The rest is absorbed by the land, oceans, or atmosphere. Heat is transferred between oceans and atmosphere and between land and atmosphere, and it is radiated back to space by land, ocean, and atmosphere (Fig. 7-6). This system of energy transfer is discussed in Chapter 7. One important feature is that the amount of solar energy that reaches the Earth must equal the amount of energy reflected or radiated back to space. If the amount of incoming energy does not equal the amount of outgoing energy, the Earth's atmosphere will either gain or lose heat continuously, and the atmospheric temperature will increase or decrease.

To understand how the chemistry of the atmosphere can affect the balance between incoming and outgoing radiation, we must understand the nature of **electromagnetic radiation** (Fig. 5-17). All objects radiate electromagnetic energy, but the **wavelength** range emitted by an object depends on its temperature. The hotter the object, the shorter the wavelengths of electromagnetic energy that it radiates. The sun, which is very hot, radiates much of its energy in the visible and ultraviolet wavelengths, whereas the Earth and oceans, which are much cooler, radiate energy in the much longer wavelengths of the infrared region of the spectrum.

We can understand this relationship of radiation wavelength to temperature from our everyday experience. If you turn the burner of an electric stove on at full power and hold your hand close to it, you quickly feel the heat that it radiates. You see no immediate change in the metal of the burner, because the burner is radiating all of its energy in long wavelengths in the (infrared) part of the electromagnetic spectrum, not visible to humans. As the burner heats further, it begins to glow red. At this higher temperature, it is radiating energy at wavelengths in the visible red and near-infrared part of the spectrum. If you could heat the burner even further, it would radiate at progressively shorter wavelengths until it appeared white, like the sun or the filament of an incandescent lightbulb.

Electromagnetic radiation from the sun must pass through the atmosphere before reaching the Earth's surface. Similarly, longer-

wavelength radiation emitted by the Earth, ocean, and clouds must pass through the atmosphere as it is radiated to space. You can feel the radiated heat by placing your hand near a rock or paved surface, especially immediately after dark in summer. As the longer wavelength radiation from the Earth passes through the atmosphere, it is partially absorbed by atmospheric gases, but these gases do not absorb all wavelengths of electromagnetic radiation equally. Each gas absorbs some wavelengths more effectively than others.

The **absorption spectrum** of the atmosphere and the gases that contribute the strongest absorption in parts of that spectrum are shown in **Figure CC9-2**. Compare this absorption spectrum with the **emission spectra** of the sun and the Earth (**Fig. CC9-2**). We see that the sun's radiated energy is strongly absorbed by the atmosphere at ultraviolet wavelengths, but there is little absorption in the visible region of the spectrum where the sun's radiant energy is concentrated. In contrast, there are strong absorption bands in the infrared portion of the spectrum in which most of the Earth's radiation is emitted.

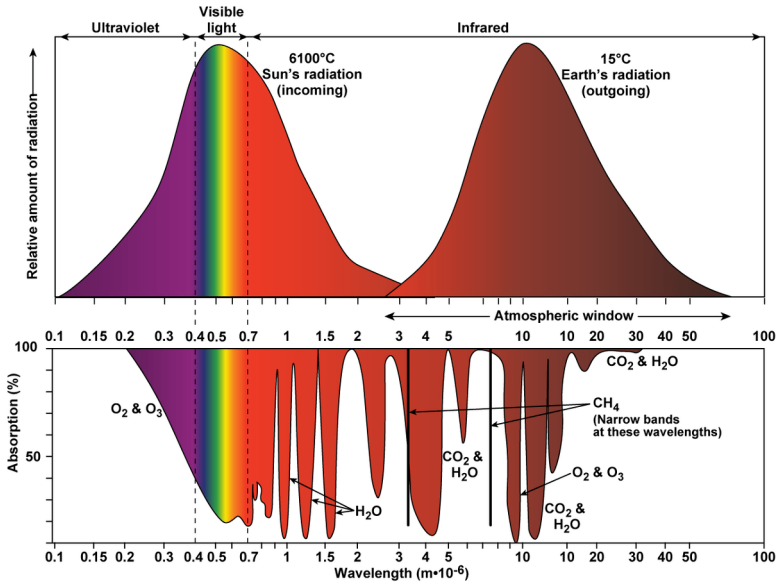


Figure CC9-2. The upper part of this figure shows the emission spectra of the Sun and the Earth. The lower part shows the absorption spectrum of the Earth's atmosphere. The sun emits most of its energy in the visible, ultraviolet, and near-infrared parts of the spectrum, whereas the much cooler Earth emits radiation at much longer, infrared wavelengths. The amount of energy that enters the atmosphere from the sun equals the amount lost from the Earth to space by radiation. The sun's radiation is partially absorbed by the Earth's atmosphere, particularly in the ultraviolet region. The Earth's radiation is also partially absorbed by the atmosphere, but at the much longer infrared wavelengths. Carbon dioxide and water vapor strongly absorb radiation at these longer wavelengths, but they do not absorb strongly at most wavelengths of the incoming solar radiation. Other greenhouse gases also absorb radiation primarily in the far-infrared wavelengths.

The atmosphere does not absorb all radiation that passes through it, even at wavelengths where the atmosphere is shown to have an absorption band (**Fig. CC9-2**). Only a fraction of the energy at such a wavelength is absorbed as it passes through the atmosphere. The rest is transmitted either to the Earth's surface or to space. The fraction absorbed is a function of the number of absorbing molecules of the gas that lie in the radiation's path as it passes through the atmosphere. If the atmospheric concentration of a gas or gases that absorb at a particular wavelength is increased, the fraction of the energy absorbed at this wavelength will increase. Consequently, the fraction transmitted to the Earth's surface or to space is decreased.

Carbon dioxide does not absorb strongly in most of the visible wavelengths that correspond to the sun's energy spectrum (**Fig. CC9-2**). However, it does absorb strongly throughout almost the entire wavelength range of the Earth's radiation. If the carbon dioxide concentration in the atmosphere is increased, there will be little effect on the quantity of solar energy passing through the atmosphere to reach the Earth's surface. However, the quantity of energy that is radiated by the Earth and atmosphere and transmitted to space will decrease. Because less energy will now escape to space, but the same amount will be received from the sun, excess energy will be retained in the atmosphere, and the global climate will warm. This process is known as the "greenhouse effect." The name *greenhouse effect* refers to the similarity of this phenomenon to the effects of the glass of a greenhouse window. Glass transmits visible light but absorbs or reflects infrared energy radiated by objects within the greenhouse, keeping the greenhouse warmer than the surrounding air.

What we normally mean when we say the greenhouse effect is actually incorrect. Anthropogenic addition of greenhouse gases should be called the "enhanced greenhouse effect" because carbon dioxide and other gases that occur naturally in the atmosphere already cause a natural greenhouse effect. The temperature at the Earth's surface would be much lower than it is now

if there were no natural greenhouse effect.

We focus on carbon dioxide as a greenhouse gas because its concentration in the atmosphere has increased more than 50% since 1800, and is now increasing at a faster rate each year (**Fig. 1-2**). This increase is the result of burning fossil fuels and deforestation (live vegetation absorbs atmospheric carbon dioxide and releases oxygen). Several other greenhouse gases may also be important, including methane, **chlorofluorocarbons** (freon and similar compounds, often called "CFCs"), nitrogen oxides, and **ozone**. Each of these gases preferentially absorbs the Earth's long-wavelength radiation more than the sun's shorter-wavelength radiation. In addition, each of these gases is released to the atmosphere in substantial quantities as a result of various human activities, including fossil fuel burning, industry, and farming. Although the quantities of carbon dioxide released to the atmosphere are much larger than the quantities of these other gases released, the other gases absorb long-wavelength radiation more effectively. Therefore, global climate change studies must also take into account these other gases.

We know that the carbon dioxide concentration in the atmosphere has increased in the past two centuries. The concentrations of other greenhouse gases have undoubtedly also increased, although we have only limited information about how much. Because more of the Earth's radiated heat energy is trapped and prevented from escaping to space while the amount of solar radiation reaching the Earth's surface has not been substantially changed, the Earth's atmosphere should be heating up. This is why the expected climate change has been called the "global warming problem." However, the atmospheric heat balance is more complicated than the simple input and output of radiated energy. Many other processes affect the balance between heat absorption and heat loss by the atmosphere, and most of these processes are, in turn, affected by changes in atmospheric temperature or chemistry.

These secondary effects can either add to the warming that would be expected from the greenhouse effect or act in the opposite direction and cause a tendency for the Earth's atmosphere to cool. Hence, they provide what are known as positive or negative **feedbacks** (often called "feedback loops") to the primary greenhouse effect. Positive feedbacks would cause faster atmospheric warming than would occur if just the simple enhanced greenhouse effect were operating. Negative feedbacks would cause slower atmospheric warming. If negative feedbacks were great enough, the net effect of the release of greenhouse gases could actually be a colder climate.

Most models of the complicated atmospheric heat balance predict that negative feedbacks will not dominate, and the net result of the enhanced greenhouse effect will be global warming. This conclusion is supported by the very good correlation in the Earth's past between high atmospheric carbon dioxide concentrations and high average climatic temperatures (**Fig. CC9-1**). However, there is considerable scientific uncertainty about the many feedback mechanisms, their magnitude, and how quickly they will respond to the extremely rapid (in geological time) increase of carbon dioxide concentrations in the atmosphere. Although most scientists believe that the release of greenhouse gases will cause global warming (estimated by models to be an increase of an additional 1°C to 6°C during the next century), this outcome is not certain. What is certain is that the release of greenhouse gases has altered the atmospheric heat balance and that the alterations will alter the Earth's climate. For this reason, scientists now refer to the issue not as the "greenhouse effect" or "global warming," but instead as "global climate change."

To see the complexity of the feedback mechanisms involving ocean-atmosphere interactions, we can briefly examine two such mechanisms, neither of which is currently well understood. First, consider a positive feedback. The oceans contribute the greenhouse gas methane to the atmosphere (**Chap. 5**). The methane is primarily a by-product of the biological activity in the oceans. The rate of methane production and its release to the atmosphere could increase if biological productivity in the oceans increased. About one-quarter of all the carbon dioxide released by human fossil fuel burning has now become dissolved in the oceans. Higher dissolved carbon dioxide concentrations may result in higher rates of **photosynthetic** growth and production in the oceans, which may lead to an increase in methane release from the oceans. Because methane is a greenhouse gas, it would add to the greenhouse effect, trapping more heat.

Second, consider a negative feedback. Clouds reflect a fraction of the sunlight that hits their upper surfaces back to space. The extent of the Earth's cloud cover is determined largely by the rate of evaporation of water from the ocean surface. The primary effect of greenhouse gases is to increase atmospheric and thus ocean surface temperatures. Higher water temperatures would lead to increased evaporation. Consequently, increasing global greenhouse gases may cause the extent of cloud cover to increase, and more solar radiation may be reflected to space to offset the additional heat trapped in the atmosphere by the greenhouse gases. This negative feedback is a particularly good example of the complexity of the climate change question, because additional clouds will not necessarily result in a greater reflection of solar heat to space. Certain types of clouds at different levels in the atmosphere are better reflectors than others. In addition, clouds not only reflect but also absorb incoming solar energy, and they both absorb and reflect the radiation outgoing from the Earth. Each of these factors may vary continuously as the rate and location of evaporation at the ocean surface and the ocean and atmospheric circulation change continuously in response to each other.

Of the many positive and negative feedbacks, some may respond almost immediately to greenhouse gas concentration changes in the atmosphere, and others may take centuries or millennia to respond to such changes. A number of the more important feedbacks involve interactions between the atmosphere and the oceans, and they are subject to intensive research because they must be understood if we are to successfully model and predict global climate change.

None of the mathematical models that were used up to about 1991 were able to include any of the relevant interactions between the atmosphere and oceans. The reasons were that too little was known about these interactions to include them in a meaningful way, and even the supercomputers that were available would have been overwhelmed by the additional computations needed to include them. As computational power and knowledge of ocean-atmosphere interactions continue to improve, and as more information about these feedbacks is included in the global climate models, the models may dramatically change predictions of the climate changes that we can expect as a result of the enhanced greenhouse effect. Nevertheless, it will be many years, or even decades, before these complex processes will be understood well enough that we can expect there to be a high degree of probability in model-based predictions of the future of Earth's climate.

Fortunately, evidence is strong that the presence of many feedbacks within a complex system tends to produce stable situations that are difficult to change. The feedbacks generally tend to return the system toward a stable configuration. However, most of the feedbacks are nonlinear responses (in other words, the response does not vary in exact proportion to the change in the stimulant). Hence, the ocean-atmosphere is a **chaotic** system (**CC11**).

One important characteristic of chaotic systems is that they tend to remain relatively stable when disturbed, until the disturbance passes a critical point at which the system suddenly shifts to a completely different, but also relatively stable, state. Possibly the beginnings and endings of ice ages are driven by such chaotic dynamics. Could it be that the enhancement of the greenhouse effect will not significantly affect the Earth's climate until a critical point is reached at which the climate will change abruptly to something as different as an ice age? Only decades of research or time can provide an answer.

The consequences of global climate change caused by the release of greenhouse gases are mostly unknown and difficult to predict. However, they are likely to be severe. For example, one effect will be the thermal expansion of ocean water as its temperature increases. The resultant sea-level rise will likely inundate many coastal cities in low-lying areas, such as Florida and some island nations. Partial melting of the polar ice sheets is likely to also contribute to sea-level rise. In addition, rainfall patterns might be changed. The bands of rainfall that now sustain the breadbaskets that feed much of the global population (**Chap. 7**) are likely to move northward, causing these regions to become deserts like those of North Africa. Because **hurricanes** are fueled by the heat energy of ocean water, another effect of a warmer atmosphere and ocean might be an increase in the intensity of hurricanes. Hurricanes might also be able to sustain damaging winds while traveling farther away from the equator than they do at present.

This page titled 17.9: CC9 The Global Greenhouse Effect is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.10: CC10 Modeling

Essential to Know

- **To understand environmental processes, we generally must create a conceptual or mathematical model of the system (or parts of the system) that describes the components of the system and their interactions. The parts are often called "cells," "boxes," or "compartments."**
- **There are many types of interaction. Examples include heat transfer between the ocean and the atmosphere, and the effect of changes in food supply on the population of an animal species.**
- **Conceptualized models must be parameterized and tested. Parameterization includes applying data to each model cell and developing equations to describe the flow of the modeled parameter between each adjacent pair of cells. These equations must be based on the physical laws that apply to each interaction. Testing consists of running the model to see how it fits the measured data. Parameters are adjusted between tests until the model fits the data set.**
- **Models must be validated by being applied to a data set that was not used in testing the model. This is often difficult to do for models for which the environmental data sets are sparse or limited to short time scales.**
- **Sensitivity analysis is performed by changing model inputs or assumptions and observing the changes in output. This type of analysis can help identify the parameters or processes that most affect the system's behavior and, thus, are important for field studies or monitoring.**
- **To facilitate model creation, interactions are often assumed to be linear. Interactions are linear if one component of a system varies in direct proportion to a change in the other. However, most, if not all, natural systems include many nonlinear interactions.**
- **The nonlinear nature of natural systems limits the degree to which models can be used to make accurate predictions of future changes in the modeled system. Models of systems that are nonlinear in nature can be used only to predict a range of likely future conditions, and these predictions become more uncertain with increasing time into the future.**

Understanding the Concept

The Earth, oceans, atmosphere, and the **biota** that inhabit these **environments** are all intimately linked in a system governed by physical laws. Each atom interacts with many others according to these laws, and it is the sum of these interactions and the atoms themselves that constitute the system. No matter how much information we obtain or how many computers we construct and program, describing this system accurately in all its details is impossible. Therefore, we must create a highly simplified system, or model, that replicates the real system in as much detail and as accurately as possible. We can create this model system either conceptually in our minds or mathematically by listing the components of the system and the mathematical relationships between each connected pair of these components.

Models that represent the entire Earth can be constructed at only a very gross level of detail. Hence, we generally construct models of parts of the system (subsystems) that we believe to be somewhat distinct from the larger system. We then represent the rest of the system by inputs and outputs to the subsystem. The descriptions of **plate tectonics**, atmospheric and ocean circulation, and marine biology, as well as other discussions in this textbook, are conceptual models of the real world.

Many different types of interactions must be described to construct models of the environment. Each model must take into account different interactions depending on which subsystem is being modeled. For example, models of atmospheric circulation must include a representation of the processes of heat transfer between ocean and atmosphere, land surface and atmosphere, and sea ice and atmosphere, and a representation of the internal **convective** processes (**CC3**) due to differential heating of air masses at different locations. Similarly, models of the changes of fish populations over time must represent the effects of varying food supply and varying abundance of predatory **species** on the growth, survival, and reproduction rates of the species involved. Models, including the two just mentioned, can be simple or extremely complicated, especially if they represent many different processes within a subsystem.

One example of a highly simplified model, and of how it can be useful, is the model describing the possible effects of changing the rate of input of a **contaminant** to an **estuary** or other body of water that is illustrated in **Figures CC8-1 and CC8-2**. At the other extreme, in coupled ocean-atmosphere global **climate** models, which are used to investigate, among other things, the

likely consequences of the greenhouse effect, the ocean is typically segmented into “cells” that are formed by division of the ocean depths into 20 to 60 horizontal layers, or slices, and subdivision of those slices into segments of 2° to 4° of **latitude** and **longitude**. The atmosphere is typically represented by up to 90 horizontal levels subdivided at 1° to 5° intervals of latitude and longitude. The result is models that can have several million to tens of millions of adjoining cells. For each interface between two cubic cells (6 per cell), the model must have equations, based either on observed data or on physical laws, that reflect the rate of transport of water, heat, water vapor, and other parameters between the two cells as the properties of the water and air in each cell change.

These massive global climate models can be run on only the fastest computers, which operate at more than a exaflop (a exaflop is a measure of computer speed and is equivalent to a billion billion or 1018 floating point operations per second). Even with these fast computers, running a model through as few as 100 years of climate can take days to months of computer time, and the output data from such a model run can consist of tens to hundreds of terabytes. A terabyte is 1024 gigabytes, or 1,099,511,627,776 (more than a million million) bytes.

The development and application of models involves three fundamental steps, each of which must be revisited and modified as needed when new information is available. The first step is to conceptualize the system under study as a series of separate but interacting cells (often called “compartments” or “boxes”). Most often, these cells are **spatial** areas, but they can be, for example, individual species within an **ecosystem**. The diagrams of the phosphorus and nitrogen cycles in **Figures 12-6 and 12-7** are models in which the cells are both spatial (positioned within the atmosphere, **mixed layer**, or deep layer) and representative of the different chemical forms (e.g., dissolved ammonia, nitrate, nitrite, dissolved organic nitrogen, nitrogen in animal tissues, nitrogen in **bacteria** and **detritus**). **Figures 12-10, 12-11, 12-12** are models in which the cells are individual species in a **food chain** or **food web**.

In developing a conceptual model, it is important to know exactly what question is being asked, so that the model can be as simple as possible. For example, if we were investigating the fate of **anthropogenic** nitrogen compounds released to the ocean, the model in **Figure 12-7** would be adequate if we were interested only in how much of the nitrogen was transported to the deep layer and how fast. However, a much more complicated (and computer-intensive) model would be needed if we were interested in how the nitrogen released from North America was distributed to each ocean basin.

The second step in the modeling process is parameterization. In this process, values of the modeled parameter (e.g., nitrogen concentration, **biomass**, or heat content) must be assigned to each cell, and equations must be written to express the flows of the parameter between each pair of adjacent cells. These flow equations can be complex if they are based on physical principles. For example, heat transfer between two ocean water cells is a combination of heat **conduction**, **diffusion**, and mixing across the cell boundary, in addition to water transport between the cells. Usually, a single equation is written for such flows on the basis of observed data, as the processes involved are too complex to express as physical laws. Often, the observed data are very limited or available only for a small number of cells, particularly in complex models such as global climate models. After parameterization, the model is tested by being run with the parameters identified to see if it reproduces the data. For example, the global climate model would be examined for how closely it reproduces changes in temperature data at locations for which real-world data are available. Next, the parameters of the model are adjusted to improve the match to the real-world data. The model may be tested and parameters adjusted in this way many times.

The third step in model development is validation to determine whether the model actually does represent the way the natural system functions. In this step, the model must be tested with a data set that is completely different from the data used for the parameterization step. This is important because the process of parameterization forces the model to fit the input data. Thus, the model parameters may represent only the input set of conditions rather than the underlying processes involved.

Validating some models is relatively easy. For example, global climate models can be parameterized with decades of climate data and then validated by the use of data from a completely different time period. This is possible because relatively detailed historical atmospheric and ocean climate data are available for a number of decades in the past. Validating other models, especially complex ecological models, is much more difficult because the amount and historical coverage of the existing data are often very limited, and parameterization requires using most or all of the available data. Models that cannot be properly validated may still be useful for many research studies, especially sensitivity analysis, which is described in the next paragraph, but they are inherently unreliable if used to extrapolate beyond the existing data set, and they cannot be relied upon for prediction.

The most familiar use of models is to predict future conditions, particularly **weather** and climate. However, the most important use of models is to determine the parameters or processes that most affect a system’s behavior and, thus, are the most important for field studies or monitoring. To do this, we subject the models to sensitivity analysis. Sensitivity analysis is a simple, albeit sometimes tedious, process of repeatedly running the model, each time making small changes in a different input parameter or combination of input parameters (the initial value in a cell, or the coefficients of the equations relating flow between cells).

This analysis can reveal which input parameters produce substantial changes in the model outputs and which produce only small changes. The former are sensitive parameters that must be accurately determined (or formulated for a flow equation) because small errors will greatly affect the accuracy of the model's representation of the real world. They are also probably the most important parameters in controlling change in the real-world system that is being modeled. As a result, the most sensitive parameters of a model are often the most important parameters to focus on in field research and monitoring programs.

Models that have been adequately tested and validated can be used as predictive tools. The best-known examples of such uses of models are the weather forecasts seen nightly on television, the predictions of the future paths of **hurricanes**, long-term (months) climate predictions that are of particular importance to farmers, and predictions of future climate change associated with the past and future releases of greenhouse gases. These predictions are all based on weather and climate models that have been parameterized, tested, and validated with extensive historical data. However, as we all know, the predictions are always somewhat general in nature and are often inaccurate.

For example, as hurricanes travel across the Caribbean, several days away from the U.S., the model predictions of the storm's path are shown as a wedge, with the future position more and more uncertain the further out in time the prediction is. The predicted range of possible landfall times and locations of the storm is narrowed as the storm approaches, but the predicted impact point often shifts and is no longer even within the range of the impact areas predicted several days earlier. In addition, predictions of strengthening or weakening of the storm as it approaches over the ocean are often incorrect, and sometimes the storm intensity suddenly and unexpectedly increases or decreases. Similar observations could be made about other model predictions.

Why are models not better predictive tools? There are several reasons. First, the models, although very sophisticated, are not the real world, and they do not account for variability within the cells of the model. Model cells often represent a large segment of the oceans or atmosphere, within which there is heterogeneity and variability that is not taken into account in the model. Second, the equations describing flow between cells are "fitted" to the testing and validation data and do not precisely describe all the factors that can affect these flows. Third, and this is the most important reason, natural systems include many nonlinear processes, and thus, these systems are **chaotic**, as discussed in more detail in **CC11**.

Even the most sophisticated models cannot make detailed and accurate long-term predictions for chaotic systems. In these systems, models will at best be able to accurately predict a range of likely future conditions. This is indeed exactly what hurricane path predictions do. These predictions are generally quite accurate for approximately a 24-h period ahead of the prediction, but they become progressively less accurate as the predictions extend out in time. Similarly, weather predictions are often quite accurate up to about 24 h ahead, but they become much less accurate farther out in time. In fact, there is a theoretical limit to future predictions in chaotic systems. Because this limit is only several days for weather predictions, these predictions will never be accurate for more than a few days into the future.

This page titled 17.10: CC10 Modeling is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.11: CC11 Chaos

Essential to Know

- **Some interactions in natural systems are linear (or almost linear over the range of values present in the environment), but many are nonlinear because the components do not vary in direct proportion to each other. For example, doubling the food supply may cause the population of an animal species to double if food is scarce, but it may result in little change in population if food is already abundant.**
- **Linear systems are predictable. If we increase one parameter (e.g., the available food supply), the system will adjust to equilibrium (the population will rise to a new stable value that is related in direct proportion to the new food supply).**
- **Nonlinear systems behave unpredictably. Changing one parameter may lead to a new equilibrium (which may be drastically different from the original system) or may cause the system to oscillate in a chaotic way.**
- **Complex nonlinear systems with many interactions are chaotic and unpredictable, but they tend to vary irregularly within an identifiable range of conditions. For example, weather is inherently unpredictable, but climate (range and average conditions) does not generally change from year to year.**
- **Complex nonlinear systems may respond to unusual changes in one or more parameters by "jumping" to oscillate around a different set of average conditions.**
- **Natural systems include many interactions, at least some of which are nonlinear. All Earth environmental processes including, atmosphere, and ocean processes are therefore chaotic.**
- **Because natural systems are chaotic, their exact state cannot ever be forecasted (predicted) accurately. These systems are inherently variable, and they may undergo sudden changes to new average conditions.**

Understanding the Concepts

In most cases, we conceptualize the world around us as behaving according to simple rules. For example, it seems reasonable that if we doubled the temperature difference between ocean water and the overlying air, we would find that the rate of heat transfer between the two also doubled; or if we doubled the rate of food production, we would find that the animal **species** eating this food doubled in population. These are linear relationships in which one parameter changes in direct proportion to another. To be linear, relationships need not be one-to-one (as just described). The only requirement is that a certain amount of change in one parameter will always result in the change of another parameter by the same amount, even if one parameter increases and the other decreases (**Fig. CC11-1**).

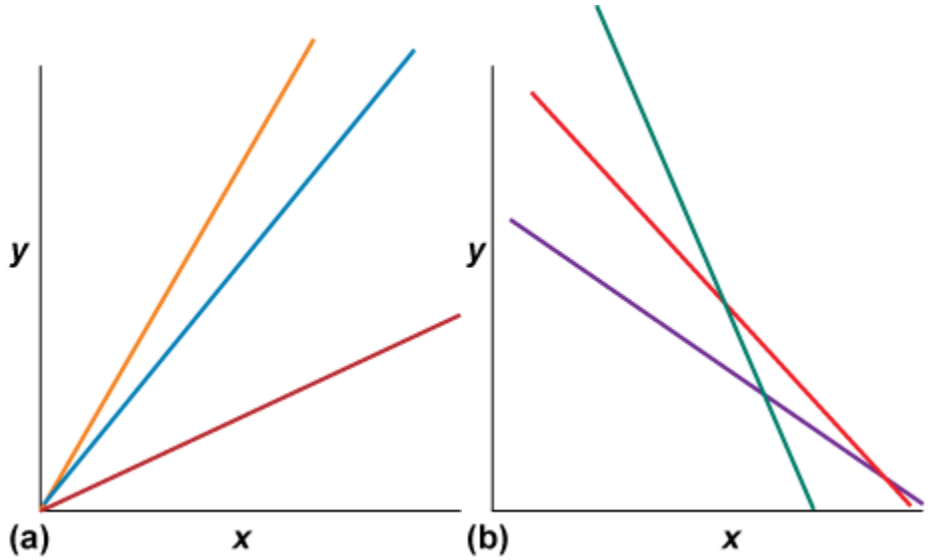


Figure CC11-1. Linear relationships can be depicted by the straight line on a graph that plots one variable (x) against another (y). The value of one variable may (a) increase with an increase in the other variable, or (b) decrease with an increase in the other variable.

Linear relationships are simple to conceptualize and to express mathematically. They are also intuitively attractive. For example, the hypothesized linear effect of changing food production rate on a consumer population seems perfectly reasonable at first. However, it does not represent the true behavior of the systems. To understand why, think about what would happen if we were to drastically increase or decrease the food supply for an animal species that has no predators. If we drastically increased the food supply, the population would increase, but at some level, the species would become so crowded that some other factor (e.g., breathing space) would prevent it from increasing further. If we drastically decreased the food supply, the population would decline. Eventually, it would reach such a low level that the species would not be able to reproduce successfully because, for example, the survivors would have to spread so far apart to obtain sufficient food that there would be too few available mates. Because the hypothesized relationships do not maintain proportionality if the value of one parameter is changed too much, they are nonlinear.

Examples of nonlinear relationships are depicted in **Figure CC11-2**. Many, if not all, of the relationships that control the real world are nonlinear, which critically affects our ability to model the real world. Fortunately, many relationships, such as the one just discussed, may be linear or nearly so over a wide range of values of the two parameters (**Fig. CC11-2**). For instance, the food–population relationship discussed earlier may be nearly linear within the range of food production rates that are observed in the real world. In these circumstances, we can use linear relationships in our models and the models will be similar enough to the real world to be useful.

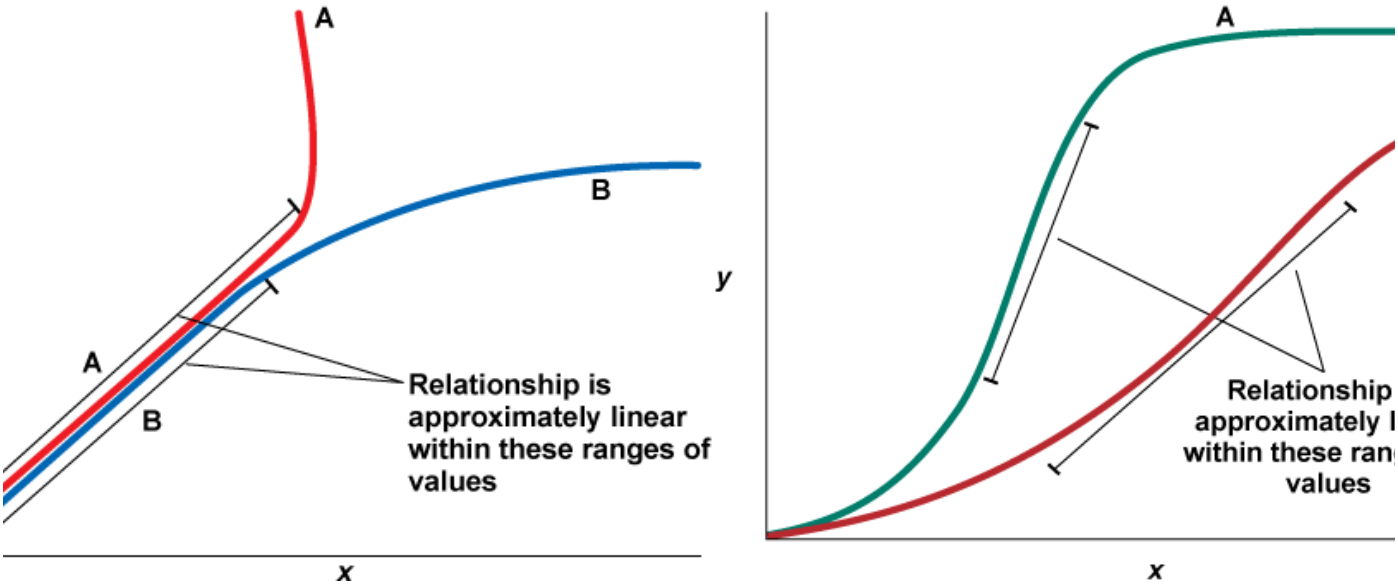


Figure CC11-2. Nonlinear relationships can be identified by the curvature of a graph that plots one variable (x) against another (y). Although the curvature can be complex, there is often a range of two values within which the relationship is approximately linear.

Not until the 1970s did scientists begin to examine the behavior of mathematical models that incorporate nonlinear relationships. What they found was startling. Models that use only linear relationships, no matter how complicated, always reach a new equilibrium if one of the parameters is changed (unless parameters are constantly changing). However, even the simplest of nonlinear models involving only two parameters may never reach an equilibrium. The system may oscillate in a seemingly random, or **chaotic**, way, never exactly repeating itself.

The mathematics of nonlinear systems can be complicated. Fortunately, we can understand chaotic behavior by considering a simple system called the "Lorenzian waterwheel," named for Edward Lorenz, who first investigated the chaotic behavior of this system (**Fig. CC11-3**). In this waterwheel, the buckets are pivot-mounted so that they always stay upright (like the chairs in a Ferris wheel). Each bucket has a small hole through which water can escape (but not drip into a bucket below). This system can be modeled by three simple but nonlinear equations, and it has also been constructed and tested as an actual physical model.

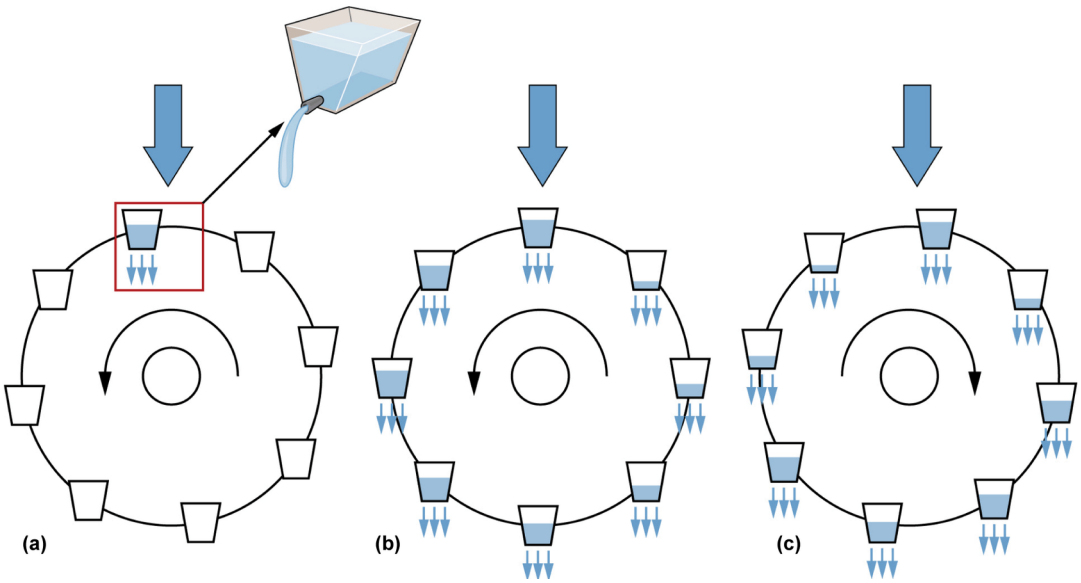


Figure CC11-3. The Lorenzian waterwheel. (a) If water is poured into the top bucket faster than it can escape from the bucket, the wheel will be set in motion. (b) If the water flow to the buckets is maintained, the wheel may settle into an equilibrium in which it rotates at constant speed. This can happen when, for example, the buckets are just empty by the time they reach the top and are refilled. (c) The acceleration and speed of the wheel depend on the amount of water in each bucket, which depends on the length of time that each bucket spends under the fill point, which depends on the speed of rotation of the wheel. These are nonlinear relationships, and the wheel may never settle into a constant rotation rate, even if the water input rate remains constant. Furthermore, the rotation rate may vary constantly, and the direction of rotation may even be reversed.

If we pour water into the top bucket, the waterwheel's behavior depends on how fast the water is poured, how fast each bucket empties, and how quickly the wheel turns (which determines how long each bucket is in the right part of the rotation to be filled).

What do you think happens if we pour water into the Lorenzian waterwheel at different rates but leave everything else unchanged? If we pour the water in at a very slow rate, we might expect it to escape from the top bucket fast enough that the bucket does not fill and the waterwheel will not turn. This is indeed what happens. If we pour the water faster, the top bucket will partially fill and set the wheel in motion, bringing the next bucket under the pouring point to fill in turn and continue the rotation (**Fig. CC11-3b**). As each bucket moves through the rotation, it will progressively empty. We might expect that, if we maintain a stable rate of filling, the rotation will be smooth and the rate of rotation will increase as the fill rate increases. This is what happens when the fill rate is relatively slow. However, at faster fill rates something interesting happens.

Consider the wheel rotating relatively slowly. One or more successive buckets fills almost to the top as the wheel rotates. Because these buckets are heavy, they accelerate the rotation as they move toward the bottom point of the wheel, causing buckets behind them to fill for a shorter time. The full buckets can pass through the low point in the wheel before they have emptied, whereas the following buckets are filled less and empty completely. Once past the low point in the rotation, a once-full bucket (now partially full) may be heavier than the once partially filled buckets (now empty) behind it, so it now tends to slow the rotation of the wheel (**Fig. CC11-3c**).

We can see that the rotation rate of the wheel will vary constantly, and the direction of rotation may even reverse. However, what is most surprising about the motion is that, even though we do not vary the fill rate, the wheel's rotation does not, as we might intuitively expect, settle into a regular back-and-forth oscillation. Instead, the rotation varies continuously back and forth in a pattern that is never repeated, a "chaotic" situation. This happens because of the nonlinearity in relationships, such as the relationship between the rate of rotation and the amount of filling of each bucket.

Because natural systems involve many nonlinear relationships, they can act like the Lorenzian waterwheel: sometimes they can appear stable or change smoothly to a new equilibrium if one of their important parameters changes, or they can oscillate chaotically. For example, a **convection** cell is similar to the waterwheel, although inverted. If heat is supplied very slowly to the lower part of a fluid and lost slowly from the top, the heat can be transferred by **conduction** within the fluid and there is no movement. At somewhat higher heating or cooling rates, smoothly flowing convection cells may form (**CC3**), and at even higher rates the convection cell motion becomes **turbulent** and chaotic.

Even seemingly simple biological systems can exhibit this type of behavior. For example, as discussed previously, populations

respond in a nonlinear way to changes in food availability. If food supply increases, then population increases accordingly, but as population continues to increase, other factors eventually limit and reduce the population. Depending on the nature of these relationships, an increase in food supply can lead to a new, higher population of the consumer species, chaotic variations of this population, or even collapse and extinction if, for example, a small increase in population drastically reduces reproduction rates because of overcrowding.

Natural systems generally depend on many interactions, some of which are nonlinear. As a result, most natural systems are chaotic, and this fact has many implications for science. Since the 1970s, a new scientific subdiscipline has developed to study chaos. The term chaos has now generally been replaced by complexity. Complexity is often the preferred term because certain characteristics of complex nonlinear systems can be deduced that enable us to understand their behavior and how their component parameters might vary under certain circumstances, even though the precise future status of the system cannot be predicted. We do not need to fully understand the behavior of nonlinear systems unless we become research scientists, but several of their characteristics are important to all of us.

Nonlinear system models are extremely sensitive to very (infinitely) small differences in initial conditions. This characteristic was discovered by Edward Lorenz when he was using a computer model of **weather** patterns, and these observations are what started the systematic study of chaos. Lorenz found that if he started his mathematical model several times from the same set of input parameters (e.g., temperature and **humidity** distribution), the predicted weather exhibited variations that, although complicated, were reproduced exactly each time the model was run.

Lorenz then decided to start his model at a time interval later than the starting point of his previous model runs. He started this new set of calculations with the values of input parameters that his previous model runs had predicted would occur at this time interval. Unexpectedly, he found that his model quickly began to predict weather patterns that differed from those that the model had predicted for this same time period when the model had been started at the earlier time (**Fig. CC11-4**). The reason for this difference, Lorenz found, was that the nonlinear equations in his model generated completely different weather patterns when the later-start model run used input data that were rounded from the values produced in the earlier-start model runs. When he ran his model from the beginning, the computer maintained each parameter's value at each time interval to six decimal places (e.g., 0.267902). When he started his model in the middle, he rounded the input data to three decimal places (e.g., 0.268).

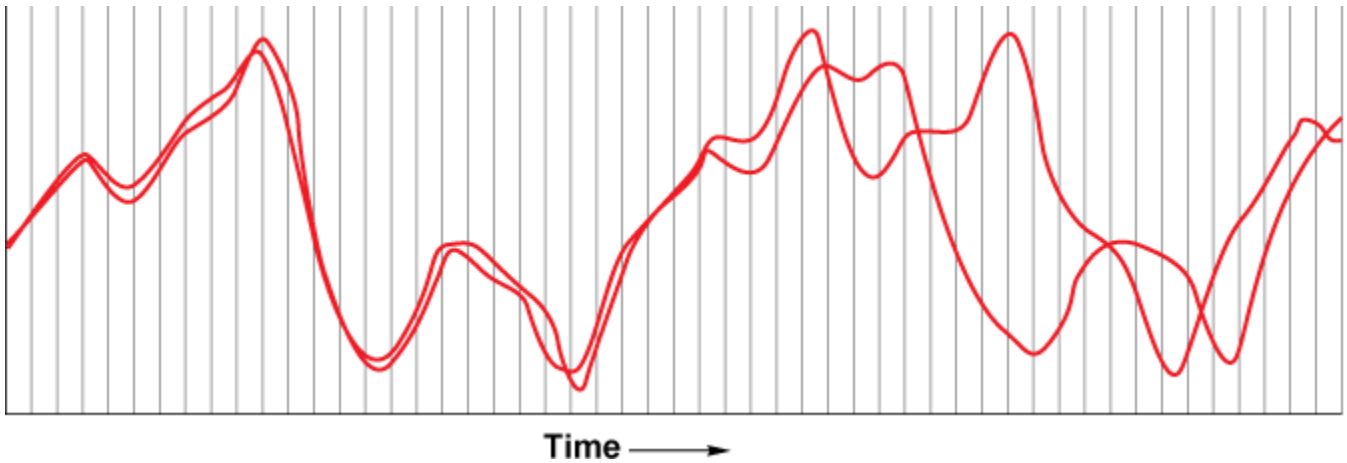


Figure CC11-4. These two plot lines depict the results obtained from the same nonlinear mathematical model that simulates weather patterns. The only difference between the two model runs is that the starting-point data were rounded to a smaller number of decimal places for one model run than for the other. The patterns grow farther and farther apart until they lose any resemblance to each other. Many natural systems show this sensitive dependence on initial conditions. As a result, it is impossible to measure and model such systems well enough to predict their precise future behaviors.

This sensitive dependence on initial conditions has often been called the "butterfly effect." This term originated from the example that a butterfly stirring the air one day in Beijing could alter the storm systems in New York the next month. As a consequence of the butterfly effect, models of nonlinear systems can never predict future behavior exactly. In addition, the further into the future a model predicts, the less likely it is to be accurate. For example, weather forecasts will never be much better than they are today. We will never be able to accurately predict the exact time or place a thunderstorm will occur on the next day, nor predict with certainty whether it will rain a few days in the future. Similarly, we will never be able to forecast or predict the exact future state of any complex nonlinear system, including **plate tectonic** movements, **ecosystem** dynamics, or fish and wildlife population variations.

Fortunately, although many complex nonlinear systems never reach equilibrium, they do tend to oscillate chaotically within a range of conditions (values of the component parameters) that models can predict. For example, weather is chaotic and unpredictable, but **climate** (the average and range of temperature, rainfall, etc., at a specific location and time of year) can be predicted with some confidence because it changes little from year to year. Hence, models of complex systems (**CC10**), if they adequately resemble the real world, can be used with reasonable success to predict the average conditions and range of variations that will occur in the future.

Unfortunately, although complex nonlinear systems do tend to oscillate chaotically within a predictable range of conditions, this dynamic "equilibrium" can be disturbed by small changes in component characteristics. In some instances, if a critical value of one or more components is changed, the system can "jump" from one set of average conditions to a completely different set of average conditions around which it oscillates chaotically. Thus, a system may appear to suddenly change drastically, even though none of the components of the system changed substantially before the jump occurred. For example, a very small increase in a parameter, such as the average temperature of the oceans and atmosphere, could have little effect on climate until a critical point is reached. At that point, the Earth's climate could suddenly become much warmer or colder. Recognition of this concept has led to the now well-known term "tipping point".

We now know that this type of sudden climate change has occurred in the past (**Chaps. 8**, **9**, **Fig. 7-21**). Similarly, sudden changes in ecosystems, such as drastic declines or blooms in certain species, may be a natural consequence of nature's nonlinearity. Sudden changes in other complex systems, such as ocean circulation and the motions of tectonic plates, may have occurred in the past for this same reason. It is important to realize that sudden changes in natural systems have occurred in the past and will occur again in the future. Furthermore, no matter how well we model such systems, it will be very difficult to reproduce these changes faithfully, and virtually impossible to develop an accurate predictive capability that will alert us to them before they occur.

This page titled 17.11: CC11 Chaos is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.12: CC12 The Coriolis Effect

Essential to Know

- **When set in motion, freely moving objects, including air and water masses, move in straight paths while the Earth continues to rotate independently.**
- **Because freely moving objects are not carried with the Earth as it rotates, they are subject to an apparent deflection called the "Coriolis effect." To an observer rotating with the Earth, freely moving objects that travel in a straight line appear to travel in a curved path on the Earth.**
- **The Coriolis effect causes an apparent deflection of freely moving objects to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The deflection is said to be *cum sole*, or "with the sun."**
- **The Coriolis deflection is greatest at the poles and decreases at lower latitudes. There is no Coriolis effect for objects that move directly east to west or west to east at the equator.**
- **Regardless of their speed, freely moving objects at the same latitude appear to complete a circle and return to their original location in the same period of time. This period, called the "inertial period," is 12 h at the poles and increases progressively at lower latitudes. The inertial period is 24 h at 30°N or 30°S, and it approaches infinity near the equator.**
- **For objects moving within 5° on either side of the equator, the Coriolis effect can often be ignored because the deflection is small and the inertial period is very long.**
- **Freely moving objects moving at the same speed appear to follow circular paths with smaller radii at higher latitudes. Freely moving objects at the same latitude appear to follow paths with larger radii if they are moving at higher speeds.**
- **The magnitude of the Coriolis deflection (the rate of increase of distance from a straight-line path) increases with increasing speed.**

Understanding the Concept

From common experience, we know that freely moving objects, such as bullets, move in a straight line unless acted on by an external force. However, all is not what it seems. If we very carefully examine the bullet's flight, we find that its path is deflected very slightly to one side (in addition to the downward deflection due to **gravity**). Curiously, this deflection is always to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. This apparent deflection, called the **Coriolis effect**, occurs because we live on a rotating Earth, and it occurs for all objects that move freely without connection to the solid Earth surface, including bullets, thrown footballs, and moving air and **water masses**.

Some simplified examples help to explain how the Coriolis deflection works.

Merry-Go-Round without Gravity

Consider a merry-go-round rotating counterclockwise when viewed from above (**Fig. CC12-1a**). This is equivalent to the Earth seen from above the North Pole (**Fig. CC12-1c**). Imagine that you are an observer sitting on the merry-go-round and that someone at the center of the merry-go-round throws a ball outward. As the ball travels through the air in a straight line, the merry-go-round continues to turn, carrying you around with it (**Fig. CC12-1a**). You do not see the ball travel in a straight line. Instead, the ball appears to travel in a curve deflected to the right (**Fig. CC12-1b**). If the merry-go-round spins clockwise (equivalent to viewing the Southern Hemisphere from above the South Pole; **Fig. CC12-1d**), the apparent deflection of the ball is to the left (**Fig. CC12-1e**).

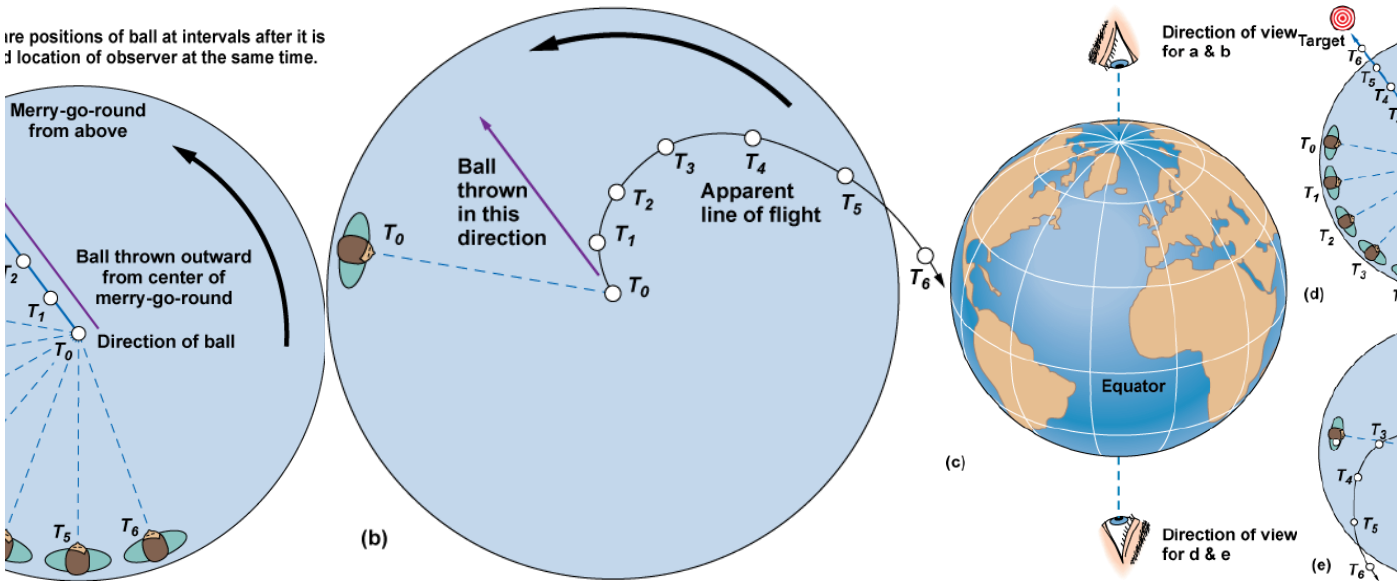


Figure CC12-1. (a) A ball thrown outward toward a target from the center of rotation of a merry-go-round has zero orbital velocity due to the merry-go-round's rotation, and will travel outward in a straight line (compared to the ground) to hit the target. However, an observer sitting away from the axis of rotation moves around with the merry-go-round such that the observer's angle of vision to the ball and target changes. (b) If the observer is not aware of the merry-go-round rotation (just as we are not aware of the Earth's rotation), the observer will see the ball appear to move in a circular path, curving to the right. To observe this phenomenon, rotate part (a) to place the observer at the observing location in part (b) for each of the time intervals T1, T2, and so on. (c) Looking down on the merry-go-round depicted in parts (a) and (b) is equivalent to looking down on the Earth from above the North Pole. If we were to look down from above the South Pole, the merry-go-round would appear to rotate clockwise instead of counterclockwise. (d) If the merry-go-round rotates clockwise (the equivalent of the Southern Hemisphere), the ball still moves in a straight line. However, (e) the apparent deflection is in the opposite direction, and the observer sees the ball curve to the left.

If you sit on the merry-go-round away from its center of rotation and throw a ball from there at a target, the ball again appears to be deflected to the right. However, it misses the target, even though it flies in a straight line. The reason is that the ball was moving with the merry-go-round when it was thrown. While in the observer's hand, the ball is constrained to move in a circle with the merry-go-round, but when released, it is no longer constrained. In fact, if the ball were simply released on the merry-go-round instead of thrown, it would roll off in the direction it was moving at the instant it was released. The same thing occurs when we twirl an object on the end of a piece of string and then let go. The object flies off in a line whose direction is determined by where it was in its circle of rotation when it was released (**Fig. CC12-2a**). If a ball is thrown from a rotating merry-go-round, it is given straight-line motion by the throw. However, because of the rotation of the merry-go-round, it is already in motion in a direction tangential to the ball's circle of rotation at the point it was released. These two straight-line motions (components) are combined, so the ball's actual direction is a straight line between the directions of the two components (**Fig. CC12-2b**). The ball's direction and speed are easily calculated from the two components, which are called "vectors" (**Fig. CC12-2c**).

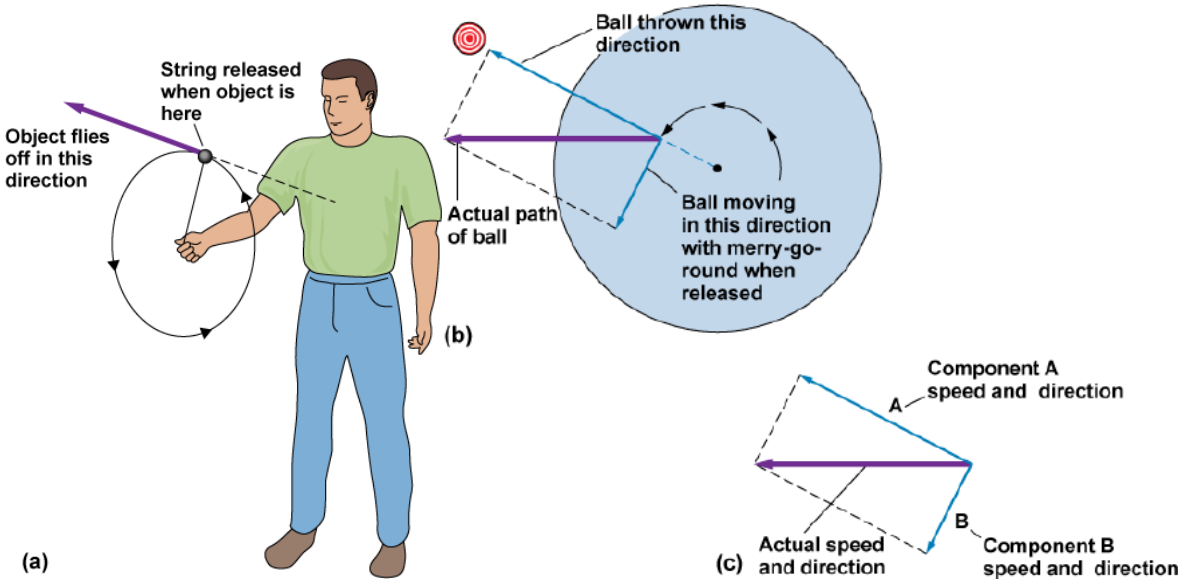


Figure CC12-2. (a) A centripetal force is needed to keep an orbiting object in its orbit. When an object is twirled on the end of a string, we supply this force with our muscles through tension in the string. If we release this tension by letting go of the string, the object will fly outward. (b) When a ball is thrown outward from a rotating merry-go-round, it flies off in a straight line (just as the object twirled on the end of a string does), but it also has an additional velocity (speed and direction) imparted to it by the throw. The actual velocity with which the ball moves is a combination of the two velocities. (c) We can determine the ball's actual velocity by geometrically combining its two component velocities. The lines A and B are drawn in the respective directions of the two velocities. The length of each line is proportional to the speed imparted in that direction. The actual direction of the ball's motion can be determined by drawing lines (the dotted lines) parallel and equal to each arrow from the head of the other to form a parallelogram (in this case, it is a rectangle). The diagonal of this parallelogram shows the actual direction of motion, and its length shows the speed.

Centripetal Force

Spinning objects, like the object on the end of a twirling string, fly off if released. This tendency is often called "centrifugal" force. However, centrifugal force does not exist. Circling objects tend to fly off when released because the force restraining them in their circular path is no longer present. A force that acts toward the center of rotation, called **centripetal force**, must be applied to maintain any orbiting object in its rotating orbital path (which can be circular or elliptical). This force may be exerted in several ways. For example, **friction** between the observer and the merry-go-round keeps the observer from flying off, whereas tension in the string prevents a twirling object from leaving its circular path. Earth's gravity is the centripetal force that prevents us from flying off the Earth.

Centripetal force must increase as the rate of rotation increases. For example, the amount of friction needed to keep an observer on the merry-go-round becomes greater as the merry-go-round speeds up. The rate of rotation is measured by the angular velocity (the rate of rotation measured in degrees of angle per unit time; **Fig. CC12-3a**). For a constant angular velocity, orbital velocity increases at points more distant from the center of rotation (**Fig. CC12-3a**). At a fixed distance from the center of rotation, angular velocity increases if orbital velocity increases, and decreases if orbital velocity decreases (**Fig. CC12-3b**).

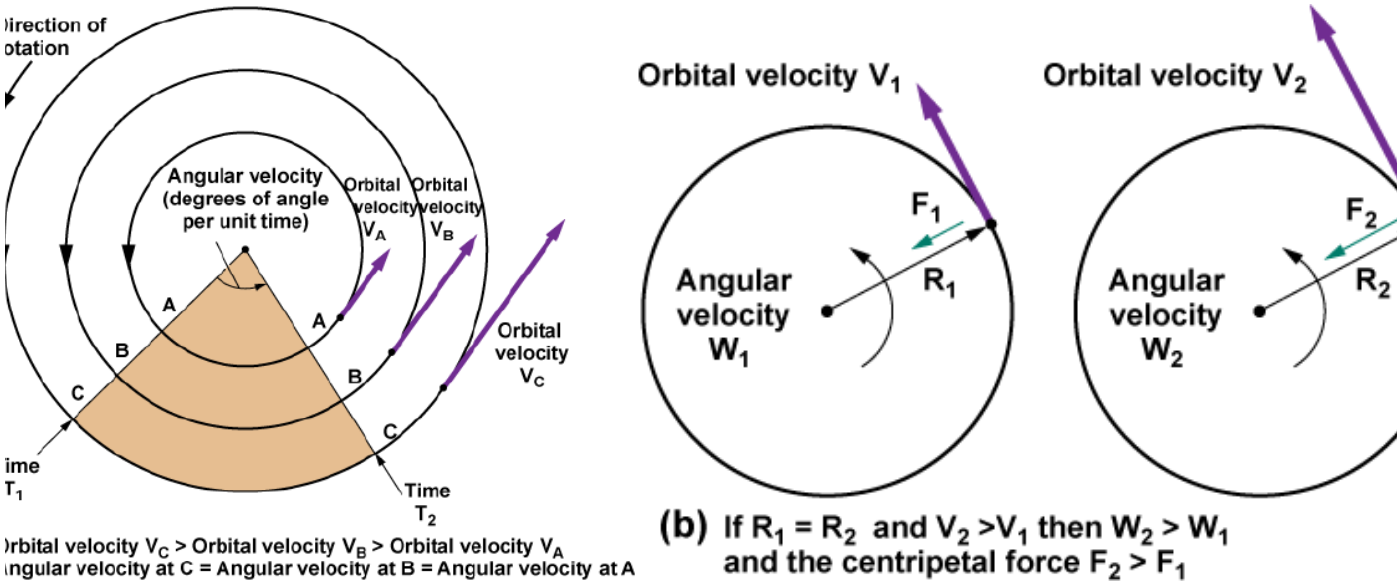


Figure CC12-3. The velocity of an orbiting object can be expressed as either angular velocity or orbital velocity. (a) The angular velocity is the number of degrees of angle moved through in unit time. The orbital velocity is the distance the object moves along its orbital path per unit time. On a rotating solid object (like a merry-go-round or the Earth), the orbital velocity of a point increases with distance from the center of rotation, and the angular velocity is the same for all points regardless of location. (b) Angular velocity, orbital velocity and centripetal force are related. For two objects at the same distance from the center of rotation, if the angular velocity (rate of rotation) is increased, the orbital velocity is also increased and a larger centripetal force is needed to maintain the object in orbit. For example, the tension in a string becomes greater as we twirl it and its attached object faster.

Merry-Go-Round with Gravity

Now imagine a merry-go-round with gravity that acts toward a point located below its center of rotation (**Fig. CC12-4a**). This merry-go-round is equivalent to a small circular area of the Earth's surface centered at the North Pole with gravity acting toward the Earth's center (**Fig. CC12-4b**).

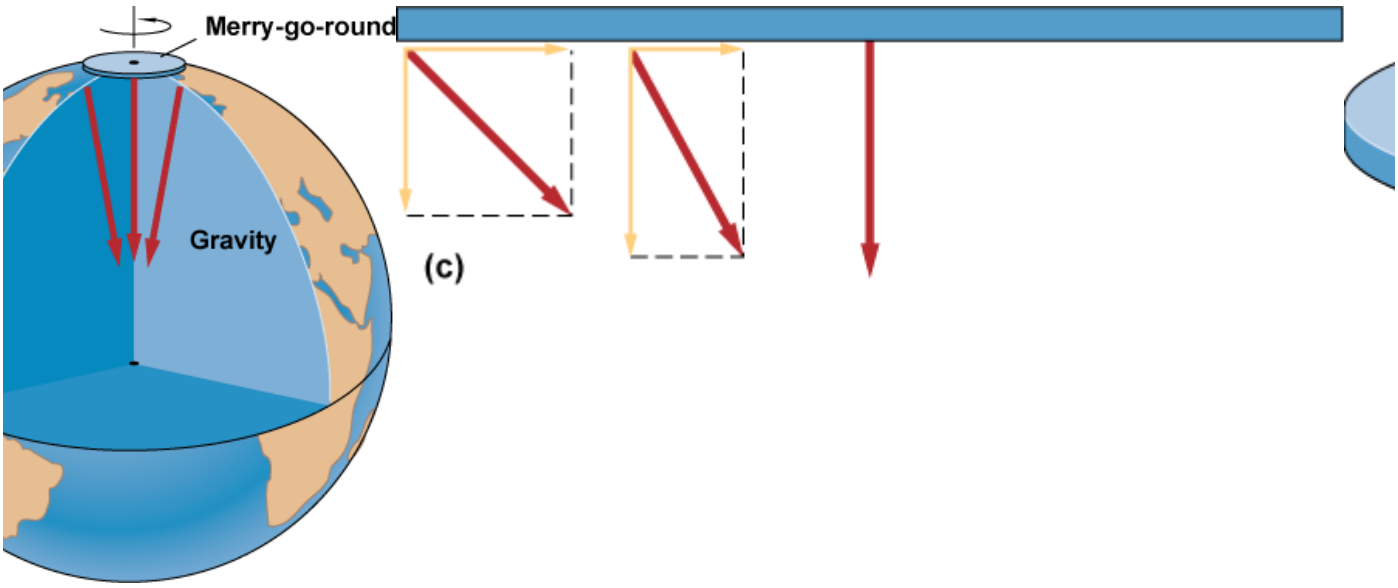


Figure CC12-4. (a) A hypothetical merry-go-round with a gravitational force acting toward a point beneath its center. This hypothetical merry-go-round is like a section of the Earth near the pole (b). However, in this example, the distance to the center of gravity is greatly reduced to help us understand the effects of changes in the directions of gravity and centripetal force with latitude on the Earth's surface. (c) The gravitational force is the same strength (or very nearly so) at all points on the merry-go-round, but it is directed vertically down only at the center of rotation. At all other locations, it can be resolved into two components (the yellow lines), one parallel to the merry-go-round surface and another vertically downward. The component parallel to the surface increases in strength with increasing distance from the center of rotation. (d) If we place a ball on the surface of the merry-go-round, hold it in place so that it rotates with the merry-go-round, and then release it, it will tend to fly outward like an object twirled on a string. However, there is a component of gravity that acts toward the merry-go-round's center. This component provides a centripetal force that, if exactly balanced with the rotation rate of the ball on the merry-go-round (its orbital velocity), will prevent the ball from flying outward.

If a ball is placed at any point on the merry-go-round other than the center, a small component of the gravitational force pulls it toward the center of rotation (**Fig. CC12-4c**). If the merry-go-round and ball are rotating and the ball is released to roll freely, the ball will move outward, away from the center of rotation, just as the object on the end of a twirling string flies outward if released. However, except at the center, the component of gravity that acts toward the center of rotation (**Fig. CC12-4c**) pulls the ball toward this center. If the ball is at the correct distance from the center of rotation, the component of gravitational force parallel to the merry-go-round surface provides exactly the centripetal force needed to balance its rotation with the merry-go-round. Thus, the ball remains in this position and rotates with the merry-go-round (**Fig. CC12-4d**).

If we set the ball in motion with respect to the merry-go-round, this balance will be disturbed. For example, if the ball is moved toward the center of rotation (on the Earth, equivalent to moving it toward the pole), the component of gravity parallel to the merry-go-round surface is reduced (**Fig. CC12-4c**). Because this component of gravity is now less than the centripetal force needed to maintain the ball in its smaller orbit, the ball tends to move back away from the center of rotation. If the ball is moved away from the center of rotation, a small additional gravitational attraction tends to pull it back toward the center.

Next, imagine that the ball is moved in the direction of rotation (equivalent to moving from west to east on Earth). This motion is in the same direction as the ball's path as it rotates with the merry-go-round, so the ball's orbital and angular velocity are increased, and a larger centripetal force is needed to keep the ball in this faster orbit. However, the gravitational force component remains unchanged. Therefore, the ball tends to move outward. Similarly, if the ball is moved in a direction exactly opposite that of the merry-go-round (east to west), the angular velocity is decreased, and the ball tends to move inward. Thus, motions of the ball directed with or against the direction of rotation (east to west or west to east) are turned away from or toward the center of rotation, respectively.

No matter which direction the ball is started in motion, it will move either toward or away from the center of rotation, and this movement will be counteracted by an imbalance between gravity and centripetal force. The counteracting force will grow as the ball moves farther inward or outward. This force will slow the ball's inward or outward motion and eventually reverse its direction. The ball will oscillate toward and away from the center of rotation (**Fig. CC12-5a**) in a motion similar to that of a pendulum (**Fig. CC12-5b**).

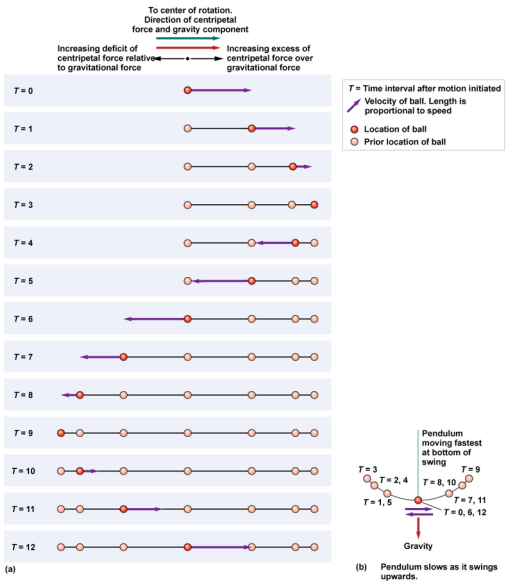


Figure CC12-5. (a) If a ball is set in motion toward the center of a hypothetical merry-go-round with a gravitational force that acts toward a point beneath its center (see **Figure CC12-4**), the ball will oscillate back and forth. As it moves toward the center, the component of gravity in this direction decreases while the required centripetal force increases (because the ball retains its orbital velocity but is now in a smaller-radius orbit). The imbalance between centripetal force and gravitational attraction slows and then reverses the ball's direction of motion. The ball is accelerated until it passes through its original location relative to the center of rotation, but it has enough momentum to continue outward. As the ball moves farther away from the center, the component of gravity increases while the required centripetal force decreases. As a result, there is an excess of gravitational force over centripetal force, and the ball is slowed, and its motion eventually reverses back toward the center of rotation. (b) These oscillating motions are very similar to the movements of a pendulum.

As the ball oscillates back and forth, it is still moving in its circular orbit with the merry-go-round. We can show mathematically that the ball would always go through one complete oscillation from its extreme innermost point outward and back to this innermost point in exactly the time it takes for the merry-go-round to complete one revolution. You need not be concerned with this calculation, but the result is important because we can now see what the path of the ball set in motion would be (**Fig. CC12-6**). To an observer anywhere on the rotating merry-go-round, the ball would appear to travel in clockwise circles. It would complete a circle twice for each time the merry-go-round completed one revolution (the equivalent of two circles every 24 h on the Earth).

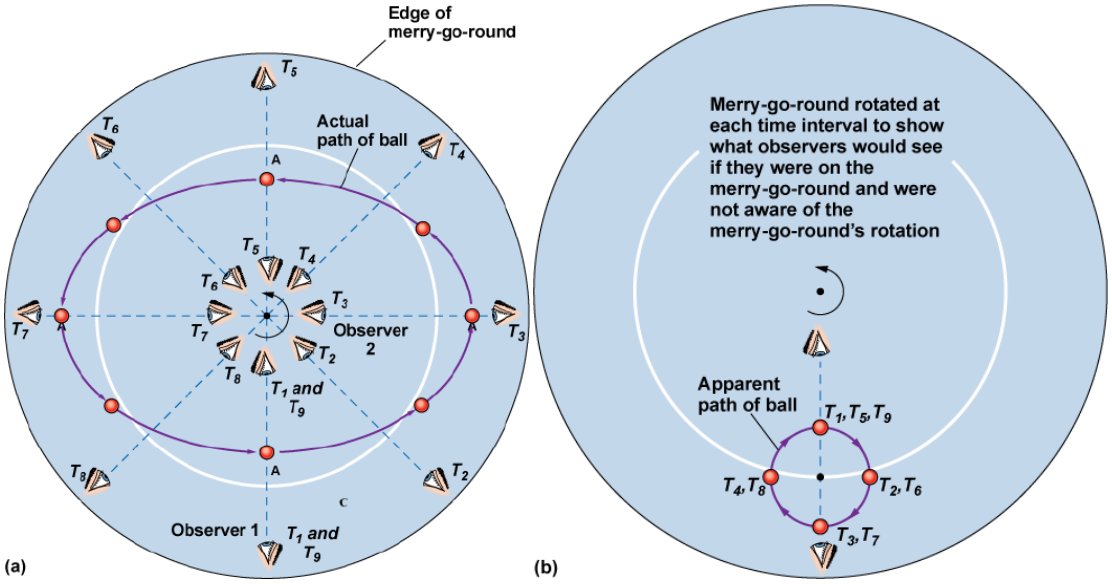


Figure CC12-6. (a) If a ball is set in motion toward the center of a hypothetical merry-go-round with a gravitational force that acts toward a point beneath its center (see **Figure CC12-4**), this motion is added to the orbital motion that the ball already has due to the rotation of the merry-go-round. The added motion and the interaction of centripetal force and gravity cause the ball to oscillate back and forth toward and away from the center, as shown in **Figure CC12-5**. When the rotational motion is added, the actual path of the ball on the merry-go-round is an ellipse, and it returns to its original position after exactly one revolution. (b) To an observer at any location on the rotating merry-go-round, the ball appears to move in a circle. Follow the ball's direction and distance from each observer's eye in part (a) of this figure and see how they plot in a circle in part (b).

The Earth, Centripetal Force, and Gravity

The Earth is more complicated than a merry-go-round because it is a sphere, although not quite an exact sphere. Gravity acts toward the Earth's center, whereas centripetal force acts toward the center of rotation (perpendicular to the Earth's axis of rotation). Therefore, gravity and centripetal force act in the same direction only at the equator, and the difference between their directions increases from zero at the equator to 90° at the poles (**Fig. CC12-7a**).

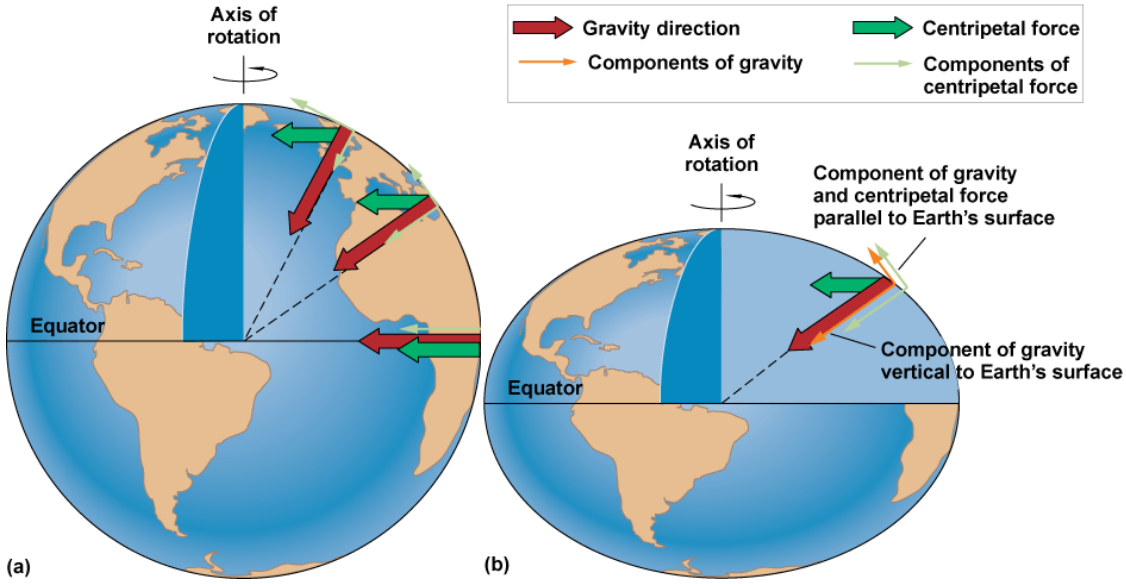


Figure CC12-7. (a) If the Earth were exactly spherical, there would be no component of gravity parallel to the Earth's surface at any latitude. However, to keep an object in orbit at the Earth's surface, there would have to be a component of a centripetal force parallel to the Earth's surface, everywhere except at the equator. Because there would be no gravitational force component to provide this component of centripetal force, an object anywhere on the Earth's surface except at the equator would move into a wider orbit (in which the required centripetal force was lower). Thus, the object would slide toward the equator. (b) In response to the force imbalance that would be created if the Earth were spherical, the Earth and all other planets that were at one time totally molten or gaseous were not formed as perfect spheres. Instead, the Earth is an oblate spheroid, a squashed spheroid shape, as shown (but much exaggerated) in the figure. Because the Earth is an oblate spheroid, gravity does not act exactly perpendicular to the Earth's surface, except at the equator and poles. Consequently, at all points other than these, there is a small component of gravity parallel to the Earth's surface that balances the required centripetal force for an object at that point.

Centripetal force can be resolved into two components: one parallel to the Earth's surface oriented north–south, and the other oriented toward the Earth's center (Fig. CC12-7a). The centripetal force needed to keep an object on the Earth's surface, and variations in this force needed to balance even very large changes in an object's speed relative to the Earth's surface, are extremely small in comparison with gravity. Therefore, variations in the component of centripetal force that acts toward the Earth's center (the same direction as gravity) are easily compensated by gravity and the **pressure gradient** (Fig. CC12-8). In other words, moving objects whose angular velocity is increased are immeasurably reduced in weight, whereas those whose angular velocity is reduced are immeasurably heavier.

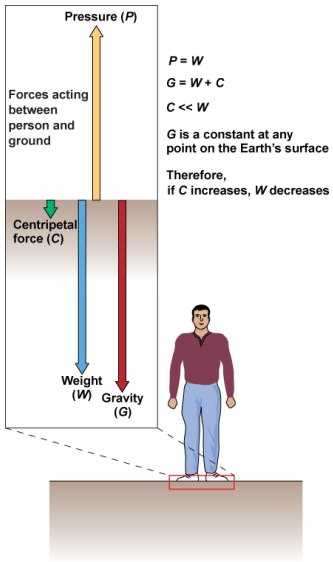


Figure CC12-8. There is a balance between the vertical components of gravity (which is invariable at any given location on the Earth), centripetal force, and the Earth's pressure gradient for objects moving over Earth's surface. The pressure gradient force is equal to the object's weight. The pressure gradient force and the object's weight must be very slightly smaller than the gravitational force because the gravitational force must provide the required centripetal force to maintain the object in its orbit with the rotating Earth. Thus, the object's weight equals the gravitational force minus the very much smaller centripetal force. The magnitude of the centripetal force has been greatly exaggerated in this figure.

If the Earth were perfectly spherical, the component of centripetal force parallel to the Earth's surface would not be compensated by gravity, because gravity would act perpendicular to the Earth's surface at all locations. Consequently, freely moving objects would "slide" across the surface toward the equator (**Fig. CC12-7a**). This explains why the Earth is not a perfect sphere, but is instead an oblate spheroid (the shape you would get if you squeezed the Earth at its poles to deform it like a squeezed basketball). The Earth's diameter as measured from the North to the South Pole is 12,714 km, slightly less than its 12,756-km diameter measured at the equator. Because the Earth is not a perfect sphere, gravity acts at a very small angle to the surface (**Fig. CC12-7b**), and there is a small component of gravity parallel to the Earth's surface everywhere except at the equator and poles. This gravity component compensates for the component of centripetal force parallel to the Earth's surface (**Fig. CC12-7b**), so freely moving objects do not slide toward the equator.

If an object is set in motion relative to the Earth, the object's angular velocity will change just as it did for the ball on the merry-go-round. On the Earth, the vertical component of the altered centripetal force is compensated by gravity and the Earth's pressure gradient. However, if the angular velocity of an object is altered, the balance between the small component of gravity acting parallel to the Earth's surface and the component of centripetal force acting parallel to the Earth's surface is upset. The component of gravity parallel to the surface cannot compensate for changes in the corresponding component of centripetal force (unless the Earth changes its shape). Consequently, altering the angular velocity of a freely moving object on the Earth's surface causes it to move across the surface, toward or away from the pole.

Motion on the Earth: North and South

Now let's look at motions on the Earth's surface. First, consider a projectile fired toward the north in the Northern Hemisphere. This projectile has a component of motion from west to east imparted to it by the Earth's rotation. The projectile retains this west-to-east velocity, but the velocity of the Earth rotating underneath it decreases as the projectile moves north (**Fig. CC12-9a**). The projectile continues eastward while the Earth below it moves eastward progressively more slowly. Thus, the projectile "leads" the Earth's rotation at an increasing rate and follows a path that appears to an observer to be a curve deflected to the right (**Fig. CC12-9b**). In addition, because gravity prevents the projectile from maintaining its original distance from the Earth's axis of rotation, the projectile's angular velocity increases as it moves north. This requires increased centripetal force and creates a force imbalance that "pushes" the projectile back to the south, away from the Earth's axis of rotation.

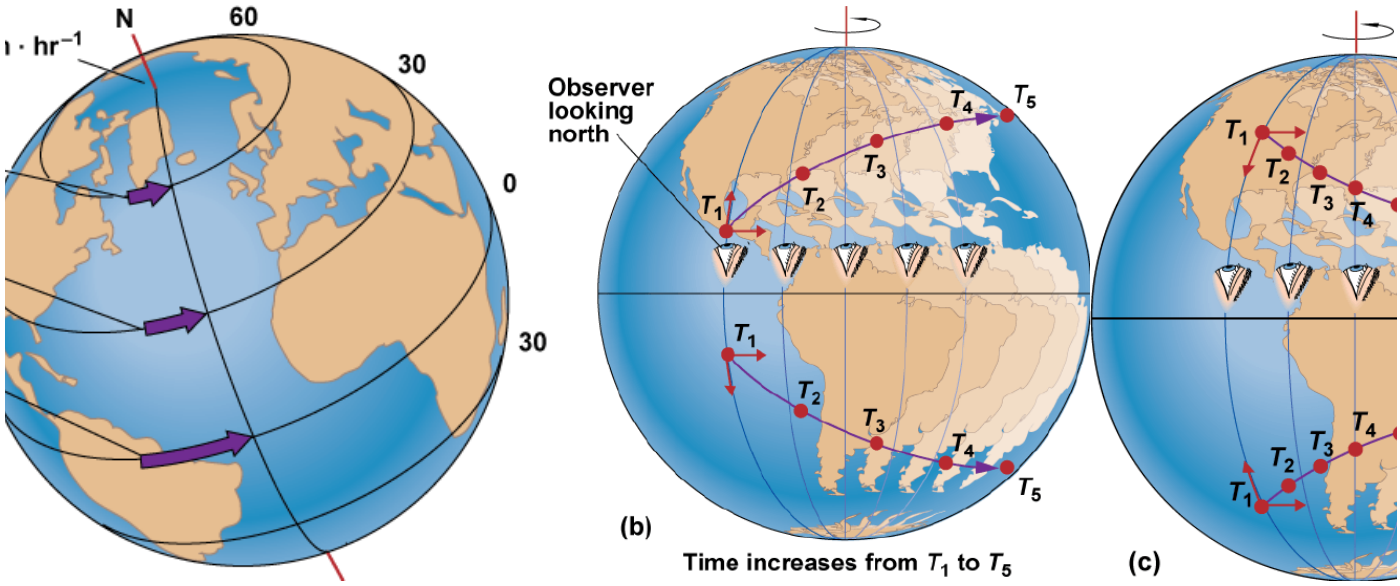


Figure CC12-9. (a) The orbital velocity due to the Earth's rotation of points on the Earth's surface decreases as latitude increases. (b) If an object is started in motion directly toward the pole in either hemisphere, it moves progressively into latitudes in which the Earth's surface is moving more slowly in its rotational orbit. However, if the object is moving freely (that is, if it is not attached by strong frictional forces to the solid Earth), it retains the orbital velocity that it had when it was started in motion. Thus, in addition to its motion to the north (or south in the Southern Hemisphere), the object continues to move to the east at its original orbital velocity and progressively moves ahead of the latitude from which it was set in motion. To an observer moving with the Earth's surface, the result is an apparent deflection in the path of the object, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. (c) If an object is started in motion directly toward the equator in either hemisphere, it moves progressively into latitudes in which the Earth's surface is moving more quickly in its rotational orbit. Thus, in addition to its motion to the south (or north in the Southern Hemisphere), it continues to move to the east at its original orbital velocity and progressively falls behind the latitude from which it was set in motion. The apparent deflection of the path of the object is, once again, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere.

Similarly, a projectile fired southward in the Northern Hemisphere passes over an Earth that moves eastward progressively faster under the projectile, and it "lags" the Earth's rotation (**Fig. CC12-9c**). It also appears to be deflected to the right and is "pushed" back north because it has a lower angular velocity. In the Southern Hemisphere, the situation is similar, but the deflection is to the left (**Fig. CC12-9b,c**).

Motion on the Earth: East and West

When an object is set in motion in a west-to-east or east-to-west direction, it moves either with or against the Earth's rotation (**Fig. CC12-10a**). Thus, the object's angular velocity is increased or decreased. When it is set in motion west to east, angular velocity is increased, and therefore the object tends to move outward away from the pole. A very small reduction in the object's weight allows gravity to compensate for the increased centripetal force and prevents the object from flying upward away from the Earth's surface. However, there is no compensating force to provide the needed increase in the component of centripetal force across the Earth's surface, and the object moves toward the equator, or to the right in the Northern Hemisphere and left in the Southern Hemisphere (**Fig. CC12-10b**).

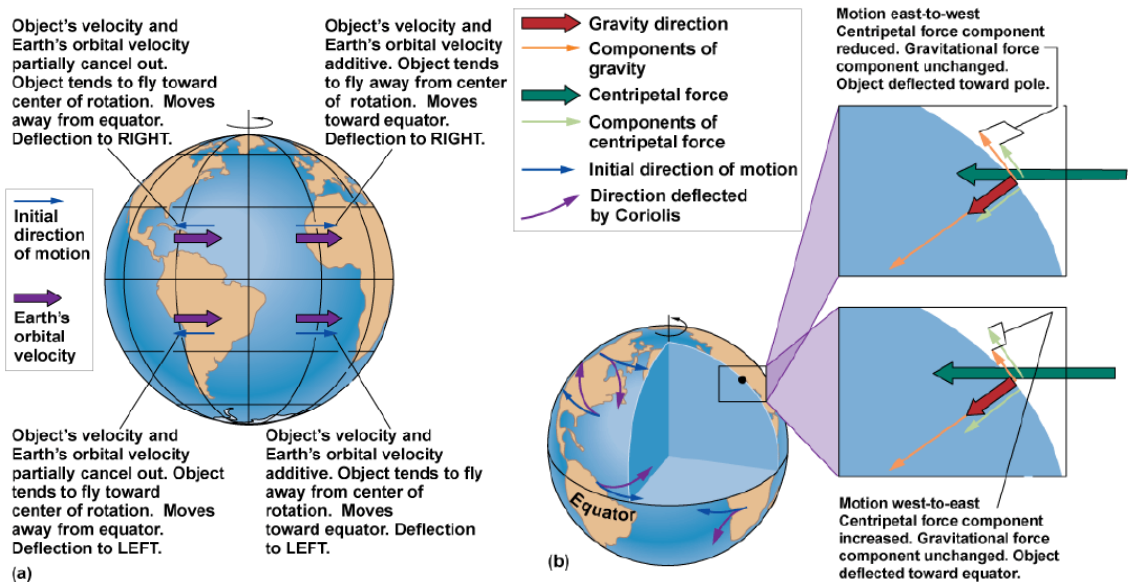


Figure CC12-10. (a) When objects are started in motion across the Earth's surface in an east-to-west direction, their velocity is in the opposite direction to that imparted by the orbital velocity of the Earth's surface due to the Earth's rotation. As a result, the effective orbital velocity of the object is reduced, and the centripetal force needed to maintain the body in its orbit with the Earth's spin is decreased. The slight decrease in the vertical component of centripetal force is readily compensated by a negligibly small increase in the object's weight (change in pressure gradient). (b) In contrast, no such compensation is possible for the component of centripetal force that acts parallel to the Earth's surface. The slight excess of gravitational force over the component of centripetal force parallel to the Earth's surface needed to maintain the body in its orbit, with the Earth's spin deflects the object's path toward the axis of rotation or toward the pole in each hemisphere. Thus, the deflection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Similarly, (a) when objects are started in motion across the Earth's surface in a west-to-east direction, their velocity is in the same direction as that imparted by the orbital velocity of the Earth's surface due to its rotation. (b) The effective orbital velocity of the object is increased, but the centripetal force provided by the component of gravity parallel to the Earth's surface is too small to maintain this orbital velocity. Therefore, the object's path is deflected away from the axis of rotation or away from the pole in each hemisphere. This deflection is also to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. For ease of illustration, the magnitude of centripetal force has been greatly exaggerated relative to the force of gravity in this diagram.

When an object is set in motion east-to-west, the angular velocity is decreased, and the centripetal force needed to maintain its orbit is slightly decreased. The object's weight increases slightly and prevents it from moving vertically downward. However, the reduction in the component of centripetal force parallel to the Earth's surface remains unbalanced, and the object is moved across the Earth's surface toward the pole. Again, the deflection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (**Fig. CC12-10b**).

If the object is moving either west-to-east or east-to-west exactly at the equator, there is no component of centripetal force parallel to the Earth's surface, a small adjustment in weight can totally compensate the changes in centripetal force, and there is no deflection (**Fig. CC12-10b**).

Coriolis Effect Characteristics

All objects moving freely relative to the Earth are deflected by the Coriolis effect unless they are moving directly east or west along the equator. The deflection is always to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. However, it is easier to say that the direction of the deflection is *cum sole*, which means "with the sun." To an observer in the Northern Hemisphere looking toward the equator and thus toward the arc of the sun across the sky, the sun moves across the sky from left to right. In the Southern Hemisphere, the sun moves from right to left.

Freely moving objects appear to move in circular paths on the Earth's surface. These circular paths are distorted if objects move large distances across the Earth's surface. This is because the Earth is equivalent to a flat merry-go-round only at the North or South Pole, where the apparent rotation occurs across a surface that is perpendicular to the axis of rotation. At all other points on the Earth, the motion occurs on a surface that is inclined at an angle other than 90° to the axis of rotation. As a result, at all points other than the poles, a component of the Coriolis deflection is directed vertically downward and "blocked" by adjustments of the object's weight. Consequently, the rate of deflection is reduced.

The proportion of centripetal force directed vertically downward increases from zero at the poles to 100% at the equator (**Fig. CC12-7a**). Therefore, the Coriolis deflection is at a maximum at the poles, is reduced progressively with decreasing latitude, and reaches zero at the equator. In addition, the time necessary for a freely moving object to complete a circle increases with decreasing latitude. This time period, called the "inertial period," is 12 h at the poles, increases to 24 h at 30° latitude, and is infinite at the equator.

The diameter of the circle in which freely moving objects appear to move is determined by both the object's speed and its latitude. Because freely moving objects at the same latitude will complete a circle in the same amount of time, a faster-moving object must move in larger circles. A bullet travels in a circle of huge diameter, and along only a very tiny part of its circular path in the extremely short time it remains airborne. Consequently, it appears to travel in a "straight" line. Air masses in the atmosphere and water masses in the oceans move much more slowly and travel in paths of much smaller radius.

For objects moving at the same speed, the inertial period and the diameter of the apparent circular path decrease with increasing latitude. Therefore, the Coriolis effect "increases" with increasing latitude. Because the diameter of the circle in which even a slowly moving object appears to travel is very large near the equator, the Coriolis deflection is often considered negligible at latitudes within 5° of the equator.

Objects moving at different speeds at the same latitude are deflected through the same angle in equal times, but the faster object is deflected farther from its original path (**Fig. CC12-11**).

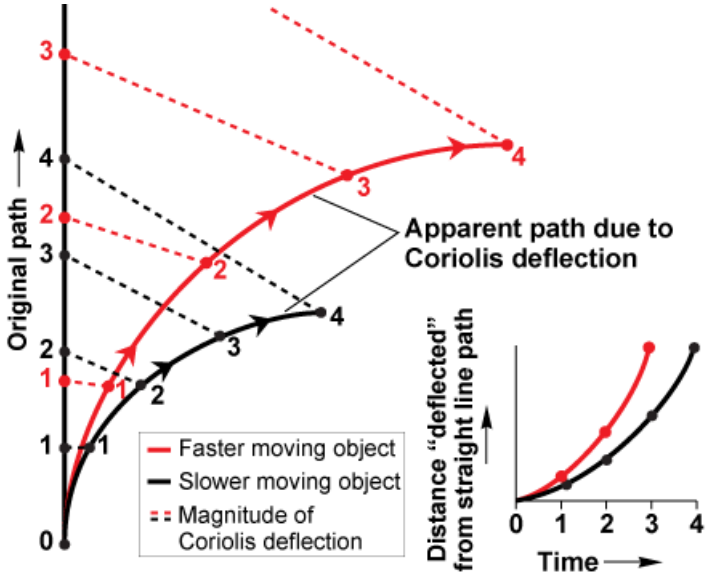


Figure CC12-11. The distance between the location of an object deflected by the Coriolis effect and the location it would be expected to occupy if it were not deflected (that is, if it followed a straight-line path) increases more quickly with time if the object's speed is greater. Thus, the magnitude of the Coriolis deflection is said to "increase" with increasing speed of the object.

Thus, the Coriolis deflection "increases" with increasing speed.

This page titled 17.12: CC12 The Coriolis Effect is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.13: CC13 Geostrophic Flow

Essential to Know

- **Horizontal pressure gradients exert a force that accelerates fluid molecules in the direction of pressure decrease on the gradient. The acceleration increases as the strength of the pressure gradient increases.**
- **The Coriolis effect deflects fluids that flow on a pressure gradient until they flow across the gradient. The flow then continues along a line of constant pressure (isobar) as a geostrophic flow.**
- **Geostrophic flow conditions occur when the pressure gradient force is balanced by the Coriolis deflection.**
- **Geostrophic wind and current speeds are determined by the steepness of the pressure gradient. Wind or current speed increases as the steepness of the gradient increases.**
- **Geostrophic wind or current speed and direction can be determined from contour maps. The direction of flow is parallel to the isobars, and the speed is higher where the pressure gradient is steeper (isobars are closer together).**
- **Geostrophic winds and currents flow counterclockwise around low-pressure zones and clockwise around high-pressure zones in the Northern Hemisphere. In the Southern Hemisphere, they flow clockwise around low-pressure zones and counterclockwise around high-pressure zones.**

Understanding the Concept

The atmosphere and ocean waters are stratified fluids (**CC1**), in each of which density decreases with increasing distance from the Earth's center. The only exceptions are in limited areas where stratification is unstable. It is in these areas of unstable stratification that density-driven vertical motions of the fluid occur.

If the oceans or atmosphere were at equilibrium, density would be uniform at any one depth or altitude. The vertical density gradient would be the same everywhere, so the total weight of the overlying water and/or air column at a specific depth or altitude would be the same everywhere. Because this total weight determines atmospheric or water pressure, pressure would be uniform at any depth or altitude.

Neither the atmosphere nor the oceans is at equilibrium, because the density of atmospheric gases and ocean water is altered locally by changes in such factors as air or water temperature, dissolved salt concentration in the water, and water vapor pressure in the air (**Chaps. 5, 7**). Consequently, the vertical distribution of density in both oceans and atmosphere varies from place to place, and there are horizontal variations of pressure at any given height in the atmosphere or depth in the oceans.

Surfaces that consist of points of equal altitude or depth are referred to as horizontal or level surfaces. However, these surfaces are actually spherical because the Earth is a sphere. Horizontal **pressure gradients** develop in the oceans as a result of density differences between **water masses** and also because winds tend to move ocean surface waters, causing the water to pile up in some locations (**Chap. 8**).

Where there is a horizontal pressure gradient, the fluid is subject to a force that tends to accelerate molecules from high-pressure areas toward low-pressure areas. The acceleration is greater when the pressure gradient is steeper. We perform one of the simplest demonstrations of acceleration along a pressure gradient every time we open a soda bottle or can. Once they are free to do so, the gas molecules in the high-pressure zone within the bottle are accelerated toward the lower pressure outside.

In our soda bottle experiment, the density gradient is extremely steep because the pressure difference between the air outside the bottle and the gas in the bottle is large, and the distance between the high-pressure zone inside the bottle and the low-pressure zone of the surrounding air is very short. When we open the bottle, gas molecules in the high-pressure zone are accelerated very rapidly and must move only a short distance to reach the low-pressure zone. As a result, the pressure equalizes almost instantaneously. In contrast, the pressure differential between horizontally separated high- and low-pressure zones in the oceans and atmosphere is very small, and these zones are separated by much greater distances. Consequently, the accelerations produced by atmospheric and oceanic horizontal pressure gradients are small. In addition, air or water does not flow directly from the high-pressure zone to the low-pressure zone, because the air and water molecules are subject to the **Coriolis effect** (**CC12**) once they have been set in motion.

To understand how motions induced by the pressure gradient and Coriolis effect interact, consider the following facts. Freely

moving objects, including air or water molecules, actually travel in straight paths unless acted on by another force, such as the pressure gradient. The Coriolis deflection is only a perceived deflection seen by an observer on the rotating Earth. However, from our rotating frame of reference, the freely moving object is deflected *cum sole*. In **Figure CC12-11**, we can see that the magnitude of the deflection and the rate of increase both increase with time. These are the characteristics of an acceleration in the direction 90° *cum sole* to the direction of motion, and this "acceleration" increases with the object's speed (**Fig. CC12-11**).

We can now examine what happens when a fluid begins to flow in response to a horizontal pressure gradient. The fluid is accelerated down the pressure gradient from the high-pressure region toward the low-pressure region. As it begins to move, it is deflected by the Coriolis effect. The faster it moves, the greater is the Coriolis deflection (**CC12**). The direction of motion thus turns away from the direct path down the pressure gradient, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. As the moving fluid is deflected, the pressure gradient acceleration continues to act toward the low-pressure region while the Coriolis deflection continues to act at 90° *cum sole* to the changed direction of flow. The speed increases as the molecules are accelerated, thus increasing the Coriolis deflection.

The molecules continue to accelerate while the direction of flow is progressively turned cum sole until the flow is directed across the pressure gradient along a **contour** of equal pressure (**Fig. CC13-1**). When the flow is in this direction, the pressure gradient, acceleration, and Coriolis deflection act in opposite directions and balance each other. The speed of the fluid at this balance point is determined by the steepness of the pressure gradient because stronger pressure gradients cause greater accelerations that must be balanced by greater Coriolis deflections, which increase with increasing speed (**CC12**).

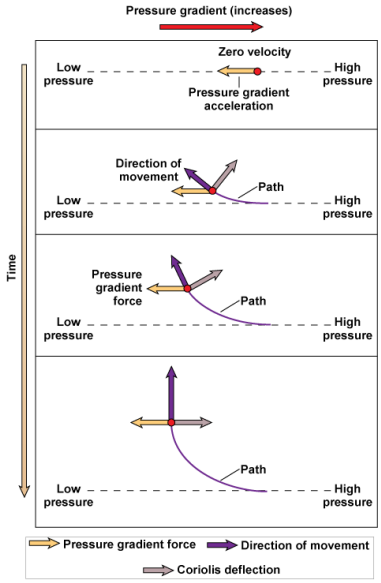


Figure CC13-1. When a current is initiated on a horizontal pressure gradient, the initial direction of motion is directly down the gradient. However, as it moves, the water mass is deflected by the Coriolis effect. It accelerates and is deflected until it flows directly across the pressure gradient, and the pressure gradient force is balanced by the Coriolis deflection. This is a "geostrophic" current. If the pressure gradient is steeper, the acceleration is greater, and the geostrophic current is faster, but it is still balanced by the greater Coriolis deflection associated with the higher speed.

This type of flow, in which the pressure gradient and Coriolis deflection are balanced, is called **geostrophic** flow, and the moving air or water masses are geostrophic winds or geostrophic **currents**. The most important features of geostrophic winds or currents are that the flow is directed along contours of equal pressure within a pressure gradient, and that wind or current speed is determined by the steepness of the pressure gradient.

Geostrophic flows are almost never aligned exactly along the contours of equal pressure on the pressure gradient. The reason is that pressure gradients are continuously changing in response to changes in the factors that create them, such as wind stress on the oceans and **convection** in the atmosphere (**CCS3**). In addition, **friction** between moving air masses and the ground or ocean surface, or between ocean water and the seafloor, reduces the Coriolis deflection. Thus, geostrophic winds at the surface and currents near the seafloor do not flow exactly along contours of equal pressure. Instead, they are aligned generally along these contours but offset toward the center of low-pressure zones and away from the center of high-pressure zones.

Horizontal pressure gradients are usually shown as maps of lines of equal pressure called **isobars**. Most **weather** maps in newspapers and on television are of this type. The isobaric maps in **Figure CC13-2** show high- and low-pressure zones and the

pressure gradients between these zones. The pressure difference between adjacent isobars is the same for all adjacent isobars at all locations on each map. The gradient is steeper where the isobars are closer together. Therefore, the spacing of the isobars reveals the steepness of the pressure gradient (the change in pressure per unit distance).

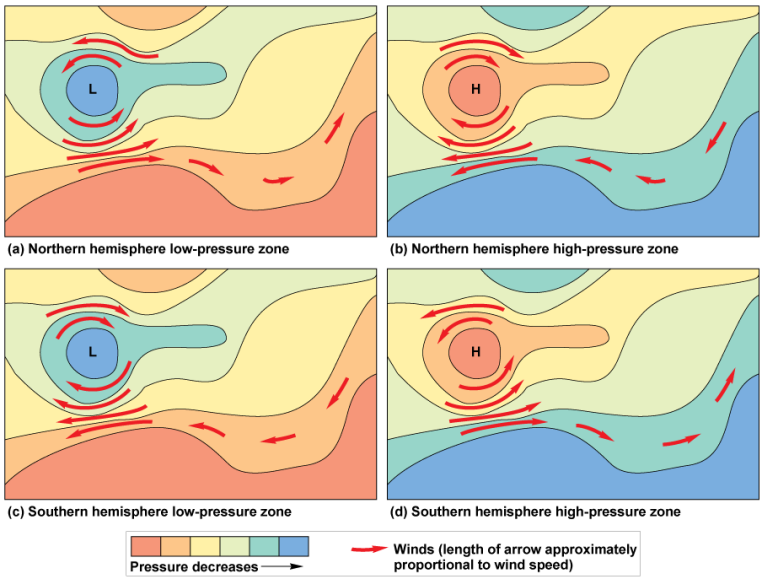


Figure CC13-2. Isobaric contour maps can be used to deduce the wind (or current) speeds and directions associated with the isobars. The contour map examples in this figure show circulation at (a) a low-pressure zone in the Northern Hemisphere, (b) a high-pressure zone in the Northern Hemisphere, (c) a low-pressure zone in the Southern Hemisphere, and (d) a high-pressure zone in the Southern Hemisphere. Note that the directions of rotation are opposite for the two hemispheres. Winds blow counterclockwise around a low-pressure zone and clockwise around a high-pressure zone in the Northern Hemisphere, and in the reverse directions in the Southern Hemisphere.

Most upper-air wind and ocean current systems are geostrophic, and their flow direction is nearly parallel to the pressure contours. As a result, pressure contour maps can be used to estimate both the speed and direction of winds or currents. The wind or current direction is parallel to the isobars, and the speed can be estimated from the pressure gradient. Because the Coriolis deflection is to the right in the Northern Hemisphere, winds and currents in this hemisphere flow counterclockwise around low-pressure zones (**Fig. CC13-2a**) and clockwise around high-pressure zones (**Fig. CC13-2b**). In the Southern Hemisphere, the deflection is to the left, and winds and currents flow clockwise around low-pressure zones (**Fig. CC13-2c**) and counterclockwise around high-pressure zones (**Fig. CC13-2d**). The wind or current speed is determined by the steepness of the pressure gradient. Accordingly, wind or current speeds are higher where isobars are closer together and slower where they are more widely separated (**Fig. CC13-2**).

This page titled 17.13: CC13 Geostrophic Flow is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.14: CC14 Phototrophy, Light, and Nutrients

Essential to Know

- **Organic matter is created from inorganic substances (primary production) by phototrophy and chemosynthesis.**
- **Phototrophy is performed by organisms that use light as their energy source to fuel the primary production process. There are two known mechanisms: photosynthesis and a recently discovered metabolic process that uses compounds called rhodopsins to capture light energy.**
- **Most primary production in Earth's ecosystems is created by oxygenic photosynthesis, but too little is known about the use of the rhodopsin-based mechanism in ocean ecosystems to assess its relative importance.**
- **Plants and algae are eukaryotes that perform oxygenic photosynthesis through a complex series of reactions that occur in pigmented cells called "chloroplasts."**
- **Photosynthesis originated and still takes place in certain Prokaryotes called Cyanobacteria. Chloroplasts are thought to have evolved from photosynthetic symbiotic cyanobacteria that became a permanent resident in a eukaryotic cell.**
- **A nitrogen-containing pigment called "chlorophyll *a*" plays the central role in transferring light energy to the oxygenic photosynthesis process. Chloroplasts contain accessory pigments that capture light energy and transfer it to chlorophyll *a*.**
- **Accessory pigments collect light energy at wavelengths at which chlorophyll *a* absorbs light poorly. Each species has a different suite of accessory pigments optimized to collect light energy available in its specific habitat.**
- **Oxygenic photosynthesis uses light energy, carbon dioxide, and water to produce oxygen (which is released) and relatively simple organic compounds. Most organic compounds produced by oxygenic photosynthesis are used for cellular respiration. Only a small proportion enters biochemical cycles and is converted to more complex organic molecules.**
- **Only a very small percentage of the light energy used in photosynthesis is used to produce biomass.**
- **Nitrogen and phosphorus, among other elements, are present in key compounds involved in photosynthesis. Hence, depletion of available nitrogen and phosphorus stops production of these compounds and limits primary productivity.**
- **Anoxygenic photosynthesis is an alternate pathway employed by anaerobic bacteria. Anoxygenic photosynthesis uses light energy, carbon dioxide, and a hydrogen donor, such as H_2S or H_2, to create relatively simple organic compounds and produces a very small amount of organic matter in comparison with algal and cyanobacterial photosynthesis in most present-day ocean ecosystems.**
- **Chemosynthetic bacteria and archaea synthesize organic compounds by using chemical energy from the oxidation of compounds such as H_2S and H_2 instead of light energy.**

Understanding the Concept

For life to exist, carbon, hydrogen, oxygen, nitrogen, phosphorus, and many other essential elements must be synthesized into organic compounds. It is believed that most organic matter on the Earth today was initially converted from inorganic matter to organic compounds by **photosynthesis**. Exceptions include a few simple organic compounds created by electrical discharges in the atmosphere, and the **chemosynthetic communities** described in **Chapters 12 and 15**, the importance of **rhodopsin**-based **phototrophy** is uncertain.

Photosynthesis is a complex process in which plants, algae, and certain **bacteria** and **archaea** use light energy to convert carbon dioxide into organic compounds. There are two major known photosynthetic pathways: oxygenic and anoxygenic photosynthesis. There are also variations of these pathways, such as those used by certain archaea. Because their photosynthetic mechanisms differ from those of plants and algae, organisms that use these variations are often not considered to use photosynthesis by biologists who prefer to restrict the definition of photosynthesis to the specific pathway used in plants. In this text, we use a broader definition of photosynthesis that includes all pathways used by any organisms that use chlorophylls to capture light energy and carbon dioxide to produce organic matter. Organisms that use light energy to provide sufficient organic matter to generate new biomass are called **phototrophs** or photoautotrophs.

The first step in all variations of photosynthesis is the capture of light energy by the photosynthetic organism. In oxygenic photosynthesis by plants and algae, light is captured in complex cell structures called "chloroplasts" located near the organism's

surface. The chloroplasts contain **chlorophyll** *a* and, in many species, the related pigments chlorophyll *b* and chlorophyll *c*. There are also a number of "accessory pigments," including carotenoids, xanthophylls, phycobilins, and chlorophyll *d* and chlorophyll *f*. Each photosynthetic species has its own unique combination of pigments. Accessory pigments capture light energy and transfer it to chlorophyll *a*, which then performs the next step in the synthesis. Chlorophylls *b* and *c* can be characterized as accessory pigments because they, too, collect light energy and pass it on to chlorophyll *a*. The transfer of energy from accessory pigments to chlorophyll *a* is an extremely efficient process. In some cases, transfer efficiency approaches 100%.

Why are accessory pigments needed? Chlorophyll *a* absorbs light effectively only toward the red and violet ends of the spectrum (**Fig. CC14-1a**). To capture light energy from the central blue–green–yellow part of the spectrum, other pigments are needed. Carotenoids, which include the group of pigments called xanthophylls, absorb energy primarily in the blue and blue-green parts of the spectrum (**Fig. CC14-1b**), whereas phycobilins absorb energy primarily in the green, yellow, and ultraviolet parts (**Fig. CC14-1c**). Chlorophylls *b* and *c* also absorb in different parts of the visible spectrum than chlorophyll *a* (**Fig. CC14-1a**). Chlorophyll *d* and chlorophyll *f* absorb in the infrared wavelengths. Thus, the chloroplast contains an array of accessory pigments to capture as much of the available light energy as possible. Accessory pigments transfer light energy that they capture to chlorophyll *a*, which performs the first steps in photosynthesis.

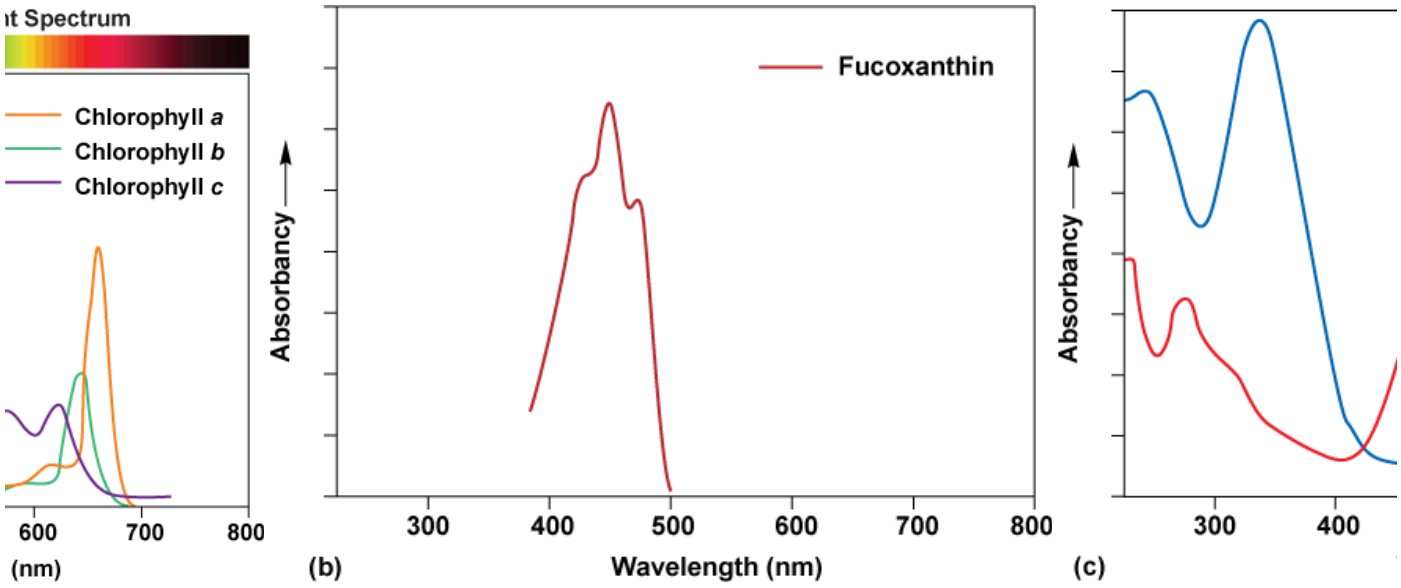


Figure CC14-1. Absorption spectra of photosynthetic pigments. (a) Chlorophylls *a*, *b*, and *c*. (b) Fucoxanthin (a typical xanthophyll). (c) Typical phycobilins. Note that the other pigments absorb energy in parts of the spectrum that are not absorbed efficiently by chlorophyll *a*.

Red and blue-violet light are absorbed more effectively by seawater than are green and blue light (**Fig. 5-17**). Hence, accessory pigments are especially important in marine algae that must photosynthesize below the immediate surface layer. **Absorption spectra** of seawater in the open ocean and coastal waters are different (**Fig. 5-18**), and the **wavelength** distribution of light changes with depth. Species of marine algae that live at different locations and depths have different accessory pigment compositions to optimize their capture of available light energy.

Photosynthetic cyanobacteria also use chlorophyll or a closely related pigment, bacteriochlorophyll, to perform oxygenic photosynthesis, but they do not have chloroplasts. Photosynthesis is believed to have originated in cyanobacteria and, over a period of about 1 billion years. They transformed Earth's atmosphere from one with almost no oxygen to the oxygen-rich atmosphere that now exists. Eukaryotes developed only toward the end of this transformation process, and chloroplasts are thought to have originated from a symbiotic cyanobacteria that became permanently located within its host Eukaryote cells.

Some archaea (and perhaps bacteria) use a complex protein, such as archaerhodopsin or halorhodopsin, to collect light energy and produce organic compounds for their metabolic needs, so these organisms employ phototrophy, but they are not photosynthesizers. At present, it is not known how important this rhodopsin-based phototrophy is in the overall primary production in the oceans, but its contribution is thought to be comparatively small. In contrast, cyanobacteria are estimated to produce about one-half of all ocean primary production. They do so because they appear to be the principal primary producers in the large areas of the oceans that are nutrient-limited. Thus, although their population and primary productivity per unit area of oceans are low, their total numbers and total primary production rivals the total primary production by algae and plants in the ocean.

The key molecule in oxygenated photosynthesis, chlorophyll *a* (**Fig. CC14-2**), is a complex molecule that contains not only carbon (C), hydrogen (H), and oxygen (O) atoms, but also a magnesium atom (Mg) and, most importantly, nitrogen atoms (N). The other forms of chlorophyll and rhodopsins also contain nitrogen. When dissolved nitrogen is depleted in ocean waters, nitrogen-containing compounds, including chlorophyll *a*, cannot be synthesized by algae and **primary production** is limited or stopped (**Chap. 12**). Thus, nitrogen as nitrate (NO^{3-}), nitrite (NO^{2-}), or ammonia (ammonium **ion** [NH_4^+] and NH_3) is essential to photosynthesis. The exception to this is that certain cyanobacteria can convert molecular nitrogen to ammonia, although the conversion process is energy intensive and takes place very slowly, especially in oxygenated environments such as most ocean water.

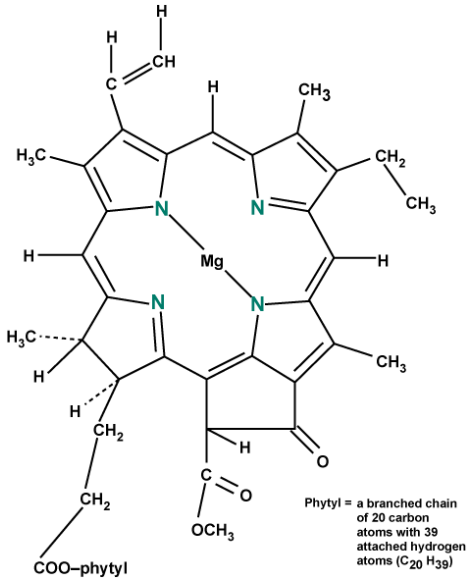


Figure CC14-2. The molecular structure of chlorophyll a, which is needed for oxygenic photosynthesis. The molecule is complex and contains four nitrogen atoms.

After chlorophyll *a* has captured light energy, it can transfer an excited electron (an electron to which excess energy has been added) to other molecules, which transfer the energy through yet other molecules in a complex series of steps. During these transfers, the raw materials of photosynthesis—water and carbon dioxide—are brought together and converted into organic compounds.

Some reactions in photosynthesis occur only in the presence of light. The complex process of oxygenated photosynthesis can be simplified and summarized as follows. Light energy is used to split a water molecule into a hydrogen atom with an excited electron and a hydroxyl ion, each of which immediately reacts with other molecules in the chloroplast. The hydroxyl ion is then converted to oxygen and water, which are released, and the hydrogen atom with its excited electron is combined with NADP (nicotinamide adenine dinucleotide phosphate) to form NADP-H. At the same time, additional light energy is transferred to ADP (adenosine diphosphate) to form ATP (adenosine triphosphate) (**Fig. CC14-3**). ATP is a versatile energy-storing molecule that can donate the considerable energy associated with one of its phosphate bonds by releasing or transferring this phosphate group, which changes ATP back into ADP. ATP is an energy provider in both photosynthesis and **respiration**. Both ATP and ADP contain phosphorus, which is therefore also an essential **nutrient** for photosynthesis.

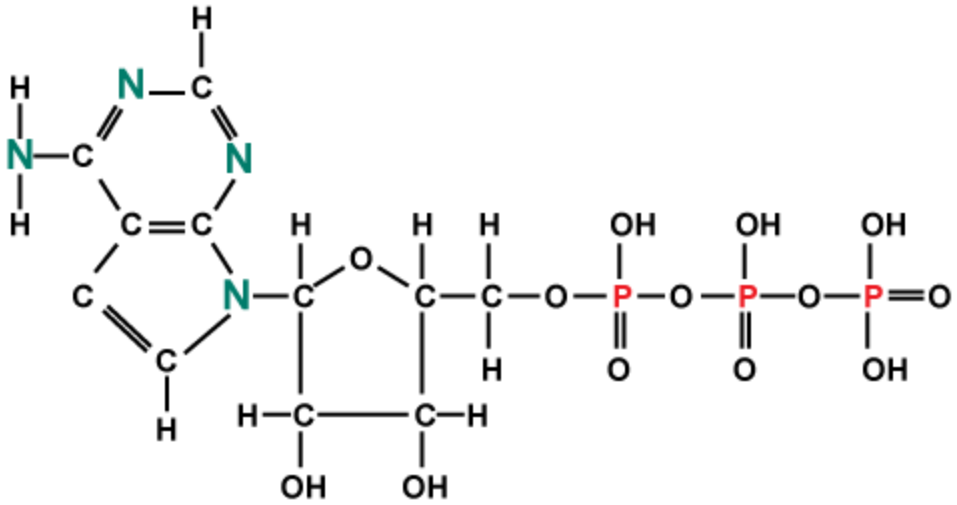


Figure CC14-3. The molecular structure of ATP, a chemical essential to life, is complex. The molecule contains both phosphorus (P) and nitrogen (N) atoms.

Once light energy is stored in ATP and NADP-H, the remaining steps in photosynthesis can continue in the dark. However, if light is removed for an extended period, the store of these molecules in the chloroplast is used up. The photosynthesis steps that can occur in the dark use hydrogen from NADP-H and energy from ATP to reduce carbon dioxide. Oxygen is removed from carbon dioxide and released, whereas the carbon is combined with other carbon atoms to form organic compounds.

There are several different pathways for these reactions, each producing a different organic compound or compounds. One of the most common pathways leads to the production of glucose ($C_6H_{12}O_6$). Glucose and other organic molecules formed by photosynthesis are subsequently used by the organism as basic building blocks for the vast array of other organic compounds produced by living organisms in their **biochemical** cycles. Large amounts of glucose and the other compounds created by photosynthesis are also converted back into carbon dioxide in the respiration processes of the plant. The overall efficiency of transferring light energy through photosynthesis to create new **biomass** is extremely low, a few percent at most.

The process of oxygenic primary production can be depicted by the following, much simplified equation:

$$6CO_2 + 6H_2O \rightarrow \text{visible light} \rightarrow C_6H_{12}O_6 + 6O_2$$

Cyanobacteria photosynthesize in a somewhat different way. First, cyanobacteria contain a unique form of chlorophyll called “bacteriochlorophyll.” In addition, they do not use water to provide hydrogen for photosynthesis, and they do not release oxygen. Cyanobacterial photosynthesis can be depicted by the following simplified equation:

$$6CO_2 + 6\ H_2[A] \rightarrow \text{visible and far red light} \rightarrow C_6H_{12}O_6 + 6[A]$$

H2[A] is a hydrogen donor and can be H_2S, H_2, or various organic compounds. Cyanobacterial photosynthesis does not make a major contribution to ocean primary production in most ocean areas, because **environments** where both light and a suitable hydrogen donor are available are severely limited.

In cyanobacterial photosynthesis, the needed energy is provided by light. However, it is noteworthy that many of the molecules that replace water as the hydrogen donor in cyanobacterial photosynthesis can also combine with oxygen to provide the energy needed to synthesize organic compounds in chemosynthesis (**Chap. 12**).

This page titled 17.14: CC14 Phototrophy, Light, and Nutrients is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.15: CC15 Food Chain Efficiency

Essential to Know

- **All organisms require food for respiration, growth, and reproduction.**
- **All organisms lose some of the food as waste products.**
- **The rate of biomass production by an organism that is available as food for the next trophic level equals the total amount of food ingested, minus the proportion used for respiration and reproduction, and the proportion lost as waste products.**
- **The percentage of ingested food converted to new biomass is the food chain (or trophic) efficiency. It is usually approximately 10% (ranging from 1% to 40%), regardless of species or food source.**
- **Approximately 90% of the available food biomass is used in respiration and reproduction or lost as waste products at each trophic level in a food chain.**
- **Long marine food chains that lead to consumers at higher trophic levels, such as tuna, utilize primary production hundreds of times less efficiently than short marine food chains, such as those that support anchovies or sardines, or most terrestrial food chains used to produce food for human consumption.**

Understanding the Concept

All organisms require food, which they can obtain by performing **photosynthesis** or **chemosynthesis** or by ingesting organic matter. In all species, food is distributed and used in four ways: (1) to provide energy through respiration for the life processes of the organism, (2) to provide the basic materials for production of the myriad compounds contained in any organism's body, (3) to provide the basic materials needed to produce offspring (including eggs and sperm), and (4) as waste products that are **excreted**. Hence, not all food is incorporated into new body tissues or **biomass**. The amount converted to new biomass equals the total food intake minus the amounts used for **respiration** and reproduction and lost as waste. Food used for reproductive processes does create new biomass, but the vast majority of eggs, **larvae**, and juveniles of most species are consumed by predators. Consequently, most new organic matter produced during reproduction does not contribute to the species' adult biomass.

The percentage of food ingested by a particular species and converted to new biomass is often referred to as food chain efficiency or "trophic efficiency." The total biomass of a species can be estimated if the total amount of available food and the **food chain efficiency** are both known. For example, studies of Loch Ness have determined that it would be virtually impossible for a viable population of "monsters" to live there, because the total amount of food available in the loch would not be enough to sustain even one very large animal, unless it converted all available food into new biomass at an impossibly high food chain efficiency.

The efficiency of converting food to biomass is an important parameter in managing food supplies for human populations. For example, much of the corn fed to beef cattle is used by the cattle in respiration, reproduction, and waste generation. Hence, although beef has a higher protein content than corn and therefore is a desirable food, the same amount of corn could feed more people if eaten directly than if first fed to cattle, which are then eaten. This concept is particularly important in many parts of the world where famine persists, but where cattle are raised on corn and other grains that could be consumed directly by people.

Terrestrial **food chains** used for human food are generally very short. Humans consume primarily plant material, which is at the **primary production** or first **trophic level**, or plant-fed animals, which are at the second trophic level. Ocean food chains that lead to human foods are, in many cases, much longer. For example, tuna generally feed at the fifth trophic level (**Fig. 12-11**). The overall efficiency of the food chain from primary production (**phytoplankton**) to tuna is very low because each step is inefficient.

Food chain efficiency is difficult to measure precisely because it varies with many factors, such as the amount of food available, food quality, the nature and timing of the reproductive cycle, and the level of physical activity required to avoid predators. As an example of this variability, consider that two human children of the same height who ingest approximately the same amount of food, but one may be much thinner than the other. We might conclude that the thinner child has a "faster metabolism." What we mean is that the thinner child uses more food for respiration (because of either greater physical activity or a genetic disposition to utilize food less efficiently in respiration) or that the thinner child sends a greater proportion of food to waste (because of **genetic** differences in the ability to digest and assimilate food). Of course, if the food ingested by these two children was not equal in quantity, the thinner child could also be thinner simply because of a limited food supply and eating less. Another illustration of

the variability of food chain efficiency is the difference in efficiency between children, whose biomass is increasing, and adults, whose biomass stays relatively constant, even though they may ingest approximately the same amount of food as children.

Food chain efficiency is variable. It has been determined that, if averaged over a population and over time, this efficiency ranges from about 1% to 40% and, in most instances, is approximately 10% between any two trophic levels. Hence, on average, all species above the primary producer trophic level convert only about 10% of their food into new biomass, which then is available to be consumed at the next trophic level. The following relationships summarize marine food chains:

1 kg of biomass at **trophic level 1** (phytoplankton)

produces **0.1 kg** of biomass at **trophic level 2** (zooplankton)

produces **0.01 kg** of biomass at **trophic level 3** (e.g., small fishes, baleen whales)

produces **0.001 kg** of biomass at **trophic level 4** (e.g., larger fishes such as mackerel, squid)

produces **0.0001 kg** of biomass at **trophic level 5** (e.g., predatory fishes such as tuna)

produces **0.00001 kg** of biomass at **trophic level 6** (e.g., killer whales)

In this example, certain types of organisms are designated as feeding at a particular trophic level. However, some of them, such as killer whales, may feed at more than one trophic level.

Many seafood species captured and eaten by people feed at a much higher trophic level than terrestrial animals used as human food. Consequently, human use of ocean primary production is grossly inefficient compared with terrestrial primary production.

This page titled 17.15: CC15 Food Chain Efficiency is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.16: CC16 Maximum Sustainable Yield

Essential to Know

- **Fishing initially reduces the size of a fish stock. However, because the reduction results in greater food availability for the fishes remaining, the rate of production normally increases.**
- **The additional biomass produced represents an excess over that needed to maintain the population. This excess can be harvested safely.**
- **As fishing increases, stock size is reduced to a critical level known as the "maximum sustainable yield," at which the production and reproduction rates of the population are just sufficient to balance the removal rate from predation and fishing. If fishing yield is increased and continued beyond the maximum sustainable yield, the population can no longer sustain itself and collapses.**
- **Maximum sustainable yield is difficult to establish because fish stocks vary as a result of year-to-year climate-induced changes and changes caused by other factors, such as diseases.**
- **Maximum sustainable yield also depends on the age structure of the population and on the degree of age selectivity in fishing methods used. Harvesting older fishes tends to increase the sustainable yield because the remaining younger fishes are faster-growing, but it also tends to reduce the breeding population because the younger fishes are sexually immature.**
- **Maximum sustainable yield is usually established by using one year's data for stock size and reproductive success to project the survival of adults and young into the next year. Unexpected events, such as disease outbreaks, can render such estimates inaccurate and inadequately protective.**
- **Most fisheries are managed at a yield 20% to 40% below the estimated maximum sustainable yield to allow a safety margin. This safety margin may not always be adequate, but any safety factor means that fishes that could be harvested will not be.**

Understanding the Concept

Although fisheries and **shellfisheries** provide only a small fraction (about 1%) of the world's total human food supply, they provide a much larger percentage of its required protein. In many areas, the oceans are the only significant source of protein.

The world's total fish catch is currently approximately 90 million tonnes per year, about one-third of the estimated global annual fish production. Some optimistic biologists believe that the global fishery catch could be raised by a factor of as much as 10. These optimistic estimates are based on the assumption that existing fisheries can be exploited to the maximum possible extent, that many new stocks of fishes will be discovered, and that many currently underutilized exotic **species**, including **invertebrates** such as **sea cucumbers** and **sea urchins**, will be fully utilized. Unfortunately, about 85% of major marine fish stocks are currently depleted, overexploited, or fished at their biological limits. Consequently, the global total fish catch has begun to decline from its previous maximum level. Many fisheries have been closed or severely restricted in attempts to reverse this damage.

The oceans do not appear to have the potential to help solve the world's food supply problem. However, if fisheries are to continue sustaining human populations in areas that have historically depended on them, and if seafood is to continue supplying the same proportion of human food, the world's fisheries must be managed carefully. The goal of such management must be to catch the maximum amount of seafood that can be taken from the oceans without damaging individual species or marine ecosystems. To fully meet this objective, we would need to harvest species selectively at low **trophic levels** instead of the currently consumer-desired species, such as tuna. Furthermore, we would have to learn to farm the sea as we do the land, eliminating undesirable plants and animals from ocean farms.

Neither approach is likely to be fully acceptable, at least in the foreseeable future. Consequently, the principal approach of fishery management, which will probably continue for many years, is to manage the fishing of each individual species that is targeted by fishers. The specific objective is to maximize the total amount of the species caught (the yield) while ensuring that the **standing stock** does not decline to levels that cannot sustain the yield in the future. Thus, the goal is to manage each species to ensure the **maximum sustainable yield**.

To determine the maximum sustainable yield, we must have a good understanding of the species' life cycle. Consider how fish stocks respond to fishing. The stock (or **biomass**) of a fish species is limited primarily by its food supply and its predators. If

human or other predators harvest more of the species than is normally taken by its natural predators, more food will be available for others of the species. If excess food is available, the species will reproduce and grow to use this food until the population is again at a size where food availability limits further growth. Exceptions occur when the excess food is consumed by competitor species, but in most cases, if we harvest a species and reduce its stock somewhat, the total production of the species will increase. If we harvest an amount equal to this increased production each year, the population will remain stable, but at a lower number than before we began to harvest. If we harvest more than the increase in production, the stock will progressively decline.

As a stock declines with increased fishing, a critical level, generally 35% to 70% of the original stock, is reached at which the remaining stock becomes so small that it is barely able to grow and reproduce fast enough to replace fishes removed from the population by predation and fishing. If fishing continues to increase beyond this critical point, the stock will begin to decline precipitously (**Fig. CC16-1a**), and unless the catch is reduced, the stock will collapse to a very low abundance and will not recover. The critical level at which the species is just able to replace the stock lost to fishing is the maximum sustainable yield.

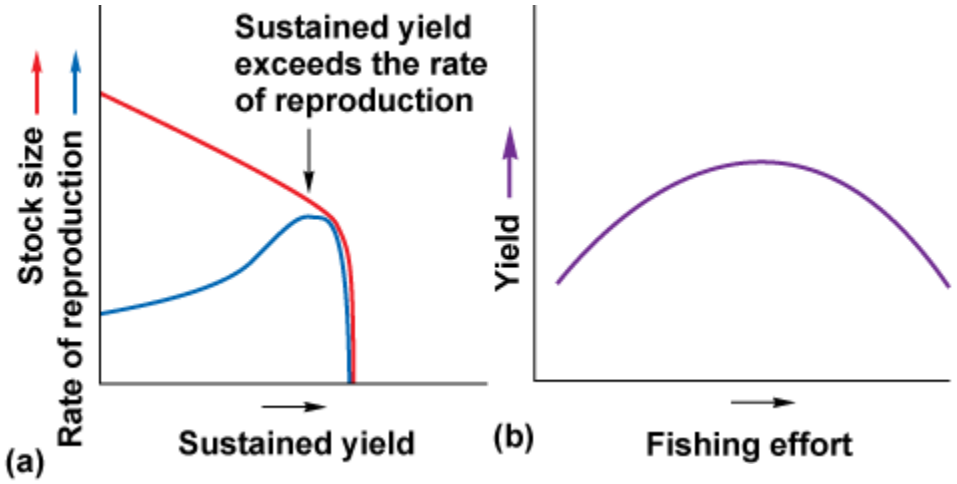


Figure CC16-1. Conceptual model showing (a) the changes in fish stock size with increasing fishing, and (b) fishing yield as a function of fishing effort. Note that moderate fishing can normally produce high sustained yields because the rate of reproduction initially increases as the stock size is reduced and more food becomes available to younger fishes. However, if the fishing effort and yield are increased further, the stock size is reduced to such a level that there are no longer enough individuals to sustain a high rate of reproduction, and the stock collapses.

One of the most important characteristics of a fishery is the "fishing effort," which is the number of boat or person days of fishing expended. As a typical fishery develops, fishing effort increases as more boats and fishers target the resource species. At first, yield increases rapidly, but as the maximum sustainable yield is approached, fishing effort increases faster than yield because stock size is reduced (**Fig. CC16-1a**). Each boat must fish longer to catch the same amount of fish. Consequently, either the cost of each fish caught rises, or the catch and income of each fisher declines. Therefore, for economic reasons, the optimum harvest level of a species may be well below its maximum sustainable yield. Fishery management often controls fishing effort to control yield (**Fig. CC16-1b**).

If the maximum sustainable yield is exceeded in an uncontrolled fishery, fishing effort often rises dramatically as stocks decline. Fishers targeting the species try to protect their livelihoods by increasing their efforts in an attempt to maintain their individual historical catch levels. This dynamic occurs both with technologically advanced fishers targeting a regional resource, such as North Atlantic cod, and with subsistence fishers in island communities, whose growing populations lead to increased fishing effort for local **reef** fishes.

Changes in the age structure of populations subject to fishing also affect maximum sustainable yield because age structure influences the population's reproductive capacity. Like children, young fishes increase their biomass with time faster than adults do, even if they consume similar amounts of food. If larger adult fishes of a particular stock are preferentially targeted by fishers, the excess food and decreased competition from larger fishes will enable a greater number of young fishes to grow successfully. Because young fishes gain weight or biomass faster than adult fishes do, a greater number of young fishes in the population will cause an increase in the rate of biomass production, even if the available food supply remains the same. If large fishes are preferentially removed, the stock gains more biomass per year, and the maximum sustainable yield is increased. Hence, fishery management commonly attempts to control the size of harvested fishes by establishing minimum mesh sizes for nets, requiring the release of captured small individuals, or other methods.

Although it is generally advantageous to reduce the average age of a fish population to increase the maximum sustainable yield,

the stock can be adversely affected if too many large fish are removed. Only the larger fish of most species are of reproductive age. Therefore, reducing the number of large fishes also reduces the breeding population.

Clearly, establishing maximum sustainable yield is a difficult task that requires knowledge of stock size, age structure, and reproductive process for each species. It is further complicated by the natural year-to-year variability of fish populations and the requirement that maximum sustainable yield be established before or early in a given year. The natural variability of fish populations from year to year is very large, at least for some species, due to **climate** variation, disease, and other factors. For example, a relatively small change in water temperature may alter the timing of a **phytoplankton bloom**. In turn, this change may cause an entire year class of **larvae** to die if they depend on the timely availability of this food. In addition, because diseases are present in fish populations, the equivalent of epidemics can occur and decimate the stock. Therefore, fish stocks can vary dramatically from year to year and on even shorter timescales, and the maximum sustainable yield will vary accordingly.

If the size and condition of the stock were continuously monitored and known, the maximum sustainable yield could be continuously adjusted to accommodate changes. However, data cannot be gathered and analyzed quickly enough to do this. Consequently, maximum sustainable yields generally are estimated from the previous year's data. Estimates of the stock of adults and number of juveniles that enter the population are made each year. In addition, estimates are made of the survival rates of both adults and juveniles through the coming winter. These data are used to project what the stock will be in the following year, and the projection is then used to estimate maximum sustainable yield and set fishing limits for the following year. If an unexpected event occurs that adversely affects the population after the maximum sustainable yield has been estimated and fishing limits established, the permitted yield may be high enough to damage the stocks. If the unexpected decrease in the stock is recognized early enough in the fishing season, emergency measures can be taken to reduce fishing efforts. However, the stock size is often not well known until after the fishing season ends.

To account for uncertainties and variability in maximum sustainable yield, most fisheries set the permissible catch 20% to 40% below the estimated maximum sustainable yield. This practice leads to conflicts because some people feel that this safety margin is not enough to ensure protection of the stock, and others feel that part of the resource is being wasted because the maximum sustainable yield is not fully used.

This page titled 17.16: CC16 Maximum Sustainable Yield is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.17: CC17 Species Diversity and Biodiversity

Essential to Know

- **Species diversity is a well-defined scientific term that expresses a combination of species richness (number of species) and evenness (degree to which the community has balanced populations with no dominant species).**
- **High species diversity is generally equated with healthy ecosystems, but there are exceptions.**
- **Biodiversity is a poorly defined term that refers to a combination of genetic diversity (genetic variation within populations of a species), species diversity (species richness and evenness within a community), ecosystem diversity (variations in communities of species within an ecosystem), and physiological diversity (variations in feeding, reproduction, and predator avoidance strategies within a community or ecosystem).**

Understanding the Concept

The term **biodiversity** entered common usage when the global community of nations recognized the need to preserve species from extinction and unique **habitats** or **ecosystems** from destruction. The term has no precise definition but is based on the much more precisely defined technical term **species diversity**, which has been in use for many years.

Species diversity is a measure of species richness and evenness. *Richness* describes the number of individual species, whereas *evenness* expresses the degree to which a **community** has balanced populations in which there is not a small number of numerically dominant **species**. Several statistical indices have been developed to express both species richness and evenness as a single species diversity number.

Species richness is important because the larger the number of species present in a community, the more robust the community is considered to be. If there are many species, a community disturbance (such as a change in temperature or **salinity**) may lead to the exclusion or even extinction of one or a few species, but most species are likely to survive. Evenness is important because if the community is dominated by only a few species and these species are excluded or rendered extinct by a disturbance, competition between the remaining species to become the new dominant species may cause severe ecosystem instability. Alternatively, an uneven community may reflect effects of a disturbance, such as **pollution**, that favors dominance by tolerant species.

Although communities are generally thought to be desirable and stable if they contain many species and have strong evenness, those characteristics may not always be ideal. For example, **coral reefs** that are undisturbed by storm wave damage for many decades become dominated by just a few species of hard corals, and most other hard coral species are excluded by competition with these dominant species. Thus, the beautiful, untouched hard coral communities present in a few sheltered locations have low species richness, and species are unevenly distributed. These reefs have low species diversity (at least of corals).

In contrast, a coral reef that was damaged several years earlier by a **hurricane** has more species of corals (and possibly other species) and greater evenness, or higher species diversity. Few of us would consider this damaged **environment** an ideal situation, but the periodic disturbance may be necessary to restore and preserve the diversity of coral reefs since high species diversity is generally considered a positive attribute of any ecosystem.

During the hundreds of millions of years that life has existed on the Earth, virtually all the species that have lived have become extinct and been replaced by others. Because this natural process still continues today, none of the species now on the Earth are likely to be living a few million years in the future. Human disturbances, including pollution, have caused many species to become endangered or extinct. The rate of species extinction due to human disturbances is estimated to be many times faster than natural extinction, and possibly is much faster than the rate at which new species can evolve to replace those that are lost.

Only recently have scientists begun to realize that probably the best way to control the extinction rates of species is to control losses of their habitat and disturbances of their community structure. In response to the need to consider habitat and community structure in protecting individual species, a new concept was developed called “biodiversity.” Biodiversity is still often interpreted as species richness, but it should include many attributes of natural ecosystems. For example, biodiversity can be separated into four components:

- Genetic diversity, or variation in the genes within a species or population. High genetic diversity is thought to maximize the potential for new species development. It is also thought to maximize the potential for species survival when the species is subjected to an environmental disturbance because some members of the species may be more resistant.
- Species diversity, or species richness and evenness. High species diversity is thought to maximize the stability of the ecosystem and its resistance to environmental disturbances.
- Ecosystem diversity, or variation in the communities of species within an ecosystem. High ecosystem diversity is thought to reflect the availability of a wide variety of ecological niches. If the range of niches is large, environmental disturbance is unlikely to alter the ecosystem in such a way that more than a few species lack a niche within which to survive.
- Physiological diversity, or variation in physiological adaptations to feeding, reproduction, and predator avoidance within a community or ecosystem. High physiological diversity is thought to reflect a greater ability of the community within an ecosystem to adjust to environmental disturbances and maintain its stability.

A global agreement was established in 1993 to preserve biodiversity. Because *biodiversity* is a poorly defined term that encompasses many different characteristics of organisms in their living environment, this agreement will continue to be very difficult to translate into management actions. Protection of biodiversity may mean very different things to different nations or individuals.

This page titled 17.17: CC17 Species Diversity and Biodiversity is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.18: CC18 Toxicity

Essential to Know

- **A chemical is toxic if it can cause death or adverse sublethal effects in organisms exposed to it at a concentration above a critical threshold.**
- **Chemicals that are toxic can have many possible sublethal adverse effects on organisms, such as inhibition of the ability to photosynthesize or feed. The most important sublethal effects appear to be those that interfere with reproductive success.**
- **All chemicals that can be toxic have a concentration threshold below which they have no sublethal or lethal toxic effects. Hence, for every chemical in the ocean environment, there is a concentration below which it is environmentally safe.**
- **Many substances that are toxic at high concentrations are also essential to life, and the growth of marine organisms may be inhibited if they are not present at concentrations above a certain threshold. For these substances, there is an optimum range between the minimum concentration that an organism requires to meet its needs and the concentration above which the substance is toxic.**
- **The range of optimum concentration can be large or small. The concentration at which toxicity occurs can be many times higher than or very close to the natural range.**
- **Anthropogenic inputs of potentially toxic substances can be assimilated safely in the oceans if the amount introduced does not cause concentrations to exceed the threshold at which sublethal toxicity occurs. The quantity that can be safely assimilated is different for each substance and is determined in part by its sublethal toxicity threshold and background concentrations.**
- **Sublethal or lethal toxicity threshold concentrations are difficult to determine because they vary among species, among substances, and with other factors, such as physical stresses and synergistic and antagonistic effects of other chemical constituents.**
- **Marine organisms bioaccumulate most toxic substances. Bioaccumulation occurs when the concentration in the organism is higher than the environmental concentration, but the concentrations are in equilibrium.**
- **A few toxic substances are biomagnified in marine organisms. Biomagnification occurs when the organism retains all the toxic substance to which it is exposed in its food or environment and does not lose any of the substance, even if its environmental concentration decreases.**
- **Carcinogenicity, mutagenicity, and teratogenicity can be considered to be lethal or sublethal effects, but there is probably no concentration threshold below which there is no effect. Each exposed organism has a small probability of suffering an effect at any specific concentration. More individuals within a population suffer the effects as the concentration increases.**
- **Carcinogenic, mutagenic, and teratogenic substances occur naturally. Anthropogenic inputs will increase the incidence of the effects of such substances. However, at least for some of these substances, anthropogenic inputs would need to be large before the increased incidence would be significant or measurable in comparison with the natural incidence of effects.**

Understanding the Concept

A chemical is toxic if it is present at high enough concentrations to cause death or adverse sublethal effects in organisms, including human beings, that are exposed to the substance. If an organism is exposed to a concentration that is lethally toxic, the organism dies. If an organism is exposed to a concentration that is sublethally toxic, the organism is not killed, but it is disadvantaged in some way.

Examples of the many possible sublethal effects are partial inhibition of **photosynthesis** in plants and partial inhibition of feeding, hunting, or prey avoidance capabilities in animals. However, the most important sublethal effects are generally believed to be those that reduce reproductive success. Reproductive success can be reduced by partial inhibition of egg or sperm production, reduction in the probability of fertilization when egg and sperm meet, reduction in the survival ability of **larvae** or juveniles, and in other ways. Because of the wide range of possible sublethal effects caused by toxic chemicals, identifying and measuring such effects is often impossible either in the **environment** or in laboratory experiments.

The effects of any chemical that is potentially toxic vary with its concentration (**Fig. CC18-1**) and with individual species. Above

a certain concentration, the chemical can cause death. At somewhat lower concentrations, the same substance may produce sublethal adverse effects. Below an even lower concentration, the substance is not toxic as it causes no sublethal effects and does not harm the organism. This concentration is the sublethal toxicity threshold (Fig. CC18-1a). Any concentration below the sublethal toxicity threshold could be considered an acceptable concentration. However, this is not true for naturally occurring potentially toxic substances, because some are also essential **nutrients**. These substances have an optimum concentration range below the sublethal toxicity threshold, within which they do not harm the organism (Fig. CC18-1b). However, below this optimum range, some essential function is limited by a lack of the substance, and the organism is disadvantaged (Fig. CC18-1b).

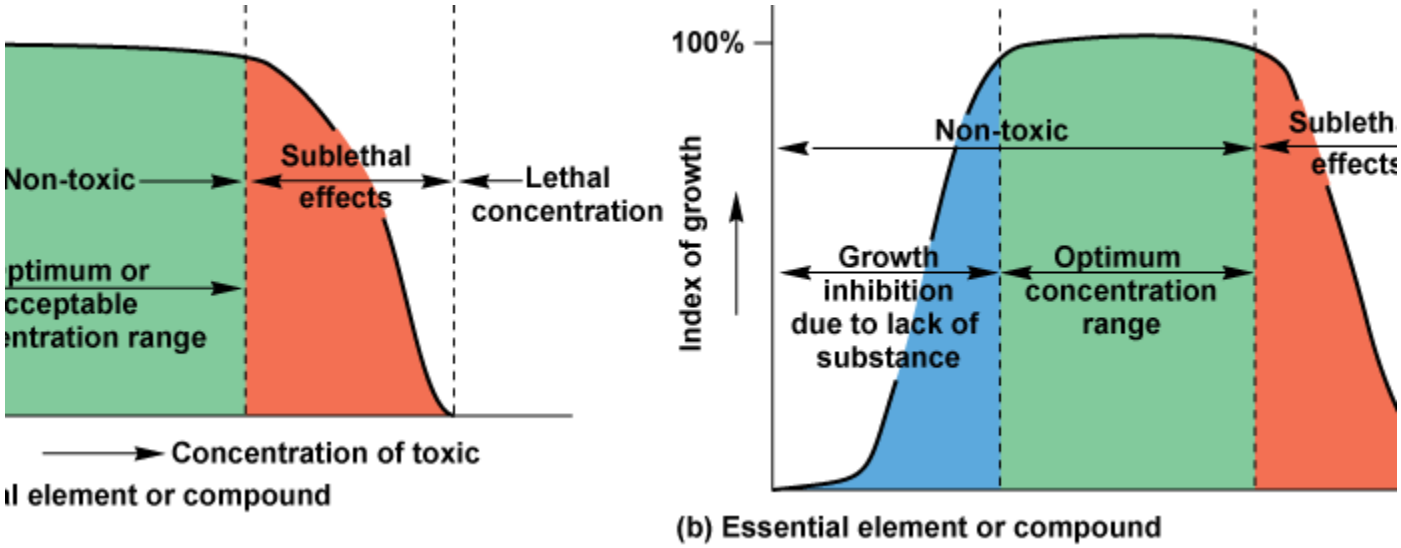


Figure CC18-1. Conceptual model of toxicity, showing the general behavior of two different types of toxic substances: (a) elements or compounds that are not essential to life, and (b) elements or compounds that are essential to living organisms for their growth and reproduction, but that become toxic to the organisms if their concentration is too high. The health index plotted on the y-axis for nonessential substances in (a) can be many different factors, such as reproductive success or incidence of disease. For essential substances (b), the index of growth can similarly be such factors as primary production rate or biomass. The concentrations at which transitions occur among growth inhibition, the optimum or nontoxic range, sublethal concentrations, and lethal concentrations vary among species, among life stages (e.g., juveniles and adults), and among toxic substances. They can even vary among individuals of a species or for a single individual, depending on the level of environmental stress that the individual is experiencing as a result of other factors.

All substances have an acceptable or optimum concentration range within which they do not harm organisms or the **ecosystem**. Therefore, **anthropogenic** additions of chemicals to the oceans are not always harmful. Such inputs can be safe if the **assimilative capacity** or acceptable concentration range in the environment is not exceeded. The assimilative capacity is determined by the sublethal toxicity threshold of the most sensitive species.

The relationships shown in Figure CC18-1 generally apply to all chemicals other than **carcinogens**, **mutagens**, and **teratogens**. However, concentrations at which the inhibition of growth ends (for essential nutrients), sublethal toxicity begins, and lethal toxicity occurs can all vary by orders of magnitude for different species, and even for different life stages of a species, exposed to a specific substance.

Most importantly, the optimum concentration range for essential nutrients may vary dramatically. For example, zinc is a toxic but essential chemical that has a wide optimum concentration range. In certain parts of the oceans, zinc may be the nutrient most likely to be depleted and hence most likely to restrict the growth of certain organisms. Because it has a large optimum concentration range, zinc becomes sublethally toxic only at concentrations that are many orders of magnitude above those normally present in the oceans. In contrast, copper has a very narrow optimum concentration range. Like zinc, copper may be growth-limiting to certain species in some parts of the oceans. However, at concentrations only a few times higher than the limiting concentration and within the range often present in **contaminated** coastal waters, copper can reduce the rate of photosynthesis by certain species of **phytoplankton**. Hence, copper is sublethally toxic at concentrations very close to those that occur naturally in the oceans.

Acceptable or optimum concentration ranges and the range of concentrations present naturally must both be considered in assessments of whether anthropogenic discharges might create **pollution** problems. For example, much smaller quantities of copper than of zinc can be discharged safely to the oceans.

Concentrations that represent the transitions between growth inhibition, nontoxicity, sublethal toxicity, and lethal toxicity are very difficult to determine for several reasons. First, members of a species differ somewhat in susceptibility to a given toxic chemical. Second, different species and life stages of a single species may have dramatically different susceptibilities to a particular chemical. Third, complicated, antagonistic, and synergistic effects occur among potentially toxic substances and other chemicals in the environment. Synergistic effects increase toxicity. For example, high concentrations of one trace metal can increase the toxicity of another trace metal. Antagonistic effects are the opposite; high concentrations of one trace metal may reduce the toxicity of another metal. Fourth, the susceptibility of a species to a chemical may be altered by stress due to other environmental factors, including temperature, **salinity**, and the extent of competition from other species. Fifth, the toxicity of many substances is determined by their chemical form. For example, **ionic** copper is substantially more toxic than copper **complexed** with organic matter.

Because of the variability among chemicals, among species, among species' life stages, among chemical states, and with synergistic, antagonistic, and stress factors, it is not possible to determine a single concentration at which a chemical is lethal or sublethal in the environment. For this reason, toxicologists often measure lethal toxic concentrations of chemicals on selected species in the laboratory by using **bioassays**. They assume that sublethal effects will occur at concentrations below the lethal concentration. They also assume that some species are more sensitive to the toxic chemical than the tested species is. A "safe" concentration in the environment is then estimated. Generally, a concentration two orders of magnitude below the minimum lethal concentration is arbitrarily assumed to be safely below the level at which any sublethal adverse effects occur. However, in some cases, such as copper, the optimum concentration range (**Fig. CC18-1b**) is less than two orders of magnitude below the lethal concentration threshold. Therefore, this approach may set the "safe" concentration below naturally occurring levels and sometimes even below the level needed to sustain growth.

Bioassays are conceptually simple. The test species is usually exposed for 96 h (a duration chosen because one set of tests can be set up and run in a 5-day workweek) to several different concentrations of the potentially toxic substances in seawater. The test species is also exposed to control seawater with none of the potentially toxic substances added. After exposure, the number of organisms that have died or exhibited a sublethal effect (e.g., failure of an **embryo** to develop or of a larval stage to **metamorphose** to the next life stage) are counted. Generally, no death or sublethal effect occurs in the control seawater (otherwise the experiment is not acceptable, because a factor other than the tested contaminant has caused a confounding effect). If the test substance is toxic, some or all individuals will have died or shown the sublethal effect in the test with the highest contaminant concentration. A progressively smaller proportion of the test population is usually affected at lower contaminant concentrations. From these tests, the contaminant concentrations at which one-half of the test organisms are killed or suffer the observed sublethal effect are calculated. These values are called the LC_{50} and EC_{50}, respectively.

Bioassay tests must be interpreted carefully because they do not use the most sensitive organism in the ecosystem being assayed. Without testing all species, there is no way to determine which is the most sensitive. In fact, because different life stages of some species have different levels of sensitivity (juvenile stages are generally more sensitive than adults), all life stages of all species would have to be tested. In addition, most marine species are difficult to keep alive in the laboratory because this alien environment stresses the organisms. Species used for bioassays must be stress-tolerant, and hence they are also likely to be tolerant of toxicant stress.

Many factors, such as salinity, temperature, light intensity, water chemistry, **sediment** characteristics, food supply, and competition or cooperation with other individuals or species, are different in bioassays than in the environment. The additional stresses of the test environment may make test organisms more susceptible to toxicant stress.

Bioassays are often conducted with samples of wastes in the form discharged (e.g., sewage or industrial **effluent**). Although this approach takes into account synergism or antagonism between components of the waste, there remains the question of whether additional such effects occur when the waste is mixed in seawater and subject to different conditions in the environment.

Most chemicals that exhibit toxicity are more highly concentrated in tissues of marine species than in the surrounding seawater. This fact has led to a popular belief that toxic chemicals are taken up continuously by marine species and that concentrations of the toxins in their tissues increase progressively throughout their lifetime, regardless of how the concentration changes in their environment. This may be true for some species and some chemicals, but it is not true for most species and most substances. Most potentially toxic chemicals are **bioaccumulated** by most marine species, but are not **biomagnified**.

Bioaccumulation occurs when the concentration of a compound is regulated by equilibria between the organism and its food and/or surrounding water. The substance is taken in by the organism from its food and/or directly from the water, but the organism is capable of **excreting** some of the substance either directly back to the water or through its urine or feces. Hence, if the food or seawater concentration of the substance increases, the concentration in the organism will increase, but if the food or seawater concentration decreases, the concentration in the organism will also decrease. However, it may take some time for equilibrium to

be reached.

Certain toxic substances are more dangerous than others in marine ecosystems because they biomagnify. Biomagnification occurs when organisms at each **trophic level** in a **food chain** retain all or almost all of a toxic substance ingested in food (**CC15**). Consequently, the concentration of the toxin increases at each trophic level.

Certain chemicals are carcinogenic (cancer-causing), mutagenic (causing **genetic** changes in the offspring by altering the parental **DNA**), or teratogenic (causing abnormal development of the embryo). Although each can be considered a sublethal effect, many scientists do not believe that these effects follow the toxicity-concentration relationships shown in **Figure CC18-1**, but rather that they may occur at any concentration of these substances.

If this hypothesis is correct, a single molecule of such a chemical could cause cancer in an exposed individual. However, if many individuals were each exposed to a single molecule, only a few (if any) would be expected to develop cancer. If the same group of individuals were each exposed to two molecules of the chemical, twice as many would be expected to develop cancer. This type of effect is said to be “probabilistic.” The probability that any one individual will contract cancer from a given concentration of the chemical is reduced proportionally as the concentration is reduced, but it never reaches zero.

There are many naturally occurring carcinogens, teratogens, and mutagens, but only a very small percentage of marine populations suffer adverse effects as a result of these naturally occurring compounds. Because such a small fraction is affected, great numbers of individuals would have to be monitored to accurately estimate the percentage of a population that suffers such effects. Furthermore, determining the extent to which a higher incidence of such effects might be caused by human contamination of the oceans, or parts thereof, would be even more difficult.

If the probabilistic-effect hypothesis is correct, any quantity of anthropogenic input of a carcinogen, mutagen, or teratogen will increase the incidence of such effects in marine species. The increase may be extremely small in comparison with the incidence of such effects due to the naturally occurring concentrations, in which case many scientists believe adverse effects on the marine ecosystem would be negligible. Hence, some believe that it would be environmentally safe and acceptable to dispose of limited quantities of anthropogenic carcinogens, mutagens, and teratogens in the ocean.

This page titled 17.18: CC18 Toxicity is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.19: CC19 Metagenomics

Essential to Know

- **Within oceans, there are diverse microbial biomes, which are ecological niches where microbes interact with one another and their surrounding environment.**
- **Metagenomics can tell us what organisms are there and what they are capable of doing, but can't tell us if they are actually doing it. Other related techniques are necessary to identify which species and genes are active and how this information impacts our understanding of marine ecosystems and ocean chemistry.**
- **Bioinformatics is used to assemble and analyze metagenomic data using powerful computers.**
- **Metagenomic studies have revealed a vast array of microbes in marine environments, many of which have never been identified, let alone cultured in the lab.**
- **Viruses are present in significantly higher numbers than any other organism in the ocean, and recent metagenomic and related studies have shed light on their impact.**

Understanding the Concept

Within the ocean, a wide array of microorganisms actively engage in biogeochemical cycling while interacting with one another as well. The complex interactions exist within smaller ecological niches known as microbial **biomes.** These biomes comprise a diverse group of organisms, including bacteria, algae, viruses, and archaea (**Fig. CC19-1**). It is important not only to understand the role these biomes play in marine ecosystems but also to identify the players involved.

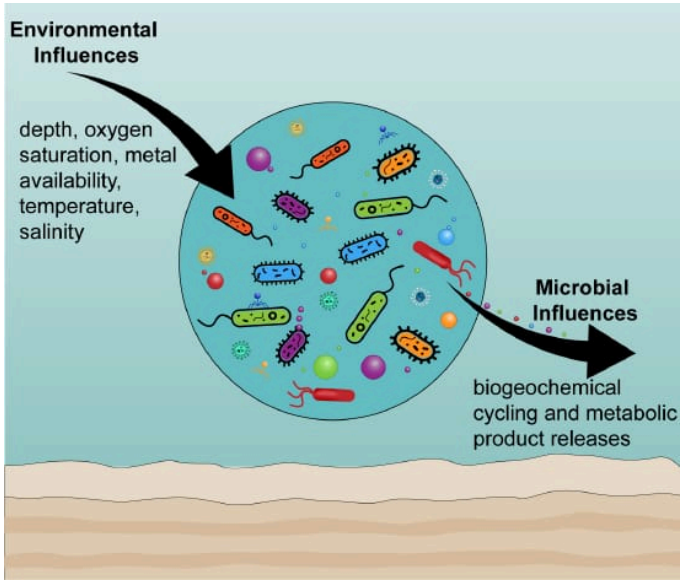


Figure CC19-1. Microbial biomes consist of thousands of microorganisms, including viruses, bacteria, and archaea. The community structure is shaped by the chemical and physical elements (sunlight, oxygen saturation, metal availability, temperature, salinity) within the environment. These diverse communities play a critical role in biogeochemical cycling and nutrient availability and can respond rapidly to environmental changes.

Investigating how microbial niches exist within marine environments has historically been challenging, primarily because many of the microbes found there cannot be cultured (grown in a lab). It was not until the discovery of **DNA sequencing** that we truly began to understand the diversity of microbes that exist in ocean biomes. **DNA**, which is the genetic makeup (code) of an organism and contains all the information needed to perform the life processes of that species, can be sequenced so that every nucleotide that makes up that code can be identified.

Thousands of novel organisms have been identified since DNA sequencing technology, primarily **metagenomics**, was discovered. Despite extensive efforts to better characterize marine microbes using DNA sequencing, a large portion of microbes in ocean

ecosystems have yet to be identified.

Metagenomics is the analysis of all the genomic DNA in an environmental sample that contains multiple species (**Fig. CC19-2**). Metagenomics data provides information on what organisms are present in the sample and what metabolic capabilities they have. This allows scientists to infer what they may be doing in the microbial biome, including their possible involvement in biogeochemical cycling. Time-lapse studies or comparative studies can be used to determine how communities change over time and space, as well as how they react to environmental changes. It is important to note that the limitation of metagenomics is that while it can tell us what genes are present, it can't tell us if or how often those genes are in use in the cell's life processes. Essentially, metagenomics provides the map of that organism, similar to a map of a community, but we can't know for sure that every business on the map is open.

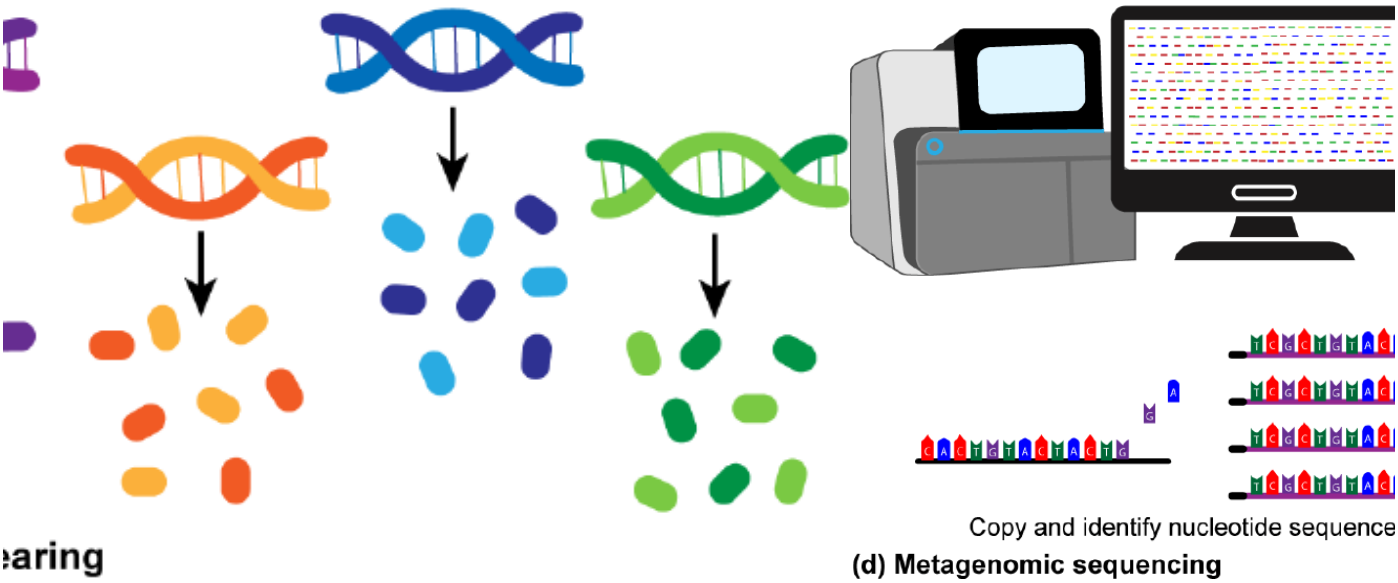


Figure CC19-2. The metagenomics process. (a) Marine samples are collected either via water collection or sediment coring. (b) All of the DNA in the sample is extracted. (c) The DNA is sheared, or broken into smaller pieces, for sequencing. (d) Samples are sequenced where millions of copies of the fragmented DNA are made, and the nucleotide bases of those copies are identified. Sequences are returned as reads that report the nucleotide sequence and quality (Phred score) of each DNA strand. (e) Reads are aligned based on overlapping nucleotides, and similar sequences are binned (grouped) together to form metagenome-assembled genomes (MAGs). (f) The MAGs are then analyzed to identify the organisms present (taxonomic data), their metabolic pathways, and the dominant functions of the organisms present (functional data), as well as the overall community interactions with one another and the environment (comparative data).

Samples can be taken from any type of environment (marine environments, soil, feces, gut) without any prior knowledge or understanding of the microorganisms present. Within marine environments, samples are either collected by seafloor coring or water collection. A sediment push core can be used to sample ocean floors and is typically then separated into about 2 cm sections before total DNA extraction is completed. Water sampling involves the collection of a large volume of water, at least 2 to 3 L, which can be collected at varying depths. Once collected, the water is processed in the lab using filtration methods to concentrate the microbes in the sample. The filter with the concentrated microbes is then processed for total DNA extraction. Once DNA has been extracted and confirmed to be of sufficient concentration and quality, the samples are ready for sequencing.

Metagenomics is the sequencing, or reading of the nucleotide bases (A, G, C, T) that make up DNA in the correct order, of all of the DNA present in a sample. In order to do this, the DNA is broken, typically through shearing, into pieces of approximately the same length. The size that the DNA is broken into depends on the sequencing technology, where the typical size for short-reads is around 400 nucleotides. There are two primary ways of shearing DNA: mechanical and enzymatic. Mechanical shearing is often completed using sonication, which transmits high-frequency, short-wavelength acoustic energy, breaking the DNA into smaller pieces at random spots in the DNA. Enzymatic shearing uses specific and non-specific enzymes to nick and cut the DNA. Shearing the DNA results in millions of sequences that need to be read then reassembled and analyzed (**Fig. CC19-2**). The sequences are typically recorded in text-based files that store DNA sequences with associated quality scores.

DNA sequences generally fall into two categories: short-read and long-read. The difference between short- and long-read sequencing refers to the length of DNA that is copied in a single sequence. Short-reads are typically about 400 nucleotides in length, while long-reads range from 1500 to >10,000 nucleotides in length. The longer read lengths are used to account for long sequences of repeats in the DNA and areas where shorter runs missed DNA sequencing. Shorter reads tend to be more

accurate, making them more reliable (e.g., higher quality) than some long-read technologies. While other long-read technologies are extremely accurate (99.99%), their cost makes their use prohibitive. Often, to maximize coverage, a mix of short- and long-read sequencing is used, balancing high throughput and affordability. Different sequencing technologies vary in how the sequencing is performed, but the overall output is similar.

Before sequences can be reassembled, it is important to check the quality of the sequences and discard any low-quality sequences. This is a crucial step because low-quality reads have a high error rate and can result in the incorrect identification of a gene or even an organism. Quality is measured using **Phred quality scores**, which reflect the confidence that a nucleotide base read is correct. Phred scores (Q-score) typically range from 10 to 50 on a logarithmic scale, where a lower Phred score indicates a higher likelihood of error. For example, a Phred score of 20 (Q20) indicates that there is a 1 in 100 (1%) chance that a nucleotide is incorrect, and a Phred score of 40 (Q40) indicates that there is a 1 in 10,000 (0.01%) chance that a nucleotide is incorrect. The generally accepted benchmark for quality is a Phred score of 30 (Q30). Phred scores are reported in the final sequencing files.

Metagenomic data is assembled and analyzed using **bioinformatics.** Because the organisms in the original sample are unknown, the sequences must be reassembled without a template; this is known as ***de novo* assembly**. To do this, the sequences are aligned, or scaffolded, by looking for areas where the reads overlap. This process is repeated until there are no more available overlaps. The assembled reads are referred to as **contigs**. The contigs then undergo a process known as **binning**, which groups contigs that likely come from the same or closely related genomes. Once grouped, genomes are assembled to create **metagenome-assembled genomes (MAGs)**. The MAGs are then analyzed to identify the organisms in the sample, and identify novel metabolic pathways and the specific genes present. This allows assessment of the potential role each microbe may have in the system.

Due to the sheer number of reads, processing them requires a high amount of computational power and is most often completed using supercomputers.

Different types of analysis of the assembled MAGs can be performed in order to answer many different questions. The data analyzed can be organized into three types:

- Taxonomic data provides insight into the identities of the organisms present in a sample. This includes matching the sequences to a database of known sequences. From there, the **relative abundance** (approximation) of specific species can be determined, and the taxonomic diversity can be examined. Additionally, novel species can be identified.
- Functional data includes annotating all known genes and identifying the metabolic pathways present for each organism. Additionally, the most abundant genes (**functional abundance**) in the system and the dominant functions in the system can be determined.
- Comparative data looks at how the MAGs interact within the biome in relation to one another and the environment as a whole. The types of analysis performed include correlation plots of interactions with the environment, such as depth, oxygen saturation, metal availability, temperature, and salinity, as well as between species (co-occurrence networks).

To organize and share sequencing data from the marine biome, the KMAP Global Ocean Gene Catalog 1.0, the largest open-source database of ocean microbes, was developed. At the time of its introduction in 2024, the catalog comprised 308.6 million gene groups from over 2100 ocean samples.

As mentioned above, metagenomics can be used to identify which microorganisms comprise the sampled biome and which genes these microorganisms possess, but it cannot tell us whether these genes are active. Many genes are inactive or are only active under certain environmental conditions. DNA is the repository for all the information needed to perform the life processes of that species. Active genes in this DNA library are passed to RNA, where they act as instructions to run the cell's life processes, including the construction of essential compounds such as amino acids, other organic acids, sugars, lipids, and proteins. One particular set of RNA, called mRNA (messenger RNA), contains the information needed for the cell to construct proteins.

Determining which genes are active in an environmental sample is difficult, but can be addressed by using one or more of three techniques: metatranscriptomics, proteomics, and metabolomics. Metatranscriptomics is performed in a similar way to metagenomics, except that it extracts and analyzes mRNA in the environmental sample, so it provides information on which genes are active in protein construction. Proteomics and metabolomics analyze environmental samples: proteomics for proteins and metabolomics for various other compounds, including amino acids, other organic acids, sugars, and lipids. Each of these three techniques involves careful sample preparation and analysis and produces a large amount of data that must be processed. Further, the results need to be carefully interpreted to obtain accurate information about which genes identified by metagenomic analysis are actually in use under the environmental conditions present when each sample was collected.

This page titled 17.19: CC19 Metagenomics is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

17.20: References

Figure CC-1 – NASA, 2007: Blue Marble – Image of the Earth from Apollo 17. Accessed 21 November 2025, Source.

Figure CC04-01 – Sundborg, Å. 1956: The River Klarälven a study of fluvial processes. *Geografiska Annaler*, 38, 125–316, Source.

Figure CC09-01 – Scott, R. C., 1996: *Introduction to Physical Geography*. West Publishing, 560 pp, Source.

Figure CC11-03 – Adapted from Gleick, J., 1987: *Chaos: Making a New Science*, Viking Penguin, 400 pp.

Figure CC11-04 – Adapted from Gleick, J., 1987: *Chaos: Making a New Science*, Viking Penguin, 400 pp.

This page titled 17.20: References is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

INDEX

A

absorbed
- 14.8: Communication and Navigation

absorption
- 12.6: Primary Production and Nutrients
- 17.14: CC14 Phototrophy, Light, and Nutrients
- 17.9: CC9 The Global Greenhouse Effect
- 5.11: Transmission of Light and Other Electromagnetic Radiation

abyssal fans
- 4.10: Present-Day Oceans

abyssal plains
- 4.10: Present-Day Oceans
- 6.7: Sediment Transport, Deposition, and Accumulation

abyssal zone
- 12.11: Biological Provinces and Zones
- 15.7: Beyond the Sun's Light
- 15.8: Hydrothermal Vents

abyssopelagic zone
- 12.11: Biological Provinces and Zones

acid rain
- 5.6: Dissolved Chemicals in Seawater
- 6.1: Sediments and Biogeochemical Cycles

acidic
- 5.6: Dissolved Chemicals in Seawater

acidification
- 1.1: The Ocean's and Earth's Environment
- 11.1: Processes that Form and Modify the Coast
- 13.5: Fisheries
- 16.6: Climate Change, Acidification, and Deoxygenation
- 16.7: Risk Assessment and Management
- 6.10: The Sediment Historical Record
- 6.4: Biogenous Sediments

acoustic thermometry
- 5.12: Transmission of Sound

acoustic tomography
- 3.5: Chemical and Physical Oceanography

adaptation
- 14.1: Ecological Requirements
- 14.4: Hunting and Defense
- 14.8: Communication and Navigation

adiabatic expansion
- 7.10: Local Weather Effects
- 7.2: Water Vapor in the Atmosphere

Adriatic Sea
- 13.4: Algal Blooms

aerobic respiration
- 12.9: Dissolved Oxygen and Carbon Dioxide

Afghanistan
- 4.4: Convergent Plate Boundaries

Africa
- 10.5: Tides in the Open Ocean
- 12.11: Biological Provinces and Zones
- 13.2: Nutrient Supply to the Coastal Photic Zone
- 2.2: Exploration and Mapping
- 4.10: Present-Day Oceans
- 4.2: Plate Tectonics
- 4.4: Convergent Plate Boundaries
- 4.5: Divergent Plate Boundaries
- 4.8: Plate Interiors
- 4.9: Sea-Level Change and Climate
- 6.9: Distribution of Surface Sediments
- 7.10: Local Weather Effects
- 7.4: Climatic Winds
- 7.6: Interannual Climate Variations
- 8.14: Ocean Circulation and Climate
- 8.5: Open-Ocean Surface Currents
- 9.9: Tsunamis

African Plate
- 4.10: Present-Day Oceans
- 4.2: Plate Tectonics

agriculture
- 11.7: Deltas
- 13.4: Algal Blooms
- 16.5: Ocean Pollution Issues

Agulhas Current
- 8.10: Eddies
- 8.5: Open-Ocean Surface Currents
- 9.5: Sea Development and Wave Height

air density
- 7.2: Water Vapor in the Atmosphere

air masses
- 7.2: Water Vapor in the Atmosphere
- 7.7: Land-Ocean-Atmosphere Interactions

Alaska
- 10.2: Characteristics of Tides
- 10.6: Tidal Currents
- 11.1: Processes that Form and Modify the Coast
- 13.6: Estuaries
- 14.6: Reproduction
- 15.3: Kelp Forests
- 16.5: Ocean Pollution Issues
- 16.9: Box 1 - Petroleum in the Marine Environment
- 2.3: National and International Law Applied to Oceans
- 2.4: Ocean Resources
- 4.10: Present-Day Oceans
- 4.4: Convergent Plate Boundaries
- 6.3: Lithogenous Sediments
- 6.9: Distribution of Surface Sediments
- 7.8: Global Climate Zones
- 9.9: Tsunamis

Alaska King crab
- 1.1: The Ocean's and Earth's Environment

Albemarle Sound
- 13.6: Estuaries

Alcatraz Island
- 13.6: Estuaries
- 16.5: Ocean Pollution Issues

Aleutian Basin
- 4.10: Present-Day Oceans

Aleutian Island Plate
- 4.4: Convergent Plate Boundaries

Aleutian Islands
- 11.1: Processes that Form and Modify the Coast
- 4.3: Plate Boundaries
- 4.4: Convergent Plate Boundaries
- 7.8: Global Climate Zones

Aleutian Trench
- 9.9: Tsunamis

algae
- 11.2: Beaches
- 12.10: Organic Carbon
- 12.5: Phototropic Primary Production and Light
- 13.1: Special Characteristics of Coastal Oceans
- 14.2: Habitats
- 14.3: Feeding
- 14.4: Hunting and Defense
- 14.7: Associations
- 15.2: Coral Reefs
- 15.4: Rocky Intertidal Communities
- 15.6: Polar Regions
- 16.2: Adverse Effects of Human Activities
- 17.19: CC19 Metagenomics

algal ridge
- 15.2: Coral Reefs

alkaline
- 5.6: Dissolved Chemicals in Seawater

Alps
- 4.4: Convergent Plate Boundaries

Aluetian Basin
- 4.10: Present-Day Oceans

Aluetian Sea
- 4.10: Present-Day Oceans

Alvin
- 15.7: Beyond the Sun's Light
- 15.8: Hydrothermal Vents
- 3.7: Scuba, Manned, and Unmanned Submersibles

Amazon River
- 13.2: Nutrient Supply to the Coastal Photic Zone
- 4.10: Present-Day Oceans
- 6.3: Lithogenous Sediments
- 6.9: Distribution of Surface Sediments
- 7.5: Climate and Ocean Surface Water Properties

American alligators
- 12.13: Nekton

American Samoa
- 2.3: National and International Law Applied to Oceans

amino acid
- 1.2: Oceans and the Origins of Life

ammonia
- 12.6: Primary Production and Nutrients
- 12.9: Dissolved Oxygen and Carbon Dioxide

Amoco Cadiz
- 16.9: Box 1 - Petroleum in the Marine Environment

amoeba
- 12.12: Plankton

amphidromic systems
- 10.4: Tides in the Earth's Oceans
- 10.5: Tides in the Open Ocean

amphipods
- 12.12: Plankton
- 15.7: Beyond the Sun's Light
- 15.8: Hydrothermal Vents

amplitude
10.7: Tide Prediction
9.1: Complexity of Ocean Waves
anadromous
13.6: Estuaries
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.8: Communication and Navigation
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
anaerobic
17.14: CC14 Phototrophy, Light, and Nutrients
anchovies
12.13: Nekton
13.5: Fisheries
7.6: Interannual Climate Variations
ancient civilizations
2.2: Exploration and Mapping
Andes
7.5: Climate and Ocean Surface Water Properties
Andes Mountains
4.4: Convergent Plate Boundaries
7.10: Local Weather Effects
anemone crabs
14.7: Associations
anemonefishes
14.4: Hunting and Defense
14.6: Reproduction
14.7: Associations
anemones
12.12: Plankton
14.3: Feeding
14.4: Hunting and Defense
14.7: Associations
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
16.6: Climate Change, Acidification, and Deoxygenation
angelfishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
angle of incidence
12.5: Phototropic Primary Production and Light
7.3: Water and Heat Budgets
anglerfishes
14.4: Hunting and Defense
Angola
12.9: Dissolved Oxygen and Carbon Dioxide
angular momentum
8.17: Box 1- Westward Intensification
angular velocity
17.12: CC12 The Coriolis Effect
animals
12.1: How do we describe life?
anions
5.3: The Dissolving Power of Water
annelid worms
14.3: Feeding
anoxic
12.9: Dissolved Oxygen and Carbon Dioxide
13.4: Algal Blooms
14.3: Feeding
15.8: Hydrothermal Vents
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
6.5: Hydrogenous Sediments
Antarctic
12.11: Biological Provinces and Zones
12.7: Food Webs
15.6: Polar Regions
16.6: Climate Change, Acidification, and Deoxygenation
17.5: CC5 Transfer and Storage of Heat by Water
4.10: Present-Day Oceans
4.2: Plate Tectonics
7.1: The Atmosphere
8.7: High-Latitude Surface Currents
Antarctic Bottom Water
8.13: Thermohaline Circulation
Antarctic Circumpolar Current
8.5: Open-Ocean Surface Currents
8.7: High-Latitude Surface Currents
8.8: Upwelling and Downwelling
Antarctic Convergence
15.6: Polar Regions
8.13: Thermohaline Circulation
Antarctic Divergence
12.8: Geographic Variation in Primary Production
Antarctic fur seals (Arctocephalus gazella)
15.6: Polar Regions
Antarctic Penisula
4.10: Present-Day Oceans
Antarctic Plante
4.2: Plate Tectonics
Antarctic Plate
4.10: Present-Day Oceans
Antarctica
10.5: Tides in the Open Ocean
12.12: Plankton
12.8: Geographic Variation in Primary Production
16.6: Climate Change, Acidification, and Deoxygenation
4.9: Sea-Level Change and Climate
6.3: Lithogenous Sediments
6.9: Distribution of Surface Sediments
7.4: Climatic Winds
8.14: Ocean Circulation and Climate
8.8: Upwelling and Downwelling
9.5: Sea Development and Wave Height
anthropogenic
12.13: Nekton
12.9: Dissolved Oxygen and Carbon Dioxide
15.6: Polar Regions
16.1: Pollution versus Contamination
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
17.10: CC10 Modeling
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.9: CC9 The Global Greenhouse Effect
6.1: Sediments and Biogeochemical Cycles
antinodes
10.4: Tides in the Earth's Oceans
10.6: Tidal Currents
9.12: Standing Waves
Apatosaurus
12.13: Nekton
aphotic zone
12.11: Biological Provinces and Zones
12.5: Phototropic Primary Production and Light
14.2: Habitats
15.7: Beyond the Sun's Light
5.11: Transmission of Light and Other Electromagnetic Radiation
Appalachian Mountains
4.4: Convergent Plate Boundaries
4.9: Sea-Level Change and Climate
aquaculture
13.5: Fisheries
Arabeian Sea
6.9: Distribution of Surface Sediments
Arabian Gulf
7.5: Climate and Ocean Surface Water Properties
Arabian Penisula
7.6: Interannual Climate Variations
Arabian Sea
12.9: Dissolved Oxygen and Carbon Dioxide
4.10: Present-Day Oceans
6.3: Lithogenous Sediments
archaea
1.2: Oceans and the Origins of Life
12.1: How do we describe life?
12.10: Organic Carbon
12.12: Plankton
12.5: Phototropic Primary Production and Light
14.2: Habitats
14.3: Feeding
14.7: Associations
15.8: Hydrothermal Vents
17.14: CC14 Phototrophy, Light, and Nutrients
17.19: CC19 Metagenomics
3.6: Sampling Living Organisms in the Sea
6.4: Biogenous Sediments
Arctic
12.11: Biological Provinces and Zones
13.3: Seasonal Cycles
15.6: Polar Regions
16.6: Climate Change, Acidification, and Deoxygenation
7.6: Interannual Climate Variations
8.14: Ocean Circulation and Climate
Arctic Ice Cap
4.5: Divergent Plate Boundaries
Arctic Ocean
12.11: Biological Provinces and Zones
13.3: Seasonal Cycles
15.6: Polar Regions
15.8: Hydrothermal Vents
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
17.5: CC5 Transfer and Storage of Heat by Water
2.4: Ocean Resources
4.10: Present-Day Oceans
6.9: Distribution of Surface Sediments
7.5: Climate and Ocean Surface Water Properties
7.8: Global Climate Zones
Arctic Oscillation
7.6: Interannual Climate Variations

Argentina
2.3: National and International Law Applied to Oceans
Argo
3.7: Scuba, Manned, and Unmanned Submersibles
armor
14.4: Hunting and Defense
arrowhead crab (Huenia heraldica)
14.7: Associations
arsenic
16.1: Pollution versus Contamination
16.3: Toxicity
asexual reproduction
12.12: Plankton
14.6: Reproduction
15.3: Kelp Forests
Asia
11.1: Processes that Form and Modify the Coast
12.13: Nekton
14.6: Reproduction
4.2: Plate Tectonics
6.3: Lithogenous Sediments
7.4: Climatic Winds
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
7.9: Weather Systems
8.5: Open-Ocean Surface Currents
aspect ratio
14.5: Selected Adaptations in Fishes
assimilative capacity
16.1: Pollution versus Contamination
16.11: Box 3 - Radioactive Wastes in the Marine Environment
17.18: CC18 Toxicity
asthenosphere
17.1: CC1 Density and Layering in Fluids
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.1: Structure of the Earth
Atlantic
7.5: Climate and Ocean Surface Water Properties
Atlantic cod (Gadus morrhua)
12.13: Nekton
Atlantic eels
14.6: Reproduction
Atlantic Ocean
10.2: Characteristics of Tides
10.5: Tides in the Open Ocean
10.8: Marine Species and the Rhythm of Tides
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons
11.7: Deltas
12.11: Biological Provinces and Zones
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.6: Estuaries
14.6: Reproduction
15.2: Coral Reefs
15.5: Sargasso Sea
15.8: Hydrothermal Vents
16.6: Climate Change, Acidification, and Deoxygenation
17.8: CC8 Residence Time
2.2: Exploration and Mapping
2.3: National and International Law Applied to Oceans
4.10: Present-Day Oceans
4.4: Convergent Plate Boundaries
4.5: Divergent Plate Boundaries
4.7: Hot Spots
4.8: Plate Interiors
6.3: Lithogenous Sediments
6.4: Biogenous Sediments
6.5: Hydrogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments
6.9: Distribution of Surface Sediments
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
7.8: Global Climate Zones
7.9: Weather Systems
8.10: Eddies
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.9: Coastal Currents
9.5: Sea Development and Wave Height
atmosphere
15.4: Rocky Intertidal Communities
17.3: CC3 Convection and Convection Cells
17.8: CC8 Residence Time
4.1: Structure of the Earth
5.1: Origins and Distribution of the Earth's Water
7.1: The Atmosphere
7.10: Local Weather Effects
7.3: Water and Heat Budgets
atmospheric circulation
7.4: Climatic Winds
7.7: Land-Ocean-Atmosphere Interactions
atmospheric pressure
7.9: Weather Systems
atolls
11.5: Coral Reefs and Atolls
4.8: Plate Interiors
9.9: Tsunamis
atoms
5.2: The Water Molecule
attractive forces
5.10: Surface Tension and Viscosity
Aurelia
12.12: Plankton
Australia
11.5: Coral Reefs and Atolls
13.2: Nutrient Supply to the Coastal Photic Zone
2.2: Exploration and Mapping
3.4: Seafloor Sediments
4.10: Present-Day Oceans
4.2: Plate Tectonics
6.10: The Sediment Historical Record
6.9: Distribution of Surface Sediments
7.4: Climatic Winds
7.6: Interannual Climate Variations
7.9: Weather Systems
8.5: Open-Ocean Surface Currents

autonumheader:yes2
1: The Ocean Planet
1.1: The Ocean's and Earth's Environment
1.2: Oceans and the Origins of Life
1.3: How to Study Ocean Data
10: Tides
10.1: Tide-Generating Forces
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
10.7: Tide Prediction
10.8: Marine Species and the Rhythm of Tides
10.9: Energy from Tides
11: Coasts
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures
11.5: Coral Reefs and Atolls
11.6: Wetlands
11.7: Deltas
12: Foundations of Life in the Oceans
12.1: How do we describe life?
12.10: Organic Carbon
12.11: Biological Provinces and Zones
12.12: Plankton
12.13: Nekton
12.14: Benthos
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
12.4: Studying Microbes in the Ocean
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.7: Food Webs
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13: Coastal Oceans and Estuaries
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.5: Fisheries
13.6: Estuaries
14: Marine Ecology
14.1: Ecological Requirements
14.2: Habitats
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.7: Associations
14.8: Communication and Navigation
15: Ocean Ecosystems
15.1: Communities and Niches
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
15.6: Polar Regions
15.7: Beyond the Sun's Light
15.8: Hydrothermal Vents
16: Impacts of Humans on the Oceans
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.4: Waste Disposal
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
16.9: Box 1 - Petroleum in the Marine Environment
2: History and Importance of Ocean Studies
2.1: What is Oceanography
2.2: Exploration and Mapping
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
3: Studying the Oceans
3.1: Difficulties of Studying the Ocean Environment
3.2: The Unseen Domain
3.3: Bathymetry
3.4: Seafloor Sediments
3.5: Chemical and Physical Oceanography
3.6: Sampling Living Organisms in the Sea
3.7: Scuba, Manned, and Unmanned Submersibles
3.8: Satellites
3.9: Computers and Modeling
4: Plate Tectonics-Evolution of the Ocean Floor
4.1: Structure of the Earth
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.5: Divergent Plate Boundaries
4.6: Transform Faults and Fracture Zones
4.7: Hot Spots
4.8: Plate Interiors
4.9: Sea-Level Change and Climate
5: Water and Seawater
5.1: Origins and Distribution of the Earth's Water
5.10: Surface Tension and Viscosity
5.11: Transmission of Light and Other Electromagnetic Radiation
5.12: Transmission of Sound
5.2: The Water Molecule
5.3: The Dissolving Power of Water
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
5.5: Salinity
5.6: Dissolved Chemicals in Seawater
5.7: Heat Properties of Water
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
5.9: Ice Formation
6: Ocean Sediments
6.1: Sediments and Biogeochemical Cycles
6.10: The Sediment Historical Record
6.2: Classification of Sediments
6.3: Lithogenous Sediments
6.4: Biogenous Sediments
6.5: Hydrogenous Sediments
6.6: Cosmogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments
6.9: Distribution of Surface Sediments
7: Ocean-Atmosphere Interactions
7.1: The Atmosphere
7.10: Local Weather Effects
7.2: Water Vapor in the Atmosphere
7.3: Water and Heat Budgets
7.4: Climatic Winds
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
7.7: Land-Ocean-Atmosphere Interactions
7.8: Global Climate Zones
7.9: Weather Systems
8: Ocean Circulation
8.1: Energy Sources
8.10: Eddies
8.11: Inertial Currents
8.12: Langmuir Circulation
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
8.17: Box 1- Westward Intensification
8.18: Box 2 - Equatorial Currents
8.2: Wind-Driven Currents
8.3: Ekman Motion
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
8.7: High-Latitude Surface Currents
8.8: Upwelling and Downwelling
8.9: Coastal Currents
9: Waves
9.1: Complexity of Ocean Waves
9.10: Internal Waves
9.11: Rossby and Kelvin Waves
9.12: Standing Waves
9.2: Progressive Waves
9.3: What is Wave Motion?
9.4: Making Waves
9.5: Sea Development and Wave Height
9.6: Wave Dissipation
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
9.9: Tsunamis
Introduction to Ocean Sciences 6e (Segar)

autotrophs
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.7: Food Webs

AUV
10.5: Tides in the Open Ocean
3.3: Bathymetry
3.7: Scuba, Manned, and Unmanned Submersibles

B

backscattered
7.3: Water and Heat Budgets

backscattering
5.11: Transmission of Light and Other Electromagnetic Radiation

backshore
11.1: Processes that Form and Modify the Coast

backwash
11.2: Beaches

bacteria
12.1: How do we describe life?
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.5: Phototropic Primary Production and Light
12.7: Food Webs
12.9: Dissolved Oxygen and Carbon Dioxide
14.2: Habitats
14.3: Feeding
14.7: Associations
15.7: Beyond the Sun's Light
15.8: Hydrothermal Vents
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.9: Box 1 - Petroleum in the Marine Environment
17.10: CC10 Modeling
17.14: CC14 Phototrophy, Light, and Nutrients
17.19: CC19 Metagenomics
3.6: Sampling Living Organisms in the Sea
6.4: Biogenous Sediments

bacteriovores
12.12: Plankton

Baffin Bay
3.4: Seafloor Sediments
4.10: Present-Day Oceans

Baha California
13.2: Nutrient Supply to the Coastal Photic Zone

Bahamas
6.3: Lithogenous Sediments
6.5: Hydrogenous Sediments
7.9: Weather Systems
8.15: Tracing Water Masses

Baja California
14.6: Reproduction

baleen
12.12: Plankton

baleen whales
12.12: Plankton
12.7: Food Webs
14.6: Reproduction
15.6: Polar Regions

Balkan Penisula
4.4: Convergent Plate Boundaries

Baltic Sea
13.1: Special Characteristics of Coastal Oceans
13.4: Algal Blooms
4.10: Present-Day Oceans
7.5: Climate and Ocean Surface Water Properties
8.11: Inertial Currents

banded coral shrimp (Stenopus hispidus)
14.3: Feeding

banded sea snake (Laticauda colubrine)
12.13: Nekton

Banggai cardinalfish (Pterapogon kauderni)
14.6: Reproduction

Bangladesh
11.7: Deltas
7.9: Weather Systems

barnacles
12.12: Plankton
14.3: Feeding
14.6: Reproduction
15.4: Rocky Intertidal Communities
3.1: Difficulties of Studying the Ocean Environment

barracudas
14.5: Selected Adaptations in Fishes

barred filefish (Cantherhines dumerilii)
14.5: Selected Adaptations in Fishes

barrier islands
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons

barrier reefs
11.5: Coral Reefs and Atolls
15.2: Coral Reefs

basaltic rocks
11.2: Beaches

basket start
14.3: Feeding

basslets
14.5: Selected Adaptations in Fishes

bathyal zone
12.11: Biological Provinces and Zones
15.7: Beyond the Sun's Light

bathymetry
1.3: How to Study Ocean Data
3.2: The Unseen Domain
3.3: Bathymetry

bathypelagic zone
12.11: Biological Provinces and Zones

Bay of Bengal
12.9: Dissolved Oxygen and Carbon Dioxide
13.1: Special Characteristics of Coastal Oceans
4.10: Present-Day Oceans
6.3: Lithogenous Sediments
6.9: Distribution of Surface Sediments

Bay of Fundy
10.6: Tidal Currents
10.9: Energy from Tides

baymouth bar
11.3: Barrier Islands and Lagoons

bays
13.1: Special Characteristics of Coastal Oceans
16.5: Ocean Pollution Issues

beach composition
11.2: Beaches

beach dynamics
11.2: Beaches

beach management
11.4: Beaches and Human Structures

beach slope
11.2: Beaches

beaches
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons
12.11: Biological Provinces and Zones
13.4: Algal Blooms
14.3: Feeding
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
7.8: Global Climate Zones
9.8: Waves in Shallow Water

beachfront
11.4: Beaches and Human Structures

beaked whales
16.5: Ocean Pollution Issues

Beaufort Sea
11.3: Barrier Islands and Lagoons
14.6: Reproduction

Benguela Current
8.5: Open-Ocean Surface Currents

Bennett's butterflyfish (Chaetodon bennetti)
14.4: Hunting and Defense

benthic
12.11: Biological Provinces and Zones
12.14: Benthos
12.8: Geographic Variation in Primary Production
13.4: Algal Blooms
14.2: Habitats
14.6: Reproduction
15.2: Coral Reefs
15.7: Beyond the Sun's Light
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
3.6: Sampling Living Organisms in the Sea

benthos
12.11: Biological Provinces and Zones
12.14: Benthos
12.8: Geographic Variation in Primary Production
13.1: Special Characteristics of Coastal Oceans
14.2: Habitats
14.3: Feeding
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.5: Ocean Pollution Issues

Bering Sea
1.1: The Ocean's and Earth's Environment
14.6: Reproduction
15.6: Polar Regions
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.6: Climate Change, Acidification, and Deoxygenation
4.10: Present-Day Oceans
4.4: Convergent Plate Boundaries
6.9: Distribution of Surface Sediments
7.8: Global Climate Zones
8.13: Thermohaline Circulation

Bering Strait
4.10: Present-Day Oceans
8.13: Thermohaline Circulation

berm
11.2: Beaches
11.3: Barrier Islands and Lagoons

bicarbonate
5.6: Dissolved Chemicals in Seawater

binary fission
14.6: Reproduction

bioaccumulation
16.3: Toxicity
17.18: CC18 Toxicity

biochemical
16.2: Adverse Effects of Human Activities
17.14: CC14 Phototrophy, Light, and Nutrients

biochemical cycles
12.6: Primary Production and Nutrients

biodiversity
1.1: The Ocean's and Earth's Environment
12.11: Biological Provinces and Zones
12.12: Plankton
12.13: Nekton
12.2: Production, Consumption, and Decomposition
13.6: Estuaries
14.2: Habitats
14.3: Feeding
14.7: Associations
14.8: Communication and Navigation
15.1: Communities and Niches
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
15.6: Polar Regions
15.8: Hydrothermal Vents
16.2: Adverse Effects of Human Activities
17.17: CC17 Species Diversity and Biodiversity
3.6: Sampling Living Organisms in the Sea

biogenous sediment
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments

biogenous sediments
6.2: Classification of Sediments
6.4: Biogenous Sediments

biogeochemical cycles
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.9: Dissolved Oxygen and Carbon Dioxide
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.4: Waste Disposal
17.8: CC8 Residence Time
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
6.1: Sediments and Biogeochemical Cycles

biogeochemical cycling
12.4: Studying Microbes in the Ocean
17.19: CC19 Metagenomics
bioluminescence
12.11: Biological Provinces and Zones
14.4: Hunting and Defense
15.7: Beyond the Sun's Light
biomagnification
16.3: Toxicity
16.5: Ocean Pollution Issues
17.18: CC18 Toxicity
biomass
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
12.8: Geographic Variation in Primary Production
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.5: Fisheries
14.2: Habitats
14.3: Feeding
15.2: Coral Reefs
15.5: Sargasso Sea
15.8: Hydrothermal Vents
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.9: Box 1 - Petroleum in the Marine Environment
17.10: CC10 Modeling
17.14: CC14 Phototrophy, Light, and Nutrients
17.15: CC15 Food Chain Efficiency
17.16: CC16 Maximum Sustainable Yield
17.8: CC8 Residence Time
biomes
12.3: Microbes in Charge
17.19: CC19 Metagenomics
biosphere
17.8: CC8 Residence Time
biota
12.11: Biological Provinces and Zones
17.8: CC8 Residence Time
bioturbation
14.3: Feeding
black corals
15.2: Coral Reefs
Black Sea
17.8: CC8 Residence Time
black smokers
15.8: Hydrothermal Vents
black turban snails (Tegula funebralis)
15.4: Rocky Intertidal Communities
blacksaddle mimic filefish (Paraluteres prionurus)
14.4: Hunting and Defense
blackstriped goatfish (Upeneus tragula)
14.3: Feeding
Bligh Reef
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
bloom
16.2: Adverse Effects of Human Activities
blooms
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
14.6: Reproduction
17.16: CC16 Maximum Sustainable Yield
blubber
12.13: Nekton
14.6: Reproduction
blue whale (Balaenoptera musculus)
12.13: Nekton
blue whales
12.12: Plankton
12.7: Food Webs
bluefin trevally (Caranx melampygus)
14.4: Hunting and Defense
bluegreen algae
12.12: Plankton
bluespotted cornetfish (Fistularia commersonii)
14.5: Selected Adaptations in Fishes
body shape
14.5: Selected Adaptations in Fishes
boiling point
5.7: Heat Properties of Water
bone skeletons
12.13: Nekton
bony fishes
12.13: Nekton
Bosporus Strait
17.8: CC8 Residence Time
Boston
7.8: Global Climate Zones
bottlenose dolphins
12.13: Nekton
13.4: Algal Blooms
bottlenose whale
12.13: Nekton
boundary currents
8.18: Box 2 - Equatorial Currents
8.8: Upwelling and Downwelling
8.9: Coastal Currents
box corers
3.4: Seafloor Sediments
boxfishes
14.4: Hunting and Defense
Brahmaputra River
13.1: Special Characteristics of Coastal Oceans
4.10: Present-Day Oceans
6.3: Lithogenous Sediments
brain coral
14.3: Feeding
Brazil
13.2: Nutrient Supply to the Coastal Photic Zone
6.9: Distribution of Surface Sediments
7.9: Weather Systems
Brazil Current
8.5: Open-Ocean Surface Currents
breakers
9.8: Waves in Shallow Water
breakwaters
11.4: Beaches and Human Structures
brines
8.13: Thermohaline Circulation
Britain
7.6: Interannual Climate Variations
British Columbia
6.9: Distribution of Surface Sediments
7.6: Interannual Climate Variations
brittle stars
14.7: Associations
15.7: Beyond the Sun's Light
broadclub cuttlefish (Sepia latimanus)
12.13: Nekton
brown cup corals (Paracyathus stearnsii)
15.3: Kelp Forests
brown pelican (Pelecanus occidentalis)
12.13: Nekton
bulkheads
16.2: Adverse Effects of Human Activities
buoyancy
12.12: Plankton
12.13: Nekton
12.5: Phototropic Primary Production and Light
14.2: Habitats
14.5: Selected Adaptations in Fishes
15.3: Kelp Forests
17.1: CC1 Density and Layering in Fluids
17.2: CC2 Isostasy, Eustasy, and Sea Level
burrfishes
14.4: Hunting and Defense
butterflyfishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.7: Associations
bycatch
2.4: Ocean Resources

C

calcareous algae
11.2: Beaches
calcareous plates
12.12: Plankton
calcareous sediment
6.4: Biogenous Sediments
6.9: Distribution of Surface Sediments
calcareous shells
12.12: Plankton
calcium
15.8: Hydrothermal Vents
5.6: Dissolved Chemicals in Seawater
calcium carbonate
12.11: Biological Provinces and Zones
12.12: Plankton
12.13: Nekton
14.4: Hunting and Defense
6.10: The Sediment Historical Record
6.4: Biogenous Sediments

California
10.2: Characteristics of Tides
10.8: Marine Species and the Rhythm of Tides
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.4: Beaches and Human Structures
11.6: Wetlands
12.13: Nekton
12.5: Phototropic Primary Production and Light
13.2: Nutrient Supply to the Coastal Photic Zone
13.4: Algal Blooms
14.3: Feeding
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.5: Ocean Pollution Issues
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.6: Transform Faults and Fracture Zones
6.3: Lithogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
7.10: Local Weather Effects
7.6: Interannual Climate Variations
7.9: Weather Systems

California Current
16.6: Climate Change, Acidification, and Deoxygenation
8.5: Open-Ocean Surface Currents
8.9: Coastal Currents

California gray whales
14.6: Reproduction

California mussel (Mytilus californianus)
14.6: Reproduction

California mussels (Mytilus californianus)
14.3: Feeding
15.4: Rocky Intertidal Communities

California sea lion (Zalophus californianus)
12.13: Nekton
14.4: Hunting and Defense

camouflage
14.4: Hunting and Defense

Canada
10.2: Characteristics of Tides
11.1: Processes that Form and Modify the Coast
11.7: Deltas
13.6: Estuaries
16.6: Climate Change, Acidification, and Deoxygenation
2.4: Ocean Resources
4.9: Sea-Level Change and Climate
6.7: Sediment Transport, Deposition, and Accumulation
7.10: Local Weather Effects
7.6: Interannual Climate Variations
8.14: Ocean Circulation and Climate

Canary Current
8.5: Open-Ocean Surface Currents

cancer
16.11: Box 3 - Radioactive Wastes in the Marine Environment

Cape Hatteras
8.9: Coastal Currents

capillary waves
9.3: What is Wave Motion?
9.4: Making Waves
9.6: Wave Dissipation

carbon
12.10: Organic Carbon
12.6: Primary Production and Nutrients
17.14: CC14 Phototrophy, Light, and Nutrients

carbon compensation depth
12.11: Biological Provinces and Zones
6.4: Biogenous Sediments

carbon dioxide
1.1: The Ocean's and Earth's Environment
12.2: Production, Consumption, and Decomposition
12.6: Primary Production and Nutrients
12.9: Dissolved Oxygen and Carbon Dioxide
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
17.14: CC14 Phototrophy, Light, and Nutrients
17.9: CC9 The Global Greenhouse Effect
6.4: Biogenous Sediments

carbonates
6.5: Hydrogenous Sediments

carbonic acid
5.6: Dissolved Chemicals in Seawater

cardinalfishes
14.6: Reproduction
14.7: Associations

Caribbean
17.10: CC10 Modeling

Caribbean Sea
10.2: Characteristics of Tides
10.5: Tides in the Open Ocean
11.1: Processes that Form and Modify the Coast
14.3: Feeding
15.2: Coral Reefs
4.10: Present-Day Oceans
6.6: Cosmogenous Sediments

carnivores
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.7: Food Webs
13.2: Nutrient Supply to the Coastal Photic Zone
14.2: Habitats
14.6: Reproduction
15.4: Rocky Intertidal Communities

Carolina
8.9: Coastal Currents

cartilage skeletons
12.13: Nekton

catadromous
13.6: Estuaries
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.8: Communication and Navigation
16.2: Adverse Effects of Human Activities

cations
5.3: The Dissolving Power of Water

Cayman Islands
11.1: Processes that Form and Modify the Coast

CCD
12.11: Biological Provinces and Zones

Celebes sweetlips (Plectorhinchus celebicus)
14.4: Hunting and Defense

Central America
4.4: Convergent Plate Boundaries
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
8.13: Thermohaline Circulation

central rift valley
4.5: Divergent Plate Boundaries

centripetal force
10.1: Tide-Generating Forces
17.12: CC12 The Coriolis Effect
7.9: Weather Systems

Cephalopoda
12.13: Nekton

cephalopods
12.13: Nekton

cetaceans
12.13: Nekton
14.8: Communication and Navigation

challenger
15.8: Hydrothermal Vents

chaotic
13.5: Fisheries
17.10: CC10 Modeling
17.11: CC11 Chaos
17.9: CC9 The Global Greenhouse Effect
4.2: Plate Tectonics
7.6: Interannual Climate Variations
7.9: Weather Systems
8.14: Ocean Circulation and Climate

charts
1.3: How to Study Ocean Data

cheeklined wrasse (Oxycheilinus digramma)
14.5: Selected Adaptations in Fishes

chemical analysis
3.5: Chemical and Physical Oceanography

chemical bonds
5.2: The Water Molecule

chemical weathering
6.3: Lithogenous Sediments

chemistry
5.2: The Water Molecule

chemosynthesis
1.2: Oceans and the Origins of Life
12.11: Biological Provinces and Zones
12.2: Production, Consumption, and Decomposition
12.5: Phototropic Primary Production and Light
14.2: Habitats
14.3: Feeding
15.8: Hydrothermal Vents
17.14: CC14 Phototrophy, Light, and Nutrients
17.15: CC15 Food Chain Efficiency
2.2: Exploration and Mapping

Chernobyl
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues

Chesapeake Bay
1.1: The Ocean's and Earth's Environment
10.6: Tidal Currents
13.4: Algal Blooms
16.1: Pollution versus Contamination
16.5: Ocean Pollution Issues
6.10: The Sediment Historical Record

chevron barracuda (Sphyraena putnamiae)
14.5: Selected Adaptations in Fishes

Chicxulub
6.10: The Sediment Historical Record
9.9: Tsunamis
Chikyu
3.4: Seafloor Sediments
Chile
11.1: Processes that Form and Modify the Coast
13.6: Estuaries
7.10: Local Weather Effects
7.6: Interannual Climate Variations
China
10.6: Tidal Currents
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
3.4: Seafloor Sediments
4.4: Convergent Plate Boundaries
6.10: The Sediment Historical Record
6.3: Lithogenous Sediments
7.9: Weather Systems
chitons
15.4: Rocky Intertidal Communities
chlorine
5.2: The Water Molecule
5.6: Dissolved Chemicals in Seawater
chlorofluorocarbons
17.9: CC9 The Global Greenhouse Effect
7.1: The Atmosphere
8.15: Tracing Water Masses
chlorophyll
12.2: Production, Consumption, and Decomposition
13.2: Nutrient Supply to the Coastal Photic Zone
17.14: CC14 Phototrophy, Light, and Nutrients
3.6: Sampling Living Organisms in the Sea
5.11: Transmission of Light and Other Electromagnetic Radiation
7.6: Interannual Climate Variations
chloroplasts
17.14: CC14 Phototrophy, Light, and Nutrients
chordates
12.12: Plankton
Christmas tree worm (Spirobranchus giganteus)
14.3: Feeding
14.4: Hunting and Defense
chromatophores
14.4: Hunting and Defense
chrysomonads
12.12: Plankton
Chukchi Sea
14.6: Reproduction
circular orbits
10.1: Tide-Generating Forces
circulation
8.15: Tracing Water Masses
clams
12.12: Plankton
12.13: Nekton
13.1: Special Characteristics of Coastal Oceans
13.4: Algal Blooms
13.6: Estuaries
14.4: Hunting and Defense
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
16.5: Ocean Pollution Issues
Clark's anemonefish (Amphiprion clarkii)
14.4: Hunting and Defense
classification
12.1: How do we describe life?
clay
6.2: Classification of Sediments
cleaner wrasses
14.4: Hunting and Defense
climate
13.1: Special Characteristics of Coastal Oceans
14.6: Reproduction
15.6: Polar Regions
17.10: CC10 Modeling
17.11: CC11 Chaos
17.16: CC16 Maximum Sustainable Yield
17.5: CC5 Transfer and Storage of Heat by Water
17.9: CC9 The Global Greenhouse Effect
5.1: Origins and Distribution of the Earth's Water
7.1: The Atmosphere
7.10: Local Weather Effects
7.2: Water Vapor in the Atmosphere
7.3: Water and Heat Budgets
7.4: Climatic Winds
7.5: Climate and Ocean Surface Water Properties
7.8: Global Climate Zones
climate change
1.1: The Ocean's and Earth's Environment
11.1: Processes that Form and Modify the Coast
11.7: Deltas
12.9: Dissolved Oxygen and Carbon Dioxide
15.3: Kelp Forests
15.6: Polar Regions
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.9: CC9 The Global Greenhouse Effect
4.9: Sea-Level Change and Climate
6.10: The Sediment Historical Record
7.9: Weather Systems
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
9.5: Sea Development and Wave Height
climate characteristics
7.7: Land-Ocean-Atmosphere Interactions
climate patterns
8.10: Eddies
climate system
16.6: Climate Change, Acidification, and Deoxygenation
climate variations
7.6: Interannual Climate Variations
climate zones
7.8: Global Climate Zones
clingfishes
14.5: Selected Adaptations in Fishes
14.7: Associations
clown anemonefish (Amphiprion percula)
14.4: Hunting and Defense
clown triggerfish (Balistoides conspicillum)
14.5: Selected Adaptations in Fishes
clownfishes
14.4: Hunting and Defense
cnidarians
12.12: Plankton
14.4: Hunting and Defense
cnidocytes
12.12: Plankton
coal reefs
10.8: Marine Species and the Rhythm of Tides
coast
13.2: Nutrient Supply to the Coastal Photic Zone
coastal currents
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
8.9: Coastal Currents
coastal development
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures
11.5: Coral Reefs and Atolls
coastal downwelling
8.8: Upwelling and Downwelling
coastal ecosystems
11.1: Processes that Form and Modify the Coast
14.2: Habitats
coastal erosion
11.1: Processes that Form and Modify the Coast
11.2: Beaches
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures
6.8: Continental Margin Sediments
9.4: Making Waves
coastal fog
7.10: Local Weather Effects
coastal management
11.3: Barrier Islands and Lagoons
coastal oceans
13.1: Special Characteristics of Coastal Oceans
13.5: Fisheries
coastal plains
11.1: Processes that Form and Modify the Coast
13.6: Estuaries
4.8: Plate Interiors
6.3: Lithogenous Sediments
coastal processes
6.7: Sediment Transport, Deposition, and Accumulation
coastal regions
10.6: Tidal Currents
12.8: Geographic Variation in Primary Production
16.2: Adverse Effects of Human Activities
7.10: Local Weather Effects
coastal structures
11.4: Beaches and Human Structures
coastal topography
9.8: Waves in Shallow Water
coastal upwelling
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
13.2: Nutrient Supply to the Coastal Photic Zone
13.4: Algal Blooms
7.6: Interannual Climate Variations
8.8: Upwelling and Downwelling
coastal zone
12.14: Benthos
coastline
8.3: Ekman Motion
coasts
11.1: Processes that Form and Modify the Coast
16.6: Climate Change, Acidification, and Deoxygenation
cobble beach
11.2: Beaches
coccolithophores
12.12: Plankton
6.4: Biogenous Sediments
cockle
14.3: Feeding

cod
- 12.13: Nekton
- 17.16: CC16 Maximum Sustainable Yield

cohesive
- 17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension

cohesive sediments
- 17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension

Coleman's shrimp (Periclimenes colemani)
- 14.7: Associations

colonial forms
- 12.12: Plankton
- 14.3: Feeding
- 14.4: Hunting and Defense
- 14.6: Reproduction

Columbia River
- 7.6: Interannual Climate Variations

comb jellies
- 12.12: Plankton
- 14.4: Hunting and Defense

commensalism
- 14.7: Associations

common eastern oyster (Crassostrea virginica)
- 14.6: Reproduction

common mussel (Mytilus edulis)
- 14.6: Reproduction

common reef squid (Sepioteuthis lessoniana)
- 12.13: Nekton
- 14.6: Reproduction

communication
- 14.8: Communication and Navigation
- 3.8: Satellites
- 5.12: Transmission of Sound

communities
- 11.5: Coral Reefs and Atolls
- 12.11: Biological Provinces and Zones
- 12.2: Production, Consumption, and Decomposition
- 13.1: Special Characteristics of Coastal Oceans
- 15.1: Communities and Niches
- 15.6: Polar Regions
- 15.7: Beyond the Sun's Light
- 15.8: Hydrothermal Vents
- 16.2: Adverse Effects of Human Activities
- 17.17: CC17 Species Diversity and Biodiversity
- 17.19: CC19 Metagenomics
- 8.14: Ocean Circulation and Climate

community structure
- 15.1: Communities and Niches

compensation depth
- 12.5: Phototropic Primary Production and Light

competition
- 14.4: Hunting and Defense
- 14.7: Associations
- 15.4: Rocky Intertidal Communities

computer enhancement
- 8.10: Eddies

computer science
- 3.9: Computers and Modeling

concealment
- 14.4: Hunting and Defense

condensation
- 17.5: CC5 Transfer and Storage of Heat by Water
- 7.2: Water Vapor in the Atmosphere

conduction
- 17.10: CC10 Modeling
- 17.11: CC11 Chaos
- 17.3: CC3 Convection and Convection Cells
- 17.5: CC5 Transfer and Storage of Heat by Water
- 17.9: CC9 The Global Greenhouse Effect
- 5.9: Ice Formation

cone snails
- 14.4: Hunting and Defense

Congo River
- 4.10: Present-Day Oceans

conservation
- 1.1: The Ocean's and Earth's Environment
- 12.13: Nekton
- 3.6: Sampling Living Organisms in the Sea

conservative properties
- 8.15: Tracing Water Masses

constituents
- 5.6: Dissolved Chemicals in Seawater

consumers
- 12.3: Microbes in Charge
- 14.2: Habitats

contaminants
- 13.1: Special Characteristics of Coastal Oceans
- 13.6: Estuaries
- 15.6: Polar Regions
- 16.1: Pollution versus Contamination
- 16.11: Box 3 - Radioactive Wastes in the Marine Environment
- 16.2: Adverse Effects of Human Activities
- 16.5: Ocean Pollution Issues
- 17.10: CC10 Modeling
- 17.18: CC18 Toxicity
- 17.8: CC8 Residence Time
- 3.5: Chemical and Physical Oceanography
- 6.1: Sediments and Biogeochemical Cycles

continental climate
- 7.8: Global Climate Zones

continental crust
- 17.2: CC2 Isostasy, Eustasy, and Sea Level
- 4.3: Plate Boundaries
- 4.4: Convergent Plate Boundaries
- 4.7: Hot Spots

continental divergent plate boundaries
- 11.1: Processes that Form and Modify the Coast

continental margins
- 4.8: Plate Interiors

continental shelf
- 10.4: Tides in the Earth's Oceans
- 11.1: Processes that Form and Modify the Coast
- 12.2: Production, Consumption, and Decomposition
- 13.2: Nutrient Supply to the Coastal Photic Zone
- 13.4: Algal Blooms
- 14.3: Feeding
- 15.2: Coral Reefs
- 16.5: Ocean Pollution Issues
- 2.3: National and International Law Applied to Oceans
- 4.10: Present-Day Oceans
- 6.8: Continental Margin Sediments
- 8.9: Coastal Currents
- 9.5: Sea Development and Wave Height

continental shelf break
- 12.11: Biological Provinces and Zones

continental slope
- 15.8: Hydrothermal Vents

convection
- 17.10: CC10 Modeling
- 17.11: CC11 Chaos
- 17.13: CC13 Geostrophic Flow
- 17.3: CC3 Convection and Convection Cells
- 4.2: Plate Tectonics
- 5.9: Ice Formation
- 7.2: Water Vapor in the Atmosphere

convection cells
- 17.3: CC3 Convection and Convection Cells
- 4.2: Plate Tectonics
- 7.2: Water Vapor in the Atmosphere
- 7.4: Climatic Winds
- 8.12: Langmuir Circulation
- 8.5: Open-Ocean Surface Currents

convergence
- 12.8: Geographic Variation in Primary Production
- 17.3: CC3 Convection and Convection Cells
- 7.4: Climatic Winds
- 8.12: Langmuir Circulation
- 8.3: Ekman Motion
- 8.5: Open-Ocean Surface Currents
- 8.8: Upwelling and Downwelling

convergent boundaries
- 4.3: Plate Boundaries
- 4.4: Convergent Plate Boundaries

convergent plate boundaries
- 11.1: Processes that Form and Modify the Coast
- 4.3: Plate Boundaries
- 4.4: Convergent Plate Boundaries
- 4.8: Plate Interiors

conveyor belt circulation
- 12.9: Dissolved Oxygen and Carbon Dioxide
- 8.14: Ocean Circulation and Climate

Cook Inlet
- 10.6: Tidal Currents

cooperation
- 14.4: Hunting and Defense

copepods
- 12.12: Plankton
- 12.7: Food Webs
- 14.3: Feeding
- 14.7: Associations
- 15.4: Rocky Intertidal Communities
- 15.7: Beyond the Sun's Light

copper
- 15.8: Hydrothermal Vents
- 16.1: Pollution versus Contamination
- 16.2: Adverse Effects of Human Activities
- 16.3: Toxicity
- 16.5: Ocean Pollution Issues
- 17.18: CC18 Toxicity

Copper River Delta
- 11.7: Deltas

copper rockfish (Sebastes caurinus)
- 15.3: Kelp Forests

coral barnacles
- 14.3: Feeding

coral bleaching
- 16.6: Climate Change, Acidification, and Deoxygenation

coral reefs
11.1: Processes that Form and Modify the Coast
11.5: Coral Reefs and Atolls
12.11: Biological Provinces and Zones
12.14: Benthos
12.6: Primary Production and Nutrients
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
14.2: Habitats
14.3: Feeding
14.7: Associations
15.2: Coral Reefs
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
17.16: CC16 Maximum Sustainable Yield
17.17: CC17 Species Diversity and Biodiversity
17.2: CC2 Isostasy, Eustasy, and Sea Level
3.6: Sampling Living Organisms in the Sea
4.8: Plate Interiors
6.9: Distribution of Surface Sediments
corals
12.12: Plankton
13.1: Special Characteristics of Coastal Oceans
14.3: Feeding
14.4: Hunting and Defense
14.6: Reproduction
14.7: Associations
15.2: Coral Reefs
15.3: Kelp Forests
16.6: Climate Change, Acidification, and Deoxygenation
17.17: CC17 Species Diversity and Biodiversity
9.9: Tsunamis
core
4.1: Structure of the Earth
corers
3.4: Seafloor Sediments
coring techniques
3.4: Seafloor Sediments
Coriolis Effect
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
13.6: Estuaries
17.12: CC12 The Coriolis Effect
17.13: CC13 Geostrophic Flow
7.4: Climatic Winds
7.9: Weather Systems
8.10: Eddies
8.11: Inertial Currents
8.13: Thermohaline Circulation
8.17: Box 1- Westward Intensification
8.18: Box 2 - Equatorial Currents
8.2: Wind-Driven Currents
8.3: Ekman Motion
8.4: Geostrophic Currents
9.11: Rossby and Kelvin Waves
9.3: What is Wave Motion?
cormorants
12.13: Nekton
cornetfishes
14.5: Selected Adaptations in Fishes
cosmogenous sediments
6.2: Classification of Sediments
covalent bond
5.2: The Water Molecule
cowries
14.3: Feeding
14.7: Associations
crabeater seals (Lobodon carcinophagus)
15.6: Polar Regions
crabs
12.12: Plankton
12.13: Nekton
13.4: Algal Blooms
13.6: Estuaries
14.4: Hunting and Defense
14.7: Associations
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
crescenttailed bigeye (Priacanthus hamrur)
14.4: Hunting and Defense
crests
10.4: Tides in the Earth's Oceans
11.2: Beaches
9.12: Standing Waves
9.2: Progressive Waves
9.9: Tsunamis
crinoid clingfish (Discotrema crinophila)
14.7: Associations
crinoid shrimp (Periclimenes amboinensi)
14.4: Hunting and Defense
crinoids
14.4: Hunting and Defense
14.7: Associations
Croatia
13.4: Algal Blooms
crocodiles
12.13: Nekton
crust
11.5: Coral Reefs and Atolls
15.8: Hydrothermal Vents
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.1: Structure of the Earth
4.2: Plate Tectonics
5.1: Origins and Distribution of the Earth's Water
crust formation
4.5: Divergent Plate Boundaries
crustaceans
12.12: Plankton
12.7: Food Webs
13.4: Algal Blooms
14.2: Habitats
14.3: Feeding
14.4: Hunting and Defense
14.7: Associations
15.7: Beyond the Sun's Light
cryptomonads
12.12: Plankton
CTD
3.5: Chemical and Physical Oceanography
5.5: Salinity
ctenophores
12.12: Plankton
14.3: Feeding
14.4: Hunting and Defense
Cuba
7.9: Weather Systems
cum sole
10.4: Tides in the Earth's Oceans
17.12: CC12 The Coriolis Effect
17.13: CC13 Geostrophic Flow
7.9: Weather Systems
8.3: Ekman Motion
8.4: Geostrophic Currents
current speed
8.4: Geostrophic Currents
cuttlefish
12.13: Nekton
14.4: Hunting and Defense
14.6: Reproduction
Cyanea
12.12: Plankton
cyanides
16.1: Pollution versus Contamination
cyanobacteria
12.12: Plankton
12.3: Microbes in Charge
13.4: Algal Blooms
14.2: Habitats
17.14: CC14 Phototrophy, Light, and Nutrients
cyclones
7.9: Weather Systems
cysts
13.2: Nutrient Supply to the Coastal Photic Zone

D

daily tidal range
10.2: Characteristics of Tides
damselfishes
14.4: Hunting and Defense
14.6: Reproduction
darkspotted moray eel (Gymnothorax fimbriatus)
14.5: Selected Adaptations in Fishes
data
1.3: How to Study Ocean Data
data analysis
3.9: Computers and Modeling
data visualization
3.9: Computers and Modeling
Davidson Current
8.9: Coastal Currents
DDT
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.5: Ocean Pollution Issues
dead zones
12.9: Dissolved Oxygen and Carbon Dioxide
13.4: Algal Blooms
16.1: Pollution versus Contamination
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
declination
10.3: Tides on an Ocean Covered Earth
decomposer
14.2: Habitats
decomposers
12.10: Organic Carbon
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
13.3: Seasonal Cycles
14.2: Habitats
15.2: Coral Reefs
15.3: Kelp Forests
16.1: Pollution versus Contamination
decomposition
12.10: Organic Carbon
12.2: Production, Consumption, and Decomposition
13.1: Special Characteristics of Coastal Oceans
13.4: Algal Blooms
16.9: Box 1 - Petroleum in the Marine Environment

decorator crabs
14.7: Associations
decorator spider crab (Camposa retusa)
14.4: Hunting and Defense
deep layer
17.10: CC10 Modeling
deep ocean
15.7: Beyond the Sun's Light
deep ocean exploration
3.1: Difficulties of Studying the Ocean Environment
deep water masses
8.13: Thermohaline Circulation
deep zone
8.13: Thermohaline Circulation
Deepwater Horizon
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
defense
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
deltas
11.1: Processes that Form and Modify the Coast
11.7: Deltas
13.6: Estuaries
6.3: Lithogenous Sediments
demersal fishes
12.13: Nekton
density
12.12: Plankton
14.2: Habitats
14.5: Selected Adaptations in Fishes
17.1: CC1 Density and Layering in Fluids
17.13: CC13 Geostrophic Flow
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.3: CC3 Convection and Convection Cells
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
17.6: CC6 Salinity, Temperature, Pressure, and Water Density
5.10: Surface Tension and Viscosity
5.9: Ice Formation
7.1: The Atmosphere
8.1: Energy Sources
8.13: Thermohaline Circulation
9.10: Internal Waves
9.3: What is Wave Motion?
deoxygenation
1.1: The Ocean's and Earth's Environment
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
6.10: The Sediment Historical Record
deposit feeders
14.3: Feeding
15.2: Coral Reefs
deposition
12.6: Primary Production and Nutrients
17.2: CC2 Isostasy, Eustasy, and Sea Level
depositional coasts
11.1: Processes that Form and Modify the Coast
depth measurement
3.3: Bathymetry
depth of frictional influence
8.3: Ekman Motion
desposition
11.1: Processes that Form and Modify the Coast
detritivores
12.2: Production, Consumption, and Decomposition
detritus
11.6: Wetlands
12.10: Organic Carbon
12.11: Biological Provinces and Zones
12.9: Dissolved Oxygen and Carbon Dioxide
13.2: Nutrient Supply to the Coastal Photic Zone
13.6: Estuaries
14.2: Habitats
14.3: Feeding
14.4: Hunting and Defense
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.7: Beyond the Sun's Light
17.10: CC10 Modeling
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
6.4: Biogenous Sediments
diagenesis
6.10: The Sediment Historical Record
diatom oozes
6.9: Distribution of Surface Sediments
diatoms
12.12: Plankton
12.5: Phototropic Primary Production and Light
12.7: Food Webs
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.6: Estuaries
14.3: Feeding
14.4: Hunting and Defense
14.6: Reproduction
6.4: Biogenous Sediments
diel
12.12: Plankton
12.6: Primary Production and Nutrients
14.2: Habitats
diffusion
12.6: Primary Production and Nutrients
14.3: Feeding
15.2: Coral Reefs
17.10: CC10 Modeling
6.10: The Sediment Historical Record
7.1: The Atmosphere
8.13: Thermohaline Circulation
dinoflagellates
11.5: Coral Reefs and Atolls
12.12: Plankton
12.5: Phototropic Primary Production and Light
13.4: Algal Blooms
15.2: Coral Reefs
dip angle
4.7: Hot Spots
disaster
9.9: Tsunamis
disaster response
9.9: Tsunamis
disease
13.5: Fisheries
displacement waves
9.4: Making Waves
displacing force
9.4: Making Waves
dissolved
5.6: Dissolved Chemicals in Seawater
dissolved compounds
12.10: Organic Carbon
dissolved oxygen
3.5: Chemical and Physical Oceanography
dissolved salts
5.5: Salinity
dissolved substances
8.10: Eddies
diurnal tides
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
diurnally
7.10: Local Weather Effects
divergence
12.8: Geographic Variation in Primary Production
17.3: CC3 Convection and Convection Cells
7.4: Climatic Winds
8.12: Langmuir Circulation
8.3: Ekman Motion
8.5: Open-Ocean Surface Currents
8.8: Upwelling and Downwelling
divergent boundaries
4.3: Plate Boundaries
divergent plate boundaries
4.3: Plate Boundaries
4.5: Divergent Plate Boundaries
4.8: Plate Interiors
DNA
12.4: Studying Microbes in the Ocean
16.3: Toxicity
17.18: CC18 Toxicity
17.19: CC19 Metagenomics
dogtooth tuna (Gymnosarda unicolor)
12.13: Nekton
doldrums
8.18: Box 2 - Equatorial Currents
8.6: Equatorial Surface Currents
dolphins
13.4: Algal Blooms
14.8: Communication and Navigation
Doppler
3.5: Chemical and Physical Oceanography
Doppler acoustic current meters
3.5: Chemical and Physical Oceanography
downwelling
12.6: Primary Production and Nutrients
13.3: Seasonal Cycles
15.6: Polar Regions
17.3: CC3 Convection and Convection Cells
7.4: Climatic Winds
8.12: Langmuir Circulation
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.8: Upwelling and Downwelling
drag
14.5: Selected Adaptations in Fishes
dredges
11.4: Beaches and Human Structures
3.4: Seafloor Sediments
dredging
11.5: Coral Reefs and Atolls
16.4: Waste Disposal
16.5: Ocean Pollution Issues
drift nets
12.13: Nekton
2.4: Ocean Resources
drifters
3.5: Chemical and Physical Oceanography
drifting ice
8.3: Ekman Motion

drilling ships
3.4: Seafloor Sediments
drogues
3.5: Chemical and Physical Oceanography
drowned river valleys
13.6: Estuaries
dugongs
12.13: Nekton
dynamic height
8.4: Geostrophic Currents
dynamic tides
10.4: Tides in the Earth's Oceans
dynamic topography
8.4: Geostrophic Currents
dynamic topography maps
8.4: Geostrophic Currents

E

Earth
10.1: Tide-Generating Forces
4.1: Structure of the Earth
Earth's rotation
7.4: Climatic Winds
earthquake
9.9: Tsunamis
earthquakes
4.6: Transform Faults and Fracture Zones
East AFrica
2.2: Exploration and Mapping
East African Rift
4.5: Divergent Plate Boundaries
4.8: Plate Interiors
East African Rift Zone
4.3: Plate Boundaries
4.5: Divergent Plate Boundaries
East Antarctic Ice Sheet
16.6: Climate Change, Acidification, and Deoxygenation
East Australia Current
8.5: Open-Ocean Surface Currents
East Australian Current
8.5: Open-Ocean Surface Currents
East China Sea
6.3: Lithogenous Sediments
East Pacific Ridge
4.3: Plate Boundaries
East Pacific Rise
4.10: Present-Day Oceans
4.5: Divergent Plate Boundaries
East Rift Zone
11.1: Processes that Form and Modify the Coast
East Timor
2.2: Exploration and Mapping
East Wind Drift
8.7: High-Latitude Surface Currents
Easter Island
4.3: Plate Boundaries
4.7: Hot Spots
eastern boundary currents
14.6: Reproduction
8.5: Open-Ocean Surface Currents
eastern oyster (Crassostrea virginica)
16.5: Ocean Pollution Issues
ebbs
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
14.3: Feeding
echinoderms
11.2: Beaches
echo sounders
3.3: Bathymetry
echolocation
14.8: Communication and Navigation
ecological adaptations
12.11: Biological Provinces and Zones
15.6: Polar Regions
ecological challenges
3.6: Sampling Living Organisms in the Sea
ecological dynamics
14.7: Associations
ecological efficiency
17.15: CC15 Food Chain Efficiency
ecological impact
16.6: Climate Change, Acidification, and Deoxygenation
ecological preservation
11.3: Barrier Islands and Lagoons
ecological research
3.6: Sampling Living Organisms in the Sea
economy
16.7: Risk Assessment and Management
2.4: Ocean Resources
ecosystem dynamics
13.2: Nutrient Supply to the Coastal Photic Zone
ecosystems
12.2: Production, Consumption, and Decomposition
13.3: Seasonal Cycles
14.3: Feeding
15.1: Communities and Niches
15.2: Coral Reefs
15.3: Kelp Forests
15.6: Polar Regions
15.8: Hydrothermal Vents
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
16.9: Box 1 - Petroleum in the Marine Environment
17.10: CC10 Modeling
17.11: CC11 Chaos
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.8: CC8 Residence Time
2.2: Exploration and Mapping
5.1: Origins and Distribution of the Earth's Water
7.8: Global Climate Zones
Ecuador
2.3: National and International Law Applied to Oceans
eddies
12.12: Plankton
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
15.2: Coral Reefs
15.6: Polar Regions
8.10: Eddies
eels
14.5: Selected Adaptations in Fishes
14.6: Reproduction
16.2: Adverse Effects of Human Activities
EEZ
2.3: National and International Law Applied to Oceans
effluents
16.1: Pollution versus Contamination
16.5: Ocean Pollution Issues
17.18: CC18 Toxicity
5.6: Dissolved Chemicals in Seawater
eggs
10.8: Marine Species and the Rhythm of Tides
12.12: Plankton
12.13: Nekton
13.2: Nutrient Supply to the Coastal Photic Zone
13.5: Fisheries
13.6: Estuaries
14.2: Habitats
14.6: Reproduction
14.7: Associations
15.8: Hydrothermal Vents
17.15: CC15 Food Chain Efficiency
Ekman spiral
8.3: Ekman Motion
Ekman transport
11.2: Beaches
13.2: Nutrient Supply to the Coastal Photic Zone
13.6: Estuaries
16.5: Ocean Pollution Issues
8.12: Langmuir Circulation
8.18: Box 2 - Equatorial Currents
8.3: Ekman Motion
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
8.7: High-Latitude Surface Currents
8.8: Upwelling and Downwelling
8.9: Coastal Currents
El Niño
12.8: Geographic Variation in Primary Production
7.4: Climatic Winds
7.6: Interannual Climate Variations
8.18: Box 2 - Equatorial Currents
8.6: Equatorial Surface Currents
El Niño Modoki
7.6: Interannual Climate Variations
El Niño/Southern Oscillation
7.6: Interannual Climate Variations
electrical conductivity
3.5: Chemical and Physical Oceanography
5.5: Salinity
electricity generation
10.9: Energy from Tides
electromagnetic radiation
3.8: Satellites
5.11: Transmission of Light and Other Electromagnetic Radiation
electromagnetic spectrum
17.9: CC9 The Global Greenhouse Effect
electrostatic
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
elements
5.6: Dissolved Chemicals in Seawater
elephant seal (Mirounga angustirostris)
12.13: Nekton
elephant seals
16.3: Toxicity
Elkhorn coral (Acropora palmata)
15.2: Coral Reefs

emperor shrimp (Periclimenes imperator)
14.7: Associations
energy
9.4: Making Waves
energy capture
10.9: Energy from Tides
energy production
12.2: Production, Consumption, and Decomposition
energy transfer
12.7: Food Webs
17.3: CC3 Convection and Convection Cells
5.7: Heat Properties of Water
8.2: Wind-Driven Currents
8.3: Ekman Motion
England
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.5: Ocean Pollution Issues
2.2: Exploration and Mapping
3.3: Bathymetry
English Channel
10.5: Tides in the Open Ocean
ENSO
7.6: Interannual Climate Variations
8.6: Equatorial Surface Currents
9.11: Rossby and Kelvin Waves
environment
1.3: How to Study Ocean Data
13.6: Estuaries
14.1: Ecological Requirements
17.14: CC14 Phototrophy, Light, and Nutrients
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.8: CC8 Residence Time
2.4: Ocean Resources
environmental
16.11: Box 3 - Radioactive Wastes in the Marine Environment
17.10: CC10 Modeling
environmental factors
12.5: Phototropic Primary Production and Light
13.2: Nutrient Supply to the Coastal Photic Zone
environmental impact
10.9: Energy from Tides
13.4: Algal Blooms
13.6: Estuaries
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.4: Waste Disposal
16.7: Risk Assessment and Management
environmental stresses
15.4: Rocky Intertidal Communities
enzymes
12.6: Primary Production and Nutrients
15.8: Hydrothermal Vents
Ephesus
13.6: Estuaries
epifauna
12.11: Biological Provinces and Zones
14.2: Habitats
epiflora
14.2: Habitats
epipelagic zone
12.11: Biological Provinces and Zones
Equatorial Countercurrent
8.18: Box 2 - Equatorial Currents
equatorial currents
8.18: Box 2 - Equatorial Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
equatorial Pacific
7.6: Interannual Climate Variations
Equatorial Undercurrent
8.18: Box 2 - Equatorial Currents
equilibrium density level
7.4: Climatic Winds
equilibrium tides
10.3: Tides on an Ocean Covered Earth
equinoxes
7.3: Water and Heat Budgets
erosion
11.1: Processes that Form and Modify the Coast
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures
12.6: Primary Production and Nutrients
13.6: Estuaries
14.3: Feeding
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
17.2: CC2 Isostasy, Eustasy, and Sea Level
6.1: Sediments and Biogeochemical Cycles
6.2: Classification of Sediments
6.3: Lithogenous Sediments
7.9: Weather Systems
erosional coasts
11.1: Processes that Form and Modify the Coast
estuaries
10.6: Tidal Currents
11.6: Wetlands
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.6: Estuaries
14.2: Habitats
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.9: Box 1 - Petroleum in the Marine Environment
17.10: CC10 Modeling
17.8: CC8 Residence Time
4.8: Plate Interiors
6.3: Lithogenous Sediments
6.8: Continental Margin Sediments
9.9: Tsunamis
estuarine circulation
13.6: Estuaries
eukarya
12.1: How do we describe life?
eukaryotes
12.1: How do we describe life?
12.12: Plankton
17.14: CC14 Phototrophy, Light, and Nutrients
euphausiids
12.12: Plankton
14.3: Feeding
Eurasia
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.4: Convergent Plate Boundaries
Eurasian Plate
4.2: Plate Tectonics
Euroasia Plate
4.10: Present-Day Oceans
Europe
10.5: Tides in the Open Ocean
14.6: Reproduction
4.10: Present-Day Oceans
4.8: Plate Interiors
6.9: Distribution of Surface Sediments
7.6: Interannual Climate Variations
8.14: Ocean Circulation and Climate
8.5: Open-Ocean Surface Currents
European eels
14.6: Reproduction
European exploration
2.2: Exploration and Mapping
European Union
2.3: National and International Law Applied to Oceans
eustasy
17.2: CC2 Isostasy, Eustasy, and Sea Level
eutrophication
12.6: Primary Production and Nutrients
13.4: Algal Blooms
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
evaporation
13.1: Special Characteristics of Coastal Oceans
15.4: Rocky Intertidal Communities
16.9: Box 1 - Petroleum in the Marine Environment
17.3: CC3 Convection and Convection Cells
17.5: CC5 Transfer and Storage of Heat by Water
17.6: CC6 Salinity, Temperature, Pressure, and Water Density
17.8: CC8 Residence Time
5.10: Surface Tension and Viscosity
5.7: Heat Properties of Water
7.3: Water and Heat Budgets
7.5: Climate and Ocean Surface Water Properties
evaporites
6.5: Hydrogenous Sediments
evolution
12.13: Nekton
14.4: Hunting and Defense
exclusive economic zone
16.5: Ocean Pollution Issues
2.3: National and International Law Applied to Oceans
excurrent opening
12.12: Plankton
14.3: Feeding
exotic terranes
4.4: Convergent Plate Boundaries
exploration
2.2: Exploration and Mapping
extinction
16.6: Climate Change, Acidification, and Deoxygenation
extratropical cyclone
7.4: Climatic Winds
7.8: Global Climate Zones
extratropical cyclones
12.8: Geographic Variation in Primary Production
7.9: Weather Systems
9.9: Tsunamis
extreme environments
15.8: Hydrothermal Vents

Exxon Valdez
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
eye
7.9: Weather Systems
eye wall
7.9: Weather Systems

F

Falco hawkfish (Cirrhitichthys falco)
14.4: Hunting and Defense
Falkland Islands
2.3: National and International Law Applied to Oceans
fall bloom
13.3: Seasonal Cycles
fan worms
14.3: Feeding
14.4: Hunting and Defense
faults
13.6: Estuaries
fauna
14.3: Feeding
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
fecal pellets
12.10: Organic Carbon
12.2: Production, Consumption, and Decomposition
12.6: Primary Production and Nutrients
13.2: Nutrient Supply to the Coastal Photic Zone
14.3: Feeding
6.4: Biogenous Sediments
feedbacks
15.6: Polar Regions
16.6: Climate Change, Acidification, and Deoxygenation
17.9: CC9 The Global Greenhouse Effect
femtoplankton
12.12: Plankton
fermentation
12.2: Production, Consumption, and Decomposition
Ferrel cells
8.5: Open-Ocean Surface Currents
fertilization
14.6: Reproduction
fetch
13.3: Seasonal Cycles
7.9: Weather Systems
9.5: Sea Development and Wave Height
Fiji
12.13: Nekton
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
15.2: Coral Reefs
filefishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
filter feeders
14.3: Feeding
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
finback whales
12.12: Plankton
fine whale
12.13: Nekton
fins
14.5: Selected Adaptations in Fishes
fire gobies
14.4: Hunting and Defense
fire goby (Nemateleotris magnifica)
14.4: Hunting and Defense
fire urchin (Asthenosoma intermedium)
14.4: Hunting and Defense
fire urchin (Asthenosoma varium)
14.7: Associations
fish flukes
14.7: Associations
fish species
10.8: Marine Species and the Rhythm of Tides
12.13: Nekton
fisheries
13.2: Nutrient Supply to the Coastal Photic Zone
13.4: Algal Blooms
13.5: Fisheries
13.6: Estuaries
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
17.16: CC16 Maximum Sustainable Yield
2.4: Ocean Resources
fishes
11.6: Wetlands
12.10: Organic Carbon
12.12: Plankton
12.13: Nekton
12.7: Food Webs
13.3: Seasonal Cycles
13.4: Algal Blooms
13.5: Fisheries
13.6: Estuaries
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.7: Associations
14.8: Communication and Navigation
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
15.6: Polar Regions
15.7: Beyond the Sun's Light
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.9: Box 1 - Petroleum in the Marine Environment
17.16: CC16 Maximum Sustainable Yield
3.6: Sampling Living Organisms in the Sea
fjord estuaries
13.6: Estuaries
fjords
11.1: Processes that Form and Modify the Coast
12.2: Production, Consumption, and Decomposition
12.9: Dissolved Oxygen and Carbon Dioxide
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
4.9: Sea-Level Change and Climate
flabellina nudibranch (Flabellina rubrolineata)
14.3: Feeding
flagella
12.12: Plankton
flagellates
12.12: Plankton
13.2: Nutrient Supply to the Coastal Photic Zone
14.3: Feeding
flamboyant cuttlefish (Metasepia pfefferi)
14: Marine Ecology
14.6: Reproduction
flasher scorpionfish (Scorpaenopsis macrochir)
14.7: Associations
flatfish
12.13: Nekton
flatworms
16.6: Climate Change, Acidification, and Deoxygenation
floats
3.5: Chemical and Physical Oceanography
3.7: Scuba, Manned, and Unmanned Submersibles
8.4: Geostrophic Currents
flood plain
11.7: Deltas
flooding
11.1: Processes that Form and Modify the Coast
11.7: Deltas
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
7.9: Weather Systems
flora
15.4: Rocky Intertidal Communities
Florida
10.2: Characteristics of Tides
11.1: Processes that Form and Modify the Coast
11.3: Barrier Islands and Lagoons
12.13: Nekton
13.4: Algal Blooms
16.6: Climate Change, Acidification, and Deoxygenation
2.4: Ocean Resources
6.3: Lithogenous Sediments
6.5: Hydrogenous Sediments
7.8: Global Climate Zones
7.9: Weather Systems
8.4: Geostrophic Currents
8.9: Coastal Currents
Florida Current
8.4: Geostrophic Currents
flounders
12.13: Nekton
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
flowery flounder (Bothus mancus)
12.13: Nekton
fluid dynamics
17.1: CC1 Density and Layering in Fluids
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.3: CC3 Convection and Convection Cells
8.17: Box 1- Westward Intensification
fluids
17.1: CC1 Density and Layering in Fluids
flyingfishes
14.5: Selected Adaptations in Fishes
fog
7.10: Local Weather Effects

food chain
12.10: Organic Carbon
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.7: Food Webs
13.3: Seasonal Cycles
13.6: Estuaries
15.5: Sargasso Sea
17.10: CC10 Modeling
17.15: CC15 Food Chain Efficiency
17.18: CC18 Toxicity

food web
12.11: Biological Provinces and Zones
12.12: Plankton
12.7: Food Webs
13.2: Nutrient Supply to the Coastal Photic Zone
14.3: Feeding
15.3: Kelp Forests
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
17.10: CC10 Modeling
7.6: Interannual Climate Variations

foraminifera
11.2: Beaches
12.12: Plankton
6.4: Biogenous Sediments

foreshore
11.1: Processes that Form and Modify the Coast

fossil fuels
11.7: Deltas
17.9: CC9 The Global Greenhouse Effect
5.6: Dissolved Chemicals in Seawater
6.1: Sediments and Biogeochemical Cycles
8.14: Ocean Circulation and Climate

fossil record
1.2: Oceans and the Origins of Life
6.10: The Sediment Historical Record

fossils
1.2: Oceans and the Origins of Life
17.7: CC7 Radioactivity and Age Dating
4.2: Plate Tectonics
4.4: Convergent Plate Boundaries
6.10: The Sediment Historical Record
9.9: Tsunamis

fouling
3.1: Difficulties of Studying the Ocean Environment

fracture zones
4.6: Transform Faults and Fracture Zones

fragmentation
14.6: Reproduction

France
10.9: Energy from Tides
16.9: Box 1 - Petroleum in the Marine Environment
2.4: Ocean Resources
3.4: Seafloor Sediments
7.6: Interannual Climate Variations

freezing
5.9: Ice Formation

freezing point
5.7: Heat Properties of Water

frequency
14.8: Communication and Navigation
5.11: Transmission of Light and Other Electromagnetic Radiation

freshwater
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.6: Estuaries
5.1: Origins and Distribution of the Earth's Water
5.9: Ice Formation

freshwater eels
13.6: Estuaries

freshwater input
8.9: Coastal Currents

friction
10.3: Tides on an Ocean Covered Earth
11.2: Beaches
13.1: Special Characteristics of Coastal Oceans
14.5: Selected Adaptations in Fishes
16.5: Ocean Pollution Issues
17.12: CC12 The Coriolis Effect
17.13: CC13 Geostrophic Flow
4.4: Convergent Plate Boundaries
7.9: Weather Systems
8.2: Wind-Driven Currents
9.4: Making Waves
9.6: Wave Dissipation

fringing reefs
11.5: Coral Reefs and Atolls
15.2: Coral Reefs
4.8: Plate Interiors

frogfishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
15.5: Sargasso Sea

fronts
13.1: Special Characteristics of Coastal Oceans
8.10: Eddies
8.9: Coastal Currents

frustule
12.12: Plankton
12.6: Primary Production and Nutrients
13.3: Seasonal Cycles
6.4: Biogenous Sediments

Fukushima
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues

fungi
12.1: How do we describe life?
12.10: Organic Carbon
14.2: Habitats
6.4: Biogenous Sediments

G

Gakkel Ridge
4.5: Divergent Plate Boundaries

Galápagos Islands
12.13: Nekton
13.2: Nutrient Supply to the Coastal Photic Zone
2.2: Exploration and Mapping

Galápagos Ridge
15.8: Hydrothermal Vents

Galveston Bay
16.5: Ocean Pollution Issues

Galveston hurricane
7.9: Weather Systems

Ganges River
13.1: Special Characteristics of Coastal Oceans
4.10: Present-Day Oceans
6.3: Lithogenous Sediments

gases
7.1: The Atmosphere

gastropods
14.3: Feeding
14.4: Hunting and Defense

gelatinous holoplankton
12.12: Plankton

genes
12.12: Plankton

genetic diversity
14.6: Reproduction

genetics
12.1: How do we describe life?
12.12: Plankton
12.4: Studying Microbes in the Ocean
15.8: Hydrothermal Vents
16.3: Toxicity
17.15: CC15 Food Chain Efficiency
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.19: CC19 Metagenomics

genome sequencing
12.1: How do we describe life?

genus
12.1: How do we describe life?

geographic information systems
3.9: Computers and Modeling

geography
7.10: Local Weather Effects
7.5: Climate and Ocean Surface Water Properties

geological features
4.7: Hot Spots

geological history
3.4: Seafloor Sediments

geological processes
4.4: Convergent Plate Boundaries
4.5: Divergent Plate Boundaries
4.8: Plate Interiors
6.2: Classification of Sediments

geological time
6.10: The Sediment Historical Record

geology
4.1: Structure of the Earth
4.2: Plate Tectonics
6.3: Lithogenous Sediments

Georges Bank
13.2: Nutrient Supply to the Coastal Photic Zone

Georgia
10.2: Characteristics of Tides
8.9: Coastal Currents

geostrophic currents
8.11: Inertial Currents
8.18: Box 2 - Equatorial Currents
8.7: High-Latitude Surface Currents

geostrophic flow
17.13: CC13 Geostrophic Flow
7.9: Weather Systems
9.11: Rossby and Kelvin Waves

geostrophic gyres
8.5: Open-Ocean Surface Currents

ghost pipefish (Solenostomus halimeda)
14.4: Hunting and Defense

ghost pipefishes
14.5: Selected Adaptations in Fishes

giant bladder kelp (Macrocystis pyrifera)
15.3: Kelp Forests

giant clam (Tridacna gigas)
14.7: Associations
16.6: Climate Change, Acidification, and Deoxygenation
giant manta ray (Manta birostris)
12.13: Nekton
giant squid
12.13: Nekton
giant tube worm (Riftia pachyptila)
15.8: Hydrothermal Vents
gills
12.13: Nekton
14.5: Selected Adaptations in Fishes
GIS
3.9: Computers and Modeling
glacial maxima
4.9: Sea-Level Change and Climate
glaciers
11.1: Processes that Form and Modify the Coast
11.7: Deltas
13.3: Seasonal Cycles
13.6: Estuaries
17.2: CC2 Isostasy, Eustasy, and Sea Level
6.3: Lithogenous Sediments
glass squid
14.4: Hunting and Defense
gliders
3.7: Scuba, Manned, and Unmanned Submersibles
Global Climate Observing System (GCOS)
3.7: Scuba, Manned, and Unmanned Submersibles
Global Ocean Observing System (GOOS)
3.7: Scuba, Manned, and Unmanned Submersibles
Global Positioning System (GPS)
3.8: Satellites
global warming
16.6: Climate Change, Acidification, and Deoxygenation
17.9: CC9 The Global Greenhouse Effect
4.9: Sea-Level Change and Climate
7.4: Climatic Winds
Glomar Challenger
3.4: Seafloor Sediments
goatfish
14.3: Feeding
goatfishes
15.2: Coral Reefs
Gobi Desert
6.3: Lithogenous Sediments
gobies
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.7: Associations
gold
15.8: Hydrothermal Vents
Gondawanaland
4.2: Plate Tectonics
Gondwanaland
4.2: Plate Tectonics
gorgonian sea fan
14.3: Feeding
gorgonian seahorse (Hippocampus bargibanti)
14.4: Hunting and Defense
GPS
3.5: Chemical and Physical Oceanography
grab samplers
3.4: Seafloor Sediments
grain size
11.2: Beaches
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
6.2: Classification of Sediments
6.8: Continental Margin Sediments
Grakkel Ridge
15.8: Hydrothermal Vents
gravel
6.2: Classification of Sediments
gravity
10.1: Tide-Generating Forces
14.5: Selected Adaptations in Fishes
17.1: CC1 Density and Layering in Fluids
17.12: CC12 The Coriolis Effect
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
4.1: Structure of the Earth
5.1: Origins and Distribution of the Earth's Water
8.13: Thermohaline Circulation
9.3: What is Wave Motion?
gravity corers
3.4: Seafloor Sediments
gravity waves
9.3: What is Wave Motion?
gray reef shark (Carcharhinus amblyrhynchos)
12.13: Nekton
14.5: Selected Adaptations in Fishes
gray whale
12.13: Nekton
14.6: Reproduction
grazers
12.6: Primary Production and Nutrients
13.2: Nutrient Supply to the Coastal Photic Zone
14.2: Habitats
14.3: Feeding
15.3: Kelp Forests
15.5: Sargasso Sea
16.2: Adverse Effects of Human Activities
Great Barrier Reef
11.5: Coral Reefs and Atolls
Great Britain
12.13: Nekton
2.3: National and International Law Applied to Oceans
Great Lakes
9.12: Standing Waves
Great Oxygenation Event
12.3: Microbes in Charge
great white shark (Carcharodon carcharias)
12.13: Nekton
Greece
13.4: Algal Blooms
green algae
12.12: Plankton
green sea turtle (Chelonia mydas)
12.13: Nekton
14.3: Feeding
greenhouse effect
1.1: The Ocean's and Earth's Environment
11.1: Processes that Form and Modify the Coast
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.9: CC9 The Global Greenhouse Effect
3.8: Satellites
4.9: Sea-Level Change and Climate
5.7: Heat Properties of Water
9.5: Sea Development and Wave Height
greenhouse gases
16.6: Climate Change, Acidification, and Deoxygenation
17.9: CC9 The Global Greenhouse Effect
7.3: Water and Heat Budgets
Greenland
16.6: Climate Change, Acidification, and Deoxygenation
4.10: Present-Day Oceans
4.5: Divergent Plate Boundaries
4.7: Hot Spots
6.3: Lithogenous Sediments
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
Greenland Ice Sheet
16.6: Climate Change, Acidification, and Deoxygenation
Greenland Sea
16.11: Box 3 - Radioactive Wastes in the Marine Environment
8.13: Thermohaline Circulation
groins
11.4: Beaches and Human Structures
16.2: Adverse Effects of Human Activities
groundwater
12.2: Production, Consumption, and Decomposition
16.3: Toxicity
17.2: CC2 Isostasy, Eustasy, and Sea Level
5.1: Origins and Distribution of the Earth's Water
growth
12.6: Primary Production and Nutrients
grunions
10.8: Marine Species and the Rhythm of Tides
14.6: Reproduction
Guarm
2.3: National and International Law Applied to Oceans
Gulf of Alaska
7.6: Interannual Climate Variations
Gulf of Maine
13.2: Nutrient Supply to the Coastal Photic Zone

Gulf of Mexico
10.2: Characteristics of Tides
10.5: Tides in the Open Ocean
11.1: Processes that Form and Modify the Coast
11.3: Barrier Islands and Lagoons
13.4: Algal Blooms
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.9: Box 1 - Petroleum in the Marine Environment
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
4.10: Present-Day Oceans
4.9: Sea-Level Change and Climate
6.10: The Sediment Historical Record
6.3: Lithogenous Sediments
6.6: Cosmogenous Sediments
6.9: Distribution of Surface Sediments
7.6: Interannual Climate Variations
7.9: Weather Systems
8.14: Ocean Circulation and Climate
8.4: Geostrophic Currents
9.5: Sea Development and Wave Height

Gulf Stream
13.6: Estuaries
14.6: Reproduction
7.5: Climate and Ocean Surface Water Properties
7.8: Global Climate Zones
7.9: Weather Systems
8.10: Eddies
8.14: Ocean Circulation and Climate
8.18: Box 2 - Equatorial Currents
8.4: Geostrophic Currents
8.5: Open-Ocean Surface Currents
8.9: Coastal Currents
9.11: Rossby and Kelvin Waves

gyres
13.2: Nutrient Supply to the Coastal Photic Zone
8.17: Box 1- Westward Intensification
8.5: Open-Ocean Surface Currents
8.8: Upwelling and Downwelling

H

habitats
10.8: Marine Species and the Rhythm of Tides
11.3: Barrier Islands and Lagoons
12.13: Nekton
13.4: Algal Blooms
14.2: Habitats
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
17.17: CC17 Species Diversity and Biodiversity
2.4: Ocean Resources
3.6: Sampling Living Organisms in the Sea
3.7: Scuba, Manned, and Unmanned Submersibles

HABs
13.4: Algal Blooms
16.2: Adverse Effects of Human Activities

hadal zone
12.11: Biological Provinces and Zones

Haddon's sea anemone (Stichodactyla haddoni)
14.4: Hunting and Defense

Hadley cells
7.4: Climatic Winds
8.5: Open-Ocean Surface Currents

hadopelagic zone
12.11: Biological Provinces and Zones

hagfishes
12.13: Nekton

halocline
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.6: Estuaries
15.6: Polar Regions
8.13: Thermohaline Circulation

harbor seals (Phoca vitulina)
12.13: Nekton
15.3: Kelp Forests

harbors
11.4: Beaches and Human Structures

hard corals
13.3: Seasonal Cycles
14.6: Reproduction
16.6: Climate Change, Acidification, and Deoxygenation
17.17: CC17 Species Diversity and Biodiversity

hard parts
12.12: Plankton
12.6: Primary Production and Nutrients
13.3: Seasonal Cycles
14.3: Feeding
15.2: Coral Reefs
16.6: Climate Change, Acidification, and Deoxygenation
6.4: Biogenous Sediments

Harlequin shrimp (Hymenocera elegans)
14.3: Feeding

harmful algal blooms
13.4: Algal Blooms

harmonic analysis
10.6: Tidal Currents

harp shell (Harpa major)
14.3: Feeding

hatchetfishes
15.7: Beyond the Sun's Light

Hawaii
11.1: Processes that Form and Modify the Coast
11.2: Beaches
13.1: Special Characteristics of Coastal Oceans
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
2.2: Exploration and Mapping
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
4.7: Hot Spots
6.7: Sediment Transport, Deposition, and Accumulation
7.10: Local Weather Effects
7.6: Interannual Climate Variations
9.9: Tsunamis

Hawaiian cleaner wrasse (Labroides pcthirophagus)
14.4: Hunting and Defense

hawkfishes
14.4: Hunting and Defense
15.2: Coral Reefs

hawksbill sea turtle (Eretmochelys imbricata)
12.13: Nekton

hazardous waste
16.3: Toxicity

headlands
11.1: Processes that Form and Modify the Coast
13.2: Nutrient Supply to the Coastal Photic Zone
9.8: Waves in Shallow Water

heart urchin
14.3: Feeding

heat budget
7.3: Water and Heat Budgets

heat capacity
16.6: Climate Change, Acidification, and Deoxygenation
17.5: CC5 Transfer and Storage of Heat by Water
5.2: The Water Molecule
5.7: Heat Properties of Water
7.10: Local Weather Effects
7.8: Global Climate Zones

heat energy
5.7: Heat Properties of Water
8.10: Eddies

heat transfer
17.3: CC3 Convection and Convection Cells
7.3: Water and Heat Budgets

height
10.7: Tide Prediction

helical vortices
8.12: Langmuir Circulation

herbivores
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.7: Food Webs
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
14.3: Feeding
15.4: Rocky Intertidal Communities

hermaphroditism
14.6: Reproduction

hermatypic corals
15.2: Coral Reefs

hermit crabs
14.7: Associations
15.4: Rocky Intertidal Communities

herring gull (Larus argentatus)
12.13: Nekton

herrings
12.13: Nekton
12.7: Food Webs
13.5: Fisheries
14.6: Reproduction

heterotrophs
12.2: Production, Consumption, and Decomposition
12.7: Food Webs

high latitudes
13.3: Seasonal Cycles
17.5: CC5 Transfer and Storage of Heat by Water

high nutrient low chlorophyll areas
12.6: Primary Production and Nutrients

high pressure
7.9: Weather Systems

high seas
2.3: National and International Law Applied to Oceans

high tide
10.2: Characteristics of Tides
10.8: Marine Species and the Rhythm of Tides

higher high water
10.2: Characteristics of Tides

higher low water
10.2: Characteristics of Tides
highs
7.9: Weather Systems
Himalaya Mountains
4.10: Present-Day Oceans
4.3: Plate Boundaries
Himalayan Mountains
7.4: Climatic Winds
HMS Beagle
2.2: Exploration and Mapping
4.8: Plate Interiors
HMS Challenger
2.2: Exploration and Mapping
3.2: The Unseen Domain
HNLC areas
12.12: Plankton
12.6: Primary Production and Nutrients
holoplankton
12.12: Plankton
Homo erectus
2.2: Exploration and Mapping
horizontal motion
8.8: Upwelling and Downwelling
horse latitudes
7.4: Climatic Winds
horseshoe crabs
10.8: Marine Species and the Rhythm of Tides
14.6: Reproduction
host
14.7: Associations
hot spots
11.1: Processes that Form and Modify the Coast
4.2: Plate Tectonics
4.7: Hot Spots
Hudson Bay
4.10: Present-Day Oceans
Hudson River
13.6: Estuaries
human health
13.4: Algal Blooms
humidity
17.11: CC11 Chaos
humpback whale (Megaptera novaeangliae)
12.12: Plankton
12.13: Nekton
14.6: Reproduction
humpbacked shrimp (Hippolyte commensalis)
14.7: Associations
Hunga Tonga
4.4: Convergent Plate Boundaries
6.3: Lithogenous Sediments
7.1: The Atmosphere
9.9: Tsunamis
hunting
14.4: Hunting and Defense
17.18: CC18 Toxicity
Hurricane Andrew
7.9: Weather Systems
Hurricane Camille
7.9: Weather Systems
Hurricane Catarina
7.9: Weather Systems
Hurricane Harvey
7.9: Weather Systems
Hurricane Helene
7.9: Weather Systems
Hurricane Hugo
7.9: Weather Systems
Hurricane Irma
7.9: Weather Systems
Hurricane Ivan
9.5: Sea Development and Wave Height
Hurricane Katrina
11.7: Deltas
7.9: Weather Systems
hurricanes
11.3: Barrier Islands and Lagoons
17.10: CC10 Modeling
17.17: CC17 Species Diversity and Biodiversity
17.5: CC5 Transfer and Storage of Heat by Water
17.9: CC9 The Global Greenhouse Effect
7.9: Weather Systems
hydration
5.3: The Dissolving Power of Water
hydrogen
17.14: CC14 Phototrophy, Light, and Nutrients
5.2: The Water Molecule
hydrogen bond
5.10: Surface Tension and Viscosity
5.2: The Water Molecule
5.7: Heat Properties of Water
hydrogenous sediments
15.8: Hydrothermal Vents
6.2: Classification of Sediments
6.5: Hydrogenous Sediments
6.8: Continental Margin Sediments
hydroids
14.3: Feeding
14.4: Hunting and Defense
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
hydrologic cycle
5.1: Origins and Distribution of the Earth's Water
hydrophones
3.4: Seafloor Sediments
hydrosphere
4.1: Structure of the Earth
hydrostatic pressure
11.7: Deltas
hydrothermal minerals
2.4: Ocean Resources
6.5: Hydrogenous Sediments
hydrothermal vents
12.11: Biological Provinces and Zones
12.2: Production, Consumption, and Decomposition
14.2: Habitats
15.8: Hydrothermal Vents
2.2: Exploration and Mapping
3.6: Sampling Living Organisms in the Sea
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
8.15: Tracing Water Masses
hypoxic
13.4: Algal Blooms
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation

I

ice
15.6: Polar Regions
17.5: CC5 Transfer and Storage of Heat by Water
5.9: Ice Formation
7.8: Global Climate Zones
ice ages
11.7: Deltas
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.9: CC9 The Global Greenhouse Effect
4.9: Sea-Level Change and Climate
8.14: Ocean Circulation and Climate
ice exclusion
13.3: Seasonal Cycles
15.6: Polar Regions
7.8: Global Climate Zones
ice sheets
4.9: Sea-Level Change and Climate
Iceland
4.5: Divergent Plate Boundaries
4.7: Hot Spots
8.14: Ocean Circulation and Climate
Icelandic Ridge
4.7: Hot Spots
impact craters
6.6: Cosmogenous Sediments
impact waves
9.4: Making Waves
in situ
8.15: Tracing Water Masses
incubation
14.6: Reproduction
incurrent opening
12.12: Plankton
14.3: Feeding
India
11.1: Processes that Form and Modify the Coast
11.7: Deltas
12.9: Dissolved Oxygen and Carbon Dioxide
2.2: Exploration and Mapping
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.3: Plate Boundaries
7.4: Climatic Winds
7.6: Interannual Climate Variations
7.9: Weather Systems
9.9: Tsunamis
Indian Ocean
12.11: Biological Provinces and Zones
12.13: Nekton
12.9: Dissolved Oxygen and Carbon Dioxide
13.2: Nutrient Supply to the Coastal Photic Zone
14.3: Feeding
15.2: Coral Reefs
15.8: Hydrothermal Vents
2.2: Exploration and Mapping
2.4: Ocean Resources
4.10: Present-Day Oceans
4.4: Convergent Plate Boundaries
6.3: Lithogenous Sediments
6.9: Distribution of Surface Sediments
7.4: Climatic Winds
7.6: Interannual Climate Variations
7.9: Weather Systems
8.10: Eddies
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
8.5: Open-Ocean Surface Currents
9.5: Sea Development and Wave Height
9.9: Tsunamis
Indian Ocean Dipole
7.6: Interannual Climate Variations

Indoensian Arc
4.10: Present-Day Oceans
Indonesia
12.13: Nekton
13.2: Nutrient Supply to the Coastal Photic Zone
14.3: Feeding
14.4: Hunting and Defense
14.6: Reproduction
14.7: Associations
15.2: Coral Reefs
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
6.10: The Sediment Historical Record
7.6: Interannual Climate Variations
9.9: Tsunamis
Indus River
4.10: Present-Day Oceans
6.3: Lithogenous Sediments
industrial waste
13.4: Algal Blooms
inertial currents
8.11: Inertial Currents
infauna
12.11: Biological Provinces and Zones
14.2: Habitats
infrared radiation
17.9: CC9 The Global Greenhouse Effect
7.3: Water and Heat Budgets
innkeeper worm (Urechis caupo)
14.3: Feeding
inorganic compounds
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
interannual climate variations
7.4: Climatic Winds
interdisciplinary studies
2.1: What is Oceanography
interglacial periods
4.9: Sea-Level Change and Climate
Intergovernmental Panel on Climate Change (IPCC)
16.6: Climate Change, Acidification, and Deoxygenation
intermediate water masses
8.13: Thermohaline Circulation
intermediate waves
9.8: Waves in Shallow Water
internal waves
13.1: Special Characteristics of Coastal Oceans
8.13: Thermohaline Circulation
international law
2.3: National and International Law Applied to Oceans
International Seabed Authority
2.3: National and International Law Applied to Oceans
intertidal community
15.4: Rocky Intertidal Communities
intertidal environments
10.8: Marine Species and the Rhythm of Tides
intertidal zone
10.8: Marine Species and the Rhythm of Tides
12.11: Biological Provinces and Zones
14.2: Habitats
intertropical convergence zone
7.4: Climatic Winds
8.18: Box 2 - Equatorial Currents
8.6: Equatorial Surface Currents
inverse estuaries
13.6: Estuaries
invertebrates
12.12: Plankton
12.13: Nekton
13.3: Seasonal Cycles
13.4: Algal Blooms
13.6: Estuaries
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
15.2: Coral Reefs
15.3: Kelp Forests
15.7: Beyond the Sun's Light
15.8: Hydrothermal Vents
16.5: Ocean Pollution Issues
17.16: CC16 Maximum Sustainable Yield
3.6: Sampling Living Organisms in the Sea
invertegrates
16.2: Adverse Effects of Human Activities
ionic bonds
5.2: The Water Molecule
5.3: The Dissolving Power of Water
ions
12.6: Primary Production and Nutrients
13.2: Nutrient Supply to the Coastal Photic Zone
14.5: Selected Adaptations in Fishes
17.14: CC14 Phototrophy, Light, and Nutrients
17.6: CC6 Salinity, Temperature, Pressure, and Water Density
5.2: The Water Molecule
5.3: The Dissolving Power of Water
iridium
6.10: The Sediment Historical Record
iron
12.6: Primary Production and Nutrients
15.6: Polar Regions
15.8: Hydrothermal Vents
Irrawaddy River
6.3: Lithogenous Sediments
island effect
7.10: Local Weather Effects
7.7: Land-Ocean-Atmosphere Interactions
islands
11.5: Coral Reefs and Atolls
13.2: Nutrient Supply to the Coastal Photic Zone
isopods
14.7: Associations
15.4: Rocky Intertidal Communities
isostasy
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.1: Structure of the Earth
4.7: Hot Spots
isothermal
13.1: Special Characteristics of Coastal Oceans
isotherms
7.5: Climate and Ocean Surface Water Properties
isotopes
12.12: Plankton
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
17.7: CC7 Radioactivity and Age Dating
6.10: The Sediment Historical Record
8.15: Tracing Water Masses
Italy
13.4: Algal Blooms
Ixotoc 1
16.9: Box 1 - Petroleum in the Marine Environment

J

Japan
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
3.3: Bathymetry
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
6.10: The Sediment Historical Record
7.8: Global Climate Zones
9.9: Tsunamis
Jason II
3.7: Scuba, Manned, and Unmanned Submersibles
Jason Jr.
3.7: Scuba, Manned, and Unmanned Submersibles
Java
9.9: Tsunamis
Java Sea
4.10: Present-Day Oceans
Java Trench
15.7: Beyond the Sun's Light
jawless fishes
12.13: Nekton
jellies
12.12: Plankton
12.13: Nekton
14.4: Hunting and Defense
16.5: Ocean Pollution Issues
2.4: Ocean Resources
jet stream
6.3: Lithogenous Sediments
7.4: Climatic Winds
9.11: Rossby and Kelvin Waves
jetties
11.4: Beaches and Human Structures
JODIS Resolution
4.1: Structure of the Earth
JOIDES Resolution
3.4: Seafloor Sediments
juveniles
11.6: Wetlands
12.12: Plankton
13.3: Seasonal Cycles
13.5: Fisheries
13.6: Estuaries
14.4: Hunting and Defense
16.9: Box 1 - Petroleum in the Marine Environment
17.15: CC15 Food Chain Efficiency
17.18: CC18 Toxicity

K

Keeling Atoll
4.8: Plate Interiors
kelp
12.12: Plankton
12.5: Phototropic Primary Production and Light
13.1: Special Characteristics of Coastal Oceans
15.3: Kelp Forests
kelp forests
12.14: Benthos
14.2: Habitats
15.3: Kelp Forests

Kelvin waves
9.11: Rossby and Kelvin Waves
Kilauea
4.7: Hot Spots
Kilauea Volcano
11.1: Processes that Form and Modify the Coast
killer whale (Orcinus orca)
12.13: Nekton
12.7: Food Webs
14.8: Communication and Navigation
15.3: Kelp Forests
kinetic energy
8.2: Wind-Driven Currents
9.3: What is Wave Motion?
9.8: Waves in Shallow Water
king penguins (Aptenodytes patagonicus)
12.13: Nekton
Klein's butterflyfish (Chaetodon kleinii)
14.5: Selected Adaptations in Fishes
Krakatau
4.4: Convergent Plate Boundaries
6.10: The Sediment Historical Record
6.3: Lithogenous Sediments
9.9: Tsunamis
krill
12.12: Plankton
12.7: Food Webs
14.3: Feeding
15.6: Polar Regions
Kuril Islands
2.3: National and International Law Applied to Oceans
Kuroshio Current
13.6: Estuaries
7.8: Global Climate Zones
8.10: Eddies
8.5: Open-Ocean Surface Currents

L

La Niña
7.6: Interannual Climate Variations
Labrador
8.9: Coastal Currents
lagoons
11.3: Barrier Islands and Lagoons
11.5: Coral Reefs and Atolls
11.6: Wetlands
11.7: Deltas
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
14.6: Reproduction
15.2: Coral Reefs
2.4: Ocean Resources
Laguna Madre
13.6: Estuaries
Lake Agassiz
8.14: Ocean Circulation and Climate
lampreys
12.13: Nekton
land breeze
7.10: Local Weather Effects
7.7: Land-Ocean-Atmosphere Interactions
land climate zones
7.8: Global Climate Zones
landmasses
7.7: Land-Ocean-Atmosphere Interactions
landslides
11.1: Processes that Form and Modify the Coast
6.3: Lithogenous Sediments
Langmuir circulation
12.12: Plankton
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
8.12: Langmuir Circulation
lanternfishes
15.7: Beyond the Sun's Light
larvae
10.8: Marine Species and the Rhythm of Tides
12.12: Plankton
13.2: Nutrient Supply to the Coastal Photic Zone
13.5: Fisheries
13.6: Estuaries
14.2: Habitats
14.6: Reproduction
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
16.6: Climate Change, Acidification, and Deoxygenation
17.15: CC15 Food Chain Efficiency
17.16: CC16 Maximum Sustainable Yield
17.18: CC18 Toxicity
latent heat
5.2: The Water Molecule
5.7: Heat Properties of Water
7.2: Water Vapor in the Atmosphere
latent heat of fusion
17.5: CC5 Transfer and Storage of Heat by Water
5.7: Heat Properties of Water
latent heat of vaporization
17.5: CC5 Transfer and Storage of Heat by Water
5.7: Heat Properties of Water
7.2: Water Vapor in the Atmosphere
7.3: Water and Heat Budgets
lateral line
14.8: Communication and Navigation
latitude
1.3: How to Study Ocean Data
11.1: Processes that Form and Modify the Coast
12.11: Biological Provinces and Zones
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
14.3: Feeding
15.2: Coral Reefs
16.2: Adverse Effects of Human Activities
17.10: CC10 Modeling
17.12: CC12 The Coriolis Effect
17.5: CC5 Transfer and Storage of Heat by Water
2.2: Exploration and Mapping
4.7: Hot Spots
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
7.3: Water and Heat Budgets
8.5: Open-Ocean Surface Currents
lava
15.4: Rocky Intertidal Communities
4.5: Divergent Plate Boundaries
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
7.10: Local Weather Effects
Law of the Sea
2.3: National and International Law Applied to Oceans
leaching
13.6: Estuaries
lead
15.8: Hydrothermal Vents
16.3: Toxicity
leatherback sea turtle
12.13: Nekton
leeward
15.2: Coral Reefs
7.10: Local Weather Effects
9.4: Making Waves
lemon damsel (Pomacentrus moluccensis)
14.7: Associations
leopard seal (Hydrurga leptonyx)
15.6: Polar Regions
levees
11.7: Deltas
13.6: Estuaries
16.2: Adverse Effects of Human Activities
lichens
15.4: Rocky Intertidal Communities
life cycles
11.6: Wetlands
13.6: Estuaries
14.1: Ecological Requirements
14.6: Reproduction
light
5.11: Transmission of Light and Other Electromagnetic Radiation
light energy
17.14: CC14 Phototrophy, Light, and Nutrients
light intensity
14.8: Communication and Navigation
15.1: Communities and Niches
15.6: Polar Regions
16.5: Ocean Pollution Issues
17.18: CC18 Toxicity
limestone
15.2: Coral Reefs
6.5: Hydrogenous Sediments
limpets
15.8: Hydrothermal Vents
limpets snails
15.4: Rocky Intertidal Communities
lionfishes
14.5: Selected Adaptations in Fishes
liquid molecules
5.10: Surface Tension and Viscosity
lithogenous sediments
12.6: Primary Production and Nutrients
6.2: Classification of Sediments
6.3: Lithogenous Sediments
6.5: Hydrogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments
6.9: Distribution of Surface Sediments
lithosphere
4.1: Structure of the Earth
4.2: Plate Tectonics
4.4: Convergent Plate Boundaries
lithospheric plates
11.1: Processes that Form and Modify the Coast
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.1: Structure of the Earth
4.3: Plate Boundaries
4.6: Transform Faults and Fracture Zones
Little Ice Age
8.14: Ocean Circulation and Climate
littoral drift
11.2: Beaches
littoral zone
11.2: Beaches
12.11: Biological Provinces and Zones

lizardfishes
14.4: Hunting and Defense
lobsters
12.12: Plankton
local effects
7.7: Land-Ocean-Atmosphere Interactions
Lock Ness monster
17.15: CC15 Food Chain Efficiency
loggerhead sea turtle
12.13: Nekton
Loihi
4.7: Hot Spots
Long Island Sound
10.6: Tidal Currents
longitude
1.3: How to Study Ocean Data
17.10: CC10 Modeling
2.2: Exploration and Mapping
longnose hawkfish (Oxycirrhites typus)
15.2: Coral Reefs
longshore bar
11.2: Beaches
11.4: Beaches and Human Structures
longshore currents
11.2: Beaches
8.11: Inertial Currents
9.6: Wave Dissipation
longshore drift
11.2: Beaches
11.3: Barrier Islands and Lagoons
11.4: Beaches and Human Structures
13.6: Estuaries
Los Angeles
16.10: Box 2 - Sewage Wastes in the Marine Environment
Louisiana
7.9: Weather Systems
low pressure
7.9: Weather Systems
low tide
10.2: Characteristics of Tides
10.8: Marine Species and the Rhythm of Tides
lower high water
10.2: Characteristics of Tides
lower low water
10.2: Characteristics of Tides
lows
7.9: Weather Systems
lugworms (Arenicola brasiliensis)
14.3: Feeding
lunar tide
10.3: Tides on an Ocean Covered Earth
10.8: Marine Species and the Rhythm of Tides
lures
14.4: Hunting and Defense

M

Macdonald Seamount
4.7: Hot Spots
Mackenzie Delta
11.7: Deltas
mackerel
12.7: Food Webs
14.5: Selected Adaptations in Fishes
macroalgae
12.12: Plankton
13.1: Special Characteristics of Coastal Oceans
14.3: Feeding
14.6: Reproduction
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
macroplankton
12.12: Plankton
Madagascar
4.10: Present-Day Oceans
7.4: Climatic Winds
magma
15.8: Hydrothermal Vents
magmatic arc
11.1: Processes that Form and Modify the Coast
4.4: Convergent Plate Boundaries
magnesium
17.14: CC14 Phototrophy, Light, and Nutrients
5.6: Dissolved Chemicals in Seawater
magnetic field
6.10: The Sediment Historical Record
magnificent sea anemone (Heteractis magnifica)
14.3: Feeding
Maine
7.10: Local Weather Effects
manatees (Trichechus manatus)
12.13: Nekton
mandarinfish (Synchiropus splendidus)
14.6: Reproduction
manganese
15.8: Hydrothermal Vents
manganese nodules
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
6.5: Hydrogenous Sediments
mangrove swamps
11.1: Processes that Form and Modify the Coast
mangroves
11.6: Wetlands
13.6: Estuaries
Manitoba
8.14: Ocean Circulation and Climate
manned submersibles
3.7: Scuba, Manned, and Unmanned Submersibles
manta rays
14.5: Selected Adaptations in Fishes
14.7: Associations
mantle
17.1: CC1 Density and Layering in Fluids
17.3: CC3 Convection and Convection Cells
4.1: Structure of the Earth
4.2: Plate Tectonics
mantle convection
4.7: Hot Spots
map projections
1.3: How to Study Ocean Data
maps
1.3: How to Study Ocean Data
marginal seas
13.1: Special Characteristics of Coastal Oceans
13.4: Algal Blooms
15.6: Polar Regions
4.10: Present-Day Oceans
4.5: Divergent Plate Boundaries
5.6: Dissolved Chemicals in Seawater
7.5: Climate and Ocean Surface Water Properties
Mariana Arc
15.8: Hydrothermal Vents
Mariana Islands
4.4: Convergent Plate Boundaries
Mariana Trench
2.2: Exploration and Mapping
3.7: Scuba, Manned, and Unmanned Submersibles
4.3: Plate Boundaries
marine
7.5: Climate and Ocean Surface Water Properties
marine animals
11.6: Wetlands
12.8: Geographic Variation in Primary Production
marine biology
11.5: Coral Reefs and Atolls
12.11: Biological Provinces and Zones
12.12: Plankton
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
14.3: Feeding
14.4: Hunting and Defense
14.7: Associations
14.8: Communication and Navigation
15.6: Polar Regions
15.7: Beyond the Sun's Light
2.1: What is Oceanography
3.6: Sampling Living Organisms in the Sea
3.7: Scuba, Manned, and Unmanned Submersibles
marine ecology
10.8: Marine Species and the Rhythm of Tides
12.14: Benthos
15.1: Communities and Niches
marine ecosystems
1.1: The Ocean's and Earth's Environment
12.13: Nekton
12.7: Food Webs
13.6: Estuaries
14.4: Hunting and Defense
14.7: Associations
15.1: Communities and Niches
16.2: Adverse Effects of Human Activities
16.3: Toxicity
17.16: CC16 Maximum Sustainable Yield
17.19: CC19 Metagenomics
6.4: Biogenous Sediments
marine geology
11.2: Beaches
3.3: Bathymetry
6.5: Hydrogenous Sediments
marine iguana (Amblyrhynchus cristatus)
12.13: Nekton
marine instruments
3.1: Difficulties of Studying the Ocean Environment

marine mammals
12.13: Nekton
13.2: Nutrient Supply to the Coastal Photic Zone
13.5: Fisheries
14.4: Hunting and Defense
14.6: Reproduction
15.3: Kelp Forests
15.6: Polar Regions
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
7.6: Interannual Climate Variations
marine organisms
12.12: Plankton
12.5: Phototropic Primary Production and Light
3.6: Sampling Living Organisms in the Sea
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
marine productivity
15.6: Polar Regions
6.4: Biogenous Sediments
marine species
12.9: Dissolved Oxygen and Carbon Dioxide
14.1: Ecological Requirements
14.6: Reproduction
maritime climate
7.8: Global Climate Zones
maritime history
2.2: Exploration and Mapping
marsh grasses
11.6: Wetlands
Maryland
13.4: Algal Blooms
mass extinction
6.10: The Sediment Historical Record
Massachusetts
10.2: Characteristics of Tides
13.1: Special Characteristics of Coastal Oceans
7.8: Global Climate Zones
Maui
11.2: Beaches
Mauna Loa
1.1: The Ocean's and Earth's Environment
4.7: Hot Spots
maximum sustainable yield
13.5: Fisheries
17.16: CC16 Maximum Sustainable Yield
2.4: Ocean Resources
McKenzie River
8.14: Ocean Circulation and Climate
meanders
7.4: Climatic Winds
8.10: Eddies
9.11: Rossby and Kelvin Waves
mechanical current meters
3.5: Chemical and Physical Oceanography
Medea
3.7: Scuba, Manned, and Unmanned Submersibles
Mediterranean
2.2: Exploration and Mapping
Mediterranean Sea
13.6: Estuaries
14.6: Reproduction
17.8: CC8 Residence Time
4.10: Present-Day Oceans
4.4: Convergent Plate Boundaries
6.5: Hydrogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
8.13: Thermohaline Circulation
Mekong River
6.3: Lithogenous Sediments
Meng Xiang
3.4: Seafloor Sediments
menhaden
12.13: Nekton
mercury
16.1: Pollution versus Contamination
16.3: Toxicity
16.5: Ocean Pollution Issues
Meridional Overturning Circulation
12.9: Dissolved Oxygen and Carbon Dioxide
13.4: Algal Blooms
16.6: Climate Change, Acidification, and Deoxygenation
8.10: Eddies
8.14: Ocean Circulation and Climate
meroplankton
12.12: Plankton
14.6: Reproduction
mesopelagic zone
12.11: Biological Provinces and Zones
mesoscale eddies
13.2: Nutrient Supply to the Coastal Photic Zone
8.10: Eddies
mesosphere
7.1: The Atmosphere
metagenomics
12.4: Studying Microbes in the Ocean
17.19: CC19 Metagenomics
metamorphose
14.6: Reproduction
17.18: CC18 Toxicity
metamorphosis
12.11: Biological Provinces and Zones
meteorites
1.2: Oceans and the Origins of Life
6.10: The Sediment Historical Record
6.6: Cosmogenous Sediments
meteorology
7.4: Climatic Winds
methane
16.6: Climate Change, Acidification, and Deoxygenation
17.9: CC9 The Global Greenhouse Effect
methane hydrates
16.6: Climate Change, Acidification, and Deoxygenation
2.4: Ocean Resources
6.7: Sediment Transport, Deposition, and Accumulation
Mexico
13.4: Algal Blooms
6.10: The Sediment Historical Record
9.9: Tsunamis
Meyer's butterflyfish (Chaetodon meyeri)
14.5: Selected Adaptations in Fishes
microalgae
13.6: Estuaries
14.3: Feeding
15.2: Coral Reefs
16.2: Adverse Effects of Human Activities
microbes
12.10: Organic Carbon
12.12: Plankton
12.3: Microbes in Charge
12.4: Studying Microbes in the Ocean
17.19: CC19 Metagenomics
Micronesia
2.2: Exploration and Mapping
micronutrient
13.2: Nutrient Supply to the Coastal Photic Zone
microplankton
12.12: Plankton
microscopic organisms
12.3: Microbes in Charge
Middle East
4.9: Sea-Level Change and Climate
middle latitudes
13.3: Seasonal Cycles
migration
12.11: Biological Provinces and Zones
12.12: Plankton
12.13: Nekton
12.6: Primary Production and Nutrients
12.9: Dissolved Oxygen and Carbon Dioxide
14.2: Habitats
14.6: Reproduction
14.8: Communication and Navigation
15.6: Polar Regions
15.7: Beyond the Sun's Light
migratory patterns
14.6: Reproduction
mimic surgeonfish (Acanthurus pyroferus)
14.4: Hunting and Defense
mimicry
14.4: Hunting and Defense
Minamata Bay, Japan
16.5: Ocean Pollution Issues
mineralogy
4.1: Structure of the Earth
minerals
6.3: Lithogenous Sediments
mink whale
12.13: Nekton
Minnesota
8.14: Ocean Circulation and Climate
Mississippi
1.1: The Ocean's and Earth's Environment
7.9: Weather Systems
Mississippi Bay
16.5: Ocean Pollution Issues
Mississippi Delta
11.7: Deltas
Mississippi River
11.1: Processes that Form and Modify the Coast
11.7: Deltas
13.4: Algal Blooms
16.5: Ocean Pollution Issues
6.3: Lithogenous Sediments

Mississippi Sound
16.5: Ocean Pollution Issues
Mississippi Valley
8.14: Ocean Circulation and Climate
mixed layer
12.6: Primary Production and Nutrients
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
17.10: CC10 Modeling
17.5: CC5 Transfer and Storage of Heat by Water
8.13: Thermohaline Circulation
mixed tides
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
10.6: Tidal Currents
14.6: Reproduction
mixing
13.1: Special Characteristics of Coastal Oceans
mixing fluids
17.1: CC1 Density and Layering in Fluids
mixotrophs
12.12: Plankton
12.2: Production, Consumption, and Decomposition
MOC
12.9: Dissolved Oxygen and Carbon Dioxide
16.6: Climate Change, Acidification, and Deoxygenation
8.14: Ocean Circulation and Climate
modeling
10.7: Tide Prediction
16.6: Climate Change, Acidification, and Deoxygenation
17.10: CC10 Modeling
17.9: CC9 The Global Greenhouse Effect
3.9: Computers and Modeling
7.6: Interannual Climate Variations
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
molecules
1.2: Oceans and the Origins of Life
5.2: The Water Molecule
mollusks
11.2: Beaches
12.12: Plankton
12.13: Nekton
14.3: Feeding
14.7: Associations
15.2: Coral Reefs
16.6: Climate Change, Acidification, and Deoxygenation
monsoons
13.1: Special Characteristics of Coastal Oceans
7.4: Climatic Winds
7.7: Land-Ocean-Atmosphere Interactions
Monterey Canyon
6.7: Sediment Transport, Deposition, and Accumulation
moon
10.1: Tide-Generating Forces
moon wrasse (Thalassoma lunare)
14.5: Selected Adaptations in Fishes
moray eels
14.5: Selected Adaptations in Fishes
mortality
14.6: Reproduction
motion dynamics
17.12: CC12 The Coriolis Effect
Mouna Loa
4.7: Hot Spots
Mount Everest
4.2: Plate Tectonics
4.7: Hot Spots
Mount Pinatubo
4.4: Convergent Plate Boundaries
6.3: Lithogenous Sediments
Mount St. Helens
4.4: Convergent Plate Boundaries
6.3: Lithogenous Sediments
mountain building
4.6: Transform Faults and Fracture Zones
mountain effect
7.10: Local Weather Effects
mountain formation
4.4: Convergent Plate Boundaries
mucus
14.3: Feeding
14.4: Hunting and Defense
14.7: Associations
mudflats
14.3: Feeding
mussels
13.4: Algal Blooms
13.6: Estuaries
14.3: Feeding
14.6: Reproduction
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
mutalism
14.7: Associations

N

Namibia
12.9: Dissolved Oxygen and Carbon Dioxide
nanoplankton
12.12: Plankton
Nansen bottles
3.5: Chemical and Physical Oceanography
Nansen Fracture Zone
4.10: Present-Day Oceans
narwhale
12.13: Nekton
nautical charts
3.3: Bathymetry
Nautilus (Nautilus pompilius)
12.13: Nekton
15.7: Beyond the Sun's Light
navigation
10.6: Tidal Currents
14.8: Communication and Navigation
2.2: Exploration and Mapping
neap tides
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
14.6: Reproduction
nekton
12.12: Plankton
12.13: Nekton
13.2: Nutrient Supply to the Coastal Photic Zone
13.4: Algal Blooms
14.3: Feeding
nematocytes
12.12: Plankton
14.4: Hunting and Defense
neritic
12.11: Biological Provinces and Zones
nets
3.6: Sampling Living Organisms in the Sea
New England
13.2: Nutrient Supply to the Coastal Photic Zone
2.4: Ocean Resources
New Jersey
11.1: Processes that Form and Modify the Coast
11.4: Beaches and Human Structures
13.2: Nutrient Supply to the Coastal Photic Zone
13.4: Algal Blooms
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
9.10: Internal Waves
New York
10.2: Characteristics of Tides
11.4: Beaches and Human Structures
13.4: Algal Blooms
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
9.10: Internal Waves
New York Harbor
13.6: Estuaries
16.1: Pollution versus Contamination
New Zealand
11.1: Processes that Form and Modify the Coast
11.2: Beaches
13.6: Estuaries
2.2: Exploration and Mapping
2.4: Ocean Resources
niches
15.1: Communities and Niches
15.2: Coral Reefs
15.4: Rocky Intertidal Communities
17.17: CC17 Species Diversity and Biodiversity
17.19: CC19 Metagenomics
Nigeria
11.7: Deltas
Nile Delta
11.7: Deltas
Ninety East Ridge
4.10: Present-Day Oceans
nitrate
12.6: Primary Production and Nutrients
12.9: Dissolved Oxygen and Carbon Dioxide
nitrite
12.6: Primary Production and Nutrients
nitrogen
12.6: Primary Production and Nutrients
13.3: Seasonal Cycles
13.4: Algal Blooms
15.6: Polar Regions
16.5: Ocean Pollution Issues
16.7: Risk Assessment and Management
17.10: CC10 Modeling
17.14: CC14 Phototrophy, Light, and Nutrients
nitrogen oxides
17.9: CC9 The Global Greenhouse Effect
nocturnal
14.4: Hunting and Defense
nodes
10.4: Tides in the Earth's Oceans
9.12: Standing Waves
noise
16.5: Ocean Pollution Issues
5.12: Transmission of Sound
noise pollution
5.12: Transmission of Sound
nonconservative properties
8.15: Tracing Water Masses

nonindigenous species
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
Normway
6.7: Sediment Transport, Deposition, and Accumulation
North America
10.2: Characteristics of Tides
10.5: Tides in the Open Ocean
10.8: Marine Species and the Rhythm of Tides
11.1: Processes that Form and Modify the Coast
13.4: Algal Blooms
14.6: Reproduction
15.3: Kelp Forests
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.4: Convergent Plate Boundaries
6.3: Lithogenous Sediments
6.9: Distribution of Surface Sediments
7.10: Local Weather Effects
7.5: Climate and Ocean Surface Water Properties
7.8: Global Climate Zones
8.5: Open-Ocean Surface Currents
8.9: Coastal Currents
9.12: Standing Waves
North American Plate
4.2: Plate Tectonics
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.6: Transform Faults and Fracture Zones
North Atlantic cod
17.16: CC16 Maximum Sustainable Yield
North Atlantic Deep Water
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
North Atlantic Oscillation
7.6: Interannual Climate Variations
North Carolina
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
7.9: Weather Systems
North Dakota
8.14: Ocean Circulation and Climate
North Equatorial Current
8.18: Box 2 - Equatorial Currents
8.5: Open-Ocean Surface Currents
8.8: Upwelling and Downwelling
North Pacific eels
13.6: Estuaries
North Pacific Gyre
13.6: Estuaries
14.6: Reproduction
North Pole
17.12: CC12 The Coriolis Effect
4.10: Present-Day Oceans
4.5: Divergent Plate Boundaries
6.10: The Sediment Historical Record
North Sea
10.5: Tides in the Open Ocean
14.6: Reproduction
15.6: Polar Regions
3.3: Bathymetry
4.10: Present-Day Oceans
northern kelp crab (Pugettia producta)
15.3: Kelp Forests
Northwest Passage
3.4: Seafloor Sediments
Norway
11.1: Processes that Form and Modify the Coast
13.4: Algal Blooms
13.6: Estuaries
4.10: Present-Day Oceans
Norwegian Sea
15.6: Polar Regions
8.13: Thermohaline Circulation
Nova Scotia
10.2: Characteristics of Tides
10.6: Tidal Currents
10.9: Energy from Tides
6.7: Sediment Transport, Deposition, and Accumulation
nuclear
16.11: Box 3 - Radioactive Wastes in the Marine Environment
nudibranchs
14.3: Feeding
14.6: Reproduction
14.7: Associations
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
nurse shark (Nebrius concolor)
12.13: Nekton
nutrient cycling
6.4: Biogenous Sediments
nutrient limitation
12.6: Primary Production and Nutrients
nutrients
11.5: Coral Reefs and Atolls
11.6: Wetlands
11.7: Deltas
12.10: Organic Carbon
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.6: Estuaries
14.2: Habitats
14.3: Feeding
14.5: Selected Adaptations in Fishes
15.1: Communities and Niches
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.5: Sargasso Sea
15.6: Polar Regions
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.7: Risk Assessment and Management
17.14: CC14 Phototrophy, Light, and Nutrients
17.18: CC18 Toxicity
7.6: Interannual Climate Variations
8.8: Upwelling and Downwelling
9.10: Internal Waves

O

Oahu
11.1: Processes that Form and Modify the Coast
4.7: Hot Spots
observations
10.7: Tide Prediction
ocean chemistry
6.5: Hydrogenous Sediments
ocean circulation
12.9: Dissolved Oxygen and Carbon Dioxide
16.6: Climate Change, Acidification, and Deoxygenation
8.14: Ocean Circulation and Climate
8.15: Tracing Water Masses
ocean climate zones
7.8: Global Climate Zones
ocean currents
10.7: Tide Prediction
11.1: Processes that Form and Modify the Coast
12.12: Plankton
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
15.2: Coral Reefs
17.13: CC13 Geostrophic Flow
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
17.5: CC5 Transfer and Storage of Heat by Water
3.1: Difficulties of Studying the Ocean Environment
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
6.7: Sediment Transport, Deposition, and Accumulation
7.2: Water Vapor in the Atmosphere
7.3: Water and Heat Budgets
7.5: Climate and Ocean Surface Water Properties
8.1: Energy Sources
8.10: Eddies
8.11: Inertial Currents
8.12: Langmuir Circulation
8.13: Thermohaline Circulation
8.17: Box 1- Westward Intensification
8.18: Box 2 - Equatorial Currents
8.2: Wind-Driven Currents
8.3: Ekman Motion
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
8.7: High-Latitude Surface Currents
8.9: Coastal Currents
9.11: Rossby and Kelvin Waves
9.5: Sea Development and Wave Height
ocean depths
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
ocean ecosystems
12.3: Microbes in Charge
12.8: Geographic Variation in Primary Production
13.5: Fisheries
ocean floor topography
4.10: Present-Day Oceans
ocean layers
3.5: Chemical and Physical Oceanography
ocean mapping
3.1: Difficulties of Studying the Ocean Environment
ocean mining
2.3: National and International Law Applied to Oceans
ocean research
2.1: What is Oceanography
ocean research difficulties
3.1: Difficulties of Studying the Ocean Environment
ocean sampling challenges
3.1: Difficulties of Studying the Ocean Environment

ocean stratification
8.3: Ekman Motion
ocean surface
9.10: Internal Waves
ocean surface salinity
7.5: Climate and Ocean Surface Water Properties
oceanic
12.11: Biological Provinces and Zones
oceanic crust
11.5: Coral Reefs and Atolls
12.2: Production, Consumption, and Decomposition
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.6: Transform Faults and Fracture Zones
4.7: Hot Spots
4.8: Plate Interiors
oceanic ridges
10.4: Tides in the Earth's Oceans
11.1: Processes that Form and Modify the Coast
12.2: Production, Consumption, and Decomposition
15.8: Hydrothermal Vents
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.1: Structure of the Earth
4.10: Present-Day Oceans
4.3: Plate Boundaries
4.5: Divergent Plate Boundaries
4.8: Plate Interiors
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
6.5: Hydrogenous Sediments
oceanography
2.1: What is Oceanography
3.2: The Unseen Domain
3.3: Bathymetry
3.5: Chemical and Physical Oceanography
3.7: Scuba, Manned, and Unmanned Submersibles
3.8: Satellites
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
8.10: Eddies
8.4: Geostrophic Currents
octopuses
14.4: Hunting and Defense
14.6: Reproduction
15.7: Beyond the Sun's Light
oil spills
16.5: Ocean Pollution Issues
16.7: Risk Assessment and Management
16.9: Box 1 - Petroleum in the Marine Environment
Old Glory goby (Amblygobius rainfordi)
14.5: Selected Adaptations in Fishes
omnivores
12.12: Plankton
12.2: Production, Consumption, and Decomposition
14.6: Reproduction
Ontario
8.14: Ocean Circulation and Climate
ooze
6.7: Sediment Transport, Deposition, and Accumulation
operculum
14.4: Hunting and Defense
orbital motion
10.1: Tide-Generating Forces
orbital velocity
11.2: Beaches
Oregon
12.9: Dissolved Oxygen and Carbon Dioxide
13.4: Algal Blooms
7.6: Interannual Climate Variations
7.8: Global Climate Zones
9.9: Tsunamis
organic carbon
12.10: Organic Carbon
organic compounds
12.2: Production, Consumption, and Decomposition
organic matter
12.10: Organic Carbon
organisms
1.2: Oceans and the Origins of Life
12.1: How do we describe life?
12.11: Biological Provinces and Zones
12.6: Primary Production and Nutrients
14.3: Feeding
Origin of Species
2.2: Exploration and Mapping
Orinoco River
13.2: Nutrient Supply to the Coastal Photic Zone
7.5: Climate and Ocean Surface Water Properties
orthogonals
9.8: Waves in Shallow Water
osmoregulation
14.5: Selected Adaptations in Fishes
osmosis
14.5: Selected Adaptations in Fishes
osmotic pressure
14.5: Selected Adaptations in Fishes
overfished
12.13: Nekton
7.6: Interannual Climate Variations
overfishing
13.2: Nutrient Supply to the Coastal Photic Zone
13.5: Fisheries
16.5: Ocean Pollution Issues
17.16: CC16 Maximum Sustainable Yield
2.4: Ocean Resources
oviparous
14.6: Reproduction
ovoviviparous
14.6: Reproduction
oxidation
12.2: Production, Consumption, and Decomposition
6.10: The Sediment Historical Record
oxygen
12.9: Dissolved Oxygen and Carbon Dioxide
13.4: Algal Blooms
14.2: Habitats
14.5: Selected Adaptations in Fishes
16.2: Adverse Effects of Human Activities
16.9: Box 1 - Petroleum in the Marine Environment
17.14: CC14 Phototrophy, Light, and Nutrients
5.2: The Water Molecule
Oxygen Minimum Zones
1.1: The Ocean's and Earth's Environment
oysters
12.12: Plankton
13.4: Algal Blooms
13.6: Estuaries
14.6: Reproduction
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
ozone depletion
7.1: The Atmosphere
ozone layer
12.5: Phototropic Primary Production and Light
17.9: CC9 The Global Greenhouse Effect
7.1: The Atmosphere

P

Pacific Decadal Oscillation
7.6: Interannual Climate Variations
Pacific Northwest
1.1: The Ocean's and Earth's Environment
Pacific Ocean
1.1: The Ocean's and Earth's Environment
10.2: Characteristics of Tides
10.5: Tides in the Open Ocean
11.2: Beaches
12.11: Biological Provinces and Zones
12.12: Plankton
12.13: Nekton
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13.3: Seasonal Cycles
13.4: Algal Blooms
13.6: Estuaries
14.3: Feeding
14.6: Reproduction
15.2: Coral Reefs
15.8: Hydrothermal Vents
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.3: Toxicity
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
2.2: Exploration and Mapping
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.8: Plate Interiors
6.3: Lithogenous Sediments
6.4: Biogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.9: Distribution of Surface Sediments
7.10: Local Weather Effects
7.4: Climatic Winds
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
7.8: Global Climate Zones
7.9: Weather Systems
8.13: Thermohaline Circulation
8.14: Ocean Circulation and Climate
8.5: Open-Ocean Surface Currents
8.8: Upwelling and Downwelling
8.9: Coastal Currents
9.11: Rossby and Kelvin Waves
9.5: Sea Development and Wave Height
9.9: Tsunamis

Pacific Plate
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
4.6: Transform Faults and Fracture Zones
4.7: Hot Spots

Pacific salmon
14.6: Reproduction

Pacific sardines (Sardinops sagax)
12.13: Nekton

Pacific spiny oyster (Spondylus varians)
14.3: Feeding

Palau
11.1: Processes that Form and Modify the Coast
12.12: Plankton
14.5: Selected Adaptations in Fishes
15.2: Coral Reefs

paleomagnetic
4.7: Hot Spots

palm trees
15.2: Coral Reefs

Pamlico Sound
13.6: Estuaries

Pangaea
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.9: Sea-Level Change and Climate

Pangea
4.4: Convergent Plate Boundaries

Papua New Guinea
12.12: Plankton
12.13: Nekton
13.1: Special Characteristics of Coastal Oceans
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
14.7: Associations
15.2: Coral Reefs

parasites
13.5: Fisheries
14.4: Hunting and Defense
14.7: Associations
16.10: Box 2 - Sewage Wastes in the Marine Environment

partially mixed estuaries
13.6: Estuaries

passive margins
11.1: Processes that Form and Modify the Coast
13.6: Estuaries
4.8: Plate Interiors

Patagonia
11.1: Processes that Form and Modify the Coast
13.6: Estuaries
6.3: Lithogenous Sediments

PCBs
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities

pealgic
12.8: Geographic Variation in Primary Production

pearlscale angelfish (Centropyge vrolikii)
14.4: Hunting and Defense

pebble beach
11.2: Beaches

pelagic
14.2: Habitats
14.6: Reproduction
15.6: Polar Regions
15.7: Beyond the Sun's Light
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues

pelagic environment
12.11: Biological Provinces and Zones
12.14: Benthos

pelagic fishes
12.13: Nekton

pelagic habitat
14.2: Habitats

pelagic sediments
6.9: Distribution of Surface Sediments

pelagic species
3.6: Sampling Living Organisms in the Sea

pelicans
13.4: Algal Blooms
16.3: Toxicity
16.5: Ocean Pollution Issues

penguins
12.13: Nekton
12.7: Food Webs
15.6: Polar Regions
7.6: Interannual Climate Variations

Pennsylvania
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues

periwinkles
15.4: Rocky Intertidal Communities

Persian Gulf
16.9: Box 1 - Petroleum in the Marine Environment

Peru
13.2: Nutrient Supply to the Coastal Photic Zone
2.3: National and International Law Applied to Oceans
7.6: Interannual Climate Variations
8.8: Upwelling and Downwelling

Peru Current
8.5: Open-Ocean Surface Currents

petroleum
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
2.4: Ocean Resources

pH
15.4: Rocky Intertidal Communities
16.6: Climate Change, Acidification, and Deoxygenation
3.5: Chemical and Physical Oceanography
5.6: Dissolved Chemicals in Seawater
6.4: Biogenous Sediments
6.5: Hydrogenous Sediments

phase changes
5.7: Heat Properties of Water

Philadephia
16.10: Box 2 - Sewage Wastes in the Marine Environment

Philippines
12.13: Nekton
13.1: Special Characteristics of Coastal Oceans
2.4: Ocean Resources
4.10: Present-Day Oceans
6.3: Lithogenous Sediments
7.9: Weather Systems

Phillippines
2.3: National and International Law Applied to Oceans

phosphate
13.3: Seasonal Cycles

phosphorite nodules
2.4: Ocean Resources
6.5: Hydrogenous Sediments

phosphorus
12.6: Primary Production and Nutrients
13.3: Seasonal Cycles
15.6: Polar Regions
16.5: Ocean Pollution Issues
17.14: CC14 Phototrophy, Light, and Nutrients

photic zone
12.10: Organic Carbon
12.11: Biological Provinces and Zones
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
14.2: Habitats
14.3: Feeding
14.4: Hunting and Defense
15.2: Coral Reefs
15.6: Polar Regions
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
5.11: Transmission of Light and Other Electromagnetic Radiation

photoautotrophs
12.2: Production, Consumption, and Decomposition

photographs
17.14: CC14 Phototrophy, Light, and Nutrients

photoheterotrophs
12.2: Production, Consumption, and Decomposition

photosynthesis
11.1: Processes that Form and Modify the Coast
11.5: Coral Reefs and Atolls
12.10: Organic Carbon
12.11: Biological Provinces and Zones
12.12: Plankton
12.2: Production, Consumption, and Decomposition
12.5: Phototropic Primary Production and Light
12.9: Dissolved Oxygen and Carbon Dioxide
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
14.2: Habitats
14.7: Associations
15.2: Coral Reefs
15.3: Kelp Forests
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
17.14: CC14 Phototrophy, Light, and Nutrients
17.15: CC15 Food Chain Efficiency
17.18: CC18 Toxicity
17.9: CC9 The Global Greenhouse Effect
6.4: Biogenous Sediments

phototrophic
1.2: Oceans and the Origins of Life
12.5: Phototropic Primary Production and Light
8.8: Upwelling and Downwelling

phototrophs
12.2: Production, Consumption, and Decomposition
14.2: Habitats

phototrophy
12.11: Biological Provinces and Zones
12.2: Production, Consumption, and Decomposition
12.6: Primary Production and Nutrients
14.3: Feeding
5.11: Transmission of Light and Other Electromagnetic Radiation
phytoplankton
12.10: Organic Carbon
12.12: Plankton
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.4: Algal Blooms
13.5: Fisheries
13.6: Estuaries
14.2: Habitats
14.3: Feeding
14.6: Reproduction
15.2: Coral Reefs
15.5: Sargasso Sea
15.6: Polar Regions
16.2: Adverse Effects of Human Activities
17.15: CC15 Food Chain Efficiency
17.16: CC16 Maximum Sustainable Yield
17.18: CC18 Toxicity
3.6: Sampling Living Organisms in the Sea
5.11: Transmission of Light and Other Electromagnetic Radiation
6.4: Biogenous Sediments
7.6: Interannual Climate Variations
8.8: Upwelling and Downwelling
8.9: Coastal Currents
9.10: Internal Waves
picoplankton
12.12: Plankton
piers
16.2: Adverse Effects of Human Activities
pilchards
12.13: Nekton
pilot whale
12.13: Nekton
pink anemonefish (Amphiprion perideraion)
14.3: Feeding
14.4: Hunting and Defense
pinnipeds
12.13: Nekton
pipefishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
pistol shrimp
14.8: Communication and Navigation
piston corers
3.4: Seafloor Sediments
planetary vorticity
8.17: Box 1- Westward Intensification
planetary waves
9.11: Rossby and Kelvin Waves
plankton
11.5: Coral Reefs and Atolls
12.10: Organic Carbon
12.12: Plankton
12.13: Nekton
12.4: Studying Microbes in the Ocean
12.5: Phototropic Primary Production and Light
13.2: Nutrient Supply to the Coastal Photic Zone
13.5: Fisheries
15.4: Rocky Intertidal Communities
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
3.6: Sampling Living Organisms in the Sea
plants
12.1: How do we describe life?
plastics
16.5: Ocean Pollution Issues
plate boundaries
4.3: Plate Boundaries
4.6: Transform Faults and Fracture Zones
plate tectonics
16.1: Pollution versus Contamination
17.10: CC10 Modeling
17.11: CC11 Chaos
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.3: CC3 Convection and Convection Cells
4.1: Structure of the Earth
4.10: Present-Day Oceans
4.2: Plate Tectonics
4.6: Transform Faults and Fracture Zones
4.8: Plate Interiors
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
plume worms
15.8: Hydrothermal Vents
poisons
14.4: Hunting and Defense
polar bears
15.6: Polar Regions
polar easterlies
8.7: High-Latitude Surface Currents
polar ecology
15.6: Polar Regions
polar front
15.6: Polar Regions
7.4: Climatic Winds
7.9: Weather Systems
polar molecules
5.3: The Dissolving Power of Water
polar ocean zone
7.8: Global Climate Zones
polar regions
13.3: Seasonal Cycles
15.6: Polar Regions
16.6: Climate Change, Acidification, and Deoxygenation
17.5: CC5 Transfer and Storage of Heat by Water
pollution
1.1: The Ocean's and Earth's Environment
12.13: Nekton
12.9: Dissolved Oxygen and Carbon Dioxide
13.5: Fisheries
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.5: Ocean Pollution Issues
16.7: Risk Assessment and Management
16.9: Box 1 - Petroleum in the Marine Environment
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.8: CC8 Residence Time
6.1: Sediments and Biogeochemical Cycles
Polynesia
2.2: Exploration and Mapping
polyps
13.3: Seasonal Cycles
14.3: Feeding
14.4: Hunting and Defense
16.6: Climate Change, Acidification, and Deoxygenation
populations
15.2: Coral Reefs
pore waters
14.3: Feeding
porpoise
12.13: Nekton
Portland
7.8: Global Climate Zones
potassium
5.6: Dissolved Chemicals in Seawater
potential energy
8.4: Geostrophic Currents
9.3: What is Wave Motion?
9.8: Waves in Shallow Water
power plants
16.2: Adverse Effects of Human Activities
practical salinity units
5.5: Salinity
precipitation
17.3: CC3 Convection and Convection Cells
7.10: Local Weather Effects
7.3: Water and Heat Budgets
7.4: Climatic Winds
7.5: Climate and Ocean Surface Water Properties
7.8: Global Climate Zones
7.9: Weather Systems
predators
13.5: Fisheries
14.1: Ecological Requirements
14.2: Habitats
14.4: Hunting and Defense
14.6: Reproduction
14.7: Associations
14.8: Communication and Navigation
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
15.8: Hydrothermal Vents
predatory behavior
14.4: Hunting and Defense
predictions
10.7: Tide Prediction

pressure
15.7: Beyond the Sun's Light
17.6: CC6 Salinity, Temperature, Pressure, and Water Density
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
7.1: The Atmosphere
pressure gradient
10.1: Tide-Generating Forces
17.12: CC12 The Coriolis Effect
17.13: CC13 Geostrophic Flow
7.4: Climatic Winds
8.4: Geostrophic Currents
9.3: What is Wave Motion?
prey
14.4: Hunting and Defense
14.7: Associations
14.8: Communication and Navigation
17.18: CC18 Toxicity
primary production
12.2: Production, Consumption, and Decomposition
12.3: Microbes in Charge
12.5: Phototropic Primary Production and Light
12.6: Primary Production and Nutrients
12.7: Food Webs
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.5: Fisheries
14.2: Habitats
15.2: Coral Reefs
15.3: Kelp Forests
15.5: Sargasso Sea
15.8: Hydrothermal Vents
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
17.14: CC14 Phototrophy, Light, and Nutrients
17.15: CC15 Food Chain Efficiency
6.4: Biogenous Sediments
Prince William Sound
11.1: Processes that Form and Modify the Coast
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
Prochlorococcus
12.12: Plankton
producers
12.3: Microbes in Charge
progressive waves
10.2: Characteristics of Tides
10.6: Tidal Currents
9.12: Standing Waves
9.2: Progressive Waves
9.3: What is Wave Motion?
prokaryotes
12.1: How do we describe life?
17.14: CC14 Phototrophy, Light, and Nutrients
protists
12.1: How do we describe life?
14.2: Habitats
16.6: Climate Change, Acidification, and Deoxygenation
protozoa
12.7: Food Webs
PSU
5.5: Salinity
pteropods
12.12: Plankton
14.3: Feeding
6.4: Biogenous Sediments
public health
16.10: Box 2 - Sewage Wastes in the Marine Environment
public perception
16.7: Risk Assessment and Management
Puerto Rico
15.2: Coral Reefs
2.3: National and International Law Applied to Oceans
7.9: Weather Systems
pufferfishes
14.4: Hunting and Defense
Puget Sound
10.8: Marine Species and the Rhythm of Tides
13.6: Estuaries
pycnocline
12.9: Dissolved Oxygen and Carbon Dioxide
13.1: Special Characteristics of Coastal Oceans
16.6: Climate Change, Acidification, and Deoxygenation
17.1: CC1 Density and Layering in Fluids
7.6: Interannual Climate Variations
8.13: Thermohaline Circulation
8.3: Ekman Motion
9.10: Internal Waves
pycnocline zone
8.13: Thermohaline Circulation

Q

Qiantang River
10.6: Tidal Currents

R

radiation
17.3: CC3 Convection and Convection Cells
5.12: Transmission of Sound
7.3: Water and Heat Budgets
radioactive
16.11: Box 3 - Radioactive Wastes in the Marine Environment
17.3: CC3 Convection and Convection Cells
radioactive dating
6.10: The Sediment Historical Record
radioactive decay
17.7: CC7 Radioactivity and Age Dating
4.2: Plate Tectonics
radioactive isotope
8.15: Tracing Water Masses
radioactivity
16.3: Toxicity
radioisotopes
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
17.7: CC7 Radioactivity and Age Dating
5.6: Dissolved Chemicals in Seawater
radiolaria
12.12: Plankton
6.4: Biogenous Sediments
radiolarian oozes
6.9: Distribution of Surface Sediments
radionuclides
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
5.6: Dissolved Chemicals in Seawater
8.15: Tracing Water Masses
raggy scorpionfish (Scorpaenopsis venosa)
14.4: Hunting and Defense
rainbow runner (Elagatis bipinnulata)
14.4: Hunting and Defense
rainfall
13.1: Special Characteristics of Coastal Oceans
Raritan Bay
13.6: Estuaries
rays
12.13: Nekton
14.6: Reproduction
records
10.7: Tide Prediction
recycling
16.3: Toxicity
Red Sea
12.13: Nekton
13.6: Estuaries
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.7: Associations
17.8: CC8 Residence Time
4.10: Present-Day Oceans
4.5: Divergent Plate Boundaries
6.9: Distribution of Surface Sediments
7.5: Climate and Ocean Surface Water Properties
red tides
13.4: Algal Blooms
redbreasted Maori wrasse (Cheilinus fasciatus)
14.5: Selected Adaptations in Fishes
reef terrace
15.2: Coral Reefs
reef whitetip shark (Triaenodon obesus)
12.13: Nekton
reflection
5.11: Transmission of Light and Other Electromagnetic Radiation
refracted
9.8: Waves in Shallow Water
refraction
14.8: Communication and Navigation
5.11: Transmission of Light and Other Electromagnetic Radiation
5.12: Transmission of Sound
regime shift
7.6: Interannual Climate Variations
relative vorticity
8.17: Box 1- Westward Intensification
relict sediments
11.1: Processes that Form and Modify the Coast
6.8: Continental Margin Sediments
remora (Echeneis naucrates)
14.5: Selected Adaptations in Fishes
14.7: Associations
remoras
12.13: Nekton
14.5: Selected Adaptations in Fishes
14.7: Associations
remotely operated vehicles
3.7: Scuba, Manned, and Unmanned Submersibles
renewable energy
10.9: Energy from Tides

reproduction
- 10.8: Marine Species and the Rhythm of Tides
- 13.6: Estuaries
- 14.1: Ecological Requirements
- 14.2: Habitats
- 14.6: Reproduction
- 17.15: CC15 Food Chain Efficiency
- 17.18: CC18 Toxicity

reproductive cycle
- 10.8: Marine Species and the Rhythm of Tides

reproductive strategies
- 14.6: Reproduction

reptiles
- 12.13: Nekton

research
- 3.5: Chemical and Physical Oceanography
- 3.8: Satellites

residence time
- 11.5: Coral Reefs and Atolls
- 12.9: Dissolved Oxygen and Carbon Dioxide
- 13.1: Special Characteristics of Coastal Oceans
- 13.4: Algal Blooms
- 15.6: Polar Regions
- 16.1: Pollution versus Contamination
- 16.6: Climate Change, Acidification, and Deoxygenation
- 17.8: CC8 Residence Time
- 5.4: Sources and Sinks of Chemicals Dissolved in Seawater
- 6.3: Lithogenous Sediments
- 7.1: The Atmosphere

residual currents
- 13.6: Estuaries

respiration
- 12.2: Production, Consumption, and Decomposition
- 12.9: Dissolved Oxygen and Carbon Dioxide
- 13.4: Algal Blooms
- 14.3: Feeding
- 15.8: Hydrothermal Vents
- 16.2: Adverse Effects of Human Activities
- 16.6: Climate Change, Acidification, and Deoxygenation
- 17.14: CC14 Phototrophy, Light, and Nutrients
- 17.15: CC15 Food Chain Efficiency

restoring forces
- 8.2: Wind-Driven Currents
- 9.3: What is Wave Motion?
- 9.4: Making Waves

rhinoceros sea star (Protoreaster nodosus)
- 15.2: Coral Reefs

rhodopsins
- 12.2: Production, Consumption, and Decomposition

ridge crest
- 7.10: Local Weather Effects

ridges
- 13.2: Nutrient Supply to the Coastal Photic Zone

rift zones
- 4.3: Plate Boundaries
- 4.5: Divergent Plate Boundaries

right whale
- 12.13: Nekton

Rio Grande Rise
- 4.7: Hot Spots

rip currents
- 9.8: Waves in Shallow Water

risk assessment
- 16.7: Risk Assessment and Management

risk management
- 16.7: Risk Assessment and Management

river discharge
- 13.1: Special Characteristics of Coastal Oceans

river dynamics
- 11.7: Deltas

river valley
- 11.7: Deltas

rivers
- 11.1: Processes that Form and Modify the Coast
- 11.3: Barrier Islands and Lagoons
- 11.7: Deltas
- 13.1: Special Characteristics of Coastal Oceans
- 13.2: Nutrient Supply to the Coastal Photic Zone
- 13.3: Seasonal Cycles
- 13.6: Estuaries
- 16.10: Box 2 - Sewage Wastes in the Marine Environment
- 16.2: Adverse Effects of Human Activities
- 16.5: Ocean Pollution Issues
- 17.8: CC8 Residence Time
- 6.3: Lithogenous Sediments
- 9.9: Tsunamis

RMS Titanic
- 3.3: Bathymetry
- 3.7: Scuba, Manned, and Unmanned Submersibles

rockfishes
- 15.3: Kelp Forests

rocky intertidal zones
- 12.11: Biological Provinces and Zones
- 15.4: Rocky Intertidal Communities

Rocky Mountains
- 4.9: Sea-Level Change and Climate
- 7.5: Climate and Ocean Surface Water Properties

rocky shore habitats
- 15.4: Rocky Intertidal Communities

rose anemone (Tealia lineata)
- 15.3: Kelp Forests

Rossby waves
- 9.11: Rossby and Kelvin Waves

rotation
- 10.1: Tide-Generating Forces
- 17.12: CC12 The Coriolis Effect

rotational systems
- 17.12: CC12 The Coriolis Effect

ROVs
- 15.7: Beyond the Sun's Light
- 3.7: Scuba, Manned, and Unmanned Submersibles

runoff
- 11.5: Coral Reefs and Atolls
- 12.6: Primary Production and Nutrients
- 13.1: Special Characteristics of Coastal Oceans
- 13.2: Nutrient Supply to the Coastal Photic Zone
- 13.3: Seasonal Cycles
- 13.4: Algal Blooms
- 13.6: Estuaries
- 15.6: Polar Regions
- 16.2: Adverse Effects of Human Activities
- 16.5: Ocean Pollution Issues
- 17.8: CC8 Residence Time
- 4.8: Plate Interiors
- 5.4: Sources and Sinks of Chemicals Dissolved in Seawater
- 5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater

Russia
- 14.6: Reproduction
- 2.3: National and International Law Applied to Oceans
- 4.4: Convergent Plate Boundaries

S

saccopharynx
- 15.7: Beyond the Sun's Light

Sacramento River
- 13.6: Estuaries

Sahara Desert
- 6.3: Lithogenous Sediments

Saint Lawrence River
- 8.14: Ocean Circulation and Climate

salinity
- 11.5: Coral Reefs and Atolls
- 12.11: Biological Provinces and Zones
- 13.1: Special Characteristics of Coastal Oceans
- 13.3: Seasonal Cycles
- 13.6: Estuaries
- 14.2: Habitats
- 14.5: Selected Adaptations in Fishes
- 15.1: Communities and Niches
- 15.4: Rocky Intertidal Communities
- 16.2: Adverse Effects of Human Activities
- 16.5: Ocean Pollution Issues
- 16.6: Climate Change, Acidification, and Deoxygenation
- 17.1: CC1 Density and Layering in Fluids
- 17.17: CC17 Species Diversity and Biodiversity
- 17.18: CC18 Toxicity
- 17.3: CC3 Convection and Convection Cells
- 17.6: CC6 Salinity, Temperature, Pressure, and Water Density
- 3.5: Chemical and Physical Oceanography
- 5.5: Salinity
- 5.6: Dissolved Chemicals in Seawater
- 5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
- 5.9: Ice Formation
- 7.5: Climate and Ocean Surface Water Properties
- 8.1: Energy Sources
- 8.13: Thermohaline Circulation
- 8.15: Tracing Water Masses

salmon
- 13.6: Estuaries
- 14.6: Reproduction
- 16.2: Adverse Effects of Human Activities

salps
- 12.12: Plankton
- 14.3: Feeding
- 14.4: Hunting and Defense

salt marshes
- 11.1: Processes that Form and Modify the Coast
- 13.6: Estuaries
- 16.9: Box 1 - Petroleum in the Marine Environment
- 4.8: Plate Interiors

salt wedge estuaries
- 13.6: Estuaries

sampling
- 3.4: Seafloor Sediments
- 3.5: Chemical and Physical Oceanography
- 3.6: Sampling Living Organisms in the Sea
- 3.7: Scuba, Manned, and Unmanned Submersibles

San Andreas Fault
- 4.6: Transform Faults and Fracture Zones

San Diego Bay
- 13.6: Estuaries

San Francisco Bay
13.6: Estuaries
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.5: Ocean Pollution Issues
17.8: CC8 Residence Time
6.3: Lithogenous Sediments
San Francisco Delta
11.7: Deltas
16.2: Adverse Effects of Human Activities
San Joaquin River
13.6: Estuaries
sand
6.2: Classification of Sediments
sand anemones
14.3: Feeding
14.4: Hunting and Defense
sand dollars
15.2: Coral Reefs
sand dunes
11.1: Processes that Form and Modify the Coast
sand grains
11.2: Beaches
sandbars
13.6: Estuaries
sandperches
14.5: Selected Adaptations in Fishes
sandspit
11.4: Beaches and Human Structures
sandy beach
11.2: Beaches
Santa Cruz Mountains
4.6: Transform Faults and Fracture Zones
sardines
12.13: Nekton
12.7: Food Webs
13.5: Fisheries
Sargasso Sea
12.12: Plankton
12.8: Geographic Variation in Primary Production
13.6: Estuaries
14.6: Reproduction
15.5: Sargasso Sea
8.10: Eddies
Sargassum
12.12: Plankton
12.8: Geographic Variation in Primary Production
15.5: Sargasso Sea
sargassumfishes
15.5: Sargasso Sea
Saronikis
13.4: Algal Blooms
Saskatchewan
8.14: Ocean Circulation and Climate
satellites
10.5: Tides in the Open Ocean
3.3: Bathymetry
3.8: Satellites
8.10: Eddies
8.4: Geostrophic Currents
9.11: Rossby and Kelvin Waves
saturation pressure
7.2: Water Vapor in the Atmosphere
saturation solubility
12.9: Dissolved Oxygen and Carbon Dioxide
saw blade shrimp (Tozeuma armatum)
14.7: Associations
scale worms
14.7: Associations
15.8: Hydrothermal Vents
scalloped hammerhead shark (Sphyrna lewini)
12.13: Nekton
scallops
14.4: Hunting and Defense
Scandinavia
4.9: Sea-Level Change and Climate
7.8: Global Climate Zones
scarp
11.2: Beaches
scattering
12.5: Phototropic Primary Production and Light
14.8: Communication and Navigation
5.11: Transmission of Light and Other Electromagnetic Radiation
scavengers
15.4: Rocky Intertidal Communities
15.7: Beyond the Sun's Light
scavenging
14.3: Feeding
schooling
12.12: Plankton
14.4: Hunting and Defense
14.6: Reproduction
scorpionfishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.7: Associations
Scotland
8.13: Thermohaline Circulation
scuba
14.5: Selected Adaptations in Fishes
14.7: Associations
14.8: Communication and Navigation
15.2: Coral Reefs
3.6: Sampling Living Organisms in the Sea
3.7: Scuba, Manned, and Unmanned Submersibles
sea breeze
7.10: Local Weather Effects
7.7: Land-Ocean-Atmosphere Interactions
sea conditions
9.5: Sea Development and Wave Height
sea cucumbers
12.12: Plankton
14.3: Feeding
14.4: Hunting and Defense
14.6: Reproduction
14.7: Associations
15.2: Coral Reefs
15.4: Rocky Intertidal Communities
15.7: Beyond the Sun's Light
17.16: CC16 Maximum Sustainable Yield
sea fans
14.4: Hunting and Defense
14.7: Associations
16.6: Climate Change, Acidification, and Deoxygenation
sea gooseberries
12.12: Plankton
sea grass
12.13: Nekton
14.3: Feeding
16.5: Ocean Pollution Issues
sea grasses
11.1: Processes that Form and Modify the Coast
12.12: Plankton
sea gulls
12.13: Nekton
sea ice
16.6: Climate Change, Acidification, and Deoxygenation
sea level
10.5: Tides in the Open Ocean
11.3: Barrier Islands and Lagoons
11.5: Coral Reefs and Atolls
13.6: Estuaries
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.9: Sea-Level Change and Climate
6.8: Continental Margin Sediments
sea lions
12.13: Nekton
13.4: Algal Blooms
15.3: Kelp Forests
16.3: Toxicity
16.5: Ocean Pollution Issues
Sea of Japan
13.4: Algal Blooms
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
sea otters
13.4: Algal Blooms
sea otters (Enhydra lutris)
12.13: Nekton
15.3: Kelp Forests
sea pens
14.3: Feeding
14.4: Hunting and Defense
15.2: Coral Reefs
15.3: Kelp Forests
sea slugs
14.3: Feeding
sea snakes
12.13: Nekton
sea star commensal shrimp (Zenopontonia noverca)
14.7: Associations
sea star shrimp (Periclimenes soror)
14.7: Associations
sea star shrimp (Zenopontonia noverca)
14.4: Hunting and Defense
sea stars
12.12: Plankton
14.3: Feeding
14.4: Hunting and Defense
14.6: Reproduction
14.7: Associations
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
sea surface height
3.8: Satellites
sea surface slope
8.4: Geostrophic Currents
sea turtles
10.8: Marine Species and the Rhythm of Tides
12.13: Nekton
14.6: Reproduction
14.8: Communication and Navigation
16.5: Ocean Pollution Issues

sea urchins
12.12: Plankton
14.3: Feeding
14.4: Hunting and Defense
14.7: Associations
15.2: Coral Reefs
15.3: Kelp Forests
15.4: Rocky Intertidal Communities
16.6: Climate Change, Acidification, and Deoxygenation
17.16: CC16 Maximum Sustainable Yield
sea walnuts
12.12: Plankton
seabirds
11.6: Wetlands
12.13: Nekton
12.7: Food Webs
13.4: Algal Blooms
13.5: Fisheries
15.3: Kelp Forests
15.6: Polar Regions
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
7.6: Interannual Climate Variations
seaborne trade
2.2: Exploration and Mapping
seafloor
12.14: Benthos
3.4: Seafloor Sediments
seafloor exploration
3.7: Scuba, Manned, and Unmanned Submersibles
seafloor massive sulfides
15.8: Hydrothermal Vents
seafloor sediments
4.8: Plate Interiors
6.1: Sediments and Biogeochemical Cycles
seafloor slumps
9.4: Making Waves
seafloor spreading
4.5: Divergent Plate Boundaries
seafloor topography
3.3: Bathymetry
4.9: Sea-Level Change and Climate
9.10: Internal Waves
9.8: Waves in Shallow Water
seagrass filefish (Acreichthys tomentosus)
14.3: Feeding
seahorses
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.6: Reproduction
seals
12.13: Nekton
12.7: Food Webs
15.6: Polar Regions
seamounts
12.11: Biological Provinces and Zones
13.2: Nutrient Supply to the Coastal Photic Zone
4.7: Hot Spots
seasonal thermoclines
8.13: Thermohaline Circulation
seasons
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
seawalls
11.4: Beaches and Human Structures
seawater chemistry
3.5: Chemical and Physical Oceanography
seaweeds
12.12: Plankton
secondary production
12.2: Production, Consumption, and Decomposition
sediment accumulation
11.6: Wetlands
sediment accumulation rate
6.7: Sediment Transport, Deposition, and Accumulation
sediment age dating
6.10: The Sediment Historical Record
sediment analysis
3.4: Seafloor Sediments
sediment classification
6.7: Sediment Transport, Deposition, and Accumulation
sediment deposition
11.7: Deltas
17.2: CC2 Isostasy, Eustasy, and Sea Level
6.5: Hydrogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
sediment distribution
6.2: Classification of Sediments
6.8: Continental Margin Sediments
sediment dynamics
6.7: Sediment Transport, Deposition, and Accumulation
sediment layers
3.4: Seafloor Sediments
6.10: The Sediment Historical Record
sediment transport
11.1: Processes that Form and Modify the Coast
11.2: Beaches
13.6: Estuaries
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
sediment transportation
6.7: Sediment Transport, Deposition, and Accumulation
sedimentary arc
11.1: Processes that Form and Modify the Coast
4.4: Convergent Plate Boundaries
sedimentation
13.1: Special Characteristics of Coastal Oceans
17.8: CC8 Residence Time
4.10: Present-Day Oceans
4.9: Sea-Level Change and Climate
6.10: The Sediment Historical Record
6.3: Lithogenous Sediments
6.9: Distribution of Surface Sediments
sedimentation rate
14.3: Feeding
15.2: Coral Reefs
16.2: Adverse Effects of Human Activities
sediments
11.1: Processes that Form and Modify the Coast
13.6: Estuaries
14.2: Habitats
15.2: Coral Reefs
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.5: Ocean Pollution Issues
17.18: CC18 Toxicity
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.8: CC8 Residence Time
3.4: Seafloor Sediments
6.1: Sediments and Biogeochemical Cycles
6.10: The Sediment Historical Record
6.2: Classification of Sediments
6.3: Lithogenous Sediments
6.5: Hydrogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
6.8: Continental Margin Sediments
7.9: Weather Systems
8.14: Ocean Circulation and Climate
9.10: Internal Waves
sei whales
12.12: Plankton
12.13: Nekton
seiches
9.12: Standing Waves
seismic activity
4.4: Convergent Plate Boundaries
seismic tomography
4.1: Structure of the Earth
semidiurnal tides
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
senses
14.8: Communication and Navigation
sensible heat
5.7: Heat Properties of Water
7.3: Water and Heat Budgets
sequencing
12.4: Studying Microbes in the Ocean
sergeant fishes
14.5: Selected Adaptations in Fishes
setae
14.3: Feeding
settling rate
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
sewage
11.5: Coral Reefs and Atolls
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.4: Algal Blooms
13.6: Estuaries
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.4: Waste Disposal
16.5: Ocean Pollution Issues
17.18: CC18 Toxicity
sexual reproduction
14.6: Reproduction
shadow zone
5.12: Transmission of Sound

sharks
- 12.13: Nekton
- 14.4: Hunting and Defense
- 14.5: Selected Adaptations in Fishes
- 14.6: Reproduction
- 14.7: Associations
- 14.8: Communication and Navigation
- 15.3: Kelp Forests
- 15.7: Beyond the Sun's Light

sharksuckers
- 12.13: Nekton
- 14.5: Selected Adaptations in Fishes

shear stress
- 9.4: Making Waves

shelf valleys
- 4.9: Sea-Level Change and Climate

shellfish
- 11.6: Wetlands
- 12.13: Nekton
- 13.3: Seasonal Cycles
- 16.2: Adverse Effects of Human Activities

shellfish poisoning
- 13.4: Algal Blooms
- 14.4: Hunting and Defense

Sheridan
- 7.8: Global Climate Zones

ships
- 2.2: Exploration and Mapping

shoaling
- 11.4: Beaches and Human Structures

shoals
- 13.2: Nutrient Supply to the Coastal Photic Zone

shoreline
- 13.1: Special Characteristics of Coastal Oceans
- 16.2: Adverse Effects of Human Activities
- 9.4: Making Waves

shrimp
- 14.3: Feeding
- 14.4: Hunting and Defense
- 14.7: Associations
- 15.3: Kelp Forests
- 15.4: Rocky Intertidal Communities
- 15.7: Beyond the Sun's Light
- 15.8: Hydrothermal Vents

Siberia
- 4.10: Present-Day Oceans

Siera Nevada
- 7.10: Local Weather Effects

Sierra Navada
- 7.5: Climate and Ocean Surface Water Properties

silica
- 12.12: Plankton
- 12.6: Primary Production and Nutrients
- 15.6: Polar Regions
- 15.8: Hydrothermal Vents
- 6.4: Biogenous Sediments

silicate
- 13.3: Seasonal Cycles
- 13.4: Algal Blooms

siliceous
- 6.4: Biogenous Sediments

siliceous sediment
- 6.9: Distribution of Surface Sediments

silicic acid
- 12.6: Primary Production and Nutrients

silicoflagellates
- 12.12: Plankton

sill
- 11.1: Processes that Form and Modify the Coast
- 13.6: Estuaries

silt
- 6.2: Classification of Sediments

silver
- 15.8: Hydrothermal Vents

skates
- 14.6: Reproduction

slack water
- 10.6: Tidal Currents

slumping
- 11.2: Beaches

snails
- 14.6: Reproduction
- 14.7: Associations
- 15.3: Kelp Forests
- 15.5: Sargasso Sea
- 15.8: Hydrothermal Vents
- 16.6: Climate Change, Acidification, and Deoxygenation

sodium
- 5.2: The Water Molecule
- 5.6: Dissolved Chemicals in Seawater

soft corals
- 14.3: Feeding
- 14.7: Associations
- 15.2: Coral Reefs

solar energy
- 7.3: Water and Heat Budgets

solar heating
- 8.1: Energy Sources

solar radiation
- 17.3: CC3 Convection and Convection Cells
- 17.5: CC5 Transfer and Storage of Heat by Water
- 17.9: CC9 The Global Greenhouse Effect
- 7.10: Local Weather Effects
- 7.3: Water and Heat Budgets

solar system
- 5.1: Origins and Distribution of the Earth's Water

solar tides
- 10.3: Tides on an Ocean Covered Earth

soles
- 12.13: Nekton
- 14.4: Hunting and Defense
- 16.9: Box 1 - Petroleum in the Marine Environment

Solomon Islands
- 14.4: Hunting and Defense
- 14.5: Selected Adaptations in Fishes

solstice
- 7.4: Climatic Winds

soluable
- 12.6: Primary Production and Nutrients

solubility
- 5.3: The Dissolving Power of Water

Somalia
- 2.3: National and International Law Applied to Oceans
- 9.9: Tsunamis

sonar
- 16.5: Ocean Pollution Issues
- 3.3: Bathymetry
- 5.12: Transmission of Sound

sorting
- 11.2: Beaches

sound
- 14.8: Communication and Navigation

sound channel
- 5.12: Transmission of Sound

sound waves
- 5.12: Transmission of Sound

soundings
- 2.2: Exploration and Mapping
- 3.3: Bathymetry
- 5.12: Transmission of Sound

South Africa
- 10.5: Tides in the Open Ocean
- 2.2: Exploration and Mapping
- 8.10: Eddies
- 9.5: Sea Development and Wave Height

South America
- 10.5: Tides in the Open Ocean
- 12.11: Biological Provinces and Zones
- 13.2: Nutrient Supply to the Coastal Photic Zone
- 4.10: Present-Day Oceans
- 4.2: Plate Tectonics
- 4.3: Plate Boundaries
- 4.5: Divergent Plate Boundaries
- 6.3: Lithogenous Sediments
- 7.10: Local Weather Effects
- 7.5: Climate and Ocean Surface Water Properties
- 7.6: Interannual Climate Variations
- 7.8: Global Climate Zones
- 8.5: Open-Ocean Surface Currents

South American Plate
- 4.2: Plate Tectonics

South Carolina
- 7.9: Weather Systems

South China Sea
- 2.3: National and International Law Applied to Oceans
- 2.4: Ocean Resources
- 4.10: Present-Day Oceans
- 6.3: Lithogenous Sediments

South Equatorial Current
- 8.18: Box 2 - Equatorial Currents
- 8.5: Open-Ocean Surface Currents
- 8.8: Upwelling and Downwelling

South Pole
- 17.12: CC12 The Coriolis Effect
- 6.10: The Sediment Historical Record
- 7.1: The Atmosphere
- 7.4: Climatic Winds

South Sandwich Trench
- 4.10: Present-Day Oceans

Southeast Asia
- 2.2: Exploration and Mapping

Southern Ocean
- 12.6: Primary Production and Nutrients
- 12.7: Food Webs
- 15.6: Polar Regions
- 16.6: Climate Change, Acidification, and Deoxygenation
- 6.9: Distribution of Surface Sediments
- 7.1: The Atmosphere

Soviet Union
- 16.11: Box 3 - Radioactive Wastes in the Marine Environment
- 16.5: Ocean Pollution Issues
- 2.3: National and International Law Applied to Oceans
- 3.4: Seafloor Sediments

Spartina
- 14.3: Feeding

spatial
- 7.9: Weather Systems

spatial analysis
- 3.9: Computers and Modeling

spawning
10.8: Marine Species and the Rhythm of Tides
13.2: Nutrient Supply to the Coastal Photic Zone
13.6: Estuaries
14.4: Hunting and Defense
14.6: Reproduction
16.5: Ocean Pollution Issues

species
12.1: How do we describe life?
12.11: Biological Provinces and Zones
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
14.1: Ecological Requirements
14.3: Feeding
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
14.8: Communication and Navigation
15.1: Communities and Niches
16.2: Adverse Effects of Human Activities
17.10: CC10 Modeling
17.11: CC11 Chaos
17.16: CC16 Maximum Sustainable Yield
17.17: CC17 Species Diversity and Biodiversity
3.6: Sampling Living Organisms in the Sea
8.10: Eddies

species distribution
12.11: Biological Provinces and Zones
15.1: Communities and Niches
15.4: Rocky Intertidal Communities

species interactions
14.7: Associations

species succession
13.3: Seasonal Cycles

speckled sandperch (Parapercis hexophthalma)
14.5: Selected Adaptations in Fishes

speed
14.4: Hunting and Defense

sperm
14.6: Reproduction

sperm whales
12.13: Nekton
12.7: Food Webs

spider crabs
14.4: Hunting and Defense

spinecheek anemonefish (Premnas biaculeatus)
14.4: Hunting and Defense

spines
14.4: Hunting and Defense

spits
11.3: Barrier Islands and Lagoons

splash zone
12.11: Biological Provinces and Zones

sponges
12.13: Nekton
14.3: Feeding
14.4: Hunting and Defense
14.6: Reproduction
15.4: Rocky Intertidal Communities

spores
13.2: Nutrient Supply to the Coastal Photic Zone
15.3: Kelp Forests

spotted seahorse (Hippocampus kuda)
14.4: Hunting and Defense

spotted sole (Phyllichths punctatus)
14.4: Hunting and Defense

Spratly Islands
2.3: National and International Law Applied to Oceans

spreading cycle
11.1: Processes that Form and Modify the Coast
12.11: Biological Provinces and Zones
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.2: Plate Tectonics

spring bloom
13.3: Seasonal Cycles

spring tides
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
10.8: Marine Species and the Rhythm of Tides
11.2: Beaches
14.6: Reproduction
15.2: Coral Reefs

squid
12.13: Nekton
12.7: Food Webs
14.2: Habitats
14.4: Hunting and Defense
14.6: Reproduction
15.7: Beyond the Sun's Light

Sri Lanka
9.9: Tsunamis

staghorn coral
14.3: Feeding

standing stock
12.8: Geographic Variation in Primary Production
13.3: Seasonal Cycles
17.16: CC16 Maximum Sustainable Yield

standing waves
10.4: Tides in the Earth's Oceans
10.6: Tidal Currents
9.12: Standing Waves
9.2: Progressive Waves
9.3: What is Wave Motion?

starry flounder
12.13: Nekton

stationary waves
9.12: Standing Waves

steady state
17.8: CC8 Residence Time
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
6.1: Sediments and Biogeochemical Cycles

steering forces
8.2: Wind-Driven Currents

stewardship
16.7: Risk Assessment and Management

stonefishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes

storm impact
11.3: Barrier Islands and Lagoons

storm surges
11.6: Wetlands
12.11: Biological Provinces and Zones
15.4: Rocky Intertidal Communities
7.9: Weather Systems
9.9: Tsunamis

storm waves
11.4: Beaches and Human Structures
9.6: Wave Dissipation

Straits of Gibraltar
17.8: CC8 Residence Time
6.5: Hydrogenous Sediments
8.13: Thermohaline Circulation

stratified
13.1: Special Characteristics of Coastal Oceans
17.1: CC1 Density and Layering in Fluids
8.13: Thermohaline Circulation

stratigraphic record
6.10: The Sediment Historical Record

stratosphere
7.1: The Atmosphere

striated frogfish (Antennarius striatus)
14.4: Hunting and Defense

striped bass
13.6: Estuaries

stumpy spined cuttlefish (Sepia bandanensis)
12.13: Nekton

subduction
11.1: Processes that Form and Modify the Coast
12.2: Production, Consumption, and Decomposition
4.2: Plate Tectonics

subduction zones
13.6: Estuaries
15.8: Hydrothermal Vents
4.3: Plate Boundaries
4.4: Convergent Plate Boundaries
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
9.9: Tsunamis

sublittoral zone
12.11: Biological Provinces and Zones

submarine canyon
11.2: Beaches

submarines
9.10: Internal Waves

submersibles
15.7: Beyond the Sun's Light
3.7: Scuba, Manned, and Unmanned Submersibles

subpolar gyres
8.7: High-Latitude Surface Currents

subpolar regions
13.3: Seasonal Cycles

subpolar zones
7.8: Global Climate Zones

subtropical convergence
8.5: Open-Ocean Surface Currents

subtropical gyres
12.6: Primary Production and Nutrients
15.5: Sargasso Sea
16.5: Ocean Pollution Issues
7.5: Climate and Ocean Surface Water Properties
8.18: Box 2 - Equatorial Currents
8.2: Wind-Driven Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
8.7: High-Latitude Surface Currents

subtropical oceans
13.3: Seasonal Cycles

Suez Canal
17.8: CC8 Residence Time

sulfate
12.9: Dissolved Oxygen and Carbon Dioxide

sulfide
12.9: Dissolved Oxygen and Carbon Dioxide
14.2: Habitats
15.8: Hydrothermal Vents
17.14: CC14 Phototrophy, Light, and Nutrients
6.10: The Sediment Historical Record

sulfur
5.6: Dissolved Chemicals in Seawater

Sumatra
9.9: Tsunamis

sun
10.1: Tide-Generating Forces
8.1: Energy Sources
supersaturated
12.9: Dissolved Oxygen and Carbon Dioxide
supralittoral zone
15.4: Rocky Intertidal Communities
surf
11.2: Beaches
surf smelt
10.8: Marine Species and the Rhythm of Tides
surf zone
11.4: Beaches and Human Structures
9.3: What is Wave Motion?
9.8: Waves in Shallow Water
surface currents
3.8: Satellites
8.3: Ekman Motion
8.5: Open-Ocean Surface Currents
surface layer
13.3: Seasonal Cycles
13.6: Estuaries
surface microlayer
14.2: Habitats
3.5: Chemical and Physical Oceanography
5.10: Surface Tension and Viscosity
surface sediments
6.9: Distribution of Surface Sediments
surface tension
14.2: Habitats
5.10: Surface Tension and Viscosity
5.2: The Water Molecule
8.2: Wind-Driven Currents
9.3: What is Wave Motion?
surface water
8.2: Wind-Driven Currents
8.8: Upwelling and Downwelling
surface zone
8.13: Thermohaline Circulation
surfing
9.8: Waves in Shallow Water
surge channels
15.2: Coral Reefs
surgeonfishes
14.4: Hunting and Defense
survival tactics
14.4: Hunting and Defense
suspended sediments
11.5: Coral Reefs and Atolls
11.6: Wetlands
11.7: Deltas
12.5: Phototropic Primary Production and Light
13.1: Special Characteristics of Coastal Oceans
15.1: Communities and Niches
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
8.9: Coastal Currents
suspension feeders
14.3: Feeding
15.2: Coral Reefs
sustainability
16.7: Risk Assessment and Management
swash
11.2: Beaches
swell
9.5: Sea Development and Wave Height
swim bladders
12.13: Nekton
14.5: Selected Adaptations in Fishes
swimming
14.5: Selected Adaptations in Fishes
15.4: Rocky Intertidal Communities
symbiosis
11.5: Coral Reefs and Atolls
14.7: Associations
symbiotic
15.2: Coral Reefs
symbiotic relationship
13.1: Special Characteristics of Coastal Oceans
synoptic observations
3.8: Satellites
8.10: Eddies

T

Tambora
6.3: Lithogenous Sediments
Tampa Bay
16.5: Ocean Pollution Issues
taxonomy
12.1: How do we describe life?
tectonic activity
4.3: Plate Boundaries
tectonic estuaries
13.6: Estuaries
tectonic history
6.10: The Sediment Historical Record
tectonic plates
4.4: Convergent Plate Boundaries
4.5: Divergent Plate Boundaries
4.7: Hot Spots
tectonic processes
11.1: Processes that Form and Modify the Coast
17.2: CC2 Isostasy, Eustasy, and Sea Level
4.8: Plate Interiors
4.9: Sea-Level Change and Climate
telegraph cables
3.2: The Unseen Domain
temperate ocean zones
7.8: Global Climate Zones
temperature
12.11: Biological Provinces and Zones
12.12: Plankton
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.5: Fisheries
14.2: Habitats
14.6: Reproduction
15.4: Rocky Intertidal Communities
15.6: Polar Regions
15.7: Beyond the Sun's Light
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
16.9: Box 1 - Petroleum in the Marine Environment
17.1: CC1 Density and Layering in Fluids
17.11: CC11 Chaos
17.17: CC17 Species Diversity and Biodiversity
17.18: CC18 Toxicity
17.3: CC3 Convection and Convection Cells
17.5: CC5 Transfer and Storage of Heat by Water
17.6: CC6 Salinity, Temperature, Pressure, and Water Density
17.9: CC9 The Global Greenhouse Effect
3.5: Chemical and Physical Oceanography
3.8: Satellites
5.1: Origins and Distribution of the Earth's Water
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
5.9: Ice Formation
7.10: Local Weather Effects
7.3: Water and Heat Budgets
7.5: Climate and Ocean Surface Water Properties
7.6: Interannual Climate Variations
7.8: Global Climate Zones
8.10: Eddies
8.13: Thermohaline Circulation
8.15: Tracing Water Masses
terrigenous sediment
6.3: Lithogenous Sediments
6.9: Distribution of Surface Sediments
territorial sea
2.3: National and International Law Applied to Oceans
Tethys Sea
4.4: Convergent Plate Boundaries
Texas
10.2: Characteristics of Tides
11.3: Barrier Islands and Lagoons
13.4: Algal Blooms
7.9: Weather Systems
Thailand
7.6: Interannual Climate Variations
9.9: Tsunamis
Thames River
16.10: Box 2 - Sewage Wastes in the Marine Environment
thermal expansion
5.2: The Water Molecule
thermal inertia
16.6: Climate Change, Acidification, and Deoxygenation
thermal properties
5.10: Surface Tension and Viscosity
5.7: Heat Properties of Water
thermal stratification
17.1: CC1 Density and Layering in Fluids

thermocline
12.6: Primary Production and Nutrients
12.8: Geographic Variation in Primary Production
12.9: Dissolved Oxygen and Carbon Dioxide
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.4: Algal Blooms
15.2: Coral Reefs
15.5: Sargasso Sea
15.6: Polar Regions
8.13: Thermohaline Circulation
thermohaline circulation
16.6: Climate Change, Acidification, and Deoxygenation
8.1: Energy Sources
8.13: Thermohaline Circulation
thermosphere
7.1: The Atmosphere
tidal bore
10.6: Tidal Currents
10.9: Energy from Tides
tidal currents
10.6: Tidal Currents
10.9: Energy from Tides
13.1: Special Characteristics of Coastal Oceans
13.6: Estuaries
8.11: Inertial Currents
tidal curves
10.2: Characteristics of Tides
tidal energy
10.9: Energy from Tides
tidal force
10.1: Tide-Generating Forces
tidal forces
10.3: Tides on an Ocean Covered Earth
9.3: What is Wave Motion?
tidal height
10.3: Tides on an Ocean Covered Earth
10.5: Tides in the Open Ocean
tidal motion
10.5: Tides in the Open Ocean
tidal patterns
10.3: Tides on an Ocean Covered Earth
tidal power plant
10.9: Energy from Tides
2.4: Ocean Resources
tidal pull
10.1: Tide-Generating Forces
tidal range
10.2: Characteristics of Tides
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
11.1: Processes that Form and Modify the Coast
11.2: Beaches
13.6: Estuaries
14.6: Reproduction
15.4: Rocky Intertidal Communities
tidal wetlands
11.6: Wetlands
tide pools
14.5: Selected Adaptations in Fishes
15.4: Rocky Intertidal Communities
tide waves
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
tides
10.1: Tide-Generating Forces
10.2: Characteristics of Tides
10.3: Tides on an Ocean Covered Earth
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
10.7: Tide Prediction
10.8: Marine Species and the Rhythm of Tides
10.9: Energy from Tides
11.1: Processes that Form and Modify the Coast
12.11: Biological Provinces and Zones
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.6: Estuaries
14.2: Habitats
15.2: Coral Reefs
15.4: Rocky Intertidal Communities
2.2: Exploration and Mapping
8.13: Thermohaline Circulation
9.10: Internal Waves
9.4: Making Waves
tiger shark
12.13: Nekton
tobies
14.4: Hunting and Defense
tomato anemonefish (Amphiprion frenatus)
14.4: Hunting and Defense
Tonga Ridge
4.5: Divergent Plate Boundaries
topography
1.3: How to Study Ocean Data
13.2: Nutrient Supply to the Coastal Photic Zone
2.2: Exploration and Mapping
8.3: Ekman Motion
tornadoes
7.9: Weather Systems
torpedo ray
12.13: Nekton
tourism
13.4: Algal Blooms
2.4: Ocean Resources
toxic
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.3: Toxicity
16.4: Waste Disposal
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
17.18: CC18 Toxicity
toxicity
16.1: Pollution versus Contamination
16.3: Toxicity
17.18: CC18 Toxicity
toxins
13.4: Algal Blooms
13.6: Estuaries
14.4: Hunting and Defense
16.3: Toxicity
trace metals
5.6: Dissolved Chemicals in Seawater
tracers
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
8.15: Tracing Water Masses
trade
2.2: Exploration and Mapping
2.4: Ocean Resources
trade winds
13.2: Nutrient Supply to the Coastal Photic Zone
7.4: Climatic Winds
7.6: Interannual Climate Variations
7.8: Global Climate Zones
8.18: Box 2 - Equatorial Currents
8.5: Open-Ocean Surface Currents
8.6: Equatorial Surface Currents
8.7: High-Latitude Surface Currents
9.5: Sea Development and Wave Height
Transantarctic Mountains
16.6: Climate Change, Acidification, and Deoxygenation
transform boundaries
4.3: Plate Boundaries
transform faults
11.1: Processes that Form and Modify the Coast
13.6: Estuaries
4.6: Transform Faults and Fracture Zones
transform plate boundary
4.3: Plate Boundaries
transportation
2.4: Ocean Resources
trash
16.5: Ocean Pollution Issues
trawling
16.5: Ocean Pollution Issues
2.4: Ocean Resources
Treaty of the High Seas
2.3: National and International Law Applied to Oceans
tree of life
12.1: How do we describe life?
trenches
10.4: Tides in the Earth's Oceans
12.11: Biological Provinces and Zones
trevalley
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
Trieste
3.7: Scuba, Manned, and Unmanned Submersibles
triggerfishes
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes
Tristan de Cunha
4.7: Hot Spots
tritium
8.15: Tracing Water Masses
trophic efficiency
13.5: Fisheries
trophic levels
12.7: Food Webs
13.5: Fisheries
17.15: CC15 Food Chain Efficiency
17.16: CC16 Maximum Sustainable Yield
17.18: CC18 Toxicity
tropical cyclones
7.9: Weather Systems
tropical Pacific
7.6: Interannual Climate Variations
tropical regions
13.3: Seasonal Cycles
tropics
15.2: Coral Reefs
17.5: CC5 Transfer and Storage of Heat by Water

troposphere
7.1: The Atmosphere
7.2: Water Vapor in the Atmosphere
troughs
10.4: Tides in the Earth's Oceans
11.2: Beaches
9.12: Standing Waves
9.2: Progressive Waves
9.9: Tsunamis
TS diagrams
8.15: Tracing Water Masses
tsunami warning
9.9: Tsunamis
tsunamis
11.1: Processes that Form and Modify the Coast
12.11: Biological Provinces and Zones
9.4: Making Waves
9.9: Tsunamis
tube worms
14.3: Feeding
15.8: Hydrothermal Vents
tuna
12.13: Nekton
12.7: Food Webs
13.5: Fisheries
14.5: Selected Adaptations in Fishes
17.16: CC16 Maximum Sustainable Yield
tunicates
12.12: Plankton
14.3: Feeding
14.4: Hunting and Defense
turbidite layers
6.7: Sediment Transport, Deposition, and Accumulation
turbidity
11.5: Coral Reefs and Atolls
12.5: Phototropic Primary Production and Light
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.6: Estuaries
15.2: Coral Reefs
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
3.5: Chemical and Physical Oceanography
4.8: Plate Interiors
7.10: Local Weather Effects
turbidity currents
15.7: Beyond the Sun's Light
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
6.3: Lithogenous Sediments
6.7: Sediment Transport, Deposition, and Accumulation
9.4: Making Waves
turbines
10.9: Energy from Tides
turbulence
10.8: Marine Species and the Rhythm of Tides
13.1: Special Characteristics of Coastal Oceans
15.6: Polar Regions
17.11: CC11 Chaos
17.3: CC3 Convection and Convection Cells
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
17.5: CC5 Transfer and Storage of Heat by Water
6.3: Lithogenous Sediments
9.4: Making Waves
9.6: Wave Dissipation
Turkey
13.6: Estuaries
turtle grass (Thalassia)
12.13: Nekton
14.3: Feeding
14.4: Hunting and Defense
twospot lizardfish (Synodus binotatus)
14.4: Hunting and Defense
typhoons
7.9: Weather Systems

U

U.S. Senate
2.3: National and International Law Applied to Oceans
U.S. Virgin Islands
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
Ukraine
16.11: Box 3 - Radioactive Wastes in the Marine Environment
16.5: Ocean Pollution Issues
ultraplankton
12.12: Plankton
ultraviolet light
7.1: The Atmosphere
underwater communication
5.11: Transmission of Light and Other Electromagnetic Radiation
underwater navigation
3.3: Bathymetry
underwater research technologies
3.7: Scuba, Manned, and Unmanned Submersibles
underwater vehicles
3.1: Difficulties of Studying the Ocean Environment
United Kingdom
16.10: Box 2 - Sewage Wastes in the Marine Environment
3.4: Seafloor Sediments
United Nations
2.3: National and International Law Applied to Oceans
United States
11.1: Processes that Form and Modify the Coast
11.6: Wetlands
12.13: Nekton
13.1: Special Characteristics of Coastal Oceans
13.4: Algal Blooms
13.6: Estuaries
16.1: Pollution versus Contamination
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.9: Box 1 - Petroleum in the Marine Environment
2.3: National and International Law Applied to Oceans
2.4: Ocean Resources
3.3: Bathymetry
3.4: Seafloor Sediments
4.9: Sea-Level Change and Climate
6.3: Lithogenous Sediments
7.10: Local Weather Effects
7.6: Interannual Climate Variations
upwelling
12.8: Geographic Variation in Primary Production
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.4: Algal Blooms
13.5: Fisheries
14.6: Reproduction
15.2: Coral Reefs
15.6: Polar Regions
17.3: CC3 Convection and Convection Cells
8.12: Langmuir Circulation
8.5: Open-Ocean Surface Currents
8.8: Upwelling and Downwelling
Ural Mountains
4.4: Convergent Plate Boundaries
urchin shrimp (Gnathophylloides mineri)
14.7: Associations
USS Ramapo
9.5: Sea Development and Wave Height
USS Thresher
9.10: Internal Waves
USS Tuscarora
3.3: Bathymetry

V

Van der Waals force
5.2: The Water Molecule
Vanuatu
14.5: Selected Adaptations in Fishes
14.6: Reproduction
vegetation
11.1: Processes that Form and Modify the Coast
7.8: Global Climate Zones
venom
14.5: Selected Adaptations in Fishes
vertebrates
12.12: Plankton
12.13: Nekton
13.4: Algal Blooms
14.5: Selected Adaptations in Fishes
vertical migration
15.7: Beyond the Sun's Light
vertical mixing
13.1: Special Characteristics of Coastal Oceans
13.2: Nutrient Supply to the Coastal Photic Zone
13.6: Estuaries
15.6: Polar Regions
17.1: CC1 Density and Layering in Fluids
vertical movement
8.8: Upwelling and Downwelling
viperfishes
15.7: Beyond the Sun's Light
Virgina
10.2: Characteristics of Tides
11.4: Beaches and Human Structures
viruses
12.10: Organic Carbon
12.12: Plankton
13.4: Algal Blooms
14.4: Hunting and Defense
16.10: Box 2 - Sewage Wastes in the Marine Environment
16.2: Adverse Effects of Human Activities
17.19: CC19 Metagenomics
3.6: Sampling Living Organisms in the Sea

viscosity
14.3: Feeding
14.6: Reproduction
17.2: CC2 Isostasy, Eustasy, and Sea Level
17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension
5.10: Surface Tension and Viscosity
9.6: Wave Dissipation
visible light
5.11: Transmission of Light and Other Electromagnetic Radiation
vision
14.8: Communication and Navigation
viviparous
14.6: Reproduction
15.4: Rocky Intertidal Communities
volcanic activity
6.1: Sediments and Biogeochemical Cycles
volcanic islands
11.5: Coral Reefs and Atolls
4.10: Present-Day Oceans
4.7: Hot Spots
4.8: Plate Interiors
volcanoes
11.5: Coral Reefs and Atolls
4.10: Present-Day Oceans
4.4: Convergent Plate Boundaries
4.5: Divergent Plate Boundaries
4.8: Plate Interiors
6.10: The Sediment Historical Record
6.5: Hydrogenous Sediments
9.9: Tsunamis
vorticity
8.17: Box 1- Westward Intensification
8.5: Open-Ocean Surface Currents

W

walrus (Odobenus rosmarus)
12.13: Nekton
15.6: Polar Regions
Walvis Ridge
4.7: Hot Spots
Washington
10.2: Characteristics of Tides
13.4: Algal Blooms
13.6: Estuaries
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
7.6: Interannual Climate Variations
waste
2.4: Ocean Resources
waste management
16.4: Waste Disposal
water budget
7.3: Water and Heat Budgets
water circulation
8.13: Thermohaline Circulation
water cycle
5.1: Origins and Distribution of the Earth's Water
water density
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
8.1: Energy Sources
8.4: Geostrophic Currents
water masses
17.12: CC12 The Coriolis Effect
17.13: CC13 Geostrophic Flow
17.6: CC6 Salinity, Temperature, Pressure, and Water Density
8.13: Thermohaline Circulation
8.15: Tracing Water Masses
water molecule
17.5: CC5 Transfer and Storage of Heat by Water
5.10: Surface Tension and Viscosity
5.2: The Water Molecule
5.7: Heat Properties of Water
5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater
9.3: What is Wave Motion?
9.6: Wave Dissipation
water molecules
14.5: Selected Adaptations in Fishes
water quality
13.6: Estuaries
16.1: Pollution versus Contamination
16.2: Adverse Effects of Human Activities
water vapor
7.2: Water Vapor in the Atmosphere
7.8: Global Climate Zones
wave action
11.2: Beaches
wave amplitude
9.10: Internal Waves
9.2: Progressive Waves
wave behavior
9.8: Waves in Shallow Water
wave breaking
9.5: Sea Development and Wave Height
wave characteristics
9.2: Progressive Waves
wave dispersion
9.5: Sea Development and Wave Height
9.7: Deep-Water Waves
wave dynamics
9.3: What is Wave Motion?
9.5: Sea Development and Wave Height
9.6: Wave Dissipation
9.8: Waves in Shallow Water
wave energy
9.2: Progressive Waves
9.7: Deep-Water Waves
wave frequency
11.2: Beaches
9.2: Progressive Waves
wave generation
9.11: Rossby and Kelvin Waves
wave height
10.2: Characteristics of Tides
10.5: Tides in the Open Ocean
11.2: Beaches
7.9: Weather Systems
9.1: Complexity of Ocean Waves
9.10: Internal Waves
9.5: Sea Development and Wave Height
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
wave interaction
9.1: Complexity of Ocean Waves
wave interference
9.1: Complexity of Ocean Waves
9.7: Deep-Water Waves
wave length
9.2: Progressive Waves
wave orbit
11.2: Beaches
9.3: What is Wave Motion?
9.5: Sea Development and Wave Height
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
wave pattern
9.1: Complexity of Ocean Waves
wave period
10.2: Characteristics of Tides
11.2: Beaches
13.2: Nutrient Supply to the Coastal Photic Zone
9.1: Complexity of Ocean Waves
9.2: Progressive Waves
wave rays
9.8: Waves in Shallow Water
wave refraction
9.8: Waves in Shallow Water
wave shape
9.5: Sea Development and Wave Height
wave speed
10.4: Tides in the Earth's Oceans
11.2: Beaches
9.2: Progressive Waves
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
wave steepness
9.2: Progressive Waves
9.5: Sea Development and Wave Height
9.7: Deep-Water Waves
wave trains
9.7: Deep-Water Waves
wavelengths
10.4: Tides in the Earth's Oceans
11.2: Beaches
13.2: Nutrient Supply to the Coastal Photic Zone
14.4: Hunting and Defense
15.7: Beyond the Sun's Light
17.14: CC14 Phototrophy, Light, and Nutrients
17.5: CC5 Transfer and Storage of Heat by Water
17.9: CC9 The Global Greenhouse Effect
5.11: Transmission of Light and Other Electromagnetic Radiation
9.12: Standing Waves
9.2: Progressive Waves
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
wavelengths of light
8.10: Eddies

waves
10.2: Characteristics of Tides
10.4: Tides in the Earth's Oceans
10.5: Tides in the Open Ocean
10.6: Tidal Currents
11.1: Processes that Form and Modify the Coast
11.2: Beaches
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
13.6: Estuaries
15.2: Coral Reefs
15.4: Rocky Intertidal Communities
16.2: Adverse Effects of Human Activities
16.6: Climate Change, Acidification, and Deoxygenation
16.9: Box 1 - Petroleum in the Marine Environment
6.3: Lithogenous Sediments
7.2: Water Vapor in the Atmosphere
8.2: Wind-Driven Currents
9.1: Complexity of Ocean Waves
9.10: Internal Waves
9.11: Rossby and Kelvin Waves
9.2: Progressive Waves
9.3: What is Wave Motion?
9.4: Making Waves
9.6: Wave Dissipation
9.7: Deep-Water Waves
9.8: Waves in Shallow Water
9.9: Tsunamis
weather
11.1: Processes that Form and Modify the Coast
15.6: Polar Regions
17.10: CC10 Modeling
17.11: CC11 Chaos
17.13: CC13 Geostrophic Flow
7.1: The Atmosphere
7.2: Water Vapor in the Atmosphere
7.4: Climatic Winds
7.8: Global Climate Zones
7.9: Weather Systems
8.10: Eddies
weather forecasting
3.8: Satellites
7.9: Weather Systems
weathering
11.2: Beaches
12.6: Primary Production and Nutrients
5.4: Sources and Sinks of Chemicals Dissolved in Seawater
6.2: Classification of Sediments
Weddell Sea
8.13: Thermohaline Circulation
weed cardinalfish (Foa brachygramma)
14.7: Associations
West Antarctic Ice Sheet
16.6: Climate Change, Acidification, and Deoxygenation
West Germany
2.3: National and International Law Applied to Oceans
westerlies
12.8: Geographic Variation in Primary Production
7.4: Climatic Winds
8.5: Open-Ocean Surface Currents
9.5: Sea Development and Wave Height
western boundary currents
13.2: Nutrient Supply to the Coastal Photic Zone
15.2: Coral Reefs
7.9: Weather Systems
8.10: Eddies
8.17: Box 1- Westward Intensification
8.5: Open-Ocean Surface Currents
westward intensification
8.17: Box 1- Westward Intensification
wetlands
11.6: Wetlands
11.7: Deltas
13.6: Estuaries
16.2: Adverse Effects of Human Activities
16.5: Ocean Pollution Issues
16.6: Climate Change, Acidification, and Deoxygenation
16.9: Box 1 - Petroleum in the Marine Environment
whale shark (Rhincodon typus)
12.13: Nekton
whales
12.12: Plankton
12.7: Food Webs
13.4: Algal Blooms
14.4: Hunting and Defense
14.8: Communication and Navigation
15.6: Polar Regions
whip corals
14.4: Hunting and Defense
14.7: Associations
whip goby (Bryaninops yongei)
14.7: Associations
white smokers
15.8: Hydrothermal Vents
willy willys
7.9: Weather Systems
wind chill
5.7: Heat Properties of Water
wind direction
8.3: Ekman Motion
wind energy
9.5: Sea Development and Wave Height
winds
11.1: Processes that Form and Modify the Coast
13.3: Seasonal Cycles
15.5: Sargasso Sea
17.13: CC13 Geostrophic Flow
6.3: Lithogenous Sediments
7.10: Local Weather Effects
7.2: Water Vapor in the Atmosphere
7.4: Climatic Winds
7.9: Weather Systems
8.1: Energy Sources
8.10: Eddies
8.12: Langmuir Circulation
8.2: Wind-Driven Currents
9.4: Making Waves
windward
15.2: Coral Reefs
7.10: Local Weather Effects
9.4: Making Waves
World Glory (tanker)
9.5: Sea Development and Wave Height
worms
14.2: Habitats
14.6: Reproduction
14.7: Associations
15.2: Coral Reefs
15.4: Rocky Intertidal Communities
15.7: Beyond the Sun's Light
wrasses
14.4: Hunting and Defense
14.5: Selected Adaptations in Fishes

Y

Yangtze River
6.3: Lithogenous Sediments
Yellow River
6.3: Lithogenous Sediments
Yellow Sea
6.3: Lithogenous Sediments
yellow shrimp goby (Cryptocentrus cinctus)
14.7: Associations
yellowstreak fusilier (Pterocaesio lativittata)
14.5: Selected Adaptations in Fishes
Yucatán Peninsula
6.10: The Sediment Historical Record
7.9: Weather Systems

Z

Zanzibar whip coral shrimp (Dasycaris zanzibarica,
14.4: Hunting and Defense
zebra lionfish (Dendrochirus zebra)
14.5: Selected Adaptations in Fishes
zinc
15.8: Hydrothermal Vents
16.1: Pollution versus Contamination
17.18: CC18 Toxicity
zooplankton
12.10: Organic Carbon
12.12: Plankton
12.5: Phototropic Primary Production and Light
12.7: Food Webs
12.8: Geographic Variation in Primary Production
13.2: Nutrient Supply to the Coastal Photic Zone
13.3: Seasonal Cycles
13.5: Fisheries
13.6: Estuaries
14.2: Habitats
14.3: Feeding
15.2: Coral Reefs
15.6: Polar Regions
16.5: Ocean Pollution Issues
3.6: Sampling Living Organisms in the Sea
6.4: Biogenous Sediments
zooxanthellae
11.5: Coral Reefs and Atolls
13.1: Special Characteristics of Coastal Oceans
13.3: Seasonal Cycles
14.7: Associations
15.2: Coral Reefs
16.6: Climate Change, Acidification, and Deoxygenation

This page titled Index is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

GLOSSARY

absolute density | The mass per unit volume of a substance (usually expressed as grams per cubic centimeter). Compare torelative density.

absorption spectrum,absorptionspectra | Wavelengths of electromagnetic energy that substances can absorb and convert to heat. Absorption spectra are plotted as the relative efficiency of absorption versus wavelength. Compare emission spectrum.

abyssal fan | Fan-shaped sediment accumulation spreading out and decreasing in thickness from a point of large sediment input. Abyssal fans are located usually off the mouth of a river or at the foot of a submarine canyon.

abyssal hill | Sediment-covered volcanic peak that rises less than 1 km above the deep-ocean floor.

abyssal plain | Flat seafloor extending seaward from the base of the continental slope and continental rise or from the seaward edge of an oceanic trench.

abyssal zone | Benthic environment between 4000 and 6000 m depth.

acid rain | Rain with acidity higher than normal because of dissolved gaseous emissions of industry and automobiles, notably sulfur compounds.

acidification | An increase of hydrogen ion concentration to make a solution more acidic or less alkaline. In ocean acidification refers to the addition of anthropogenic carbon dioxide to the oceans that reduces the pH and renders calcium carbonate more soluble.

acoustic | Pertaining to sound.

adiabatic expansion | Expansion of a gas without the addition of external heat. Adiabatic expansion causes the temperature of the gas to decrease. Adiabatic compression causes the temperature of the gas to increase.

adsorption, adsorbed | Attraction and adhesion of ions to a solid surface.

aerobic | In the presence of oxygen.

algae, alga | Simple single-celled or many-celled photosynthetic organisms that have no root, stem, or leaf systems.

algal ridge | Irregular ridge located on the wave-exposed seaward edge of many coral reefs. The ridge is composed largely of encrusting algae.

alkalinity,alkaline | The opposite of acidity. Measure of the degree to which the concentration of hydroxyl ions (OH^-) is greater than the concentration of hydrogen ions (H^+) in a solution. Seawater is slightly alkaline.

amino acids | Members of a group of compounds that combine in varied proportions to form proteins.

amoebas | Microscopic single-celled animals that move by making continuous protrusions of the body and that feed by engulfing bits of food.

amphidromic system | Tide wave that rotates around an ocean basin during one tidal period. An amphidromic system rotates around an amphidromic point (a node) at which the tidal range is zero.

amphipods | Members of an order of the subphylum Crustacea, phylum Arthropoda, that includes laterally compressed species such as the “sand hoppers.”

amplitude | Vertical distance from the mean level of a wave (the still sea surface for an ocean wave) to the crest or to the trough. A wave’s amplitude equals one-half of the wave height.

anadromous | Pertaining to species of fishes that are spawned in freshwater, migrate to the ocean to live until they reach maturity, then return to freshwater to spawn. Compare catadromous.

anaerobic | In the absence of oyxgen.

anemones | Multicelled animals of the class Anthozoa, phylum Cnidaria. There are both single and colonial polypoid forms.

angle of approach | The angle of incidence is important in refraction.

angle of incidence | Angle between a wave front (usually an electromagnetic wave) or a wave ray and the plane of an interface that it meets.

anion | Negatively charged ion. Compare to cation.

annelids | Members of the phylum Annelida. Elongated segmented worms.

anoxia, anoxic | Total absence of dissolved or free molecular oxygen.

anthropogenic | Produced by humans and their activities.

antinode | The part of a standing wave where the vertical motion is at a maximum. Compare to node.

aphotic zone | The part of the ocean in which light is insufficient for phototrophy (which includes photosynthesis). The aphotic zone comprises all the oceans below about 1000 m in areas where ocean water is clear and extends to shallower depths in more turbid waters. Compare to photic zone.

aragonite | A mineral form of calcium carbonate. Aragonite is less common than calcite, but it comprises shells of many pteropod species.

archaea, archean, archeal | Microorganisms of the domain Archaea (previously called Archaeobacteria) that belong to an ancient group of organisms that are separate from bacteria and from which eukaryotes (multicelled organisms) may have evolved. Many archaeal species are chemosynthetic and live in extreme environments.

aspect ratio | Index of the propulsive efficiency of a fish species that is obtained by dividing the square of the height of the caudal fin by the area of the caudal fin.

assimilative capacity | Maximum rate at which a particular segment of the environment or ocean can accommodate the input of a substance. The term usually applies to anthropogenic waste materials.

asteroid | Rocky object orbiting the sun. An asteroid is smaller than a planet. Most asteroids are found in a belt between the orbits of Mars and Jupiter, but some have orbits that take them close to the sun and across the orbits of the planets.

asthenosphere | Plastic or partially molten layer of the Earth's upper mantle upon which the continents "float." The asthenosphere varies in thickness, with its upper boundary at depths of 10 km or more and its lower boundary at depths as great as 800 km.

atoll | Ring-shaped coral reef that grows upward from a submerged volcanic peak and encloses a lagoon. Atolls may support low-lying islands composed of coral debris. Compare barrier reef and fringing reef.

autotrophs, autotrophic | Algae, bacteria, archaea, or plants that can synthesize organic compounds from inorganic nutrients by phototrophy or chemosynthesis. Compare to heterotrophs.

back-arc basin | Generally shallow sea created on the non-subducting plate at an oceanic convergent plate boundary where the subduction is fast enough to stretch the edge of the nonsubducting plate. The back-arc basin lies behind the volcanic island arc. Also called "back-island basin."

backscatter,backscattered | The portion of the light or sound randomly reflected off particles suspended in a fluid that is scattered back in the general direction of the light or sound source. Compare to scatter.

backshore | Inner portion of the shore that is landward of the mean spring-tide high-water line. Waves act on the backshore only during exceptionally high tides and severe storms. Compare to foreshore.

backwash | Water flowing down a beach after the swash of one wave has stopped and before the swash of the next wave arrives.

bacteria,bacterium,bacterial | Microscopic single-celled organisms that comprise one of the three recognized biological domains. Other than cyanobacteria, most bacteria species are autotrophs and decomposers, and most are parasitic. They reproduce by cell division.

baleen | Horny material consisting of numerous plates with fringed edges that grows down from the upper jaw of plankton-feeding whales. Baleen is used to filter (strain) the plankton food from the water.

bar-built estuary | Shallow estuary (lagoon) separated from the open ocean by a bar such as a barrier island. Water in bar-built estuaries is usually well mixed vertically.

barnacles | Members of an order of the subphylum Crustacea, phylum Arthropoda, that attach to the substrate, secrete a protective covering of calcareous plates, and strain food particles from the water with a weblike structure.

barrier island | Long, narrow island built of wave-transported sand, separated from the mainland by a usually shallow lagoon.

barrier reef | Coral reef that parallels the shore but is separated from the landmass by open water. Compare to fringing reef and atoll.

basalt,basaltic | Dark-colored volcanic rock, rich in iron, magnesium, and calcium, characteristic of the oceanic crust.

bathyal zone | Benthic environment from 200 to 2000 m depth.

bathymetry, bathymetric | Seafloor mapping; the study of landform features beneath the water surface. A bathymetric map gives the depth contours of the seafloor.

baymouth bar | Shallow bar, usually of sand, that extends partially or completely across the mouth of a bay.

beach | Sand- or sediment-covered zone between the seaward limit of permanent vegetation and the mean low-water line. Beaches sometimes include the seafloor from the mean low-water line to the surf zone.

benthic | Pertaining to the seafloor, or to organisms that live on or in the seafloor.

benthos | Organisms that live on or in the ocean bottom.

berm | Nearly horizontal portion of a beach backshore at the seaward edge of which the beach slopes abruptly seaward. Compare to scarp.

binning | The process of grouping DNA sequences from the same organism together from a mixed sample

bioaccumulation,bioaccumulate | Process by which dissolved chemicals are taken up by organisms until the concentrations in the tissues are at equilibrium with those in the solution.See CC18.

bioassay | Test in which organisms are exposed to different concentrations of a substance to determine the concentration at which the substance is toxic to that species. See CC18.

bioavailable | Existing in a chemical form that is suitable for uptake by organisms.

biochemical | Pertaining to the synthesis, conversion, use, and decomposition of organic chemicals in the life processes of organisms.

biodegradable | Capable of being decomposed by organisms in the environment. The term generally refers to decomposition to harmless substances.

biodiversity | Poorly defined term that refers to a combination of genetic diversity, species diversity, ecosystem diversity, and physiological diversity within a community or ecosystem. See CC17.

biogenous | Pertaining to material of a biological origin. Usually applied to sediments in which the hard-part remains of organisms (e.g., diatom tests, radiolarian shells) constitute a high proportion of the grains. Compare to cosmogenous, hydrogenous, and lithogenous.

biogeochemical cycle | Transfers of compounds among the living and nonliving components of an ecosystem.

bioinformatics | The use of computer and software programming to analyze biological data, especially when large amounts of data are involved like in genome sequencing. See CC19

bioluminescent,bioluminescence | Light-producing. Referring to organisms that use chemical reactions to produce light.

biomagnification, biomagnify | Process by which dissolved chemicals are taken up by organisms and continuously accumulated to higher concentrations throughout the life cycle. Tissue concentrations are not in equilibrium with concentrations in the surrounding environment. See CC18.

biomass | The total amount of living matter, expressed in units of weight in the entire water column per unit area of water surface, or weight per unit of water volume.

biome | Any defined area or volume of the environment that is characterized chiefly by the dominant species present and the prevailing physical and chemical environmental conditions (e.g., topical rain forest, or kelp forest, or tundra).

biota | All living organisms in a given ecosystem, including archaea, bacteria, protists, fungi, algae, plants, and animals.

bioturbation | Reworking (churning and mixing) of sediments by organisms that burrow in them.

bloom | Very dense aggregation of phytoplankton, resulting from a rapid rate of reproduction.

blubber | Outer fat layer of whales.

bottom trawling | A widely used fishing technique where a net is dragged across the sea floor to catch bottom dwelling species.

brackish | Pertaining to water that is a mixture of freshwater and seawater and has a low salinity.

breakwater | Artificial structure constructed in the ocean to protect a shore from the action of ocean waves.

brine | Water that has a higher salinity (generally much higher) than normal seawater.

brittle stars | Organisms of the class Ophiuroidea, phylum Echinodermata, that have long, slender arms covered in bristles attached to a small central disk.

bulkhead | Structure constructed to separate land and water areas at the shoreline and designed to reduce earth slides and slumps or to lessen wave erosion at the base of a cliff.

buoyancy, buoyant | Ability of an object to float (or rise through a fluid) that results from a density difference between the object and the fluid in which it is immersed.

bycatch | Fishes and other marine animals caught in fishers' nets that are not the target of the fishing. Bycatch is generally thrown overboard as waste.

calcareous, calcium carbonate | Composed of or containing calcium carbonate.

calcite | The most common mineral form of calcium carbonate. Compare to aragonite.

caldera | Crater in a volcano, usually at the center, created by an explosion or collapse of the volcanic cone.

calorie | Obsolete unit of heat energy defined as the amount of heat needed to raise the temperature of 1 g of water by 1°C.

capillary wave | Small ocean wave that has a wavelength less than 1.5 cm. The primary restoring force for capillary waves is surface tension.

carbonate compensation depth, CCD | Depth below which all calcium carbonate particles falling from above are dissolved before they can be incorporated in the sediments.

carcinogens, carcinogenic | Compounds that cause cancer in animals. Compare to mutagens and teratogens.

carnivores, carnivorous | Animals that depend solely on other animals for their food supply. Compare herbivores and omnivores.

cartilaginous | Pertaining to vertebrate animals that have skeletons made of cartilage, a tough elastic tissue.

catadromous | Pertaining to species of fishes that are spawned at sea, migrate to a freshwater stream or lake where they live until they reach maturity, then return to sea to spawn. Compare to anadromous.

cation | Positively charged ion. Compare to anion.

CC1 | Density and Layering in Fluids: Water in the oceans is arranged in layers according to the water density. Many movements of ocean water masses are driven by differences in water density.

CC2 | Isostasy, Eustasy, and Sea Level: Earth's crust floats on the asthenosphere, rising and falling isostatically as climate-driven temperature or mass changes alter its density, affecting local sea levels. Sea levels also change eustatically worldwide due to fluctuations in ocean temperature or shifts in continental ice volume. Eustatic changes occur synchronously and more rapidly than isostatic changes.

CC3 | Convection and Convection Cells: Fluids heated from below, like Earth's mantle, oceans, or atmosphere, rise as their density decreases. Upon cooling at higher levels, they become denser and sink. This creates convection cells in which material rises in upwelling zones, spreads, cools, and descends in downwelling zones.

CC4 | Particle Size, Sinking, Deposition, and Resuspension: Suspended particles in water or air sink at rates based on size, with larger particles descending faster. While high currents or winds can resuspend deposited material, large particles are typically harder to lift, though cohesive fine particles also resist resuspension. These dynamics primarily dictate the grain size of local sediments.

CC5 | Transfer and Storage of Heat by Water: Water's thermal properties are vital for climate regulation. Its high heat capacity enables oceans to store and release vast amounts of thermal energy while maintaining stable temperatures. High latent heat of vaporization enables heat transport via vapor, while the latent heat of fusion allows ice to buffer climate extremes in high latitudes.

CC6 | Salinity, Temperature, Pressure, and Water Density: Sea water density is controlled by temperature, salinity, and, to a lesser extent, pressure. Density is higher at lower temperatures, higher salinities, and higher pressures. Movements of water below the ocean's surface layer are driven primarily by density differences.

This page titled Glossary is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

Appendix 1 - Units and Converstion Factors

Wherever possible, this text uses metric units of measurement. The metric system of measurement is used by all major countries but the United States, and it is the universal system used by all scientists. At present, an additional process is under way directed toward using only those metric units that are standard units approved in the International System of units (SI units). However, some of these standard units are unfamiliar even to many scientists, so they are not yet universally used. SI units are used in this text if they are generally in wide usage. In some instances where nonstandard units are much more widely used in the United States, the nonstandard units are used in this text. Over time, there will likely be a migration to the universal and exclusive use of SI units. The SI unit system has only seven base units from which all other units are derived (derived units are combinations of the base units). Tables listing the base units of the SI system, some derived SI units, and all of the units used in this text are included in this appendix. Some conversion factors to relate the SI units and other metric units to other units commonly used in the United States are also included.

The scientific community almost universally uses exponential notation for numbers. Exponential notation is explained in Chapter 1 and is used throughout this text. This appendix includes a table ("Exponential Notation and the Decimal System") for converting exponential numbers to their nonexponential equivalents. To multiply numbers expressed exponentially, add the exponents (superscript numbers). For example, $10^{-1} \times 10^{3} = 10^{2}$ (one-tenth of 1000 is equal to 100). To divide, subtract the exponents. The "SI Base Units" table also lists the common terms used to identify certain exponent values. For example, 1000 is equivalent to 10^{3}, or 1×10^{3}, and 1000 can be expressed as "one thousand," while other units can be prefixed by "kilo-" to express one thousand of the units; for example, 1 kilogram is equal to 1000 grams.

Note that the naming system for large and small numbers is different in different parts of the world. In many non-English-speaking countries (and formerly in England), the term billion does not mean 10^{9}. Instead, in this alternate system billion is 10^{12} and trillion is 10^{18}. This is one reason why it is always better to use exponential notation for large or small numbers.

SI Base Units

SI Base Units

Parameter	Unit	Symbol
Length	meter	m
Mass	gram	g
Time	second	s
Electrical current	ampere	A
Thermodynamic temperature	kelvin	K
Amount of substance	mole[a]	mol
Luminous intensity	candela	cd

[a]1 mol is equal to the molecular weight in grams

Examples of Derived SI Units

Derived SI Units

Derived Quantity	Name	Symbol
Area	square meter	m^2
Volume	cubic meter	m^3

Derived Quantity	Name	Symbol
Speed, velocity	meter per second	$m \cdot s^{-1}$
Acceleration	meter per second squared	$m \cdot s^{-2}$
Mass density	kilogram per cubic meter	$kg \cdot m^{-3}$
Specific volume	cubic meter per kilogram	$m^3 \cdot kg^{-1}$
Current density	ampere per square meter	$A \cdot m^{-2}$
Magnetic field strength	ampere per meter	$A \cdot m^{-1}$
Amount-of-substance concentration	mole per cubic meter	$mol \cdot m^{-3}$
Luminance	candela per square meter	$cd \cdot m^{-2}$
Mass fraction	kilogram per kilogram, which may be represented by the number 1	$kg \cdot kg^{-1} = 1$

SI Derived Units with Special Names and Symbols

Special Names and Symbols

Derived Quantity	Name	Symbol	Expression in Terms of Othe SI Units
Plane angle	radian	rad	—
Frequency	hertz	Hz	—
Force	newton	N	—
Pressure, stress	pascal	Pa	$N \cdot m^{-2}$
Energy, work, quantity of heat	joule	J	$N \cdot m$
Power, radiant flux	watt	W	$J \cdot s$
Celsius temperature	degree Celsius	°C	—
Luminous flux	lumen	lm	$cd \cdot sr$
Illuminance	lux	lx	$lm \cdot m^2$
Activity (of a radionuclide)	becquerel	Bq	—

For ease of understanding and convenience, a few SI derived units have been given special names and symbols. These are some that are relevant to this text.

Exponential Notation and the Decimal System

Exponential Notation

Value	Exponential Expression	Name	Prefix
1,000,000,000,000	10^{12}	1 trillion	tera-
1,000,000,000	10^9	1 billion	giga-
1,000,000	10^6	1 million	mega-

Value	Exponential Expression	Name	Prefix
1,000	10^3	1 thousand	kilo-
100	10^2	1 hundred	centa-
10	10^1	ten	deca-
1	10^0	one	uni-
0.1	10^{-1}	1 tenth	deci-
0.01	10^{-2}	1 hundredth	centi-
0.001	10^{-3}	1 thousandth	milli-
0.000,001	10^{-6}	1 millionth	micro-
0.000,000,001	10^{-9}	1 billionth	nano-
0.000,000,000,001	10^{-12}	1 trillionth	pico-

Conversions between Basic Units of Length, Mass, and Time

Length

1 kilometer = 1000 meters (m)
1 meter = 100 centimeters (cm)
1 centimeter = 10 millimeters (mm)
1 millimeter = 10 micrometers (mm)
1 micrometer = 10 nanometers (nm)

Mass

1 tonne (metric ton) = 1000 kilograms (kg)
1 kilogram = 1000 grams (g)
1 gram = 10 milligrams (mg)
1 milligram = 10 micrograms (mg)
1 microgram = 10 nanograms (ng)
1 nanogram = 10 picograms (pg)

Time

1 hour = 3600 seconds (s)

Units of Measurement Used in this Text

- Length/distance
  - **meter (m)** — ***SI base unit***
  - millimeter (mm)
  - centimeter (cm)
  - kilometer (km)
  - foot (ft.) — *obsolete nonmetric unit*
  - nautical mile (n.m.) — *equals 1.853 km (1.15 statute mi.) or 1 min (1/60 of a degree) of longitude*
- Mass
  - **kilogram (kg) — *SI base unit***
  - gram (g)
  - milligram (mg)
  - microgram (mg)

    - tonne (t) — *metric ton, equal to 1000 kg*
- Time
    - **second (s) — *SI base unit***
    - year (yr)
    - day (day)
    - hour (h)
    - minute (min)
- Area
    - meter squared (m^2) — *derived SI unit*
    - centimeter squared (cm^2)
    - kilometer squared (km^2)
- Volume
    - meter cubed, cubic meters (m^3) — *derived SI unit*
    - centimeter cubed (cm^3)
    - kilometer cubed (km^3)
    - liter (l) — *equals 100 cm^3, so 1 m^3 = 1000 l*
- Pressure
    - pascal (Pa) — *derived SI unit (not used in this text)*
    - kilogram per centimeter squared ($kg \cdot cm^{-2}$) — *equals 9.56 × 106 Pa*
    - atmosphere (atm) — *equals 1.03 $kg \cdot cm^{-2}$*
- Mass density (absolute density)
    - kilogram per cubic meter ($kg \cdot m^{-3}$) — *derived SI unit*
    - gram per cubic centimeter ($g \cdot cm^{-3}$)
- Speed/velocity
    - meters per second ($m \cdot s^{-1}$) — *derived SI unit*
    - centimeters per second ($cm \cdot s^{-1}$)
    - kilometers per hour ($km \cdot h^{-1}$) — *1 $km \cdot h^{-1}$ = 0.28 $m \cdot s^{-1}$= 28 $cm \cdot s^{-1}$*
    - kilometers per day ($km \cdot day^{-1}$)
- Acceleration
    - meters per second per second ($m \cdot s^{-2}$) — *derived SI unit*
- Temperature
    - degrees Celsius (°C)
- Energy and related unites
    - joule (J) — *derived SI unit*
    - joule per meter squared ($J \cdot m^{-2}$) — *used as measure of wave energy*
    - joule per gram ($J \cdot g^{-1}$) — *used as measure of latent heat*
    - joule per gram per degree Celsius ($J \cdot g^{-1} \cdot °C^{-1}$) — *used as measure of heat capacity*
    - calorie (cal) — *obsolete measure of energy*
    - calories per gram ($cal \cdot g^{-1}$) — *obsolete, used as measure of latent heat*
    - calories per gram per degree Celsius ($cal \cdot g^{-1} \cdot °C^{-1}$) — *obsolete, used as measure of heat capacity*
- Concentration
    - mole per cubic meter ($mol \cdot m^3$) — *derived SI unit (not used in this text)*
    - milligram per kilogram ($mg \cdot kg^{-1}$)
    - microgram per kilogram ($mg \cdot kg^{-1}$)
    - Notes:
        - *Concentrations are expressed as mass of dissolved (or constituent) substance in one unit mass of the solution (or combined mixture)*
        - *1 mol is defined as the molecular weight of a substance expressed in grams. To convert concentration in $mg \cdot kg^{-1}$ to $mol \cdot m^{-3}$, divide by the molecular weight of the dissolved substance (or constituent); then divide by the solution (or mixture) density in $kg \cdot m^3$ and multiply by 1000*
- Miscellaneous

- grams per year ($g \cdot yr^{-1}$) — *mass transport rate*
- cubic meters per second ($m^3 \cdot s^{-1}$) — *volume transport rate*
- practical salinity unit — *dimensionless ratio; no abbreviation, but sometimes listed as PSU*
- millions of years ago (mya) — *millions of years before the present date*
- before the common era (BCE) — *number of years before the year 1 of the Christian calendar, equivalent to BC*
- the common era (CE) — *number of years after the year 1 of the Christian calendar, equivalent to AD*

Additional Conversations: Metric to Other Units

Length

1 km = 0.62 mi.

1 km = 0.54 n.m.

1 km = 1093 yard

1 m = 39.4 in.

1 m = 3.28 ft.

1 cm = 0.394 in.

Mass

1 kg = 2.2 lb

1 t = 2205 lb

1 t = 1.10 U.S. tons

1 g = 0.035 ounce

Area

1 km^2 = 0.386 $mi.^2$

1 km^2 = 247.1 acres

1 m^2 = 10.7 $ft.^2$

1 cm^2 = 0.155 $in.^2$

Volume

1 m^3 = 25.3 $ft.^3$

1 m^3 = 264 U.S. gallons

1 cm^3 = 0.061 $in.^3$

1 l = 2.12 pints

Pressure

1 $kg \cdot cm^{-2}$ = 14.2 $lb \cdot in.^{-2}$

$1\ kg \cdot cm^{-2} = 0.97\ atm$

$1\ kg \cdot cm^{-2} = 956\ millibars$

$1\ kg \cdot cm^{-2} = 0.956\ bar$

$1\ kg \cdot cm^{-2} = 956{,}000\ Pa$

$1\ kg \cdot cm^{-2} = 29.0$ in. of mercury at 0°C

$1\ kg \cdot cm^{-2} = 737$ mm of mercury at 0°C

Speed

$1\ km \cdot h^{-1} = 0.62\ mi. \cdot h$

$1\ km \cdot h^{-1} = 0.54\ knot$

$1\ cm \cdot s^{-1} = 1.97\ ft. \cdot min^{-1}$

$1\ cm \cdot s^{-1} = 0.033\ ft. \cdot s^{-1}$

Temperature

0°C = 32°F

10°C = 50°F

20°C = 68°F

30°C = 86°F

40°C = 104°F

100°C = 212°F

This page titled Appendix 1 - Units and Converstion Factors is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

Appendix 2 - Dimensions of Earth and Ocean

The Earth

Earth's Dimensions

Space	Distance (km)
Radius at equator	6,378
Radius at poles	6,357
Average radius	6,371
Circumference at equator	40,077

Areas of Land and Oceans

Earth's Areas

Space	Area (km^2)
Land (29.22% of total)	149,000,000
Oceans and seas (70.78%)	361,000,000
Pacific Ocean (marginal seas included)	180,000,000
Atlantic Ocean (marginal seas included)	107,000,000
Indian Ocean (marginal seas included)	74,000,000
Ice sheets and glaciers	15,600,000
Land plus continental shelf	177,400,000
Land in the Northern Hemisphere	100,200,000
Ocean in the Northern Hemisphere	154,800,000
Land in the Southern Hemisphere	48,700,000
Ocean in the Southern Hemisphere	206,300,000

Percentage of Ocean Area Occupied by Ocean Provinces

Area of Earth's Ocean Features

Space	Percentage
Continental shelf and slope	15.30%
Continental rise	5.3%
Abyssal seafloor	41.80%
Volcanoes and volcanic ridges	3.1%
Oceanic ridges and rises	32.70%
Trenches	1.7%

Density and Mass of Earth's Spheres

Earth's Density and Mass

Space	Average Thickness or Radius (km)	Mean Density ($g \cdot cm^{-1}$)
Atmosphere	—	—
Oceans and seas	3.8	1.03
Ice sheets and glaciers	1.6	0.9
Continental crust (includes continental shelves)	35	2.8
Oceanic crust (excludes continental shelves)	8	2.9
Mantle	2881	4.53
Core	3473	10.7

Elevations and Depths

Earth's Highs and Lows

Space	Distance (m)
Average elevation of land	840
Average depth of oceans	3,800
Average depth of Pacific Ocean (marginal seas excluded)	3,940
Average depth of Atlantic Ocean (marginal seas excluded)	3,310
Average depth of Indian Ocean (marginal seas excluded)	3,840
Highest elevation of land (Mount Everest)	8,848
Greatest depth of oceans (Mariana Trench)	11,035

This page titled Appendix 2 - Dimensions of Earth and Ocean is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

Appendix 3 - Classification of Marine Organisms

Understanding life on the Earth and how this life has evolved and may evolve in the future in response to human influences requires that we understand the differences between the many forms of life on the Earth, and how and why the differences have developed. Fundamental to such studies is taxonomy, the classification of organisms to express their relationship to each other.

One fundamental concept of taxonomy is that organisms can be identified as belonging to a species. The definition of a species is generally accepted to be a population of organisms whose members interbreed under natural conditions and produce fertile offspring that are reproductively isolated from other such groups. A new definition based on genetics, the composition of DNA, may be developed in the future. The current definition has generally worked well, although there are difficulties, for example, when two populations that are capable of interbreeding are geographically separated and do not interbreed. In some such cases, scientists disagree as to whether these separate populations are of the same species or are different enough to be considered separate species. Populations of king salmon that breed in different rivers along the coasts of the United States and Canada are an example. Other difficulties arise because the members of a single species may look very different. For example, bulldogs, terriers, and poodles all belong to the same species. Conversely, organisms that appear to be very similar may belong to different species; red squirrels and gray squirrels are an example.

Millions of species on the Earth have been identified and studied, and certainly many more millions remain to be found, particularly in tropical rainforests and coral reefs. Species are arranged in a hierarchical classification similar to a human family tree. Taxonomists have established a series of levels for this hierarchy, from species at the bottom to kingdom, which was the top level until recent years, when a new top level—domain—was added. This new top level became needed when archaea were discovered. Initially placed in a kingdom along with bacteria, archaea were subsequently found to be entirely different from bacteria. In fact, archaea are different enough that they do not fit into any of the five kingdoms that were previously recognized and that are still listed in many texts. Although the new top level (domain) is not yet universally accepted, the generally accepted classification now identifies three such domains:

Three Domains of Living Organisms

- **Domain Bacteria – Most of the Known Prokaryotes**
  - *Although it is clear that the bacteria need to be separated into kingdoms, they are not well enough studied for this to have been done yet. At present, they are separated into five phyla.*
  - Phylum Proteobacteria: Nitrogen-fixing bacteria
  - Phylum Cyanobacteria: Photosynthetic bacteria (formerly blue-green algae)
  - Phylum Eubacteria: True gram-positive bacteria
  - Phylum Spirochetes: Spiral bacteria
  - Phylum Chlamydiae: Intracellular parasites
- **Domain Archaea – Prokaryotes of Extreme Environments**
  - Kingdom Crenarchaeota: Thermophiles
  - Kingdom Euryarchaeota: Methanogens and halophiles
  - Kingdom Korarchaeota: Some hot-springs microbes
- **Domain Eukarya – Eukaryotic Cells**
  - Protista (Informal Group): Single-celled organisms that have a nucleus and other internal structure in the cell (e.g., diatoms, dinoflagellates, foraminifera, radiolaria, marine algae, seaweeds)
  - Kingdom Fungi: Multicelled organisms that are not able to photosynthesize (or chemosynthesize)
  - Kingdom Plantae (or Metaphyta): Multicelled photosynthetic autotrophs (e.g., kelp)
  - Kingdom Animalia (or Metazoa): Multicelled heterotrophs (invertebrates and vertebrates)

Viruses have traditionally been considered to be non-living (since they cannot reproduce without a host).However, this view is now changing, and viruses may soon be recognized as life.In addition, there is now evidence that viruses may have preceded all of the other existing kingdoms, that they stand somewhere at the base of the tree of life, and may have evolved into cellular organisms.Some may even have devolved from a member of the 3 domains, losing (as no longer necessary) their prior cellular structure and ability to breed without a host, two of the characteristics recognized as generally defining life. As a result, it is likely that viruses will have to be found a place within the taxonomic hierarchy. Whether this will require reorganization of the 3 Domain structure shown in the table above to include a new Domain, a new Kingdom or lower order, or even a new classification level above Domain is not yet clear.

The taxonomic hierarchy generally has nine levels, although sometimes one or another level is omitted, sometimes intermediate levels are added within the hierarchy, and there is, as yet, no universal agreement on the use of the top level domains. The major levels in the hierarchy are as follows:

Domain

Kingdom

Phylum

Subphylum

Class

Order

Family

Genus

Species

The principal phyla and classes to which all common marine species belong, and some examples of each type are listed at the end of this appendix.

Scientific names of organisms are important for many reasons, one of which is that common names are often confusing. For example, the common name "red snapper" is used in the Gulf of Mexico for the species Lutjanus campechanus. In the Pacific Ocean the same common name, "red snapper," is used for at least three different species (*Lutjanus bohar, Lutjanus malabaricus*, and *Lutjanus gibbus*) that are related but very different from one another. This difference does not matter to the local populations who eat these fishes (indeed, when you order red snapper in a restaurant, you may be served any one of many species that may not even be snappers). However, scientists must be able to distinguish them in the literature, and they do so by using scientific names.

The scientific names given to species always consist of two parts: a genus name followed by a species name. For example, mussels are classified as follows:

- **Domain**: Eukarya
- **Kingdom**: Animalia (Metazoa)
- **Phylum**: Mollusca
- **Class**: Bivalvia
- **Order**: Mytiloida
- **Superfamily**: Mytilacea (this is good example of an intermediate level used in the taxonomic hierarchy for only some orders)
- **Family**: Mytilidae
- **Genus**: Mytilus

The common blue mussel of the east coast of North America is *Mytilus edulis*, and a different species, *Mytilus californianus*, is present only on the west coast.

These and other scientific names may seem complicated, but they are either Latin or Greek words modified to explain something that the taxonomist considers important in identifying the organism. For example, "*Mytilus*" is derived from the Greek word for sea mussels, "*californianus*" expresses that this species is present only on the West Coast; and "*edulis*" means "edible," a suitable term for this species that is so good to eat. Similarly, the genus name *Sargassum* is applied to species of algae present in the Sargasso Sea, and Enterococcus is the genus name for species of bacteria ("*-coccus*") that are present in vertebrate intestines ("*Entero-*").

Scientific names can be complicated and sometimes obscure, but in general, they follow an internationally accepted set of rules. Note that the convention is to use italics for scientific names and to capitalize the genus name only. In addition, shortening the genus name is acceptable when clear. For example, this text refers to mussels as *M. edulis* and *M. californianus* after the full names have been used once. There are other conventions as well. For example, "*sp.*" after the genus name refers to a single species of the genus whose species name is not known, and "*spp.*" after the genus name indicates that all (or many) of the

species of that genus are being referred to (e.g., "*Mytilus sp.*" refers to an unstated or unknown species of mussel, and "*Mytilus spp.*" refers to all or any species of mussel). The "*sp.*" and "*spp.*" are often omitted, as is done for simplicity in many cases in this text. Sometimes the genus name is followed by abbreviations such as "*cf.*" and "*aff.*" The first of these abbreviations ("*cf.*") means that the genus listed is thought to be correct and that the species is very similar to the species identified by the name following the genus name and "*cf.*" but there are enough differences to lead the investigator to believe that this may turn out to be a new, but very closely related, species or a subspecies. The second abbreviation ("*aff.*") means that the species is very similar to the species identified by the name following the genus name and "*aff.*" but it is sufficiently different that it is most likely an undescribed species similar to the one named.

Taxonomic Classification

Phyla that have no members in the marine environment, as well as phyla in which all species are extinct, are omitted from the following list. The list includes general descriptions of members of each classification and, in some cases, examples of common species and a listing of figures in this text that show a photograph of a member of this classification. The classification presented is not accepted by all taxonomists. Classifications, including the distinctions between phyla, continually change as more is learned.

Domain Archaea. Organisms that have no nucleus, predominantly single-celled. Occur primarily in extreme environments.

Kingdom Crenarchaeota. Thermophiles.

Kingdom Euryarchaeota. Methanogens and halophiles.

Kingdom Korarchaeota. Some hot-springs microbes.

Domain Bacteria. Organisms that have no nucleus, predominantly single-celled.

Phylum Proteobacteria. Nitrogen-fixing bacteria.

Phylum Cyanobacteria. Photosynthetic bacteria (formerly blue-green algae).

Phylum Eubacteria. True gram-positive bacteria.

Phylum Spirochetes. Spiral bacteria.

Phylum Chlamydiae. Intracellular parasites.

Domain Eukarya, Protista (*Informal*). Organisms with a nucleus confined by a membrane, predominantly single-celled. This group may eventually be broken up and reclassified.

Phylum Chrysophyta. Diatoms (Figs. 6-6; 12-16a), coccolithophores (Figs. 6-7; 12-16c), silicoflagellates.

Phylum Pyrrophyta. Autotrophic dinoflagellates (Fig. 12-16b), zooxanthellae.

Phylum Chlorophyta. Green algae (Fig. 13-4a,d; 15-13f).

Phylum Phaeophyta. Brown algae (Fig. 13-4a). Kelps (Figs. 12-4a,b; 15.3a,c; 15-7c), Sargassum (Figs. 15-8; 15-15), and others.

Phylum Rhodophyta. Red algae (Fig. 13-4a,b,e).

Phylum Zoomastigophora. Flagellated protozoa, including heterotrophic dinoflagellates.

Phylum Sarcodina. Amoebas and relatives, including foraminifera (Figs. 6-8a; 12-19d) and radiolaria (Figs. 6-9; 12-19e).

Phylum Ciliophora. Protozoa with cilia.

Domain Eukarya, Kingdom Fungi. Fungi, lichens.

Phylum Mycophyta. Marine fungi are primarily benthic decomposers. Marine lichens are primarily intertidal.

Domain Eukarya, Kingdom Plantae (Metaphyta). Autotrophic multicelled plants.

Phylum Tracheophyta. Plants with roots, stems, leaves, and special cells that transport nutrients and water.

Class Angiospermae. Flowering plants with seeds contained in a closed seedpod. Turtle grass (*Thalassia*; Fig. 14-12b), marsh grasses (*Spartina*; Fig. 14-12c).

Domain Eukarya, Kingdom Animalia (Metazoa). Multicelled heterotrophs (animals).

Phylum Placozoa. Amoeba-like multicelled animals.

Phylum Mesozoa. Wormlike parasites of cephalopods.

Phylum Porifera. Sponges (Fig. 14-28m).

Phylum Cnidaria (Coelenterata). Jellies and related organisms, all of which have stinging cells.

Class Hydrozoa. Polypoid colonial animals, most with a medusa-like stage. Includes Portuguese man-of-war (Fig. 12-20d).

Class Scyphozoa. Jellies with no (or reduced) polyp stage in life cycle. Medusa stage dominates (Fig. 12-20a,b).

Class Anthozoa. Zooanthids (Fig. 14-8h), anemones (Figs. 14-6c; 14-8g,i; 14-20a; 14-35a; 15-9b), sea pens (Fig. 14-6a,b), and corals (Figs. 13-12; 14-8a-f; 14-28a; 15-5e,f; 15-7; 15-9a) that have only a polypoid body form.

Phylum Ctenophora. Comb jellies (Fig. 12-20d), “sea gooseberries.” Predatory, predominantly planktonic.

Phylum Platyhelminthes. Flatworms, flukes, tapeworms. Many are parasitic; many others are free-living predators.

Phylum Nemertea. Ribbon worms. Benthic and pelagic.

Phylum Gnathostomulida. Microscopic, wormlike, meiofaunal (small organisms that live in the spaces between sediment grains).

Phylum Gastrotricha. Microscopic, ciliated, meiofaunal.

Phylum Rotifera. Ciliated, less than about 2 mm long. Most species freshwater. Planktonic or epibenthic.

Phylum Kinorhyncha. Small, spiny, segmented worms; meiofaunal.

Phylum Acanthocephala. Spiny-headed worms. All are intestinal parasites of vertebrates.

Phylum Entoprocta. Small polyplike suspension feeders.

Phylum Nematoda. Roundworms. Most are 1 to 3 mm; marine species are infaunal.

Phylum Bryozoa. Small moss animals that form encrusting or branching colonies on the seafloor.

Phylum Phoronida. Suspension-feeding tube worms. Infaunal; inhabit shallow and temperate sediments.

Phylum Brachiopoda. Lamp shells. Bivalve animals that superficially resemble clams. Mostly deep-water.

Phylum Mollusca. Soft-bodied animals with a mantle and muscular foot. Most species secrete a calcium carbonate shell.

Class Monoplacophora. Limpetlike shells, segmented bodies, abyssal only.

Class Polyplacophora. Chitons (Fig. 15-13d). Oval, flattened body covered by eight overlapping plates.

Class Aplacophora. Worm-shaped mollusks; soft-sediment infauna.

Class Gastropoda. Snails (Figs. 14-3b; 14-19b; 14-28h; 15-13a), limpets (Fig. 15-13b), abalones, pteropods (Figs. 6-8b; 14-3a), nudibranchs (Figs. 3-1c; 14-11c,d; 14-28d–g; 15-9f), and many others (Figs. 14-11b; 14-19c; 14-34s).

Class Bivalvia. Clams (Fig. 14-35d), scallops, oysters (Fig. 14-7b), mussels (Figs. 14-7a; 15-13e; 15-14). Bivalves (shell with two fitted segments). Mostly filter-feeding.

Class Scaphopoda. Tusk shells. Soft-sediment infauna.

Class Cephalopoda. Squid (Figs. 12-24a, 14-28i), octopi (Figs. 12-1, 14-18e; 15-18), cuttlefish (Figs. 12-24b–d, 14-1; 14-28j), *Nautilus* (Figs. 12-24e,f). No external shell except in *Nautilus* sp.

Phylum Priapulida. Small, wormlike, subtidal.

Phylum Sipuncula. Peanut worms, benthic, exclusively marine.

Phylum Echiura. Spoon worms. Spoon-shaped proboscis. Infaunal or live under rocks.

Phylum Annelida. Segmented worms. Mostly benthic. Includes feather duster worms (Fig. 14-6d), scale worms (Fig. 14-34r), and Christmas tree worms (Figs. 14-10c,d).

Phylum Tardigrada. Meiofaunal. Eight-legged. Can hibernate for long periods.

Phylum Pentastoma. Tongue worms. Parasites of vertebrates.

Phylum Pogonophora. Tube-dwelling worms with no digestive system. Absorb food through body wall. All species are marine. Mostly deep-water (Fig. 15-20b).

Phylum Echinodermata. Spiny-skinned animals. Most are benthic epifaunal or infaunal.

Class Asteroidea. Sea stars (Figs. 12-21c; 15-5b, 15-9e).

Class Ophiuroidea. Brittle stars, basket stars (Fig. 14-33e).

Class Echinoidea. Sea urchins (Figs. 14-10b; 14-11a; 14-18d; 14-35b; 15-5c; 15-10b), sand dollars (Fig. 15-5d), sea biscuits.

Class Holothuroidea. Sea cucumbers (Figs. 14-10a; 14-28l; 15-5a).

Class Crinoidea. Sea lilies and feather stars (Figs. 14-16a,b; 14-34g).

Phylum Chaetognatha. Arrowworms, stiff-bodied, mostly planktonic.

Phylum Hemichordata. Unsegmented infauna with primitive nerve cord. Acorn worms. All marine.

Phylum Arthropoda. Jointed, legged animals with segmented body covered by an exoskeleton.

Subphylum Crustacea. Copepods (Figs. 12-19a, 14-32a), krill (Fig. 14-2a), barnacles (Figs. 14-7c,d; 15-13c), amphipods, shrimp (Figs. 14-2b,c; 14-16a,b,j-l; 14-18a; 14-34h-q; 14-35c,e; 15-9d), lobster, crabs (Figs. 12-21a; 14-16f,g; 14-18c; 14-34a-e; 14-35a,b; 15-9c; 15-13g), euphausiids (Figs. 12-19b; 14-2a), isopods (Fig. 14-32b,c).

Subphylum Chelicerata. Horseshoe crabs, sea spiders.

Subphylum Uniramia. Insects. Only five species of a single genus are present in the ocean.

Phylum Chordata. Animals with a nerve cord and gills, gill slits, or lungs.

Subphylum Urochordata. Tunicates (Fig. 14-9), sea squirts, salps (Fig. 12-20e; 14-4).

Subphylum Cephalochordata. Lancelets, Amphioxus. Found in coarse temperate and tropical sediments.

Subphylum Vertebrata. Spinal column of vertebrae, internal skeleton, brain.

Class Agnatha. Jawless fishes. Cartilaginous skeletons. Lampreys, hagfishes.

Class Chondrichthyes. Cartilaginous skeletons. Rays (Fig. 12-23f,h,i), sharks (Fig. 12-23a-e,g; 14-25e), skates, sawfishes, chimeras.

Class Ostreichthyes. Bony fishes (Figs. 12-22; 14-13d; 14-14; 14-15; 14-16c-e,h,i; 14-17; 14-18b; 14-19a,d-f; 14-20b-f; 14-21; 14-22; 14-23; 14-24; 14-25a–d; 14-26; 14-27; 14-28b,c,k,n; 14-32; 14-34f,g; 14-35c; 15-7; 15-9g).

Class Amphibia. No marine species, but one species (Asian mud frog) is known to tolerate marine water. Frogs, toads, salamanders.

Class Reptilia. Turtles (Figs. 12-28a,b; 14-12d), sea snakes (Fig. 12-28c,d). One species of iguana (Fig. 12-28e,f) and one species of crocodile are marine.

Class Aves. Birds (Figs. 12-29; 16-9b). Many species live on and feed in the ocean, but all must return to land to breed.

Class Mammalia. Warm-blooded animals that have mammary glands and hair, bear live young.

Order Cetacea. Whales (Fig. 12-25a-d), porpoises, dolphins (Fig. 12-25e).

Order Sirenia. Sea cows. Manatee (Fig. 12-26), dugong.

Order Carnivora. Marine species in two suborders.

Suborder Pinnipedia. Seals (Figs. 12-27a,b; 15-16; 16-9a), sea lions (Fig. 12-27c), walrus (Fig. 12-27d).

Suborder Fissipedia. Sea otters (Figs. 12-27e; 15-10a).

Order Primates. Primates. Apes, human beings. Only one species is known to tolerate marine water—surfers, swimmers, scuba divers.

The phyla Porifera, Cnidaria, Platyhelminthes, Nematoda, Mollusca, Arthropoda, Arthropoda, and Chordata each have more than 10,000 known member species. The phyla Cyanobacteria, Chrysophyta, Pyrrophyta, Chlorophyta, Phaeophyta, Rhodophyta, Zoomastigophora. Sarcodina, Ciliophora, Mycophyta, and Echinodermata each have between approximately 1000 and 10,000 known member species. All other phyla listed have fewer than about 1000 known species. However, it is estimated that there may be between 1 and 10 million species in the oceans, of which only a few hundreds of thousands have yet been identified.

This page titled Appendix 3 - Classification of Marine Organisms is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

Appendix 4 - Milestones in the History of Ocean Study

4000 BCE Egyptians developed shipbuilding and ocean piloting capabilities, as well as trading on the Nile.

3800 BCE First maps showing water (river charts).

3000 BCE Evidence for seafaring between China and Taiwan that may have spurred colonization of the tropical Pacific Ocean islands by the Polynesians.

2000-500 BCE The Polynesians voyaged across the Pacific Ocean and settled all the major islands, first inhabiting Tonga, Samoa, ca. 1000 BCE. Micronesian stick chart used.

1200-850 BCE Phoenicians explored entire Mediterranean Sea and sailed into the Atlantic to West Africa and to Cornwall, England for trading.

900 BCE Greeks first use the term *okeanos*, a mythical god, to describe the great river that flowed in a circle around the Earth and the root of the present word ocean.

800 BCE First graphic aids to marine navigation.

600 BCE The Greek Pythagoreans assumed a spherical Earth.

> Thales of Miletus, a Greek natural philosopher, developed the origin of science as we define it today and he is alleged to have written a book on navigation.

450 BCE The Greek Herodotus compiled a map of the known world centered on the Mediterranean region.

325 BCE The Greek Pytheas explored the coasts of England, Norway, and perhaps Iceland. He developed a means of determining latitude from the angular distance of the North Star and proposed a connection between the phases of the moon and the tides.

The Greek philosopher, Aristotle, published *Meteorologica,* which described the geography of the Greek world, and *Historia Animalium* the first known treatise on marine biology—a catalog of marine organisms.

300 BCE Library founded at Alexandria by Alexander the Great, which served as the repository for scrolls from ships and land caravans; for the next 600 years, it was a maritime studies center, fostering advances in science, celestial navigation by a variety of Greek and Egyptian scholars.

276-192 BCE The Greek Eratosthenes, a scholar and librarian at Alexandria, calculated the circumference of a spherical Earth with remarkable accuracy using trigonometry noting the specific angle of sunlight that occurred at Alexandria and Syene (now called Aswan), Egypt. He also invented latitude and longitude lines based on landmarks.

127 BCE Hipparchus of Nicaea, a Greek mathematician, developed a 360 degree system and arranged latitude and longitude in regular grid by degrees.

54 BCE - 30 CE The Roman Seneca devised the hydrologic cycle to show that, despite the inflow of river water, the level of the ocean remained stable because of evaporation.

1st century CE Chinese invent first compass.

150 CE The Greek-Egyptian scientist, Claudius Ptolemy, added minutes and seconds to the latitude-longitude system and compiled a map of the Roman World, but erred in estimating Earth's circumference, which remained uncorrected for hundreds of years.

415 CE The Alexandrian library with an estimated 700,000 scrolls destroyed and the last librarian, Hypatia, murdered by religious mob; Earth considered flat again.

673-735 CE The English monk Bede published *De Temporum Ratione,* which discussed the lunar control of the tides and

recognized monthly tidal variations and the effect of wind drag on tidal height.

780 Viking raids begin.

800-1000 Hawaii colonized by Polynesians

950-1250 Medieval Warm Period

982 Eric the Red, a Norse chieftain, completed the first transatlantic voyage, reaching present-day Baffin Island, Canada.

995 Leif Ericson, son of Eric the Red, established the settlement of Vinland in what is now Newfoundland, Canada.

1000 Norwegian colonies established in North America

1050 The astrolabe, a navigation instrument used to measure the height of celestial bodies above the horizon, first arrived in Europe from the East.

1400-1900 The Little Ice Age.

1405 Admiral Cheng Ho, the commander of the Treasure Fleet of the Dragon Throne began his voyages for Emperor Chu Ti, ultimately ruling the South Pacific and Indian Ocean until 1433.

1420 Prince Henry the Navigator, of Portugal, founded a naval observatory and the first school for teaching navigation, astronomy, and cartography.

1452-1519 Leonardo da Vinci, the famous Italian scientist, observed, recorded and interpreted characteristics of currents and waves and noted that fossils in Italian mountains implied that sea level had been higher in the ancient past.

1492 Christopher Columbus rediscovered North America, sailing to the islands of the West Indies.

1497–1499 Vasco da Gama, a Portuguese navigator, journeyed to India by sea by rounding the Cape of Good Hope with four ships, thereby establishing the first all water trade route between Europe and India.

1500 The Portuguese navigator and explorer, Pedro Alvares Cabral, traveled westward in search of a route to India, but reached and explored Brazil.

1513 Juan Ponce de Leon described the swift and powerful Florida Current.

1513-1518 Vasco de Nunez de Balboa crossed the Isthmus of Panama and sailed in the Pacific Ocean.

1515 Peter Martyr proposed an origin for the Gulf Stream.

1519-1522 Ferdinand Magellan embarked on the first circumnavigation of the globe; the crew led by Sebastian del Cano completed the voyage. On the voyage, the Pacific Ocean was named.

1569 Geradus Mercator, the Flemish mapmaker, constructed a map projection of the world that was adapted to navigational charts.

1605 Bishop Resen of Copenhagen drew the first map of the Gulf Stream.

1609 Hugo Grotius, a Dutch legal scholar, published *Mare Liberum,* the foundation for all modern law of the sea.

1674 The British scientist Robert Boyle investigated the relation among temperature, salinity, and pressure of seawater with depth and reported his findings in *Observations and Experiments on the Saltiness of the Sea*.

1687 Sir Isaac Newton published *Principia Mathematica*, which includes an explanation of the operation of gravity and that tides are the result of gravitational attraction of the moon and sun.

1714 Gabriel D. Fahrenheit invents a temperature scale.

1725 Count Luigi Marsigli from Bologna, Italy compiled *Histoire Physique de la Mer*, the first book pertaining entirely to

oceanography, examining the geological formation of ocean basins along with the marine life that lived in the sea.

1733 English clockmaker John Harrison invented the first of his five accurate and durable nautical chronometers for determining longitude at sea.

1735 George Hadley proposed an explanation for the trade wind regime that involved a consideration of the Earth's rotation.

1740 Leonhard Euler, noted Swiss mathematician, developed an approximation of the three-body problem to calculate the magnitude of the lunar and solar attractive forces that generate ocean tides.

1742 Anders Celsius invents a temperature scale.

1758 Carolus Linnaeus publishes 10th edition of *Systema Naturae*, formalizing biological nomenclature.

1760 John Harrison invented the Number Four chronometer in his quest to solve the longitude problem.

Joseph Black proposed the concept of specific heat.

1768-1775 Captain James Cook commanded three major ocean voyages gathering extensive data on 1776-1780 geography, geology, biota, currents, tides, and water temperatures of all of the principal oceans.

1769 or 1770 The American scientist, Benjamin Franklin, published the first chart of the Gulf Stream (by Folger), which was used by ships to cross the North Atlantic Ocean faster.

1779 James Cook died in Hawaii.

1802 Nathaniel Bowditch of Massachusetts published the *New American Practical Navigator,* a navigational resource that continues to be revised and published to this day.

1805 Admiral Sir Francis Beaufort of the Royal Navy developed his 12-category wind force scale to aid mariners estimate wind speed; this scale was later modified.

1807 President Thomas Jefferson mandated coastal charting of the entire United States and established the U.S. Coast and Geodetic Survey, now known as the National Geodetic Survey.

1811 British chemist Sir Humphry Davy made the first gas hydrates in his laboratory.

1817-1818 Sir John Ross ventured into the Arctic Ocean to explore Baffin Island, where he sounded the bottom successfully (took the first deep-water and sediment samples) and recovered starfish and mud worms from a depth of 1.8 km.

1820 Alexander Marcet, a London physician, noted that the proportion of the chemical ingredients in seawater is unvarying in all ocean basins.

1831-1836 The epic five-year journey of Charles Darwin aboard the British research ship HMS *Beagle* led to a theory of atoll formation and later the theory of evolution by natural selection.

1835 Gaspard-Gustav de Coriolis, a French mathematician, published the first papers on an object's horizontal motion across the surface of the rotating Earth.

1836 William Henry Harvey, an Irish botanist, devised a taxonomy of seaweeds.

1838-1842 Departure of the United States Exploring Expedition led by Lieutenant Charles Wilkes on the flagship *Vincennes*. Six vessels mapping coastal areas, collected specimens.

1839-1843 Sir James Clark Ross led an expedition to Antarctica, recovering samples of deep-sea benthos down to a depth of 4.9 km.

1841, 1854 Sir Edward Forbes published *The History of British Star-Fishes* (1841) and then his *Distribution of Marine Life* (1854), in which he argued that sea life cannot exist below a depth of about 600 m (the so called azoic hypothesis).

1847 Hans Christian Oersted observed plankton.

1851 The first telegraph cable was laid across the Straits of Dover, stimulating new technologies to protect, raise and lower cables to the sea floor.

1853 The first international conference convened in Brussels by Lt. Matthew Fontaine Maury, USN for the purpose of establishing a uniform means of meteorological observations at sea.

1855 Matthew Fontaine Maury, USN compiled and standardized the wind and current data recorded in U.S. ship logs and summarized his findings in *The Physical Geography of the Sea*, credited as the first textbook of modern oceanography.

1859 Darwin's *Origin of Species* published.

1865 Fr. Pietro Angelo Secchi, a papal scientific advisor, perfected the Secchi disc for determination of the transparency of Mediterranean Sea water to sunlight.

1868-1870 Charles Wyville Thomson, aboard HMS *Lightning* and HMS *Porcupine*, made the first series of deep-sea temperature measurements and collected marine organisms from great depths, disproving Forbes' azoic hypothesis.

1870s William Ferrel, an American meteorologist, drew attention to the deflecting effect of the Earth's rotation on ocean currents (the Coriolis Effect).

1871 The U.S. Fish Commission was established with a modern laboratory at Woods Hole, MA.

1872-1876 Lead by Charles Wyville Thomson, *HMS Challenger* conducted worldwide scientific expeditions, collecting data and specimens that were later analyzed in the more than 50 volumes of the *Challenger Reports*.

1873 Charles Wyville Thomson published a general oceanography book called the *Depths of the Sea*.

1877-1880 Alexander Agassiz, an American naturalist, founded the first U.S. marine station, the Anderson School of Natural History, on Penikese Island, Buzzards Bay, MA.

1880 William Dittmar determined major salts in seawater.

1884-1901 USS *Albatross* was designed and constructed specifically to conduct scientific research at sea and undertook numerous oceanographic cruises.

1880s Development of the civil time zones and the ultimate specification of the Greenwich Mean Time (GMT) as the world-wide reference.

1888 The Marine Biological Laboratory was established at Woods Hole, MA.

1890 Alfred Thayer Mahan completed *The Influence of Sea Power upon History*.

1891 Sir John Murray and Alphonse Renard classified marine sediments.

1893 Discovery of Maunder sunspot minimum by astronomer E. Walter Maunder.

1893-1896 The Norwegian Fridtjof Nansen, while aboard the *Fram,* studied the circulation pattern of the Arctic Ocean and confirmed that a northern continent did not exist.

1900 On 8 September, a hurricane storm surge struck Galveston, TX. More than 8000 people died, mostly by drowning, in the most deadly natural disaster in U.S. history.

1902 Danish scientists with government backing established the International Council for the Exploration of the Sea (ICES) to investigate oceanographic conditions that affect North Atlantic fisheries. Council representatives were from Great Britain, Germany, Sweden, Norway, Denmark, Holland, and the Soviet Union.

1903 The Friday Harbor Oceanographic Laboratory was established at the University of Washington in Seattle.

The Laboratory that became the Scripps Institution of Biological Research, and later the Scripps Institution of Oceanography, was founded in San Diego, CA.

1905 Marconi perfected wireless telegraphy, which became important for communication on the ocean.

The Swedish oceanographer, V.W. Ekman, published his classic work on the cause of the spiral nature of wind driven ocean currents near the surface.

1906 The first Sonar type listening device was invented by Lewis Nixon in order to detect icebergs.

Prince Albert I of Monaco established the Muscée Océanographique.

1907 Bertram Boltwood calculated age of Earth by radioactive decay.

1911 Roald Amundsen was first at South Pole.

1912 Following the sinking of the *Titanic*, the International Ice Patrol was formed to monitor icebergs.

German meteorologist Alfred Wegener proposed his theory of continental drift.

1912 Scripps Institution allied with the University of California.

1918 Vilhelm Bjerknes and colleagues at the Bergen (Norway) school formulated theories of air masses, atmospheric fronts, and midlatitude storm systems.

1921 International Hydrographic Bureau founded.

1924 Discovery of the Southern Oscillation by Sir Gilbert Walker.

1925-1927 A German expedition aboard the research vessel *Meteor* studied the physical oceanography of the Atlantic Ocean, using an echo sounder extensively for the first time.

1929 British geologist Arthur Holmes proposed that convection in Earth's mantle was the driving force for continental drift. This explanation was revived in the 1960s by Harry Hess and Robert S. Dietz.

1930 The Woods Hole Oceanographic Institution was established on the southwestern shore of Cape Cod, MA.

1931 *Atlantis* launched.

1932 The International Whaling Commission was organized to collect data on whale species and to enforce voluntary regulations on whaling.

1937 *E.W. Scripps* launched.

1938 Serbian astronomer Milutin Milankovitch began work on his astronomical theory of the Ice Age.

1942 Harald Sverdrup, Richard Fleming, and Martin Johnson published the first modern oceanographic reference text, *The Oceans*, which is still consulted today.

1943 Jacques Cousteau and Emile Gagnan invented the scuba regulator and tank combination, the "aqualung" for diving underwater.

1946 On 1 April, a tsunami destroyed most of the waterfront of Hilo, Hawaii causing 159 deaths.

1949 Maurice Ewing formed the Lamont (later changed to Lamont Doherty) Geological Observatory at Columbia University in New York.

1957 Systematic monitoring of atmospheric carbon dioxide levels began at the Mauna Loa Observatory, Hawaii, under the direction of Charles D. Keeling.

1957-1958 The International Geophysical Year (IGY) was organized as an international effort to coordinate geophysical

investigations of the Earth, including the ocean.

1958 The first U.S. nuclear submarine, USS *Nautilus*, also made the first submerged transit of the Arctic ice pack, passing the geographic North Pole.

1959 Antarctic Treaty System adopted by 12 nations.

1959-1965 The International Indian Ocean Expedition was established under United Nations auspices to intensively investigate Indian Ocean oceanography.

1960 The bathyscaphe *Trieste* carrying Jacques Piccard and Don Walsh reached the bottom of the deepest (Mariana) trench (10,912 m, or 35,800 ft).

1960s Jacob Bjerknes developed concept of teleconnections involving El Niño events and the Southern Oscillation.

1962 Rachel Carson's book *Silent Spring* initiated the U.S. environmental movement.

1963 F.J. Vine and D.H. Matthews of Cambridge University interpreted sea floor magnetic field anomaly patterns in support of sea floor spreading.

1964 *Alvin*, the world's first deep-ocean research submersible was launched.

On 28 March, a tsunami struck the Seward, Alaska area, claiming more than 100 lives.

1965 Discovery of a natural deposit of methane hydrates under permafrost in Siberia.

1966 The U.S. Congress adopted the Sea Grant College and Programs Act to provide nonmilitary funding for marine science education and research.

1968 *Glomar Challenger* returns first deep-sea cores, indicating the age of the Earth's crust. The cores support theories of plate tectonics.

The U.S. National Science Foundation organized the Deep Sea Drilling project (DSDP) to core through the sediments and crust of the ocean bottom.

1969 Santa Barbara, CA, oil well blowout captured national attention.

Stratton Commission developed the first national ocean policy for the U.S.

1970 The U.S. government created the National Oceanic and Atmospheric Administration (NOAA) to oversee and coordinate government activities related to oceanography and meteorology.

Completion of the Aswan High Dam on the Nile River.

1970s The United Nations initiated the International Decade of Ocean Exploration (IDOE) to improve scientific knowledge of the ocean.

Saffir-Simpson Hurricane Intensity Scale rated hurricanes from 1 to 5 corresponding to increasing intensity. It was revised in 2010 and is now known as the Saffir Simpson Hurricane Wind Scale.

1970 John Tuzo Wilson, a Canadian geologist, proposed a cyclic model of tectonic revolution.

1971 Jason Morgan of Princeton University proposed hotspot volcanism.

1972 The Geochemical Ocean Sections Study (GEOSECS) was organized to study seawater chemistry, ocean circulation and mixing, and the biogeochemical recycling of chemical substances.

Marine Mammal Protection Act established a moratorium on the hunting of marine mammals in U.S. waters and by U.S. citizens on the high seas (reauthorized in 1994).

Provisions of the 1972 Marine Protection, Research and Sanctuaries Act authorized the President to designate national marine

sanctuaries in coastal waters of the continental shelf and in the Great Lakes (reauthorized in 2000).

1972 London Dumping Convention banned the discharge of waste from land into the ocean by ships or aircraft.

Coastal Zone Management Act passed into law whereby the federal government assists coastal states in managing and protecting coastal resources.

1973 U.S. Endangered Species Act became law.

1974 Project FAMOUS (French-American Mid-Ocean Undersea Study) mapped and sampled the Mid-Atlantic Ridge, a zone of seafloor spreading.

1974 F.S. Rowland and M.J. Molina first warned of the threat of CFCs to the stratospheric ozone shield.

1975 Sinking, on 10 November, of the ore carrier Edmund Fitzgerald on Lake Superior with the loss of her 29-member crew.

1976 Scientists discover indications of Milankovitch cycles in deep-sea sediment cores.

Magnuson-Stevens Fisheries and Conservation Act established fishery councils to manage the fishery resources in each region of the nation.

1977 The submersible *Alvin* finds hydrothermal vents in the Galápagos rift.

1978 *Seasat-A,* the first satellite dedicated to the remote sensing of the ocean, was launched.

Coastal Zone Color Scanner (CZCS) flown aboard NASA's Nimbus-7 satellite.

1979 NOAA's SLOSH numerical model developed to predict the location and height of a storm surge.

1981 The Burgess Shale declared a UNESCO World Heritage Site.

1982 The UN Convention on the Law of the Sea established, which granted jurisdiction over an exclusive economic zone (EEZ) to 151 coastal nations.

1983 The Chesapeake Bay Program became the nation's largest federal-state environmental restoration project; the nation's first estuary selected for restoration and protection.

1985 The scientific research vessel *JOIDES Resolution* replaced *Glomar Challenger* in Deep Sea Drilling Project.

Tropical Ocean-Global Atmosphere (TOGA) commenced a 10-year experiment in the equatorial Pacific, with the deployment of the TOGA Tropical Atmosphere Ocean (TAO) Array.

R. D. Ballard located wreck of *Titanic*.

1986 Moratorium on commercial whaling issued by the International Whaling Commission (IWC).

1987 Joint Global Ocean Flux Study (JGOFSS) founded to assess processes controlling carbon fluxes.

1988 The Intergovernmental Panel on Climate Change (IPCC) formed by the World Meteorological Organization and the UN Environmental Programme to evaluate the state of climate science as the basis for policy action.

Heinrich layers first described by German researcher Hartmut Heinrich. Sediment released to the floor of the North Atlantic during the melting of fleets of ice bergs.

1989 On 24 March, the supertanker *Exxon Valdez* ran aground on a reef in Prince William Sound, Alaska, producing the 2nd largest oil spill in U.S. history.

1990 The *JOIDES Resolution* retrieved a sediment sample estimated to be 170 million years old.

1991 JOI researchers bore to a depth of 2 km (1.24 mi) beneath the seafloor near the Galápagos Islands.

The NOAA National Weather Service Forecast Office in Miami, FL, began to issue outlooks for rip currents.

1992 The U.S.-French *TOPEX/Poseidon* satellite launched to monitor global ocean topography.

The Tropical Ocean Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (TOGA COARE) was conducted in western Pacific.

1993 Beginning of a series of ocean iron fertilization experiments.

1994 The Law of the Sea Treaty entered into force.

1995 *Keiko*, a small remotely controlled Japanese submersible, sets a new depth record, reaching 10,978 m (36,008 ft) in the Challenger Deep.

1998 UN declared the "Year of the Oceans" to increase awareness of the importance of the ocean.

Galileo spacecraft found possible evidence of an ocean on Jupiter's moon Europa.

Scientists from 30 nations began deploying Argo profiling floats in the world ocean.

2000 Beginning of 10-year international project, *Census of Marine Life*.

2004 On 26 December, tsunami originated off the west coast of Sumatra, Indonesia; death toll is estimated at 227,900.

U.S. Commission on Ocean Policy made recommendations for a new national ocean policy.

2005 On 28 August, Hurricane Katrina struck the Gulf coast of Louisiana and Mississippi. At least 1500 people lost their lives in the most destructive hurricane in U.S. history.

2007-09 Fourth International Polar Year (IPY).

2007 Modern Japanese drill ship, *Chikyu*, became fully operational.

Argo array is completed with over 3000 active profiling floats deployed in the world ocean.

2008 The U.S.-French Jason-2 satellite launched to continue monitoring global ocean topography.

2009 IceBridge, a six-year NASA mission, begins the largest airborne survey of polar ice.

2010 An explosion on 20 April at the Deepwater Horizon drilling platform, leased by BP, releases almost 5 million barrels of oil into the Gulf of Mexico before it is capped in mid-July.

The Census of Marine Life completes a decade of ocean life research.

The National Ocean Council is established by Executive Order.

2011 On 11 March, a tsunami originates off the east coast of Japan with a death toll of 15,093 in Japan and 9,093 persons missing.

2020 Discovery of microbial life deep within Earth's crust beneath ocean sediments.

2026 The UN High Seas Treaty, formally known as the Biodiversity Beyond National Jurisdiction (BBNJ) Agreement, officially enters into force in January.

This page titled Appendix 4 - Milestones in the History of Ocean Study is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

DETAILED LICENSING

Overview

Title: Introduction to Ocean Sciences 6e (Segar)

Webpages: 266

Applicable Restrictions: Noncommercial, No Derivatives

All licenses found:

- CC BY-NC-ND 4.0: 99.2% (264 pages)
- Undeclared: 0.8% (2 pages)

By Page

- Introduction to Ocean Sciences 6e (Segar) - *CC BY-NC-ND 4.0*
  - Front Matter - *CC BY-NC-ND 4.0*
    - TitlePage - *CC BY-NC-ND 4.0*
    - InfoPage - *CC BY-NC-ND 4.0*
    - Preface - *CC BY-NC-ND 4.0*
    - Table of Contents - *Undeclared*
    - Licensing - *CC BY-NC-ND 4.0*
  - 1: The Ocean Planet - *CC BY-NC-ND 4.0*
    - 1.1: The Ocean's and Earth's Environment - *CC BY-NC-ND 4.0*
    - 1.2: Oceans and the Origins of Life - *CC BY-NC-ND 4.0*
    - 1.3: How to Study Ocean Data - *CC BY-NC-ND 4.0*
    - 1.4: Chapter Summary - *CC BY-NC-ND 4.0*
      - 1.4.1: Key Terms - *CC BY-NC-ND 4.0*
      - 1.4.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 1.4.3: References - *CC BY-NC-ND 4.0*
  - 2: History and Importance of Ocean Studies - *CC BY-NC-ND 4.0*
    - 2.1: What is Oceanography - *CC BY-NC-ND 4.0*
    - 2.2: Exploration and Mapping - *CC BY-NC-ND 4.0*
    - 2.3: National and International Law Applied to Oceans - *CC BY-NC-ND 4.0*
    - 2.4: Ocean Resources - *CC BY-NC-ND 4.0*
    - 2.5: Chapter Summary - *CC BY-NC-ND 4.0*
      - 2.5.1: Key Terms - *CC BY-NC-ND 4.0*
      - 2.5.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 2.5.3: References - *CC BY-NC-ND 4.0*
  - 3: Studying the Oceans - *CC BY-NC-ND 4.0*
    - 3.1: Difficulties of Studying the Ocean Environment - *CC BY-NC-ND 4.0*
    - 3.2: The Unseen Domain - *CC BY-NC-ND 4.0*
    - 3.3: Bathymetry - *CC BY-NC-ND 4.0*
    - 3.4: Seafloor Sediments - *CC BY-NC-ND 4.0*
    - 3.5: Chemical and Physical Oceanography - *CC BY-NC-ND 4.0*
    - 3.6: Sampling Living Organisms in the Sea - *CC BY-NC-ND 4.0*
    - 3.7: Scuba, Manned, and Unmanned Submersibles - *CC BY-NC-ND 4.0*
    - 3.8: Satellites - *CC BY-NC-ND 4.0*
    - 3.9: Computers and Modeling - *CC BY-NC-ND 4.0*
    - 3.10: Chapter Summary - *CC BY-NC-ND 4.0*
      - 3.10.1: Key Terms - *CC BY-NC-ND 4.0*
      - 3.10.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 3.10.3: References - *CC BY-NC-ND 4.0*
  - 4: Plate Tectonics-Evolution of the Ocean Floor - *CC BY-NC-ND 4.0*
    - 4.1: Structure of the Earth - *CC BY-NC-ND 4.0*
    - 4.2: Plate Tectonics - *CC BY-NC-ND 4.0*
    - 4.3: Plate Boundaries - *CC BY-NC-ND 4.0*
    - 4.4: Convergent Plate Boundaries - *CC BY-NC-ND 4.0*
    - 4.5: Divergent Plate Boundaries - *CC BY-NC-ND 4.0*
    - 4.6: Transform Faults and Fracture Zones - *CC BY-NC-ND 4.0*
    - 4.7: Hot Spots - *CC BY-NC-ND 4.0*
    - 4.8: Plate Interiors - *CC BY-NC-ND 4.0*
    - 4.9: Sea-Level Change and Climate - *CC BY-NC-ND 4.0*
    - 4.10: Present-Day Oceans - *CC BY-NC-ND 4.0*
    - 4.11: Chapter Summary - *CC BY-NC-ND 4.0*
      - 4.11.1: Key Terms - *CC BY-NC-ND 4.0*
      - 4.11.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 4.11.3: References - *CC BY-NC-ND 4.0*
  - 5: Water and Seawater - *CC BY-NC-ND 4.0*
    - 5.1: Origins and Distribution of the Earth's Water

- *CC BY-NC-ND 4.0*
  - 5.2: The Water Molecule - *CC BY-NC-ND 4.0*
  - 5.3: The Dissolving Power of Water - *CC BY-NC-ND 4.0*
  - 5.4: Sources and Sinks of Chemicals Dissolved in Seawater - *CC BY-NC-ND 4.0*
  - 5.5: Salinity - *CC BY-NC-ND 4.0*
  - 5.6: Dissolved Chemicals in Seawater - *CC BY-NC-ND 4.0*
  - 5.7: Heat Properties of Water - *CC BY-NC-ND 4.0*
  - 5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater - *CC BY-NC-ND 4.0*
  - 5.9: Ice Formation - *CC BY-NC-ND 4.0*
  - 5.10: Surface Tension and Viscosity - *CC BY-NC-ND 4.0*
  - 5.11: Transmission of Light and Other Electromagnetic Radiation - *CC BY-NC-ND 4.0*
  - 5.12: Transmission of Sound - *CC BY-NC-ND 4.0*
  - 5.13: Chapter Summary - *CC BY-NC-ND 4.0*
    - 5.13.1: Key Terms - *CC BY-NC-ND 4.0*
    - 5.13.2: Chapter Questions - *CC BY-NC-ND 4.0*
    - 5.13.3: References - *CC BY-NC-ND 4.0*
- 6: Ocean Sediments - *CC BY-NC-ND 4.0*
  - 6.1: Sediments and Biogeochemical Cycles - *CC BY-NC-ND 4.0*
  - 6.2: Classification of Sediments - *CC BY-NC-ND 4.0*
  - 6.3: Lithogenous Sediments - *CC BY-NC-ND 4.0*
  - 6.4: Biogenous Sediments - *CC BY-NC-ND 4.0*
  - 6.5: Hydrogenous Sediments - *CC BY-NC-ND 4.0*
  - 6.6: Cosmogenous Sediments - *CC BY-NC-ND 4.0*
  - 6.7: Sediment Transport, Deposition, and Accumulation - *CC BY-NC-ND 4.0*
  - 6.8: Continental Margin Sediments - *CC BY-NC-ND 4.0*
  - 6.9: Distribution of Surface Sediments - *CC BY-NC-ND 4.0*
  - 6.10: The Sediment Historical Record - *CC BY-NC-ND 4.0*
  - 6.11: Chapter Summary - *CC BY-NC-ND 4.0*
    - 6.11.1: Key Terms - *CC BY-NC-ND 4.0*
    - 6.11.2: Chapter Questions - *CC BY-NC-ND 4.0*
    - 6.11.3: References - *CC BY-NC-ND 4.0*
- 7: Ocean-Atmosphere Interactions - *CC BY-NC-ND 4.0*
  - 7.1: The Atmosphere - *CC BY-NC-ND 4.0*
  - 7.2: Water Vapor in the Atmosphere - *CC BY-NC-ND 4.0*
  - 7.3: Water and Heat Budgets - *CC BY-NC-ND 4.0*
  - 7.4: Climatic Winds - *CC BY-NC-ND 4.0*
  - 7.5: Climate and Ocean Surface Water Properties - *CC BY-NC-ND 4.0*
  - 7.6: Interannual Climate Variations - *CC BY-NC-ND 4.0*
  - 7.7: Land-Ocean-Atmosphere Interactions - *CC BY-NC-ND 4.0*
  - 7.8: Global Climate Zones - *CC BY-NC-ND 4.0*
  - 7.9: Weather Systems - *CC BY-NC-ND 4.0*
  - 7.10: Local Weather Effects - *CC BY-NC-ND 4.0*
  - 7.11: Chapter Summary - *CC BY-NC-ND 4.0*
    - 7.11.1: Key Terms - *CC BY-NC-ND 4.0*
    - 7.11.2: Chapter Questions - *CC BY-NC-ND 4.0*
    - 7.11.3: References - *CC BY-NC-ND 4.0*
- 8: Ocean Circulation - *CC BY-NC-ND 4.0*
  - 8.1: Energy Sources - *CC BY-NC-ND 4.0*
  - 8.2: Wind-Driven Currents - *CC BY-NC-ND 4.0*
  - 8.3: Ekman Motion - *CC BY-NC-ND 4.0*
  - 8.4: Geostrophic Currents - *CC BY-NC-ND 4.0*
  - 8.5: Open-Ocean Surface Currents - *CC BY-NC-ND 4.0*
  - 8.6: Equatorial Surface Currents - *CC BY-NC-ND 4.0*
  - 8.7: High-Latitude Surface Currents - *CC BY-NC-ND 4.0*
  - 8.8: Upwelling and Downwelling - *CC BY-NC-ND 4.0*
  - 8.9: Coastal Currents - *CC BY-NC-ND 4.0*
  - 8.10: Eddies - *CC BY-NC-ND 4.0*
  - 8.11: Inertial Currents - *CC BY-NC-ND 4.0*
  - 8.12: Langmuir Circulation - *CC BY-NC-ND 4.0*
  - 8.13: Thermohaline Circulation - *CC BY-NC-ND 4.0*
  - 8.14: Ocean Circulation and Climate - *CC BY-NC-ND 4.0*
  - 8.15: Tracing Water Masses - *CC BY-NC-ND 4.0*
  - 8.16: Chapter Summary - *CC BY-NC-ND 4.0*
    - 8.16.1: Key Terms - *CC BY-NC-ND 4.0*
    - 8.16.2: Chapter Questions - *CC BY-NC-ND 4.0*
    - 8.16.3: References - *CC BY-NC-ND 4.0*
  - 8.17: Box 1- Westward Intensification - *CC BY-NC-ND 4.0*
  - 8.18: Box 2 - Equatorial Currents - *CC BY-NC-ND 4.0*
- 9: Waves - *CC BY-NC-ND 4.0*
  - 9.1: Complexity of Ocean Waves - *CC BY-NC-ND 4.0*
  - 9.2: Progressive Waves - *CC BY-NC-ND 4.0*
  - 9.3: What is Wave Motion? - *CC BY-NC-ND 4.0*
  - 9.4: Making Waves - *CC BY-NC-ND 4.0*
  - 9.5: Sea Development and Wave Height - *CC BY-NC-ND 4.0*
  - 9.6: Wave Dissipation - *CC BY-NC-ND 4.0*
  - 9.7: Deep-Water Waves - *CC BY-NC-ND 4.0*
  - 9.8: Waves in Shallow Water - *CC BY-NC-ND 4.0*
  - 9.9: Tsunamis - *CC BY-NC-ND 4.0*
  - 9.10: Internal Waves - *CC BY-NC-ND 4.0*

- 9.11: Rossby and Kelvin Waves - *CC BY-NC-ND 4.0*
- 9.12: Standing Waves - *CC BY-NC-ND 4.0*
- 9.13: Chapter Summary - *CC BY-NC-ND 4.0*
  - 9.13.1: Key Terms - *CC BY-NC-ND 4.0*
  - 9.13.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 9.13.3: References - *CC BY-NC-ND 4.0*

◦ 10: Tides - *CC BY-NC-ND 4.0*
- 10.1: Tide-Generating Forces - *CC BY-NC-ND 4.0*
- 10.2: Characteristics of Tides - *CC BY-NC-ND 4.0*
- 10.3: Tides on an Ocean Covered Earth - *CC BY-NC-ND 4.0*
- 10.4: Tides in the Earth's Oceans - *CC BY-NC-ND 4.0*
- 10.5: Tides in the Open Ocean - *CC BY-NC-ND 4.0*
- 10.6: Tidal Currents - *CC BY-NC-ND 4.0*
- 10.7: Tide Prediction - *CC BY-NC-ND 4.0*
- 10.8: Marine Species and the Rhythm of Tides - *CC BY-NC-ND 4.0*
- 10.9: Energy from Tides - *CC BY-NC-ND 4.0*
- 10.10: Chapter Summary - *CC BY-NC-ND 4.0*
  - 10.10.1: Key Terms - *CC BY-NC-ND 4.0*
  - 10.10.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 10.10.3: References - *CC BY-NC-ND 4.0*

◦ 11: Coasts - *CC BY-NC-ND 4.0*
- 11.1: Processes that Form and Modify the Coast - *CC BY-NC-ND 4.0*
- 11.2: Beaches - *CC BY-NC-ND 4.0*
- 11.3: Barrier Islands and Lagoons - *CC BY-NC-ND 4.0*
- 11.4: Beaches and Human Structures - *CC BY-NC-ND 4.0*
- 11.5: Coral Reefs and Atolls - *CC BY-NC-ND 4.0*
- 11.6: Wetlands - *CC BY-NC-ND 4.0*
- 11.7: Deltas - *CC BY-NC-ND 4.0*
- 11.8: Chapter Summary - *CC BY-NC-ND 4.0*
  - 11.8.1: Key Terms - *CC BY-NC-ND 4.0*
  - 11.8.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 11.8.3: References - *CC BY-NC-ND 4.0*

◦ 12: Foundations of Life in the Oceans - *CC BY-NC-ND 4.0*
- 12.1: How do we describe life? - *CC BY-NC-ND 4.0*
- 12.2: Production, Consumption, and Decomposition - *CC BY-NC-ND 4.0*
- 12.3: Microbes in Charge - *CC BY-NC-ND 4.0*
- 12.4: Studying Microbes in the Ocean - *CC BY-NC-ND 4.0*
- 12.5: Phototropic Primary Production and Light - *CC BY-NC-ND 4.0*
- 12.6: Primary Production and Nutrients - *CC BY-NC-ND 4.0*
- 12.7: Food Webs - *CC BY-NC-ND 4.0*
- 12.8: Geographic Variation in Primary Production - *CC BY-NC-ND 4.0*
- 12.9: Dissolved Oxygen and Carbon Dioxide - *CC BY-NC-ND 4.0*
- 12.10: Organic Carbon - *CC BY-NC-ND 4.0*
- 12.11: Biological Provinces and Zones - *CC BY-NC-ND 4.0*
- 12.12: Plankton - *CC BY-NC-ND 4.0*
- 12.13: Nekton - *CC BY-NC-ND 4.0*
- 12.14: Benthos - *CC BY-NC-ND 4.0*
- 12.15: Chapter Summary - *CC BY-NC-ND 4.0*
  - 12.15.1: Key Terms - *CC BY-NC-ND 4.0*
  - 12.15.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 12.15.3: References - *CC BY-NC-ND 4.0*

◦ 13: Coastal Oceans and Estuaries - *CC BY-NC-ND 4.0*
- 13.1: Special Characteristics of Coastal Oceans - *CC BY-NC-ND 4.0*
- 13.2: Nutrient Supply to the Coastal Photic Zone - *CC BY-NC-ND 4.0*
- 13.3: Seasonal Cycles - *CC BY-NC-ND 4.0*
- 13.4: Algal Blooms - *CC BY-NC-ND 4.0*
- 13.5: Fisheries - *CC BY-NC-ND 4.0*
- 13.6: Estuaries - *CC BY-NC-ND 4.0*
- 13.7: Chapter Summary - *CC BY-NC-ND 4.0*
  - 13.7.1: Key Terms - *CC BY-NC-ND 4.0*
  - 13.7.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 13.7.3: References - *CC BY-NC-ND 4.0*

◦ 14: Marine Ecology - *CC BY-NC-ND 4.0*
- 14.1: Ecological Requirements - *CC BY-NC-ND 4.0*
- 14.2: Habitats - *CC BY-NC-ND 4.0*
- 14.3: Feeding - *CC BY-NC-ND 4.0*
- 14.4: Hunting and Defense - *CC BY-NC-ND 4.0*
- 14.5: Selected Adaptations in Fishes - *CC BY-NC-ND 4.0*
- 14.6: Reproduction - *CC BY-NC-ND 4.0*
- 14.7: Associations - *CC BY-NC-ND 4.0*
- 14.8: Communication and Navigation - *CC BY-NC-ND 4.0*
- 14.9: Chapter Summary - *CC BY-NC-ND 4.0*
  - 14.9.1: Key Terms - *CC BY-NC-ND 4.0*
  - 14.9.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 14.9.3: References - *CC BY-NC-ND 4.0*

◦ 15: Ocean Ecosystems - *CC BY-NC-ND 4.0*
- 15.1: Communities and Niches - *CC BY-NC-ND 4.0*
- 15.2: Coral Reefs - *CC BY-NC-ND 4.0*
- 15.3: Kelp Forests - *CC BY-NC-ND 4.0*
- 15.4: Rocky Intertidal Communities - *CC BY-NC-ND 4.0*
- 15.5: Sargasso Sea - *CC BY-NC-ND 4.0*
- 15.6: Polar Regions - *CC BY-NC-ND 4.0*
- 15.7: Beyond the Sun's Light - *CC BY-NC-ND 4.0*

- 15.8: Hydrothermal Vents - *CC BY-NC-ND 4.0*
- 15.9: Chapter Summary - *CC BY-NC-ND 4.0*
  - 15.9.1: Key Terms - *CC BY-NC-ND 4.0*
  - 15.9.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 15.9.3: References - *CC BY-NC-ND 4.0*

◦ 16: Impacts of Humans on the Oceans - *CC BY-NC-ND 4.0*
- 16.1: Pollution versus Contamination - *CC BY-NC-ND 4.0*
- 16.2: Adverse Effects of Human Activities - *CC BY-NC-ND 4.0*
- 16.3: Toxicity - *CC BY-NC-ND 4.0*
- 16.4: Waste Disposal - *CC BY-NC-ND 4.0*
- 16.5: Ocean Pollution Issues - *CC BY-NC-ND 4.0*
- 16.6: Climate Change, Acidification, and Deoxygenation - *CC BY-NC-ND 4.0*
- 16.7: Risk Assessment and Management - *CC BY-NC-ND 4.0*
- 16.8: Chapter Summary - *CC BY-NC-ND 4.0*
  - 16.8.1: Key Terms - *CC BY-NC-ND 4.0*
  - 16.8.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 16.8.3: References - *CC BY-NC-ND 4.0*
- 16.9: Box 1 - Petroleum in the Marine Environment - *CC BY-NC-ND 4.0*
- 16.10: Box 2 - Sewage Wastes in the Marine Environment - *CC BY-NC-ND 4.0*
- 16.11: Box 3 - Radioactive Wastes in the Marine Environment - *CC BY-NC-ND 4.0*

◦ 17: Critical Concepts - *CC BY-NC-ND 4.0*
- 17.1: CC1 Density and Layering in Fluids - *CC BY-NC-ND 4.0*
- 17.2: CC2 Isostasy, Eustasy, and Sea Level - *CC BY-NC-ND 4.0*
- 17.3: CC3 Convection and Convection Cells - *CC BY-NC-ND 4.0*
- 17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension - *CC BY-NC-ND 4.0*
- 17.5: CC5 Transfer and Storage of Heat by Water - *CC BY-NC-ND 4.0*
- 17.6: CC6 Salinity, Temperature, Pressure, and Water Density - *CC BY-NC-ND 4.0*
- 17.7: CC7 Radioactivity and Age Dating - *CC BY-NC-ND 4.0*
- 17.8: CC8 Residence Time - *CC BY-NC-ND 4.0*
- 17.9: CC9 The Global Greenhouse Effect - *CC BY-NC-ND 4.0*
- 17.10: CC10 Modeling - *CC BY-NC-ND 4.0*
- 17.11: CC11 Chaos - *CC BY-NC-ND 4.0*
- 17.12: CC12 The Coriolis Effect - *CC BY-NC-ND 4.0*
- 17.13: CC13 Geostrophic Flow - *CC BY-NC-ND 4.0*
- 17.14: CC14 Phototrophy, Light, and Nutrients - *CC BY-NC-ND 4.0*
- 17.15: CC15 Food Chain Efficiency - *CC BY-NC-ND 4.0*
- 17.16: CC16 Maximum Sustainable Yield - *CC BY-NC-ND 4.0*
- 17.17: CC17 Species Diversity and Biodiversity - *CC BY-NC-ND 4.0*
- 17.18: CC18 Toxicity - *CC BY-NC-ND 4.0*
- 17.19: CC19 Metagenomics - *CC BY-NC-ND 4.0*
- 17.20: References - *CC BY-NC-ND 4.0*

◦ Back Matter - *CC BY-NC-ND 4.0*
- Index - *CC BY-NC-ND 4.0*
- Glossary - *CC BY-NC-ND 4.0*
- Appendix 1 - Units and Converstion Factors - *CC BY-NC-ND 4.0*
- Appendix 2 - Dimensions of Earth and Ocean - *CC BY-NC-ND 4.0*
- Appendix 3 - Classification of Marine Organisms - *CC BY-NC-ND 4.0*
- Appendix 4 - Milestones in the History of Ocean Study - *CC BY-NC-ND 4.0*
- Detailed Licensing - *CC BY-NC-ND 4.0*
- Detailed Licensing - *Undeclared*

This page titled Introduction to Ocean Sciences 6e (Segar) is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.

DETAILED LICENSING

Overview

Title: Introduction to Ocean Sciences 6e (Segar)

Webpages: 266

Applicable Restrictions: Noncommercial, No Derivatives

All licenses found:

- CC BY-NC-ND 4.0: 99.2% (264 pages)
- Undeclared: 0.8% (2 pages)

By Page

- Introduction to Ocean Sciences 6e (Segar) - *CC BY-NC-ND 4.0*
  - Front Matter - *CC BY-NC-ND 4.0*
    - TitlePage - *CC BY-NC-ND 4.0*
    - InfoPage - *CC BY-NC-ND 4.0*
    - Preface - *CC BY-NC-ND 4.0*
    - Table of Contents - *Undeclared*
    - Licensing - *CC BY-NC-ND 4.0*
  - 1: The Ocean Planet - *CC BY-NC-ND 4.0*
    - 1.1: The Ocean's and Earth's Environment - *CC BY-NC-ND 4.0*
    - 1.2: Oceans and the Origins of Life - *CC BY-NC-ND 4.0*
    - 1.3: How to Study Ocean Data - *CC BY-NC-ND 4.0*
    - 1.4: Chapter Summary - *CC BY-NC-ND 4.0*
      - 1.4.1: Key Terms - *CC BY-NC-ND 4.0*
      - 1.4.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 1.4.3: References - *CC BY-NC-ND 4.0*
  - 2: History and Importance of Ocean Studies - *CC BY-NC-ND 4.0*
    - 2.1: What is Oceanography - *CC BY-NC-ND 4.0*
    - 2.2: Exploration and Mapping - *CC BY-NC-ND 4.0*
    - 2.3: National and International Law Applied to Oceans - *CC BY-NC-ND 4.0*
    - 2.4: Ocean Resources - *CC BY-NC-ND 4.0*
    - 2.5: Chapter Summary - *CC BY-NC-ND 4.0*
      - 2.5.1: Key Terms - *CC BY-NC-ND 4.0*
      - 2.5.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 2.5.3: References - *CC BY-NC-ND 4.0*
  - 3: Studying the Oceans - *CC BY-NC-ND 4.0*
    - 3.1: Difficulties of Studying the Ocean Environment - *CC BY-NC-ND 4.0*
    - 3.2: The Unseen Domain - *CC BY-NC-ND 4.0*
    - 3.3: Bathymetry - *CC BY-NC-ND 4.0*
    - 3.4: Seafloor Sediments - *CC BY-NC-ND 4.0*
    - 3.5: Chemical and Physical Oceanography - *CC BY-NC-ND 4.0*
    - 3.6: Sampling Living Organisms in the Sea - *CC BY-NC-ND 4.0*
    - 3.7: Scuba, Manned, and Unmanned Submersibles - *CC BY-NC-ND 4.0*
    - 3.8: Satellites - *CC BY-NC-ND 4.0*
    - 3.9: Computers and Modeling - *CC BY-NC-ND 4.0*
    - 3.10: Chapter Summary - *CC BY-NC-ND 4.0*
      - 3.10.1: Key Terms - *CC BY-NC-ND 4.0*
      - 3.10.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 3.10.3: References - *CC BY-NC-ND 4.0*
  - 4: Plate Tectonics-Evolution of the Ocean Floor - *CC BY-NC-ND 4.0*
    - 4.1: Structure of the Earth - *CC BY-NC-ND 4.0*
    - 4.2: Plate Tectonics - *CC BY-NC-ND 4.0*
    - 4.3: Plate Boundaries - *CC BY-NC-ND 4.0*
    - 4.4: Convergent Plate Boundaries - *CC BY-NC-ND 4.0*
    - 4.5: Divergent Plate Boundaries - *CC BY-NC-ND 4.0*
    - 4.6: Transform Faults and Fracture Zones - *CC BY-NC-ND 4.0*
    - 4.7: Hot Spots - *CC BY-NC-ND 4.0*
    - 4.8: Plate Interiors - *CC BY-NC-ND 4.0*
    - 4.9: Sea-Level Change and Climate - *CC BY-NC-ND 4.0*
    - 4.10: Present-Day Oceans - *CC BY-NC-ND 4.0*
    - 4.11: Chapter Summary - *CC BY-NC-ND 4.0*
      - 4.11.1: Key Terms - *CC BY-NC-ND 4.0*
      - 4.11.2: Chapter Questions - *CC BY-NC-ND 4.0*
      - 4.11.3: References - *CC BY-NC-ND 4.0*
  - 5: Water and Seawater - *CC BY-NC-ND 4.0*
    - 5.1: Origins and Distribution of the Earth's Water

- *CC BY-NC-ND 4.0*
    - 5.2: The Water Molecule - *CC BY-NC-ND 4.0*
    - 5.3: The Dissolving Power of Water - *CC BY-NC-ND 4.0*
    - 5.4: Sources and Sinks of Chemicals Dissolved in Seawater - *CC BY-NC-ND 4.0*
    - 5.5: Salinity - *CC BY-NC-ND 4.0*
    - 5.6: Dissolved Chemicals in Seawater - *CC BY-NC-ND 4.0*
    - 5.7: Heat Properties of Water - *CC BY-NC-ND 4.0*
    - 5.8: Effects of Pressure, Temperature, and Dissolved Salts on Seawater - *CC BY-NC-ND 4.0*
    - 5.9: Ice Formation - *CC BY-NC-ND 4.0*
    - 5.10: Surface Tension and Viscosity - *CC BY-NC-ND 4.0*
    - 5.11: Transmission of Light and Other Electromagnetic Radiation - *CC BY-NC-ND 4.0*
    - 5.12: Transmission of Sound - *CC BY-NC-ND 4.0*
    - 5.13: Chapter Summary - *CC BY-NC-ND 4.0*
        - 5.13.1: Key Terms - *CC BY-NC-ND 4.0*
        - 5.13.2: Chapter Questions - *CC BY-NC-ND 4.0*
        - 5.13.3: References - *CC BY-NC-ND 4.0*
- 6: Ocean Sediments - *CC BY-NC-ND 4.0*
    - 6.1: Sediments and Biogeochemical Cycles - *CC BY-NC-ND 4.0*
    - 6.2: Classification of Sediments - *CC BY-NC-ND 4.0*
    - 6.3: Lithogenous Sediments - *CC BY-NC-ND 4.0*
    - 6.4: Biogenous Sediments - *CC BY-NC-ND 4.0*
    - 6.5: Hydrogenous Sediments - *CC BY-NC-ND 4.0*
    - 6.6: Cosmogenous Sediments - *CC BY-NC-ND 4.0*
    - 6.7: Sediment Transport, Deposition, and Accumulation - *CC BY-NC-ND 4.0*
    - 6.8: Continental Margin Sediments - *CC BY-NC-ND 4.0*
    - 6.9: Distribution of Surface Sediments - *CC BY-NC-ND 4.0*
    - 6.10: The Sediment Historical Record - *CC BY-NC-ND 4.0*
    - 6.11: Chapter Summary - *CC BY-NC-ND 4.0*
        - 6.11.1: Key Terms - *CC BY-NC-ND 4.0*
        - 6.11.2: Chapter Questions - *CC BY-NC-ND 4.0*
        - 6.11.3: References - *CC BY-NC-ND 4.0*
- 7: Ocean-Atmosphere Interactions - *CC BY-NC-ND 4.0*
    - 7.1: The Atmosphere - *CC BY-NC-ND 4.0*
    - 7.2: Water Vapor in the Atmosphere - *CC BY-NC-ND 4.0*
    - 7.3: Water and Heat Budgets - *CC BY-NC-ND 4.0*
    - 7.4: Climatic Winds - *CC BY-NC-ND 4.0*
    - 7.5: Climate and Ocean Surface Water Properties - *CC BY-NC-ND 4.0*
    - 7.6: Interannual Climate Variations - *CC BY-NC-ND 4.0*
    - 7.7: Land-Ocean-Atmosphere Interactions - *CC BY-NC-ND 4.0*
    - 7.8: Global Climate Zones - *CC BY-NC-ND 4.0*
    - 7.9: Weather Systems - *CC BY-NC-ND 4.0*
    - 7.10: Local Weather Effects - *CC BY-NC-ND 4.0*
    - 7.11: Chapter Summary - *CC BY-NC-ND 4.0*
        - 7.11.1: Key Terms - *CC BY-NC-ND 4.0*
        - 7.11.2: Chapter Questions - *CC BY-NC-ND 4.0*
        - 7.11.3: References - *CC BY-NC-ND 4.0*
- 8: Ocean Circulation - *CC BY-NC-ND 4.0*
    - 8.1: Energy Sources - *CC BY-NC-ND 4.0*
    - 8.2: Wind-Driven Currents - *CC BY-NC-ND 4.0*
    - 8.3: Ekman Motion - *CC BY-NC-ND 4.0*
    - 8.4: Geostrophic Currents - *CC BY-NC-ND 4.0*
    - 8.5: Open-Ocean Surface Currents - *CC BY-NC-ND 4.0*
    - 8.6: Equatorial Surface Currents - *CC BY-NC-ND 4.0*
    - 8.7: High-Latitude Surface Currents - *CC BY-NC-ND 4.0*
    - 8.8: Upwelling and Downwelling - *CC BY-NC-ND 4.0*
    - 8.9: Coastal Currents - *CC BY-NC-ND 4.0*
    - 8.10: Eddies - *CC BY-NC-ND 4.0*
    - 8.11: Inertial Currents - *CC BY-NC-ND 4.0*
    - 8.12: Langmuir Circulation - *CC BY-NC-ND 4.0*
    - 8.13: Thermohaline Circulation - *CC BY-NC-ND 4.0*
    - 8.14: Ocean Circulation and Climate - *CC BY-NC-ND 4.0*
    - 8.15: Tracing Water Masses - *CC BY-NC-ND 4.0*
    - 8.16: Chapter Summary - *CC BY-NC-ND 4.0*
        - 8.16.1: Key Terms - *CC BY-NC-ND 4.0*
        - 8.16.2: Chapter Questions - *CC BY-NC-ND 4.0*
        - 8.16.3: References - *CC BY-NC-ND 4.0*
    - 8.17: Box 1- Westward Intensification - *CC BY-NC-ND 4.0*
    - 8.18: Box 2 - Equatorial Currents - *CC BY-NC-ND 4.0*
- 9: Waves - *CC BY-NC-ND 4.0*
    - 9.1: Complexity of Ocean Waves - *CC BY-NC-ND 4.0*
    - 9.2: Progressive Waves - *CC BY-NC-ND 4.0*
    - 9.3: What is Wave Motion? - *CC BY-NC-ND 4.0*
    - 9.4: Making Waves - *CC BY-NC-ND 4.0*
    - 9.5: Sea Development and Wave Height - *CC BY-NC-ND 4.0*
    - 9.6: Wave Dissipation - *CC BY-NC-ND 4.0*
    - 9.7: Deep-Water Waves - *CC BY-NC-ND 4.0*
    - 9.8: Waves in Shallow Water - *CC BY-NC-ND 4.0*
    - 9.9: Tsunamis - *CC BY-NC-ND 4.0*
    - 9.10: Internal Waves - *CC BY-NC-ND 4.0*

- 9.11: Rossby and Kelvin Waves - *CC BY-NC-ND 4.0*
- 9.12: Standing Waves - *CC BY-NC-ND 4.0*
- 9.13: Chapter Summary - *CC BY-NC-ND 4.0*
  - 9.13.1: Key Terms - *CC BY-NC-ND 4.0*
  - 9.13.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 9.13.3: References - *CC BY-NC-ND 4.0*

◦ 10: Tides - *CC BY-NC-ND 4.0*
- 10.1: Tide-Generating Forces - *CC BY-NC-ND 4.0*
- 10.2: Characteristics of Tides - *CC BY-NC-ND 4.0*
- 10.3: Tides on an Ocean Covered Earth - *CC BY-NC-ND 4.0*
- 10.4: Tides in the Earth's Oceans - *CC BY-NC-ND 4.0*
- 10.5: Tides in the Open Ocean - *CC BY-NC-ND 4.0*
- 10.6: Tidal Currents - *CC BY-NC-ND 4.0*
- 10.7: Tide Prediction - *CC BY-NC-ND 4.0*
- 10.8: Marine Species and the Rhythm of Tides - *CC BY-NC-ND 4.0*
- 10.9: Energy from Tides - *CC BY-NC-ND 4.0*
- 10.10: Chapter Summary - *CC BY-NC-ND 4.0*
  - 10.10.1: Key Terms - *CC BY-NC-ND 4.0*
  - 10.10.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 10.10.3: References - *CC BY-NC-ND 4.0*

◦ 11: Coasts - *CC BY-NC-ND 4.0*
- 11.1: Processes that Form and Modify the Coast - *CC BY-NC-ND 4.0*
- 11.2: Beaches - *CC BY-NC-ND 4.0*
- 11.3: Barrier Islands and Lagoons - *CC BY-NC-ND 4.0*
- 11.4: Beaches and Human Structures - *CC BY-NC-ND 4.0*
- 11.5: Coral Reefs and Atolls - *CC BY-NC-ND 4.0*
- 11.6: Wetlands - *CC BY-NC-ND 4.0*
- 11.7: Deltas - *CC BY-NC-ND 4.0*
- 11.8: Chapter Summary - *CC BY-NC-ND 4.0*
  - 11.8.1: Key Terms - *CC BY-NC-ND 4.0*
  - 11.8.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 11.8.3: References - *CC BY-NC-ND 4.0*

◦ 12: Foundations of Life in the Oceans - *CC BY-NC-ND 4.0*
- 12.1: How do we describe life? - *CC BY-NC-ND 4.0*
- 12.2: Production, Consumption, and Decomposition - *CC BY-NC-ND 4.0*
- 12.3: Microbes in Charge - *CC BY-NC-ND 4.0*
- 12.4: Studying Microbes in the Ocean - *CC BY-NC-ND 4.0*
- 12.5: Phototropic Primary Production and Light - *CC BY-NC-ND 4.0*
- 12.6: Primary Production and Nutrients - *CC BY-NC-ND 4.0*
- 12.7: Food Webs - *CC BY-NC-ND 4.0*
- 12.8: Geographic Variation in Primary Production - *CC BY-NC-ND 4.0*
- 12.9: Dissolved Oxygen and Carbon Dioxide - *CC BY-NC-ND 4.0*
- 12.10: Organic Carbon - *CC BY-NC-ND 4.0*
- 12.11: Biological Provinces and Zones - *CC BY-NC-ND 4.0*
- 12.12: Plankton - *CC BY-NC-ND 4.0*
- 12.13: Nekton - *CC BY-NC-ND 4.0*
- 12.14: Benthos - *CC BY-NC-ND 4.0*
- 12.15: Chapter Summary - *CC BY-NC-ND 4.0*
  - 12.15.1: Key Terms - *CC BY-NC-ND 4.0*
  - 12.15.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 12.15.3: References - *CC BY-NC-ND 4.0*

◦ 13: Coastal Oceans and Estuaries - *CC BY-NC-ND 4.0*
- 13.1: Special Characteristics of Coastal Oceans - *CC BY-NC-ND 4.0*
- 13.2: Nutrient Supply to the Coastal Photic Zone - *CC BY-NC-ND 4.0*
- 13.3: Seasonal Cycles - *CC BY-NC-ND 4.0*
- 13.4: Algal Blooms - *CC BY-NC-ND 4.0*
- 13.5: Fisheries - *CC BY-NC-ND 4.0*
- 13.6: Estuaries - *CC BY-NC-ND 4.0*
- 13.7: Chapter Summary - *CC BY-NC-ND 4.0*
  - 13.7.1: Key Terms - *CC BY-NC-ND 4.0*
  - 13.7.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 13.7.3: References - *CC BY-NC-ND 4.0*

◦ 14: Marine Ecology - *CC BY-NC-ND 4.0*
- 14.1: Ecological Requirements - *CC BY-NC-ND 4.0*
- 14.2: Habitats - *CC BY-NC-ND 4.0*
- 14.3: Feeding - *CC BY-NC-ND 4.0*
- 14.4: Hunting and Defense - *CC BY-NC-ND 4.0*
- 14.5: Selected Adaptations in Fishes - *CC BY-NC-ND 4.0*
- 14.6: Reproduction - *CC BY-NC-ND 4.0*
- 14.7: Associations - *CC BY-NC-ND 4.0*
- 14.8: Communication and Navigation - *CC BY-NC-ND 4.0*
- 14.9: Chapter Summary - *CC BY-NC-ND 4.0*
  - 14.9.1: Key Terms - *CC BY-NC-ND 4.0*
  - 14.9.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 14.9.3: References - *CC BY-NC-ND 4.0*

◦ 15: Ocean Ecosystems - *CC BY-NC-ND 4.0*
- 15.1: Communities and Niches - *CC BY-NC-ND 4.0*
- 15.2: Coral Reefs - *CC BY-NC-ND 4.0*
- 15.3: Kelp Forests - *CC BY-NC-ND 4.0*
- 15.4: Rocky Intertidal Communities - *CC BY-NC-ND 4.0*
- 15.5: Sargasso Sea - *CC BY-NC-ND 4.0*
- 15.6: Polar Regions - *CC BY-NC-ND 4.0*
- 15.7: Beyond the Sun's Light - *CC BY-NC-ND 4.0*

- 15.8: Hydrothermal Vents - *CC BY-NC-ND 4.0*
- 15.9: Chapter Summary - *CC BY-NC-ND 4.0*
  - 15.9.1: Key Terms - *CC BY-NC-ND 4.0*
  - 15.9.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 15.9.3: References - *CC BY-NC-ND 4.0*

◦ 16: Impacts of Humans on the Oceans - *CC BY-NC-ND 4.0*
- 16.1: Pollution versus Contamination - *CC BY-NC-ND 4.0*
- 16.2: Adverse Effects of Human Activities - *CC BY-NC-ND 4.0*
- 16.3: Toxicity - *CC BY-NC-ND 4.0*
- 16.4: Waste Disposal - *CC BY-NC-ND 4.0*
- 16.5: Ocean Pollution Issues - *CC BY-NC-ND 4.0*
- 16.6: Climate Change, Acidification, and Deoxygenation - *CC BY-NC-ND 4.0*
- 16.7: Risk Assessment and Management - *CC BY-NC-ND 4.0*
- 16.8: Chapter Summary - *CC BY-NC-ND 4.0*
  - 16.8.1: Key Terms - *CC BY-NC-ND 4.0*
  - 16.8.2: Chapter Questions - *CC BY-NC-ND 4.0*
  - 16.8.3: References - *CC BY-NC-ND 4.0*
- 16.9: Box 1 - Petroleum in the Marine Environment - *CC BY-NC-ND 4.0*
- 16.10: Box 2 - Sewage Wastes in the Marine Environment - *CC BY-NC-ND 4.0*
- 16.11: Box 3 - Radioactive Wastes in the Marine Environment - *CC BY-NC-ND 4.0*

◦ 17: Critical Concepts - *CC BY-NC-ND 4.0*
- 17.1: CC1 Density and Layering in Fluids - *CC BY-NC-ND 4.0*
- 17.2: CC2 Isostasy, Eustasy, and Sea Level - *CC BY-NC-ND 4.0*
- 17.3: CC3 Convection and Convection Cells - *CC BY-NC-ND 4.0*
- 17.4: CC4 Particle Size, Sinking, Deposition, and Resuspension - *CC BY-NC-ND 4.0*
- 17.5: CC5 Transfer and Storage of Heat by Water - *CC BY-NC-ND 4.0*
- 17.6: CC6 Salinity, Temperature, Pressure, and Water Density - *CC BY-NC-ND 4.0*
- 17.7: CC7 Radioactivity and Age Dating - *CC BY-NC-ND 4.0*
- 17.8: CC8 Residence Time - *CC BY-NC-ND 4.0*
- 17.9: CC9 The Global Greenhouse Effect - *CC BY-NC-ND 4.0*
- 17.10: CC10 Modeling - *CC BY-NC-ND 4.0*
- 17.11: CC11 Chaos - *CC BY-NC-ND 4.0*
- 17.12: CC12 The Coriolis Effect - *CC BY-NC-ND 4.0*
- 17.13: CC13 Geostrophic Flow - *CC BY-NC-ND 4.0*
- 17.14: CC14 Phototrophy, Light, and Nutrients - *CC BY-NC-ND 4.0*
- 17.15: CC15 Food Chain Efficiency - *CC BY-NC-ND 4.0*
- 17.16: CC16 Maximum Sustainable Yield - *CC BY-NC-ND 4.0*
- 17.17: CC17 Species Diversity and Biodiversity - *CC BY-NC-ND 4.0*
- 17.18: CC18 Toxicity - *CC BY-NC-ND 4.0*
- 17.19: CC19 Metagenomics - *CC BY-NC-ND 4.0*
- 17.20: References - *CC BY-NC-ND 4.0*

◦ Back Matter - *CC BY-NC-ND 4.0*
- Index - *CC BY-NC-ND 4.0*
- Glossary - *CC BY-NC-ND 4.0*
- Appendix 1 - Units and Converstion Factors - *CC BY-NC-ND 4.0*
- Appendix 2 - Dimensions of Earth and Ocean - *CC BY-NC-ND 4.0*
- Appendix 3 - Classification of Marine Organisms - *CC BY-NC-ND 4.0*
- Appendix 4 - Milestones in the History of Ocean Study - *CC BY-NC-ND 4.0*
- Detailed Licensing - *CC BY-NC-ND 4.0*
- Detailed Licensing - *Undeclared*

This page titled Introduction to Ocean Sciences 6e (Segar) is shared under a CC BY-NC-ND 4.0 license and was authored, remixed, and/or curated by Douglas A. Segar.