OCEAN-101: FUNDAMENTALS OF OCEANOGRAPHY



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Fundamentals of Oceanography (Keddy)

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Glossary

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CHAPTER OVERVIEW

1: Introduction to Oceanography

1.1: Why should we study the Oceans?

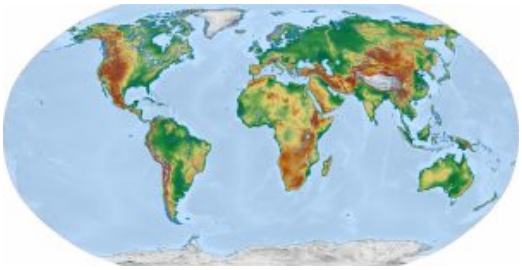
1.2: Types of Oceanographic Study

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1.1: Why should we study the Oceans?

Welcome to oceanography, or the study of the oceans. There are other terms for this science, including marine science or ocean science. Regardless of the term used, the study of the oceans is a vast science that encompasses all aspects of the largest geographic feature on our planet- our ocean! Let's start out by considering the importance of our oceans.

Recently, many people have been making a huge fuss about the health of our ocean, from cleaning it up to climate change. They talk about all of these different things: ocean warming, ocean acidification, overfishing, pollution, and coral bleaching. They tell us the oceans are in grave danger. So, yeah, all of this sounds really bad but honestly what does it all mean? More importantly, why should it matter to you?



Physical Map of the World under CC 4.0

Looking at a map you can tell that the ocean takes up the majority of our planets surface area, and when calculated it covers a whopping 72% of the earths surface! The ocean also provides over 50% of the earth's oxygen and stores 50 times more carbon dioxide than the atmosphere. It also plays a huge role in regulating our planet's climate and weather patterns. If you want to read a little more about all of this check this article out.

If all of that doesn't convince you of the importance of the ocean, think about all the fun that people can have involving the ocean and the beach. There's swimming, surfing, snorkeling, SCUBA diving, boating, parasailing, the list goes on and on!

The ocean also plays a pretty big part of the economies of the world. You may not realize how important the ocean is for coastal communities in developing countries. For example just in the United States the ocean produces about 282 million dollars per year and creates about 3 million jobs. It also provides a mode for 90% of all trade transportation in the world. Lastly, we use the ocean for a large portion of our protein source.



People swimming and diving in the water by GoodFreePhotos CC by 2.0

The biodiversity of oceans is critical. Coral reefs (a tropical marine ecosystem) are often considered the "tropical rainforest of the sea" due to their diversity in marine life and varying animal species. There is still a lot to discover about the ocean, however we do know it's important to maintain because the ocean is the largest ecosystem we have on the planet. This ecosystem maintains homeostasis (a steady state) via regulation by abiotic and biotic factors. The oceans are often overlooked especially since we don't see the dynamics between marine organisms and our own environment as we don't share this habitat.

Although the ocean is somewhat overlooked it is a truly remarkable ecosystem. Unfortunately, there has been a rise in ecological crisis in aquatic ecosystems mostly at the hands of humans. For example, overfishing can alter predator/prey balance, and pollution can significantly impact the health of the environment. If the oceans continue to be in crisis at a global level, it could result in mass species extinction. Therefore, it is really important to care for our oceans.

As we embark on this journey into the study of the oceans, please remember why our understanding is so important.

The information in this chapter is thanks to content contributions from Marisa Benjamin, Maddie Ouellette, and Allie Tolles

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1.2: Types of Oceanographic Study

Oceanography can be divided into four branches: Biological Oceanography, Chemical Oceanography, Geological Oceanography, and Physical Oceanography.

Biological oceanography is concerned with studying the living aspects of the ocean. Organisms that live in the ocean include the tiniest organisms, or plankton, and the largest organisms, or whales. Examples of topics studied by biological oceanographers include the distribution of organisms in the ocean, the types of organisms found in specific locations, the interactions between living organisms, and the abundance of specific types of organisms.

Chemical oceanography focuses on the chemicals found in water. As you will learn later, there is a wide variety of chemicals dissolved in seawater, from common salt, or sodium chloride, to even gold. Chemical oceanographers may study the concentration of various chemicals and how they are distributed both horizontally and vertically. These chemicals could be naturally occurring or may enter the water through man-made processes.

Geological oceanography focuses on the geological materials and processes that occur in our oceans. Geological oceanographers can study the constituents that make up the seafloor, the shape of the seafloor, and the processes that create the oceans. As you'll learn, geological oceanography encompasses a wide variety of research, which can span millions of years of natural processes.

Physical oceanography focuses on the physical movement of the oceans. This includes the study of waves, tides, and water currents. Physical oceanographers could study deep water currents, physical processes occurring along the shore, and even interactions between the ocean and the atmosphere.

As you can probably tell, these fields of study are closely related to one another. Oceanographic research may span multiple branches of oceanography, and some of the most interesting investigations include scientists from all four branches.

Check out this article to see what oceanographers have learned in the last decade: https://www.smithsonianmag.com/science-nature/biggest-stories-ocean-science-last-decade-180973793/

Go to this website to learn about the tools oceanographers use to study the ocean: https://www.nsf.gov/geo/oce/whatis/tools.jsp

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CHAPTER OVERVIEW

2: Introduction to Science

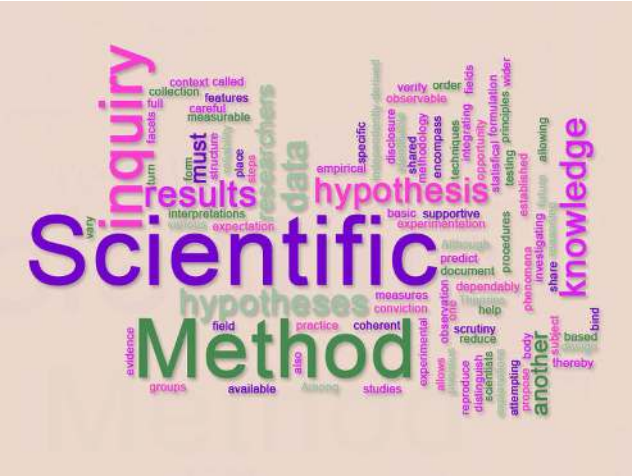
2.1: Scientific Investigation

2.2: Science Experiments

2.3: Scientific Theories

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2.1: Scientific Investigation



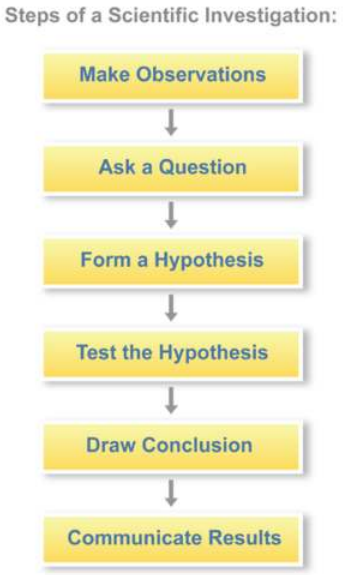
Chances are you've heard of the scientific method. But what exactly is the scientific method?

Is it a precise and exact way that all science must be done? Or is it a series of steps that most scientists generally follow, but may be modified for the benefit of an individual investigation?

The Scientific Method

"We also discovered that science is cool and fun because you get to do stuff that no one has ever done before." In the article *Blackawton bees*, published by eight to ten year old students: Biology Letters (2010) http://rsbl.royalsocietypublishing.org/content/early/2010/12/18/rsbl.2010.1056.abstract.

There are basic methods of gaining knowledge that are common to all of science. At the heart of science is the scientific investigation, which is done by following the **scientific method**. A **scientific investigation** is a plan for asking questions and testing possible answers. It generally follows the steps listed in **Figure** below. See http://www.youtube.com/watch?v=KZaCy5Z87FA for an overview of the scientific method.



Steps of a Scientific Investigation. A scientific investigation typically has these steps. Scientists often develop their own steps they follow in a scientific investigation. Shown here is a simplification of how a scientific investigation is done.

Making Observations

A scientific investigation typically begins with observations. You make observations all the time. Let's say you take a walk in the woods and observe a moth, like the one in **Figure** below, resting on a tree trunk. You notice that the moth has spots on its wings that look like eyes. You think the eye spots make the moth look like the face of an owl.



Figure 2: Marbled emperor moth Heniocha dyops in Botswana. (CC-SA-BY-4.0; Charlesjsharp).

Does this moth remind you of an owl?

Asking a Question

Observations often lead to questions. For example, you might ask yourself why the moth has eye spots that make it look like an owl's face. What reason might there be for this observation?

Forming a Hypothesis

The next step in a scientific investigation is forming a hypothesis. A **hypothesis** is a possible answer to a scientific question, but it isn't just any answer. A hypothesis must be based on scientific knowledge, and it must be logical. A hypothesis also must be falsifiable. In other words, it must be possible to make observations that would disprove the hypothesis if it really is false. Assume you know that some birds eat moths and that owls prey on other birds. From this knowledge, you reason that eye spots scare away birds that might eat the moth. This is your hypothesis.

Testing the Hypothesis

To test a hypothesis, you first need to make a prediction based on the hypothesis. A **prediction** is a statement that tells what will happen under certain conditions. It can be expressed in the form: If A occurs, then B will happen. Based on your hypothesis, you might make this prediction: If a moth has eye spots on its wings, then birds will avoid eating it.

Next, you must gather evidence to test your prediction. **Evidence** is any type of data that may either agree or disagree with a prediction, so it may either support or disprove a hypothesis. Evidence may be gathered by an **experiment**. Assume that you gather evidence by making more observations of moths with eye spots. Perhaps you observe that birds really do avoid eating moths with eye spots. This evidence agrees with your prediction.

Drawing Conclusions

Evidence that agrees with your prediction supports your hypothesis. Does such evidence prove that your hypothesis is true? No; a hypothesis cannot be proven conclusively to be true. This is because you can never examine all of the possible evidence, and someday evidence might be found that disproves the hypothesis. Nonetheless, the more evidence that supports a hypothesis, the more likely the hypothesis is to be true.

Communicating Results

The last step in a scientific investigation is communicating what you have learned with others. This is a very important step because it allows others to test your hypothesis. If other researchers get the same results as yours, they add support to the hypothesis. However, if they get different results, they may disprove the hypothesis.

When scientists share their results, they should describe their methods and point out any possible problems with the investigation. For example, while you were observing moths, perhaps your presence scared birds away. This introduces an error into your

investigation. You got the results you predicted (the birds avoided the moths while you were observing them), but not for the reason you hypothesized. Other researchers might be able to think of ways to avoid this error in future studies.

The Scientific Method Made Easy explains scientific method: http://www.youtube.com/watch?v=zcavPAFiG14 (9:55).



> *As you view The Scientific Method Made Easy, focus on these concepts:*
>
> 1. *the relationship between evidence, conclusions and theories,*
> 2. *the "ground rules" of scientific research,*
> 3. *the steps in a scientific procedure,*
> 4. *the meaning of the "replication of results,"*
> 5. *the meaning of "falsifiable,"*
> 6. *the outcome when the scientific method is not followed.*

Discovering the Scientific Method

A summery video of the scientific method, using the identification of DNA structure as an example, is shown in this video by MIT students: https://www.youtube.com/watch?v=5eDNgeEUtMg.

Why I do Science

Dan Costa, Ph.D. is a professor of Biology at the University of California, Santa Cruz, and has been studying marine life for well over 40 years. He is a leader in using satellite tags, time and depth recorders and other sophisticated electronic tags to gather information about the amazing depths to which elephant seals dive, their migration routes and how they use oceanographic features to hunt for prey as far as the international dateline and the Alaskan Aleutian Islands. In the following KQED video, Dr. Costa discusses why he is a scientist:http://science.kqed.org/quest/video/why-i-do-science-dan-costa/.



Summary

- At the heart of science is the scientific investigation, which is done by following the scientific method. A scientific investigation is a plan for asking questions and testing possible answers.
- A scientific investigation typically begins with observations. Observations often lead to questions.
- A hypothesis is a possible logical answer to a scientific question, based on scientific knowledge.
- A prediction is a statement that tells what will happen under certain conditions.
- Evidence is any type of data that may either agree or disagree with a prediction, so it may either support or disprove a hypothesis. Conclusions may be formed from evidence.
- The last step in a scientific investigation is the communication of results with others.

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2.2: Science Experiments



So what exactly is an experiment?

At first you may picture a science laboratory with microscopes and chemicals and people in white lab coats. But do all experiments have to be done in a lab? And do all scientists have to wear lab coats?

Experiments

Figure below shows a laboratory experiment involving plants. An **experiment** is a special type of scientific investigation that is performed under controlled conditions, usually in a laboratory. Some experiments can be very simple, but even the simplest can contribute important evidence that helps scientists better understand the natural world. An example experiment can be seen here http://www.youtube.com/watch?v=dVRBDRAsP6U or here http://www.youtube.com/watch?v=F10EyGwd57M. As many different types of of experiments are possible, an experiment must be designed to produce data that can help confirm or reject the hypothesis.



A laboratory experiment studying plant growth. What might this experiment involve?



In this experiment, a scientist is conducting research (and taking notes) while looking through a microscope.

Medicine From the Ocean Floor

Scientists at the University of California, Santa Cruz are looking to perhaps the largest resource yet to be explored for its medical potential: the ocean. And they are taping this resource with some state-of-the-art technology. These scientists are using robots to sort through thousands of marine chemicals in search of cures for diseases like cholera, breast cancer, and malaria. These experiments are described in the following KQED links:

- www.kqed.org/quest/blog/2009/...e-ocean-floor/
- www.kqed.org/quest/radio/medicine-from-the-ocean-floor
- science.kqed.org/quest/slides...oor-slideshow/

Variables

An experiment generally tests how one **variable** is affected by another. The affected variable is called the **dependent variable**. In the plant experiment shown above, the dependent variable is plant growth. The variable that affects the dependent variable is called the **independent variable**. In the plant experiment, the independent variable could be fertilizer—some plants will get fertilizer, others will not. The scientists change the amount of the independent variable (the fertilizer) to observe the effects on the dependent variable (plant growth). An experiment needs to be run simultaneously in which no fertilizer is given to the plant. This would be known as a **control experiment.** In any experiment, other factors that might affect the dependent variable must be controlled. In the plant experiment, what factors do you think should be controlled? (*Hint:* What other factors might affect plant growth?)

Sample Size and Repetition

The sample in an experiment or other investigation consists of the individuals or events that are studied, and the size of the sample (or **sample size**) directly affects the interpretation of the results. Typically, the sample is much smaller than all such individuals or events that exist in the world. Whether the results based on the sample are true in general cannot be known for certain. However, the larger the sample is, the more likely it is that the results are generally true.

Similarly, the more times that an experiment is repeated (which is known as **repetition**) and the same results obtained, the more likely the results are valid. This is why scientific experiments should always be repeated.

Bio-Inspiration: Nature as Muse

For hundreds of years, scientists have been using design ideas from structures in nature. Now, biologists and engineers at the University of California, Berkeley are working together to design a broad range of new products, such as life-saving milli-robots modeled on the way cockroaches run and adhesives based on the amazing design of a gecko's foot. This process starts with making observations of nature, which lead to asking questions and to the additional aspects of the scientific process. *Bio-Inspiration: Nature as Muse* can be observed at www.kqed.org/quest/television...nature-as-muse.



Super Microscopes

Microscopes are arguably one of the most important tools of the biologist. They allow the visualization of smaller and smaller biological organisms and molecules. With greatly magnified powers, these instruments are becoming increasingly important in modern day research. See the following KQED videos for additional information on these remarkable tools.

- *Super Microscope* at http://science.kqed.org/quest/video/super-microscope/.
- *The World's Most Powerful Microscope* at http://www.youtube.com/watch?v=sCYX_XQgnSA.



Summary

- An experiment is a special type of scientific investigation that is performed under controlled conditions, usually in a laboratory.
- An experiment generally tests how one variable is affected by another.
- The sample size in an experiment directly affects the interpretation of the results.
- Repetition is the repeating of an experiment, validating the results.

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2.3: Scientific Theories



Theory vs. theory. Is a scientific *theory* different from the everyday use of the word *theory*?

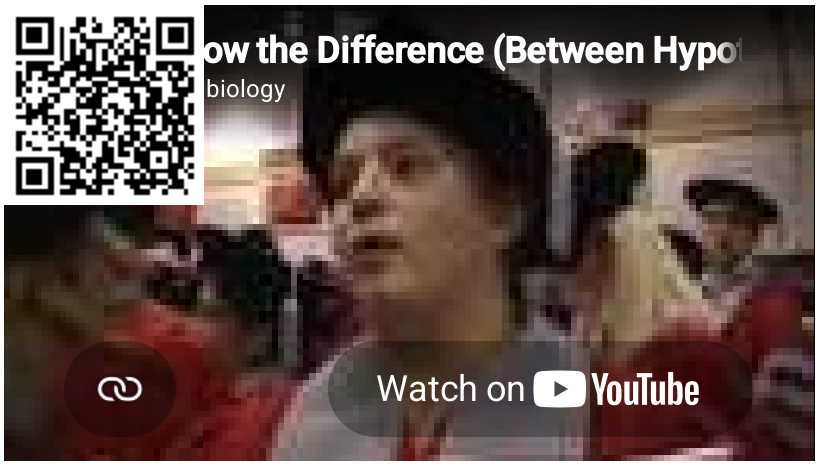
A scientific theory is accepted as a scientific *truth*, supported by evidence collected by many scientists. The theory of evolution by natural selection is a classic scientific theory.

Scientific Theories

With repeated testing, some hypotheses may eventually become scientific theories. Keep in mind, a hypothesis is a possible answer to a scientific question. A **scientific theory** is a broad explanation for events that is widely accepted as true. To become a theory, a hypothesis must be tested over and over again, and it must be supported by a great deal of evidence.

People commonly use the word *theory* to describe a guess about how or why something happens. For example, you might say, "I think a woodchuck dug this hole in the ground, but it's just a theory." Using the word *theory* in this way is different from the way it is used in science. A scientific theory is more like a fact than a guess because it is so well-supported. There are several well-known theories in biology, including the theory of evolution, cell theory, and germ theory.

Two videos explaining scientific theories can be seen at http://www.youtube.com/watch?v=S5YGhprR6KE and http://www.youtube.com/watch?v=jdWMcMW54fA.



> *As you view Know the Difference (Between Hypothesis and Theory), focus on these concepts:*
>
> 1. *the controversy surrounding the words "hypothesis" and "theory",*
> 2. *the scientific use of the words "hypothesis" and "theory",*
> 3. *the criteria for a "hypothesis,"*

> *4. the National Academy of Sciences definition of "theory",*
> *5. the meaning of the statement, "theories are the bedrock of our understanding of nature".*

The Theory of Evolution

The theory of evolution by natural selection is a scientific theory. **Evolution** is a change in the characteristics of living things over time. Evolution occurs by a process called **natural selection**. In natural selection, some living things produce more offspring than others, so they pass more genes to the next generation than others do. Over many generations, this can lead to major changes in the characteristics of living things. The theory of evolution by natural selection explains how living things are changing today and how modern living things have descended from ancient life forms that no longer exist on Earth. No evidence has been identified that proves this theory is incorrect. More on the theory of evolution will be presented in additional concepts.

The Cell Theory

The cell theory is another important scientific theory of biology. According to the **cell theory**, the cell is the smallest unit of structure and function of all living organisms, all living organisms are made up of at least one cell, and living cells always come from other living cells. Once again, no evidence has been identified that proves this theory is incorrect. More on the cell theory will be presented in additional concepts.

The Germ Theory

The **germ theory** of disease, also called the pathogenic theory of medicine, is a scientific theory that proposes that microorganisms are the cause of many diseases. Like the other scientific theories, lots of evidence has been identified that supports this theory, and no evidence has been identified that proves the theory is incorrect.

Summary

- With repeated testing, some hypotheses may eventually become scientific theories. A scientific theory is a broad explanation for events that is widely accepted as true.
- Evolution is a change species over time. Evolution occurs by natural selection.
- The cell theory states that all living things are made up of cells, and living cells always come from other living cells.
- The germ theory proposes that microorganisms are the cause of many diseases.

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CHAPTER OVERVIEW

3: Ocean Exploration

3.1: Early Trade and Navigation

3.2: Explorers and Oceanographers

3.3: Voyages for Oceanography

3.4: Modern Ocean Science

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3.1: Early Trade and Navigation

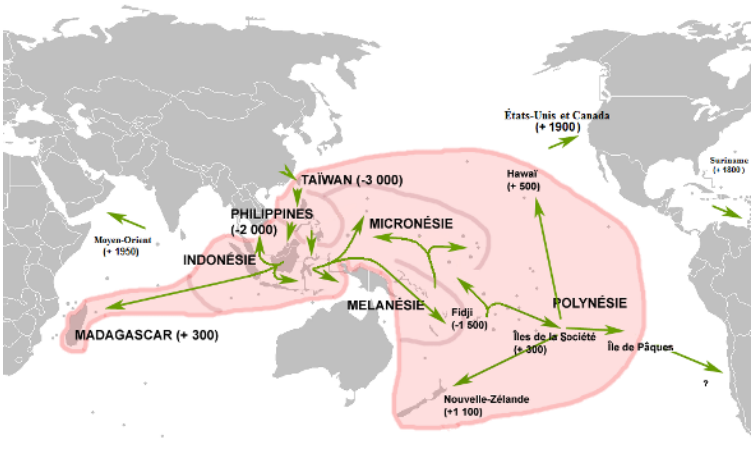
The history of ocean exploration is one filled with danger, discovery, and romance. For nearly 100,000 years intrepid individuals have traversed the oceans and crossed the seas of this planet. The first oceanographers were traders, navigators, conquers, and explorers who dared to cross the sea.



A depiction of an Egyptian boat, from c. 1420 B.C. (Wikipedia)

THE FIRST OCEAN GOERS

Ocean exploration has slowly, but steadily gained more and more interest as the desire to expand, trade, and explore have become more enticing and necessary to understand the world around us. Contrary to popular belief, the European Age of Exploration was not the start of this tradition. The earliest known explorers originated from Southeast Asia over 100,000 years ago during what is known as the Melanesian exploration. Inhabitants of what is now known as Indonesia began discovering and exploring islands to the East. These people did not stray far from land but stayed close to coastal areas. They used small simple watercraft, designed primarily for littoral fishing and travel. This early period of exploration was superseded by superior navigation and cartography techniques developed by later cultures such as the Polynesians.



Dates of human migration in the Indian and Pacific Oceans. (Wikimedia)

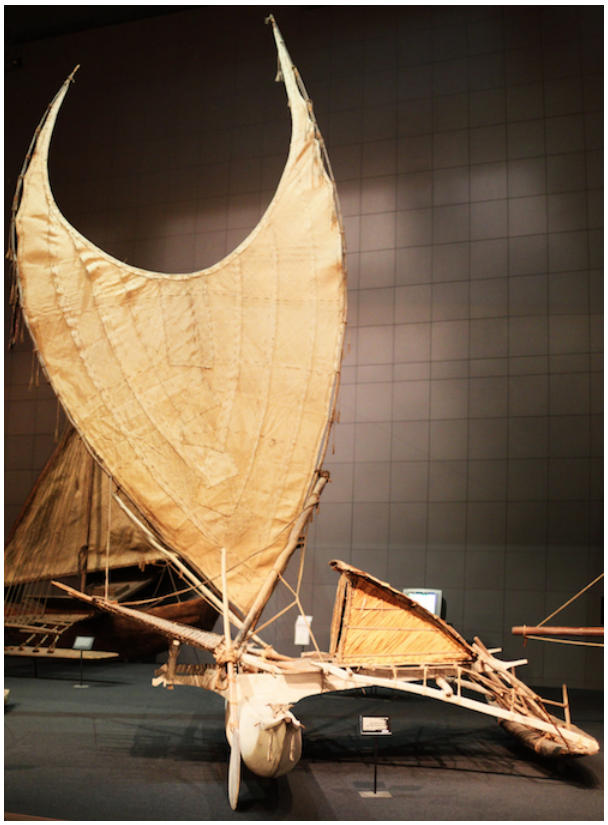
The Polynesians

The next wave of ocean exploration was the Polynesian Period. They were the first to develop open ocean exploration and navigation techniques. They consistently traveled across much of the South Pacific, passing New Zealand, Easter Island, and many others, and eventually making their way to Hawaii. This period also overlaps with the Greek exploration of the Mediterranean.

There exist many conflicting theories as to how this culture was able to disperse across such a vast area or why they choose to do so, but they developed larger, stronger, and more stable boats that allowed them to endure the harsh conditions of the open ocean.

The Polynesians relied upon a complex and rich oral tradition to pass on knowledge of their navigation techniques, known as wayfinding. Historically, this knowledge has been confined to local navigation guilds on each island, but renewed interest has seen that these techniques are recorded for all time. These oral traditions have shown not only a detailed knowledge of natural phenomena such as wind patterns but also an extensive and precise knowledge of the locations of islands all across Polynesia. There is still discourse on whether early Polynesian colonizers made use of these techniques. However, there is a strong academic consensus that the Islands of Eastern Polynesia, such as Easter Island, were explored and colonized purposively using this knowledge.

Initially, researchers thought that the Polynesians and their culture originated from an Asiatic country and migrated to the Polynesian Triangle. However, researchers were able to use pottery to date the migration of Polynesian settlements and found that the Polynesian culture started on the Polynesian islands. Specifically, they started from the southeastern Solomons and northern Vanuatu before expanding to the Fiji archipelago which is west of the Polynesian Triangle around 1100-1000 B.C. They then settled in the Lau Islands then Tonga around 896-880 B.C. and further traveled east. Colonization of eastern and southern Polynesia occurred later. A study suggests explaining the delay in colonization was due to the unique climate patterns which finally allowed off wind sailing routes to the east. Gradual improvements in their sailing technology also attributed to the delay. The invention of the double-hulled canoe became beneficial in their colonization of eastern and southern Polynesia.



A tepukei, an ancient type of boat used by the inhabitants of Taumako in Melanesia. (Wikimedia)

WESTERN NAVIGATORS

Some of the first ocean-faring people were the Minoan, Greek, and the Phoenician civilizations of the ancient Mediterranean. They utilized the Mediterranean for both trade and war, at first staying within sight of shore, but eventually using the sun, moon, and stars as navigational aids. The Phoenicians were some of the first to use celestial bodies to take them beyond the sight of land, but the knowledge eventually spread throughout the region to facilitate maritime commerce and navigation. The most famous Phoenician explorers were Hanno the Navigator and Himilco, both from Carthage. In the sixth or fifth centuries B.C., Hanno sailed

from Carthage, in modern day Tunisia, out of the Mediterranean and along the coast of Africa, reaching as far as Cameroon. Himilco, in the fifth century, sailed from North Africa to the British Isles. These journeys sought to establish and control trade routes.

Homer wrote of Odysseus using Ursa major to find his way home from Troy, and Greek navigators were using nautical charts as early as 600 BC. Some Greeks are known to have sailed all the way to India after Alexander's conquest of the east.



A Phoenician ship depicted on a sarcophagus, 2nd century A.D. (Wikimedia)

NAVIGATION IN THE EAST

After the Polynesian people had migrated across the South Pacific, the first cultures to successfully navigate the waters around Asia were the Indians in the 4th century BC and the navy of the Chinese Qin dynasty around 200 BC. Much of the ocean navigation for these groups relied on the seasonal monsoon winds, which limited travel direction and time. Regardless, both the Indians and Chinese cultures were able to sail and explore much of Southeast Asia, and even to the eastern coast of Africa. The Qin dynasty fielded a large navy to control the South China sea as early as 200 BC. The ships were as long as 100 feet and could carry up to 30 tons of cargo.



Pottery boat from the Han Dynasty, from the Hong Kong Museum of History. (Wikimedia)

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3.2: Explorers and Oceanographers

Prince Henry the Navigator (1394 - 1460)

Henry the Navigator lived from 1394 to 1460. Prince Henry of Portugal was interested in sailing and commerce, and studied navigation and mapmaking. He established a naval observatory for the teaching of navigation, astronomy, and cartography around 1450. From 1419 to his death, Prince Henry made many expeditions south along the west coast of Africa to secure trade routes and establish colonies.

Prince Henry of Portugal organized and financed many voyages that went south from Portugal and eventually rounded the African continent. His goals were to create maps of the West African coastline, establish trade routes, and spread Christianity. He encouraged voyages of expeditions and the scientific study of navigation. Furthermore, he ran an observatory and school of navigation. Prince Henry played a vital role in the development of more accurate maps and the engineering of a new ship that was more ideal for exploring rough seas.



Prince Henry the Navigator, a key figure in Portuguese exploration. (Wikipedia)

Zheng He (1371 - 1433)

Zheng He [pronounced as JUNG HUH] lived from 1371 to 1433. He was born in Yunnan at the foothills of the Himalaya Mountains. As a child, his name was Ma He. He grew up a Hui, a Chinese Muslim. When he was only 10 years old, Chinese soldiers, under the orders of the Ming Dynasty, invaded Yunnan in effort to overtake one of the last Mongol holds. They killed his father and he was captured. Along with many other boys who were taken, he was castrated and forced to serve for a Chinese prince known as Zhu Di. In 1402, Zhu Di took the throne as Emperor Yongle. He made Ma He the chief of staff for all the servants and changed his name to Zheng He. The Yongle Emperor proved to be one of the most ambitious emperors of his time and chose Zheng He to be the commander in chief of a series of missions across the Indian Ocean in order to increase China's influence.

From 1405 to 1433, Zheng He led seven naval expeditions throughout the Chinese Seas and the Indian Ocean, and he reached locations from Taiwan, the Persian Gulf, and Africa. He saw the commission of 3,500 ships and commanded at least 62 ships and 27,800 men (more than half of London’s population at the time). Zheng He’s led nine-masted flagships that measured about 400 feet long (Christopher Columbus’s Santa Maria was only 85 feet long). These were some of the largest wooden ships ever built in this time period. These naval expeditions did not serve the purpose to colonize or conquer but rather served as “treasure hunts” that brought back items of great value as tribute to the Yongle Emperor. Zheng He brought back gold, jewelry, and other delicacies. He even brought Zebras, Rhinos, and Giraffes. In 1424, the Yongle emperor died and his successor suspended all naval expeditions abroad. Zheng He went on his seventh and final voyage from 1431 to 1433. He died at sea and was buried off the coast of India.

Zheng He and his naval voyages had a great impact on the status of China at the time. These voyages increased maritime and commercial influence of China throughout the Indian ocean up until the 19th century. Foreign goods, medicines, and geological knowledge flowed through China at an unprecedented rate even though these ships only served as treasure ships. Many historians argue that China could have become a great colonial power many years before the age of great exploration if the leadership had decided to use their technology for outreach rather than for isolation.

Here is a good map of the travels of Zheng He: https://cdn.kastatic.org/KA-share/BigHistory/KU8.1.8-4_Zheng_He-Map.pdf



A statue of Zheng He in Stadthuys, Melaka. (Wikipedia)

Christopher Columbus (1451 - 1506)

Christopher Columbus lived from 1451 to 1506. He made four voyages across the Atlantic Ocean in an effort to find a new route to the East Indies by traveling west rather than east. Having inaccurate estimates of Earth's magnitude, he underestimated the distances necessary for the voyage and believed he had found islands off the coast of Asia when, in reality, he had reached the "New World."

Christopher Columbus believed sailing west would be a faster way to get to India and to all the spices and riches it held. When his idea was rejected by Portugal, Columbus went to Spain where the King and Queen agreed to sponsor him. Columbus was granted three ships (the Niña, Pinta, and Santa Maria) and funds to finance his voyage across the ocean. After two months of sailing, Columbus and his crew arrived in the present-day Bahamas. Due to his excellent navigation records, he was able to sail back to Spain with proof, in the form of gold and other materials, of his success. However, the rest of Columbus' voyages were unsuccessful. Although he died bitter and believing he had found Asia, Columbus' discovery opened up a whole new world for his successors to explore.

Ferdinand Magellan

Ferdinand Magellan lived from 1480 to 1521. He left Spain in September 1519 with 270 men and five vessels in search of a westward passage to the Spice Islands. The expedition eventually lost two ships even before finally discovering and passing through the Strait of Magellan and rounding the tip of South America in November 1520. Magellan crossed the Pacific Ocean and arrived in the Philippines in March 1521, where he was killed in a battle with the natives on April 27, 1521. Magellan's skill as a navigator makes his voyage probably the most outstanding single contribution to the early charting of the oceans.

Magellan had the idea to sail west across the ocean to get to Asia. Contrary to his expectations, Magellan sailed past modern-day Argentina and found a route to the Pacific Ocean. He was the first person to sail across the Pacific Ocean. It took him six long months to arrive in Asia though he had been looking for a shorter route. Eventually, Magellan and his crew sailed into charted waters in East Asia. Unfortunately, Magellan was killed in battle in the Philippines. His crew continued without him and reached Spain a total of three years after they had first set out on their voyage. Though Magellan perished during the voyage, his belief that the Earth is round was proven to be true. This important discovery altered the way people thought about the world and had a significant impact on future voyages.

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3.3: Voyages for Oceanography

The Challenger Expedition

In December 1872 the HMS Challenger sailed out of Portsmouth England to begin its four-year voyage circumnavigating the globe. The voyage was the first of its kind, as the ship had recently been re-purposed from military service into a research vessel. Much of the interest in the voyage was to gain understanding in the composition and structure of the seafloor, as in 1851 the first submarine telegraph cable had been laid across the English Channel, and many government agencies and investors realized that further knowledge was the key to expanding in that enterprise.

The vessel was outfitted with both a natural history laboratory, as well as a chemistry laboratory, and carried a variety of scientific instruments. Funded by the Royal Society of London, and captained by George Nares, the ship would travel 127,580 km around the Earth over a four year period. Over the course of the journey, they performed hundreds of depth soundings, open sea trawls, and collected well over 4500 new species specimens. The voyage of the Challenger was the largest oceanographic research expedition of its time and established a new order of scientific and natural research.



Sir George Strong Nares (Wikimedia)

The Albatross

The Albatross was a ship that made three expeditions designed to collect specimens and explore the depths of the ocean. It was built by the US Fish Commission in 1882. It is a scientific research vessel used to study fish population and hydrographic surveys. It is the first ship designed for this type of research, outfitted with laboratories, large storage spaces for specimens, and more technologically advanced dredge systems for studying sediments on the seafloor.

The scientific expeditions investigated the eastern Pacific Ocean, islands and atolls in the South Pacific, and the eastern tropical Pacific. It collected different marine life forms and described over 170 new species on its first expedition. During the second expedition, the ship sounded and collected specimens at a new record depth at 4,137 fathoms. On the third expedition, it collected a wide range of marine life in areas relatively unknown.

Early Arctic Exploration

In 1868, Carl Koldewey led Germany's first Arctic expedition on the GRÖNLAND sailing vessel for the east coast of Greenland. Unfortunately, the ice stopped them from reaching Greenland, so they ended up at Spitsbergen instead. The purpose of this expedition was that the geographer August Petermann, the person who initiated the first German expedition to the Arctic in 1868, wanted to confirm his theory that the Arctic Ocean was completely ice-free and therefore navigable. Even though they did not reach their desired location, they still collected data on current velocity, temperatures, and elevations in the deep sea, and magnetism.

In 1869, the second German expedition to the arctic reached their desired location. In both of these expeditions, they learned more about ice cover in the arctic, local ecosystems, Arctic flora and fauna, mineral deposits and the relevance of the Arctic to our climate. Because of these two expeditions, Germany launched many further expeditions to the polar regions, including Antarctica.



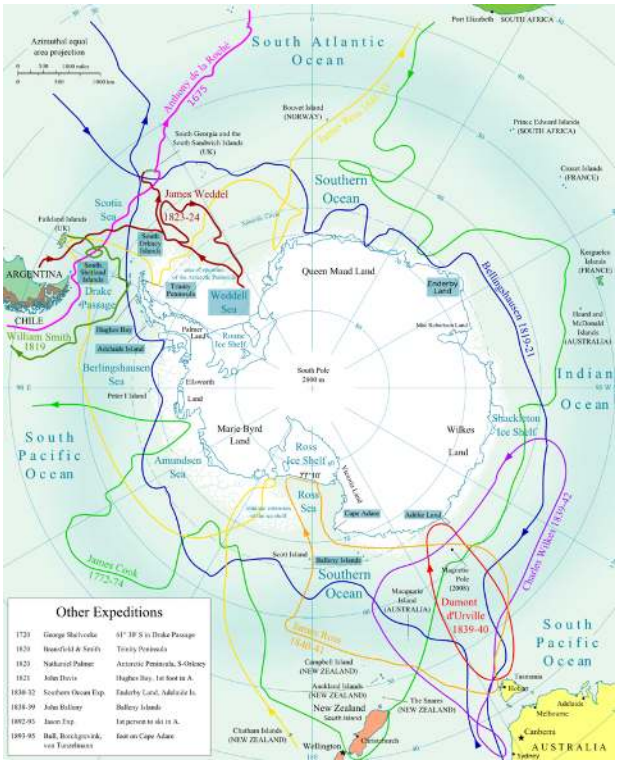
Carl Koldeway (Wikimedia)

Early Antarctic Exploration & History

Before the age of modern science and discovery, the continent of Antarctica was merely a concept postulated by ancient philosophers. In the 4th century BC, the Greek philosopher Aristotle reasoned that the large, northern, frigid landmass (the Arctic) would be balanced by a southern, frozen landmass. He named this unknown landmass “Antarktikos” which means “opposite to the north”.

Fast forward to the 1400s and many explorers began to make voyages south to try and discover this mysterious land. Bartolomeu Dias, a Portuguese sailor, kicked off the “Heroic Age of Exploration” by sailing down the west coast of Africa in 1487, but was turned back north after encountering storms near South Africa. Several more voyages were made by sailors all around the world, however, firm contact (visual and physical) with the mysterious southern continent would not come until the early 1800s.

A lot of early Antarctic exploration was driven by the search for hunting ground for seals. William Smith, a sealer, discovered the South Shetland Islands off the west coast of the Antarctic Peninsula and was then hired by the British Admiralty to survey the islands the following summer. Although William Smith may have been the first recorded to have sighted the continent, it is thought that the first to set foot on the Antarctic continent was John Davis and his crew from the Cecelia, a group of American explorers searching for these hunting grounds. Although, this is not accepted by all historians.



Major Antarctic expeditions taking place before 1897 (all pre-dating the Heroic Age of Antarctic Exploration) (Wikipedia)

Here is a helpful timeline of ocean exploration by National Geographic: https://www.nationalgeographic.org/media/ocean-exploration-timeline/

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3.4: Modern Ocean Science

Though the methods and purposes at the center of what thrusts humanity into oceanic exploration have greatly varied throughout human history, such exploration has allowed for a better understanding of the vast space that is our oceans. It wasn't until the early 1800s, that President Jefferson funded the United States Coast survey for ocean exploration (NOAA). In 1872, Challenger Expedition conducts the first oceanic expedition. On this mission, salinity, density, temperature and hundreds of new species were discovered and looked at for the first time (NOAA). In 1912, with the sinking of the Titanic in the deep seas, prompted the need to develop ways of exploration using acoustics, so objects could be detected before making contact with a ship. This was accelerated during WWI, with both sides devising ways of being able to detect enemy submarines (National Geographic).



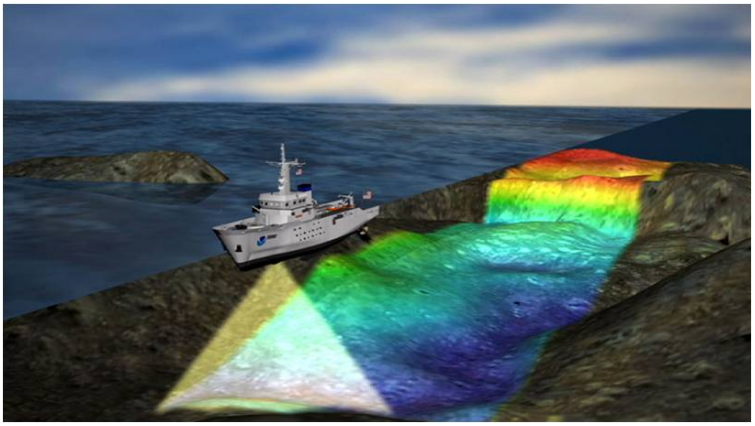
The Challenger at Juan Fernandez (Wikimedia)

Mapping the Ocean Floor

Mapping of the seafloor is typically done by measuring the water depth at a certain point, which was historically done using a "lead line". This process was extremely time consuming and inaccurate. Advanced seafloor mapping became possible in the 1920s due to the advent of underwater sound projectors (called "sonar") (NOAA). The technology was created in order to allow allied ships to combat German U-boats, precursors to modern submarines, by projecting sound through the ocean and measuring amounts of echolocation (OIC). This could give users distance and direction measurements of underwater objects.

Post World War 1, the Coast and Geodetic Survey (which would later become NOAA) turned the military tech down into the depths to allow them to begin mapping the western coast of text Atlantic, producing the first detailed maps of the ocean floor (NOAA). Even with the advance of this technology, mapping the seafloor was still a difficult task, as early sonar systems could only take depth measurements directly below the vessel, making mapping a tiresome process of moving back and forth over the surface and connecting individual data points together.

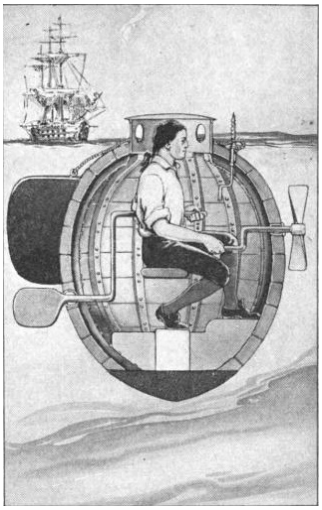
As countries throughout the globe plunged into a second world war, significant advancements in sonar technology that allowed for more accurate and improved measurements of the ocean floor. The data that was acquired from such systems was crucial in the construction of maps that included key features such as mid-ocean ridges and deep-sea trenches (NOAA). In the 1970s, methods became more accurate and efficient when the US Navy declassified "multi-beam" sonar, which allowed research vessels to make more accurate and larger images of the ocean floor. This, along with the increase of computer processing power, has allowed researchers to collect large amounts of data and quickly compile them into detailed and accurate maps.



Multibeam echosounder (Wikimedia)

Autonomous Underwater Vehicles (AUV's)

The first American submersible vehicle was called "Turtle." In Saybrook, Connecticut 1775, David Bushnell and his brother constructed a small, wooden egg-shaped submarine that was effectively used in naval combat (the first to do so) in the New York Harbor (1776). Over a century of technological advancements later, in November 1879, the Reverend George W. Garrett designed the "Resurgam;" the first modern submarine was powered by a Lamm steam engine and could travel for nearly ten hours using power stored in an insulated tank.



Turtle Submarine (Wikimedia)

With submarine evolution came the development of AUV's, which commenced in the 1960s. During the 1970s, many testbeds were developed in hopes of experimenting with technology to explore and define the potential of AUV's. For example, the University of New Hampshire's Marine System Engineering Laboratory collaborated with efforts at the US Navy's facility in San Diego to develop an open space-frame AUV, the EAVE. Subsequently, the Institute of Marine Technology Problems, Russian Academy of Sciences initiated their AUV program with the development of SKAT vehicles and the first deep-diving AUVs L1 and L2. Generally, there are two types of AUV's, manned and unmanned systems, both of which are primarily used for military missions and underwater investigations and assessment, and data gathering. As software systems and engineering advanced, funding expanded, and technologists converged (like at the 1980 international symposium for submersible of technology), AUVs finally were able to grow into first-generation operational systems capable of accomplishing defined objectives. New systems including the Autonomous Oceanographic Sampling System provided resources essential to the AUV market commercialization that flourished from 2000-2010. In terms of AUV technology, systems have constantly changed in accordance with addressing new technology problems or findings.

Over the past three decades, the industry has majorly focused on the following technologies: autonomy, energy systems/energy management, 3D imaging, navigation, processing, sensors, and communications. In the area of ocean science, vehicles such as ABE, AUTOSUB, and others have successfully gathered scientifically significant data. The technology constantly evolves, as new sensors for oceanography are being developed for future data acquisition programs. In addition, the US Navy is supporting a coordinated effort called the AOSN, with the goal of creating a network of multiple AUVs to obtain oceanographic data and information in temporal and spatial resolutions.



Autonomous Benthic Explorer in operation (Wikimedia)

The Deep Sea Drilling Project

In 1966, a collaboration between the National Science Foundation and The Regents, and the University of California began. The goal was to start an oceanic drilling project for the purpose of studying the ocean floor and finding any evidence of plate tectonics. The result of this collaboration was the *Golmar Challenger*, which left the port of Orange, Texas in 1967. The four hundred foot long ship was equipped with enough laboratories, living quarters, and mess halls to accommodate hundreds of crew members. However, the most notable part of the ship was its drill. It was capable of reaching depths of 20,000 feet underwater, which gave scientists a chance to obtain samples that could not have been collected before.



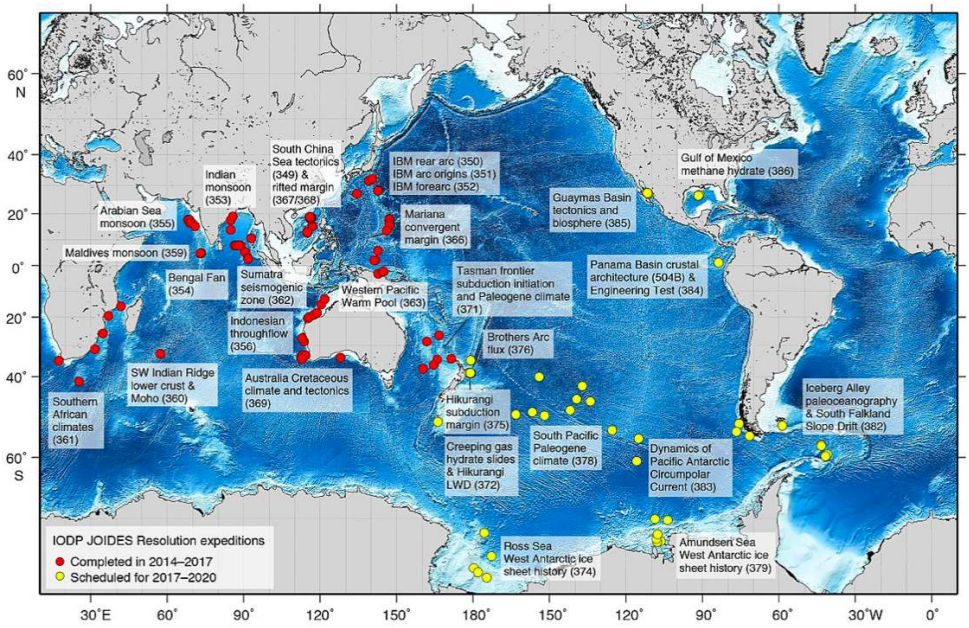
The *Golmar Challenger*. (Wikipedia)

The ship and its crew began its first thirty months coring and drilling in all major oceans and as well as the Red and Mediterranean Seas. The core samples themselves were 30 feet long and 2.5 inches in diameter. With being such an advanced means of sampling at the time, the crew found some of the most important discoveries in oceanography. Their efforts allowed the discovery of salt domes, and in turn, oil at the bottom of the ocean. When the crew drilled at mid-ocean ridges, scientists on board found that parts of

the ocean floor were only 200 million years old. This meant that the seafloor was being recycled since Earth's oceans formed, providing substantial evidence for seafloor spreading and continental drift. With over 300,000 nautical miles traveled, 19,000 cores collected, and approximately 600 sites investigated, The *Golmar Challenger* retired in 1983. However, in 1985, the project continued under the new titles of the Ocean Drilling Program and the International Ocean Drilling Program. In place of the *Golmar Challenger* was the *JOIDES Resolution*, an even larger ship with more advanced drilling equipment. To this day both the programs and *JOIDES Resolution* continue to operate, taking samples from all over Earth's oceans. Sample cores from these ships are stored in research centers all over the world and are considered priceless by many oceanographers.



The *JOIDES Resolution*, still in use to this day. (Wikipedia)



JOIDES Resolution expedition map of 2013-2020 (Wikipedia)

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CHAPTER OVERVIEW

4: Origin of the Earth and Ocean

4.1: Origin of Earth and the Solar System

4.2: Origin of the Oceans

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4.1: Origin of Earth and the Solar System

According to the **Big Bang theory**, the universe blinked violently into existence 13.77 billion years ago (Figure 4.1.1). The Big Bang is often described as an explosion, but imagining it as an enormous fireball isn't accurate. The Big Bang involved a sudden expansion of matter, energy, and space from a single point. The kind of Hollywood explosion that might come to mind involves expansion of matter and energy *within* space, but during the big bang, space *itself* was created.

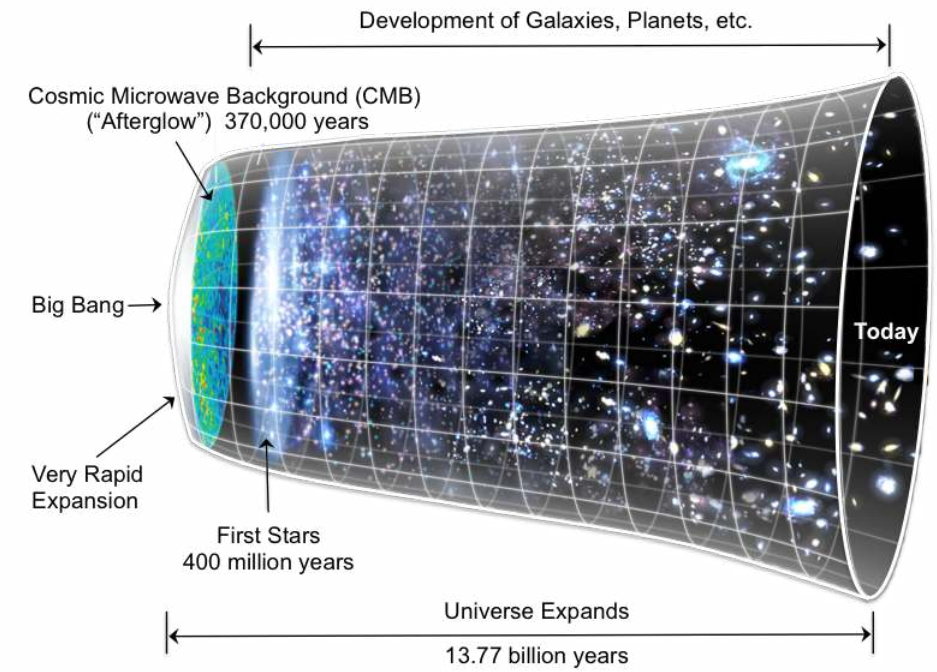


Figure 4.1.1 The Big Bang and development of the universe (Steven Earle, "Physical Geology").

At the start of the Big Bang, the universe was too hot and dense to be anything but a sizzle of particles smaller than atoms, but as it expanded, it also cooled. Eventually some of the particles collided and stuck together. Those collisions produced hydrogen and helium, the most common elements in the universe, along with a small amount of lithium. Gravity caused clouds of these early elements to coalesce into stars, and it was inside these stars that heavier elements were formed

Our solar system began to form around 5 billion years ago, roughly 8.7 billion years after the Big Bang. A **solar system** consists of a collection of objects orbiting one or more central stars. All solar systems start out the same way. They begin in a cloud of gas and dust called a **nebula**. Nebulae are some of the most beautiful objects that have been photographed in space, with vibrant colors from the gases and dust they contain, and brilliant twinkling from the many stars that have formed within them (Figure 4.1.2). The gas consists largely of hydrogen and helium, and the dust consists of tiny mineral grains, ice crystals, and organic particles.



Figure 4.1.2 Photograph of a nebula. The Pillars of Creation within the Eagle Nebula viewed in visible light (left) and near infrared light (right). Near infrared light captures heat from stars, and allows us to view stars that would otherwise be hidden by dust. This is why the picture on the right appears to have more stars than the picture on the left [NASA, ESA, and the Hubble Heritage Team (STScI/AURA) http://bit.ly/1Dm2X5a].

A solar system begins to form when a small patch within a nebula (small by the standards of the universe, that is) begins to collapse upon itself. Exactly how this starts isn't clear, although it might be triggered by the violent behavior of nearby stars as they progress through their life cycles. Energy and matter released by these stars might compress the gas and dust in nearby neighborhoods within the nebula. Once it is triggered, the collapse of gas and dust within that patch continues for two reasons. One of those reasons is that gravitational force pulls gas molecules and dust particles together. But early in the process, those particles are very small, so the gravitational force between them isn't strong. So how do they come together? The answer is that dust first accumulates in loose clumps for the same reason dust bunnies form under your bed: static electricity. As the small patch within a nebula condenses, a star begins to form from material drawn into the center of the patch, and the remaining dust and gas settle into a disk that rotates around the star. The disk is where planets eventually form, so it's called a **protoplanetary disk**. In Figure 4.1.3 the image in the upper left shows an artist's impression of a protoplanetary disk, and the image in the upper right shows an actual protoplanetary disk surrounding the star HL Tauri. Notice the dark rings in the protoplanetary disk. These are gaps where planets are beginning to form. The rings are there because incipient planets are beginning to collect the dust and gas in their orbits. There is an analogy for this in our own solar system, because the dark rings are akin to the gaps in the rings of Saturn (Figure 4.1.3, lower left), where moons can be found (Figure 4.1.3, lower right).

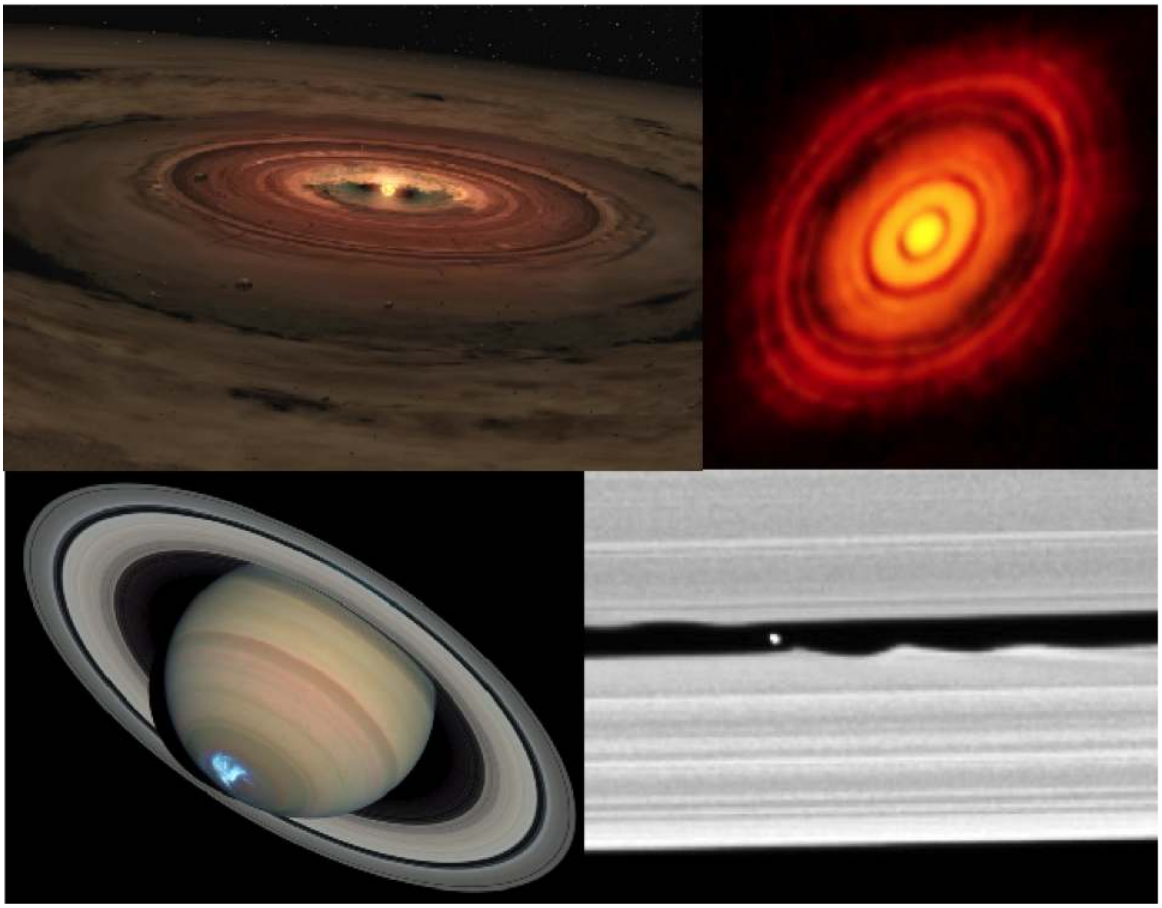


Figure 4.1.3 Protoplanetary disks and Saturn's rings. Upper left: An artists impression of a protoplanetary disk containing gas and dust, surrounding a new star. [NASA/ JPL-Caltech, http://1.usa.gov/1E5tFJR] Upper right: A photograph of the protoplanetary disk surrounding HL Tauri. The dark rings within the disk are thought to be gaps where newly forming planets are sweeping up dust and gas. [ALMA (ESO/NAOJ/NRAO) http://bit.ly/1KNCq0e]. Lower left: A photograph of Saturn showing similar gaps within its rings. The bright spot at the bottom is an aurora, similar to the northern lights on Earth. [NASA, ESA, J. Clarke (Boston University), and Z. Levay (STScI) http://bit.ly/1IfSCX5] Lower right: a close-up view of a gap in Saturn's rings showing a small moon as a white dot. [NASA/JPL/Space Science Institute, http://1.usa.gov/1g2EeYw].

In general, planets can be classified into three categories based on what they are made of (Figure 4.1.4). **Terrestrial planets** are those planets like Earth, Mercury, Venus, and Mars that have a core of metal surrounded by rock. **Jovian planets** (also called **gas giants**) are those planets like Jupiter and Saturn that consist predominantly of hydrogen and helium. **Ice giants** are planets such as Uranus and Neptune that consist largely of water ice, methane (CH_4) ice, and ammonia (NH_3) ice, and have rocky cores. Often, the ice giant planets Uranus and Neptune are grouped with Jupiter and Saturn as gas giants; however, Uranus and Neptune are very different from Jupiter and Saturn.

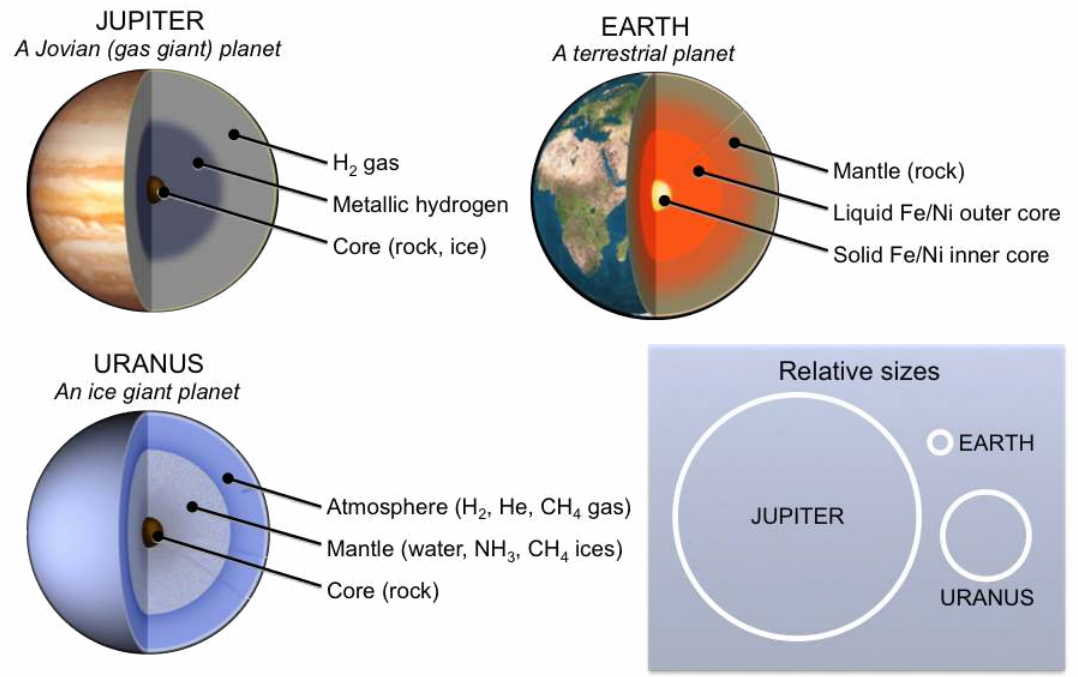


Figure 4.1.4 Three types of planets. Jovian (or gas giant) planets such as Jupiter consist mostly of hydrogen and helium. They are the largest of the three types. Ice giant planets such as Uranus are the next largest. They contain water, ammonia, and methane ice. Terrestrial planets such as Earth are the smallest, and they have metal cores covered by rocky mantles. [KP, after public domain images by Francesco A, Wolfman SF (http://bit.ly/1eP75P4), and NASA (http://1.usa.gov/1gFVsf6, http://1.usa.gov/1M89jI3)].

These three types of planets are not mixed together randomly within our solar system. Instead they occur in a systematic way, with terrestrial planets closest to the sun, followed by the Jovian planets and then the ice giants. Part of the reason for this arrangement is the **frost line** (also referred to as the **snow line**). The frost line separated the inner part of the protoplanetary disk closer to the sun, where it was too hot to permit anything but silicate minerals and metal to crystallize, from the outer part of the disk farther from the Sun, where it was cool enough to allow ice to form. As a result, the objects that formed in the inner part of the protoplanetary disk consist largely of rock and metal, while the objects that formed in the outer part consist largely of gas and ice. The young sun also blasted the solar system with raging **solar winds** (winds made up of energetic particles), which helped to drive lighter molecules toward the outer part of the protoplanetary disk.

The objects in our solar system formed by **accretion**. Early in this process, mineral and rock particles collected in fluffy clumps because of static electricity. As the mass of the clumps increased, gravity became more important, pulling material from farther away and growing these solid masses into larger and larger bodies. Eventually the mass of the objects became large enough that their gravity was strong enough to hang onto gas molecules, because gas molecules are very light.

Our Earth formed though this process of accretion about 4.6 billion years ago. The early Earth was very hot and had a molten, fluid composition, with lost of geological and volcanic activity on the surface. The Earth's heat came from a variety of processes:

- Heat came from the decay of radioactive elements within the Earth, specifically the decay of 235U, 238U, 40K, and 232Th, which are primarily present in the mantle. The total heat produced that way has been decreasing over time (because these isotopes are getting used up), and is now roughly 25% of what it was when Earth formed. This means that Earth's interior is slowly becoming cooler.
- Heat came from the thermal energy already contained within the objects that accreted to form the Earth.
- Heat came from collisions. When objects hit Earth, some of the energy from their motion went into deforming Earth, and some of it was transformed into heat. (The very worst collision that Earth experienced was with a planet named Theia, which was approximately the size of Mars. Not long after Earth formed, Theia struck Earth. When Theia slammed into Earth, Theia's metal core merged with Earth's core, and debris from the outer silicate layers was cast into space, forming a ring of rubble around Earth. The material within the ring coalesced into a new body in orbit around Earth, giving us our moon. Remarkably, the debris may have coalesced in 10 years or fewer! This scenario for the formation of the moon is called the **giant impact hypothesis**.)

- As Earth became larger, its gravitational force became stronger. This increased Earth's ability to draw objects to it, but it also caused the material making Earth to be compressed, rather like Earth giving itself a giant gravitational hug. Compression causes materials to heat up.

Heating had a very important consequence for Earth's structure. As Earth grew, it collected a mixture of silicate mineral grains as well as iron and nickel. These materials were scattered throughout Earth. That changed when Earth began to heat up: it got so hot that both the silicate minerals and the metals melted. The metal melt was much denser than the silicate mineral melt, so the metal melt sank to Earth's center to become its core, and the silicate melt rose upward to become Earth's crust and mantle. In other words, Earth unmixed itself. The separation of silicate minerals and metals into a rocky outer layer and a metallic core, respectively, is called **differentiation**. Gravity has since pulled Earth into an almost spherical shape with a radius of 6371 km, and a circumference of about 40,000 km. However, it is not a perfect sphere, as the Earth's rotation causes an equatorial bulge, so that the Earth's circumference is 21 km (0.3%) wider at the equator than it is pole to pole. Thus it is technically an "oblate spheroid."

If we were to take an inventory of the elements that make up Earth, we would find that 95% of Earth's mass comes from only four elements: oxygen, magnesium, silicon, and iron. Most of the remaining 5% comes from aluminum, calcium, nickel, hydrogen, and sulphur. We know that the Big Bang made hydrogen, helium, and lithium, but where did the rest of the elements come from? The answer is that the other elements were made by stars. The heat and pressure within stars cause smaller atoms to smash together and fuse into new, larger atoms. For example, when hydrogen atoms smash together and fuse, helium is formed. Large amounts of energy are released when some atoms fuse and that energy is what causes stars to shine.

It takes larger stars to make elements as heavy as iron and nickel. Our Sun is an average star; after it uses up its hydrogen fuel to make helium, and then some of that helium is fused to make small amounts of beryllium, carbon, nitrogen, oxygen, and fluorine, it will be at the end of its life. It will stop making atoms and will cool down and bloat until its middle reaches the orbit of Mars. In contrast, large stars end their lives in spectacular fashion, exploding as supernovae and casting off newly formed atoms —including the elements heavier than iron — into space. It took many generations of stars creating heavier elements and casting them into space before heavier elements were abundant enough to form planets like Earth.

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4.2: Origin of the Oceans

So how did the oceans form in the first place? The early Earth was formed through the accretion of various materials, and that a period of melting and intense volcanic activity followed. The materials that accreted on the early Earth contained the components that would eventually become our oceans and atmosphere. Under the high pressures found in the Earth's interior, gases remain dissolved in magma. As these magmas rise to the surface through volcanic activity, the pressure is reduced and the gases are released through a process called **outgassing**. Volcanic activity releases many different gases, including water vapor, carbon dioxide (CO_2), sulfur dioxide (SO_2), carbon monoxide (CO), hydrogen sulfide (H_2S), hydrogen gas, nitrogen, and methane (CH_4). Lighter gases such as hydrogen and helium dissipated into space, but the heavier gases remained and formed Earth's early atmosphere.

The rise of atmospheric oxygen

It's worth noting that the early atmosphere lacked free oxygen (O_2), the form of oxygen that we breathe. We know this in part because prior to 2 billion years ago, there were no sedimentary beds stained red from oxidized iron minerals. Iron minerals were present, but not in oxidized form. At that time, O_2 was produced in the atmosphere when the Sun's ultraviolet rays split water molecules apart; however, chemical reactions removed the oxygen as quickly as it was produced. It wasn't until the appearance of life that Earth's atmosphere began to become oxygenated. Photosynthetic organisms used the abundant CO_2 in the atmosphere to manufacture their food, and released O_2 as a by-product. At first all of the oxygen was consumed by chemical reactions, but eventually the organisms released so much O_2 that it overwhelmed the chemical reactions and oxygen began to accumulate in the atmosphere, although present levels of 21% oxygen didn't occur until about 350 Ma. Today the part of our atmosphere that isn't oxygen consists largely of nitrogen (78%). The oxygen-rich atmosphere on our planet is life's signature. If geologic process were the only processes controlling our atmosphere, it would consist mostly of carbon dioxide, like the atmosphere of Venus.

As the early Earth cooled, the water vapor in the atmosphere condensed and fell as rain. By about 4 billion years ago, the first permanent accumulations of water were present on Earth, forming the oceans and other bodies of water. Water moves between these different reservoirs through the **hydrological cycle**. Water is evaporated from the oceans, lakes, streams, the surface of the land, and plants (transpiration) by solar energy (Figure 4.2.1). It is moved through the atmosphere by winds and condenses to form clouds of water droplets or ice crystals. It comes back down as rain or snow and then flows through streams and rivers, into lakes, and eventually back to the oceans. Water on the surface and in streams and lakes infiltrates the ground to become groundwater. Groundwater slowly moves through the rock and surface materials; some returns to other streams and lakes, and some goes directly back to the oceans.

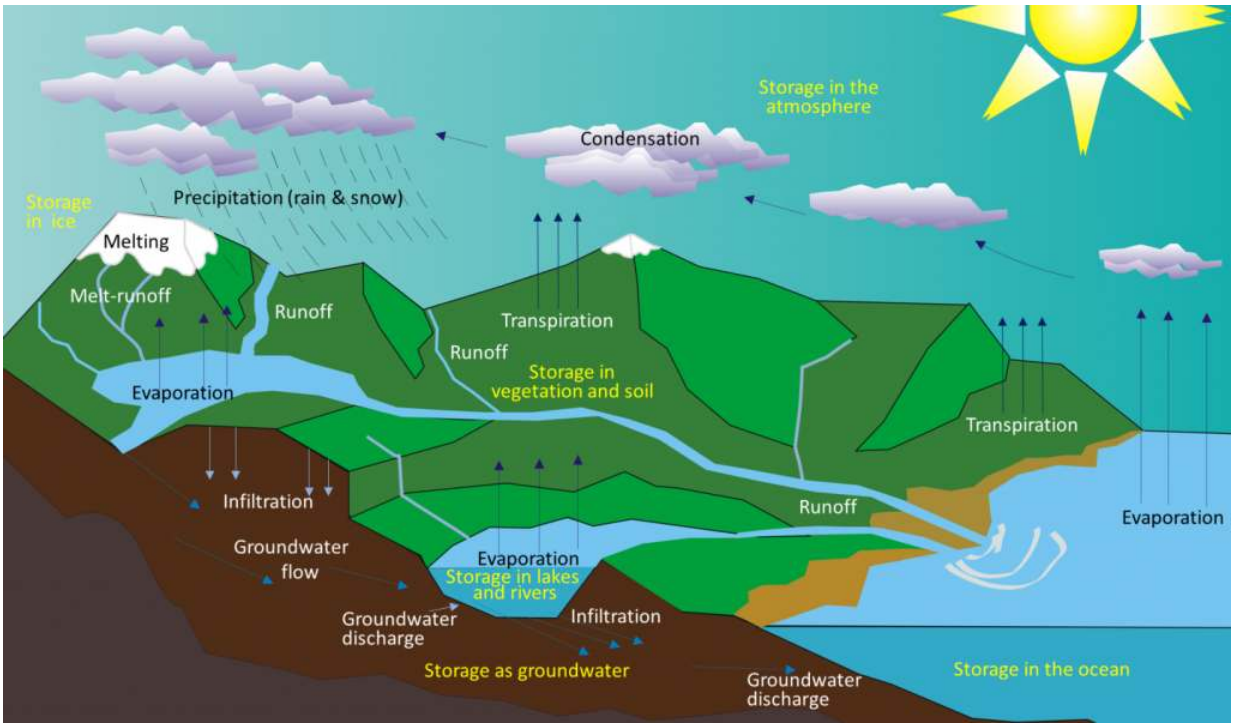


Figure 4.2.1 Earth's hydrological cycle (Steven Earle, "Physical Geology").

Water is stored in various reservoirs as it moves through this cycle. The largest, by far, is the oceans, accounting for 97% of the volume (Figure 4.2.2). Of course, that water is salty. The remaining 3% is fresh water. Two-thirds of our fresh water is stored in the ground and one-third is stored in ice. The remaining fresh water — about 0.03% of the total — is stored in lakes, streams, vegetation, and the atmosphere.

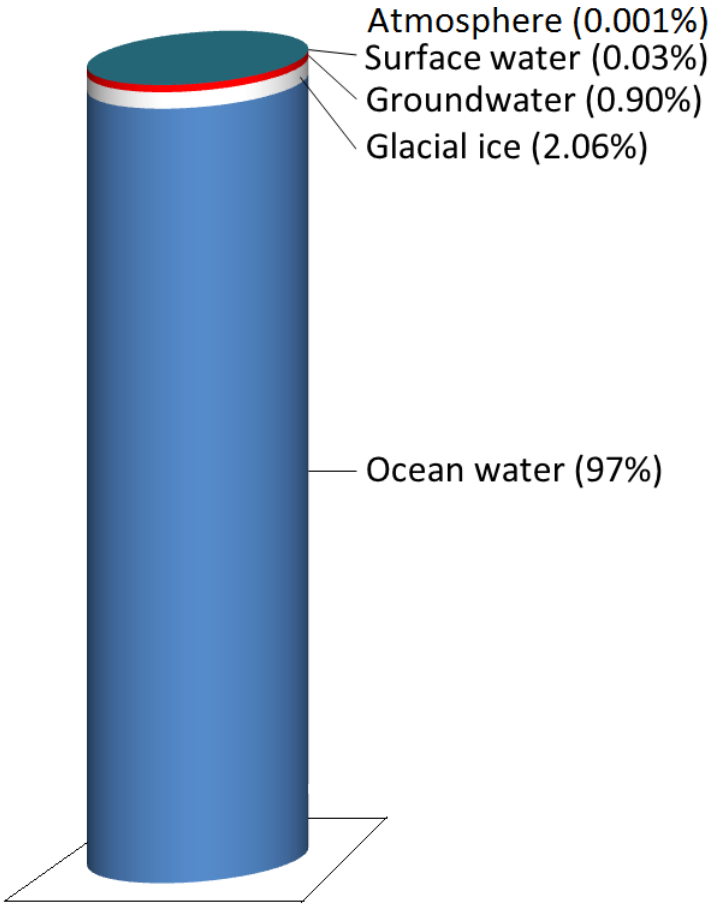


Figure 4.2.2 Proportions of Earth's water found in the various reservoirs (Steven Earle, "Physical Geology").

To put that in perspective, let's think about putting all of Earth's water into a 1 L jug. We start by almost filling the jug with 970 ml of water and 34 g of salt. Then we add one regular-sized (~20 mL) ice cube (representing glacial ice) and two teaspoons (~10 mL) of groundwater. All of the water that we see around us in lakes and streams and up in the sky can be represented by adding three more drops from an eyedropper.

Although the proportion of Earth's water that is in the atmosphere is tiny, the actual volume is huge. At any given time, there is the equivalent of approximately 13,000 km^3 of water in the air in the form of water vapor and water droplets in clouds. Water is evaporated from the oceans, vegetation, and lakes at a rate of 1,580 km^3 per day, and just about exactly the same volume falls as rain and snow every day, over both the oceans and land. The precipitation that falls on land goes back to the ocean in the form of stream flow (117 km^3/day) and groundwater flow (6 km^3/day).

How did the oceans get salty?

Outgassing was responsible for ocean formation, but how did the ocean water get salty? Most of the salts and dissolved elements in the ocean were probably outgassed along with the water vapor, so the ocean has probably always been about as salty as it is now. But we know that rainfall and other processes weather rocks on the Earth's surface, and runoff carries dissolved substances into the ocean, contributing to its salinity. Yet despite this constant input, the ocean's salt composition remains essentially the same. Therefore, the rate of input of new material must be balanced by the rate of removal; in other words, the oceans are in a **steady state** in regards to salinity.

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CHAPTER OVERVIEW

5: Basic Geography

5.1: Latitude and Longitude

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5.1: Latitude and Longitude

Any point on Earth can be defined by the intersection of its lines of latitude and longitude. **Latitude** is measured as the angle from the equator, to the Earth's center, to your position on the Earth's surface (Figure 5.1.1). It is expressed as degrees north or south of the equator (0°), with the poles at a latitude of 90°. Thus the poles are referred to as high latitude, while the equatorial region is considered low latitude. Lines of equal latitude are always the same distance apart, and so they are called **parallels** of latitude; they never converge. However, the parallels of latitude do get shorter as they approach the poles.

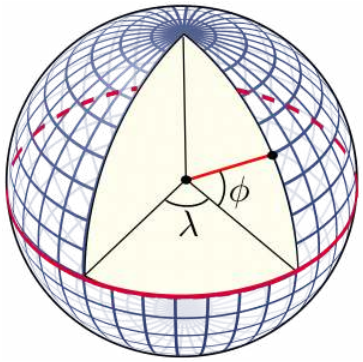


Figure 5.1.1 The latitude of a point on the Earth's surface is determined by the angle (ø) between the point and the equator, passing through Earth's center (Peter Mercator [Public domain], via Wikimedia Commons).

One degree of latitude is divided into 60 minutes (‘). One minute of latitude equals one **nautical mile**, which is equal to 1.15 land miles (1.85 km). Each minute of latitude is further divided into 60 seconds (“). So traditionally, positions have been expressed as degrees/minutes/seconds, e.g. 36° 15′ 32″ N. However, with modern digital technology, positions are increasingly expressed as decimals, such as 36° 15.25 N’, or 36.2597° N.

In the Northern Hemisphere, latitude can be determined by the angle of the North Star (Polaris) from the horizon. The North Star always sits over the North Pole. Here, if a person looks straight ahead towards the horizon, the star would be directly overhead, creating a 90° angle; thus the latitude at the North Pole is 90° N. At the equator looking north, the star is in the same direction as the horizon, so the angle between them is 0°, and thus the equatorial latitude is also 0°. At any other point in the Northern Hemisphere, the angle between the horizon and the star will give the latitude.

Early mariners used an instrument called an astrolabe to calculate this angle. Later the sextant was developed, which allowed more accurate measurements (Figure 5.1.2).

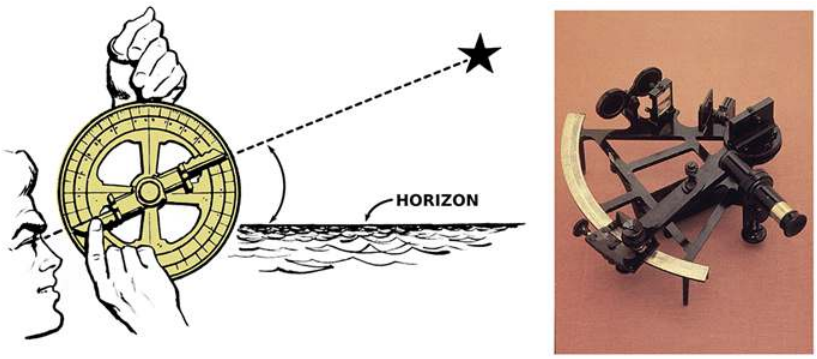


Figure 5.1.2 An astrolabe (left) and sextant (right) (Public domain via Wikimedia Commons).

There is no direct analogue to the North Star in the Southern Hemisphere that is useful for determining latitude. However, the Southern Cross and Centaurus constellations can be used to find the south celestial pole. If a line is drawn through the long axis of the Southern Cross, and another line is drawn between the two brightest stars in Centaurus, the two lines will intersect at the south celestial pole.

Longitude measures the distance east or west of an imaginary reference point, the **prime meridian** (0°), which is now defined as the line passing through Greenwich, England (although throughout history the prime meridian has also been located in Rome,

Copenhagen, Paris, Philadelphia, the Canary Islands, and Jerusalem; unlike the equator, the prime meridian's location is fairly arbitrary). Your longitude represents the angle east or west between your location, the center of the Earth, and the prime meridian (Figure 5.1.3).

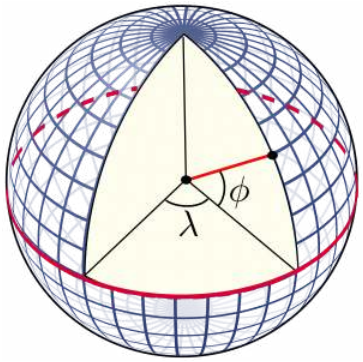


Figure 5.1.3 Longitude is determined as the angle (λ) between the prime meridian and your position (Peter Mercator [Public domain], via Wikimedia Commons).

As you move east and west from the prime meridian, eventually you reach 180° E and W on the opposite side of the globe from Greenwich. This point is the International Date Line. Lines of longitude are called **meridians** of longitude, or great circles. All circles of longitude are the same length, and are not parallel like lines of latitude; they converge as they near the poles. Therefore, while one minute of latitude always equals one nautical mile, the length of one minute of longitude will decline from the equator to to poles, where it will ultimately decline to zero.

Measuring longitude requires accurate time at your current location, and also the time at some distant point like a home port at the same instant. The time difference can be used to calculate longitude. This is because the Earth takes 24 hours for a complete 360° rotation. So in one hour, the Earth rotates through 1/24 of 360°, or 15°. Therefore, for each hour of time difference between two locations, there is a 15° difference in longitude.

Accurate measurements of latitude using the North Star have been made since at least the third century B.C.E. Because longitude measurements required accurate timekeeping, it wasn't until the mid-18th century that longitude was easily and precisely measured at sea. Before then, sailors would often sail north or south to get to the desired latitude, then just head east or west until they reached the target longitude. Solving the longitude problem was so important that the British government passed the Longitude Act in 1714, offering a £20,000 prize to anyone who could devise a method of measuring longitude at sea to within half a degree. Many unsuccessful solutions were proposed, including astronomical observations, but it was a clock maker, John Harrison, who developed a series of clocks that eventually satisfied the criteria. The first version (the H1) weighed over 80 lbs, but his final timepieces, the H4 and H5, could be held in the palm of one hand. Ironically, even though his clocks satisfied the criteria, Harrison was never named as the winner of the longitude prize, and in fact no winner was ever officially determined. With accurate timepieces now available, a ship could have one clock set for Greenwich time (or some other home location), and another clock set to local time, which could be reset each day by observing the sun. The time difference between the two clocks could be used to calculate longitude.

Today we use **GPS** (Global Positioning System) technology to determine latitude and longitude, and even the smallest smart phones and smart watches can use GPS to calculate position. GPS works through a system of orbiting satellites that constantly emit signals containing the time and their position. A GPS receiver receives these signals from multiple satellites, and triangulates the signals to calculate position. The system needs 24 satellites to be functional at one time; as of 2015, the system consisted of about 32 operational satellites, able to give a position with an accuracy of 9 meters (30 feet) or less.

Additional links for more information:

- For more about John Harrison and the longitude problem, see http://www.rmg.co.uk/explore/astronomy-and-time/time-facts/harrison
- For more about the workings of GPS technology, visit https://www.gps.gov/

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CHAPTER OVERVIEW

6: Ocean Basins

6.1: Overview of the Oceans
6.2: Continental Margins
6.3: Marine Provinces
6.4: Mapping the Seafloor

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6.1: Overview of the Oceans

Let's begin by looking at a few basic facts about the oceans. We often think of Earth in terms of its land area, but in reality 71% of the Earth's surface is covered by oceans, while only 29% is land. Oceans cover an area of 139 million miles^2 or 361 million km^2, and contain a volume of about 1.37 billion km^3 of water. All of this water is not distributed equally over the Earth; 61% of the Northern Hemisphere is covered by oceans, while in the Southern Hemisphere the oceans cover 81% of the surface area (Figure 6.1.1).

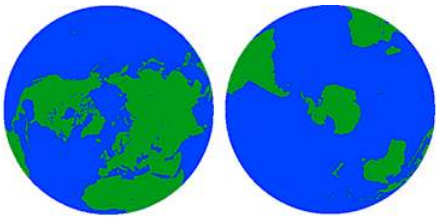


Figure 6.1.1 Ocean cover in the Northern (left) and Southern (right) Hemispheres.

Various sources differ in the number of recognized ocean basins. Historically the major oceans were recognized as the Pacific, Atlantic, Indian, and Arctic Oceans. More recently, the Southern Ocean has been recognized as fifth named ocean, comprising all of the water from the coast of Antarctica to 60° S (Figure 6.1.2). In 2000 these boundaries were submitted to the International Hydrographic Organization for official recognition, but several countries do not recognize it as a separate ocean, but rather as the southern extension of the other major oceans. The Southern Ocean has its own unique characteristics, so for the purposes of this book we will include it as a separate ocean.

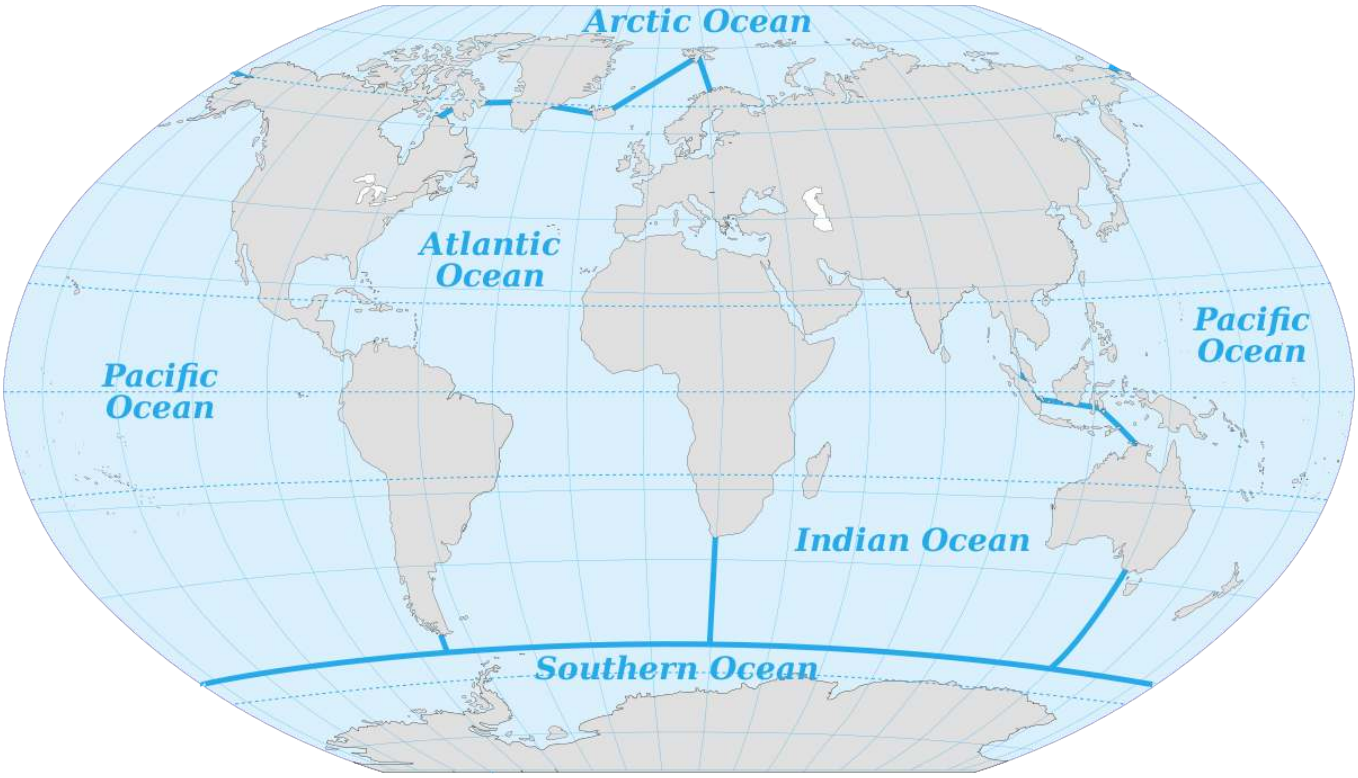


Figure 6.1.2 Map of the world oceans (By Pinpin [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 3.0], via Wikimedia Commons).

The oceans account for vast amounts of water, containing 97% of the water on Earth's surface, with over half of the water in the Pacific alone (Table 1.1.1).

Table 6.1.1 **Percentage of Earth's water in various locations**

Pacific	52%
Atlantic	25%
Indian	20%
Ice	2%
Ground water	0.6%
Atmosphere, lakes & rivers	0.01%

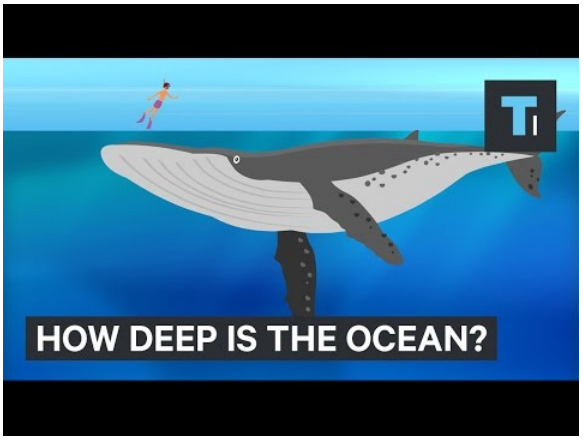
The average depth of the world ocean is about 3800m (12,500 ft), which is about four times deeper than the average land elevation is high (840m or 2800 ft). In fact Mt. Everest, the highest point on land, is 8848m (29,028 ft) high, while the deepest part of the ocean, the Challenger Deep of the Marianas Trench is 11,022m (36,200 ft) deep. So you could submerge Mt. Everest in the Marianas Trench and it would still be covered by over 2 km of water! Because there is so much more water on Earth than there is land, if you could smooth out the land elevation the entire Earth would still be covered by water about 2700 m deep.

Of the major ocean basins, the Pacific is the largest (almost as large as all of the others combined), and is the deepest (Table 1.1.2).

Table 6.1.2 **Area and depth of the major oceans**

	Area (million km^2)	Average depth (m)
Pacific	166	4282
Atlantic	87	3926
Indian	73	3963
Arctic	14	1205
Southern	20	4000

Watch the video below for some perspective on the size and depth of the oceans.



A YouTube element has been excluded from this version of the text. You can view it online here: https://rwu.pressbooks.pub/webboceanography/?p=24

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6.2: Continental Margins

Continental margins refer to the region of transition from the land to the deep seafloor, i.e. between continental and oceanic crust. In an **active continental margin**, the boundary between the continent and the ocean is also a tectonic plate boundary, so there is a lot of geological activity around the margin. The west coast of the United States is an example of an active margin, where the coastline corresponds with the boundary between the Pacific and North America Plates. A **passive continental margin** occurs where the transition from land to sea is not associated with a plate boundary. The east coast of the United States is a good example; the plate boundary is located along the mid Atlantic ridge, far from the coast. Passive margins are less geologically active. Figure 6.2.1 shows an idealized passive margin. When examining this figure, and others like it, note that there is significant vertical exaggeration; the depth scale covers approximately 5000 m, while the horizontal scale extends around 300 km. This makes the features look much steeper than they actually are. The bar at the bottom of Figure 6.2.1 shows what a passive margin would look like without this exaggeration; there is a much more gradual transition to depth.

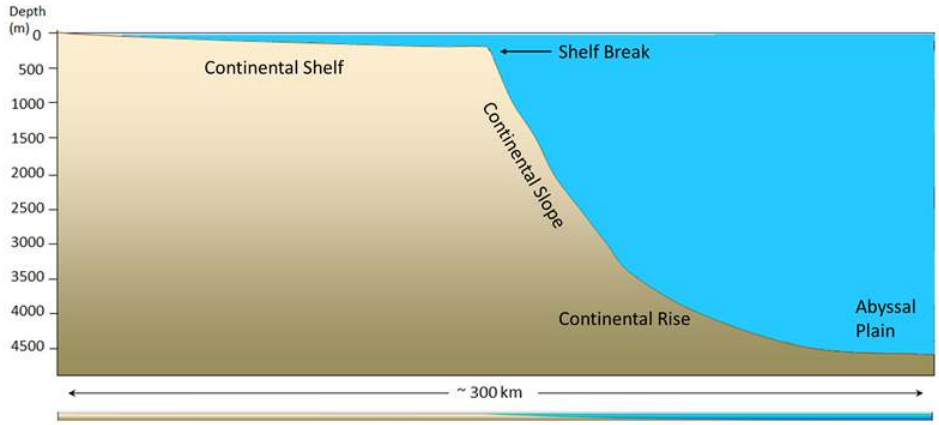


Figure 6.2.1 Features of a passive continental margin (modified by PW from Steven Earle, “Physical Geology”).

The **continental shelf** is the shallow, flooded edge of the continent. Geologically the shelf is still part of the continental crust, but it is often overlaid with marine sediments. On average, the shelf extends about 80 km from the coast; some margins have very little shelf, while the Siberian Shelf in the Arctic extends roughly 1500 km. The depth of the shelf generally remains below about 150 m, and the floor of the shelf is fairly flat. The flat topography is the result of changes in sea level; throughout history the shelves have been both submerged and exposed, and as sea level rose and fell, wave action, ice sheets, and other erosional processes smoothed out the shelf surface. Wave action and the movement of sediments over the shelf have continued this smoothing process. Continental shelves only make up about 6% of the ocean’s surface area, but they are biologically one of the richest parts of the ocean; their shallow depth prevents nutrients from sinking out, and their proximity to the coast provides significant nutrient input. The continental shelf ends at the **shelf break**, which is the point where the angle of the seafloor begins to get steeper. The shelf break averages about 135 m deep.

After the shelf break, the seafloor takes on a steeper angle (about 4°) as it descends to the deep ocean. This steeper portion of the margin is the **continental slope**, and it extends from the shelf break down to 3000-5000m. In some parts of the ocean, large submarine canyons have been carved into the continental slope; for example, Monterey Canyon in Monterey Bay, California, is a submarine canyon similar in size to the Grand Canyon! These canyons may be carved out by **turbidity currents**, which are essentially landslides of sediment, rocks, and other debris down the face of the slope.

At the bottom of the slope is the **continental rise**. This area represents where the continental crust meets the oceanic crust, as the slope begins to level off to become the deep ocean floor. The rise consists of a thick layer of accumulated sediment coming from the continent, so it is difficult to tell where the slope ends and the rise begins.

After the rise comes the **abyssal plain**, or the deep ocean floor, lying between 4500 – 6000 m. The abyssal plain includes most of the ocean floor, and is the flattest region on Earth. It is flat due to millions of years of sediment accumulation on the bottom, which buries many bottom features (Figure 6.2.2).

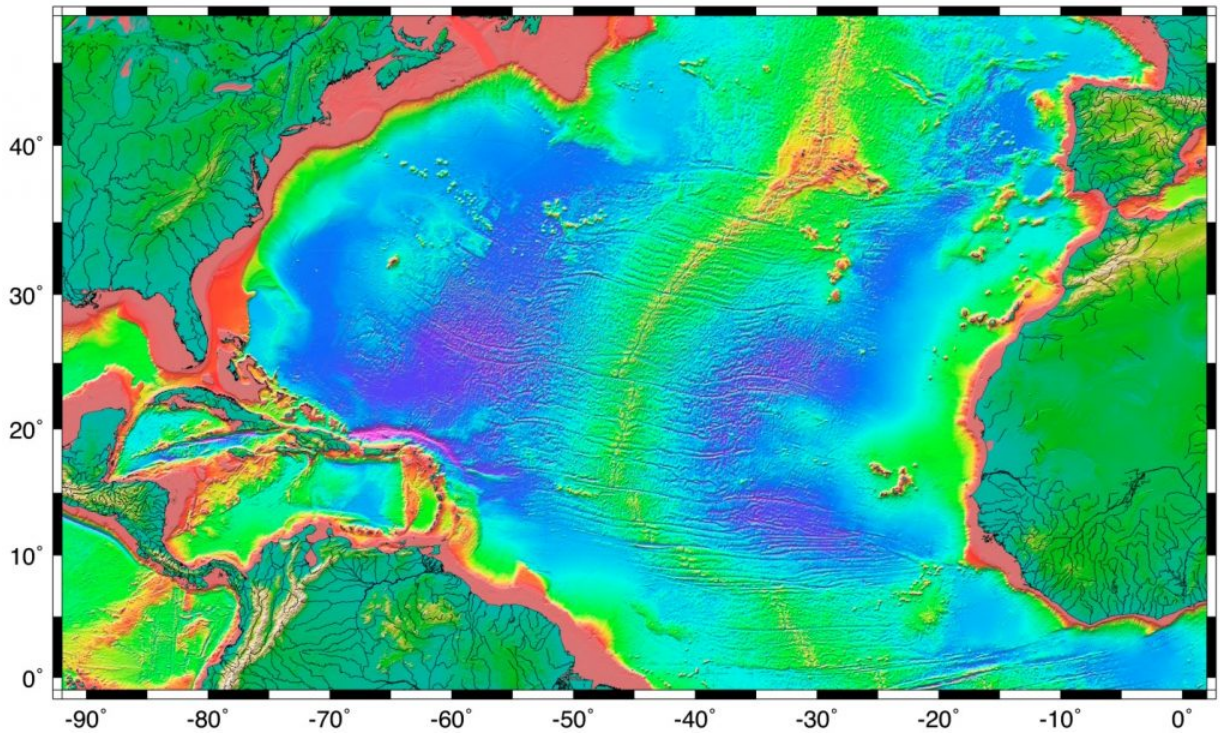


Figure 6.2.2 Topography of the North Atlantic. Shallow continental shelf regions are shown in red, and the abyssal plain is shown in blue. Along the east coast of the United States the continental slope can be seen in green (Steven Earle, "Physical Geology").

Passive margins, as described above, have wide shelves, gentle slopes, and a well-developed rise. Since passive margins are not plate boundaries, they experience long periods of relative stability which can lead to the development of these features. Active margins have similar features to passive margins, but the plate boundary affects the properties of the features. Active margins, like the Pacific coast of North America, have narrower shelves, steeper slopes, and little to no rise, particularly in convergent boundaries. Trenches associated with subduction zones act as sediment traps, preventing the accumulation of a continental rise, and keeping sediments off of the abyssal plains.

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6.3: Marine Provinces

In the previous section, we learned about the regions that make up the continental margins. So before we leave this topic, we will look at some of the other ways we can categorize the ocean environments.

The first major distinction is between the pelagic and benthic zones. The **pelagic** zone refers to the water column, where swimming and floating organisms live. The **benthic** zone refers to the bottom, and organisms living on and in the bottom are known as the **benthos**.

The pelagic zone is divided into two provinces: the **neritic** province corresponds to all of the water from the low tide line to the shelf break, while the **oceanic** province represents all of the other water in the open ocean regions.

The oceanic province is divided into depth zones (Figure 6.3.1):

- 0-200 m is the **epipelagic** zone ("epi" = "upon", as in on top of the pelagic zone). This is the region where enough light penetrates the water to support photosynthesis (see section 7.3), so it is also called the **euphotic** or **photic** zone.
- 200-1000 m is the **mesopelagic** zone ("meso" = "middle"). There is some light here, but not enough for photosynthesis, so it is called the **dysphotic zone**, or the **twilight zone**.
- 1000-4000 m is the **bathypelagic** zone ("bathy" = "deep"). There is no light at these depths, so it is referred to as the **aphotic** zone. About 75% of the living space in the ocean lies at these depths or deeper.
- 4000-6000 m is the **abyssopelagic** or **abyssalpelagic** zone, which extends to the seafloor in most areas.
- 6000 m and below is the **hadopelagic** or **hadalpelagic** zone (named for Hades or "hell"). This refers to the water in deep ocean trenches.

Inhabitants of these regions are referred to according to their habitat, for example mesopelagic fish, or epipelagic squid.

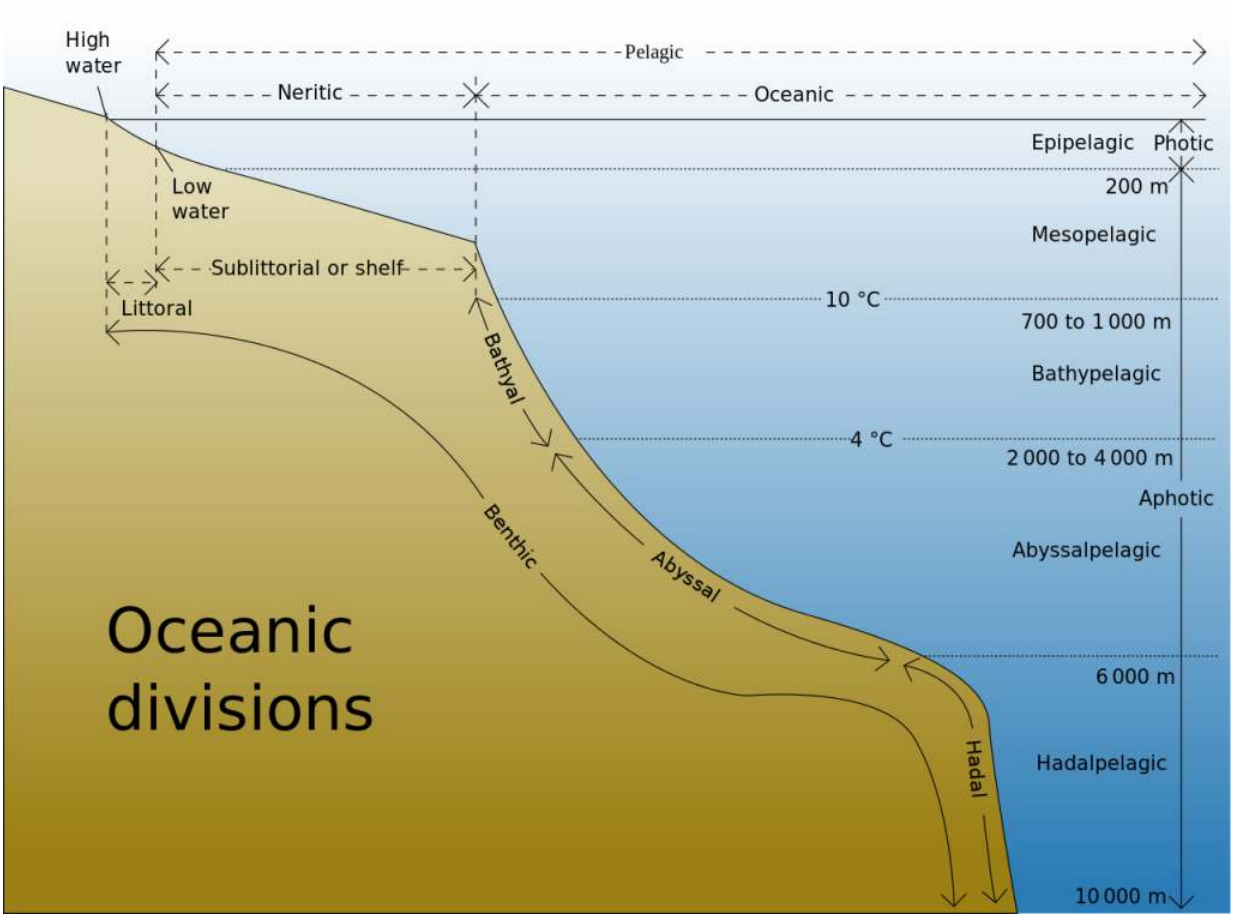


Figure 6.3.1 The major benthic and pelagic oceanic divisions (K. Aainsqatsi at en.wikipedia [Public domain, GFDL (http://www.gnu.org/copyleft/fdl.html) or CC-BY-SA-3.0], via Wikimedia Commons).

The benthic environment is also divided into zones, most of which correspond to the pelagic divisions:

- The **supralittoral** zone lies above the high tide line. Also called the spray zone, it is only submerged during storms or unusually high waves.
- The **littoral** zone is the region between the high and low tides. Thus it is also referred to as the **intertidal** zone.
- Below the littoral zone is the **sublittoral** (shelf) zone, extending from the low tide mark to the shelf break, essentially covering the continental shelf.
- The **bathyal** zone extends along the bottom from the shelf break to 4000m, so it generally includes the continental slope and rise.
- The **abyssal** zone is found between 4000-6000 m, including most of the abyssal plains. The abyssal zone represents about 80% of the benthic environment.
- The **hadal** zone includes all benthic regions deeper than 6000 m, such as in the bottom of trenches.

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6.4: Mapping the Seafloor

The previous sections included some information on the depths of the oceans in various places. So how are we able to map the ocean floor to ascertain these vast depths?

To map the ocean floor we need to know the depth at a number of places. The process of measuring the depths is known as **bathymetry**. These measurements were first made through **soundings**, where a weighted line (lead line) was let out by hand until it touched the bottom, and the depth could be recorded from the length of the line (Figure 6.4.1). This technique led to the fathom as a unit of depth; as sailors hauled in the sounding line they would stretch it out to cover their arm span. The average arm span of a sailor was about six feet, so one fathom equals six feet, and the sailors could simply count the number of "arm spans" as they pulled in the line.



Figure 6.4.1 Lead line survey from a catamaran hull in Alaska, 1942 (http://celebrating200years.noaa.gov/...hy/image7.html).

This technique had a number of drawbacks, and was usually limited to shallower water. It was very time consuming, and only gave depth data for a single point, so many individual soundings were needed to map an area. It could also be error-prone; in deep water it could be difficult to determine when the weight hit the bottom as the weight of the line itself could cause the line to keep sinking, and currents could deflect the line away from vertical, thus overestimating the depth. In later years, winches and heavy steel cables were used for deeper water, but this did not solve all of the problems inherent in the sounding method, and also added the constraint of excessive weight of the equipment.

In the 19th century, a number of modifications were made to this simple design. In 1802 the British clockmaker Edward Massey invented a mechanical device that was attached to the sounding line; as the device sank, a rotor turned a dial which locked in place when the line hit bottom (Figure 6.4.2). The line could then be reeled in and the depth read from the dial. In 1853 American sailor John Mercer Brooke developed a cannonball weight attached to a twine reel. The cannonball was dropped over the side and allowed to free-fall to the bottom; by timing the fall rate (the rate at which the twine unspooled) and noting when the rate changed as the cannonball hit the bottom, the water depth could be calculated. When it hit bottom, the cannonball was released and the line could be hauled back in, bringing with it a sample of mud in the iron bar that held the cannonball, thus confirming that the bottom had been reached.

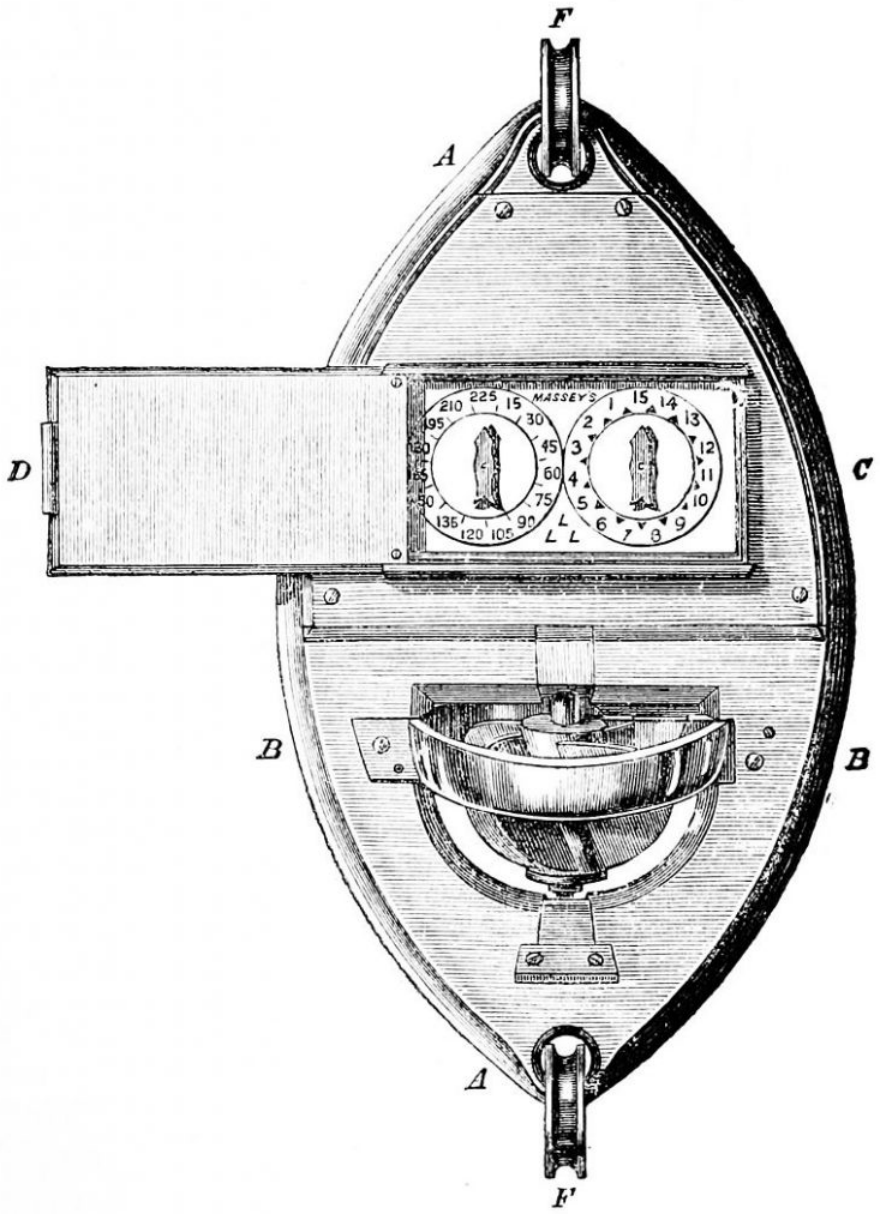


Figure 6.4.2 Massey's sounding machine (Public domain, via Wikimedia Commons).

After the *Titanic* disaster in 1912, there was an effort to develop better methods of detecting icebergs from a ship. This led to the development of **sonar** (SOund Navigation And Ranging) technology, which was soon applied to mapping bathymetry. A sonar device called an echosounder sends out a pulse of sound, then listens for the returning echo. The timing of the returning echo is used to calculate depth. We know that the speed of sound in water is approximately 1500 m/s (see section 6.4). Since the returning echo traveled to the bottom and back, the water depth corresponds to half the time it takes for an echo to return, multiplied by the speed of sound in water (Figure 6.4.3):

depth=1/2*(two-way travel time)*(speed of sound in water)

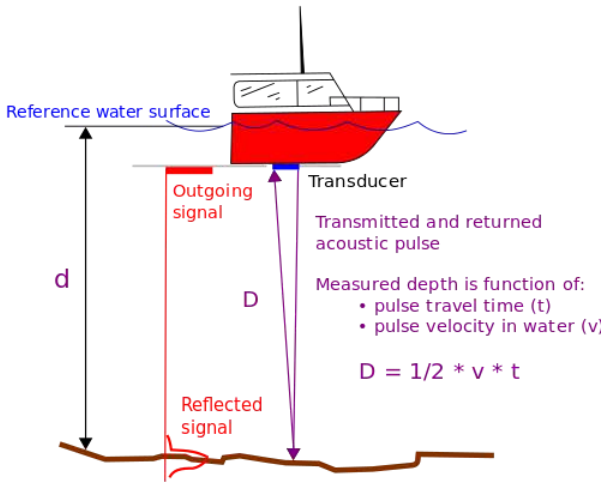


Figure 6.4.3 Measuring depth using an echosounder (Public domain via Wikimedia Commons).

Echosounders allowed a fast, continuous record of bathymetry under a moving ship. However, they only give the depth directly under a ship's path. Today, high resolution seafloor maps are made through multibeam or side scan sonar, either from a ship or from a towed transmitter (Figure 6.4.4). Multibeam sonar produces a fan-shaped acoustic field allowing a much a wider area (>10 km wide) to be mapped simultaneously.

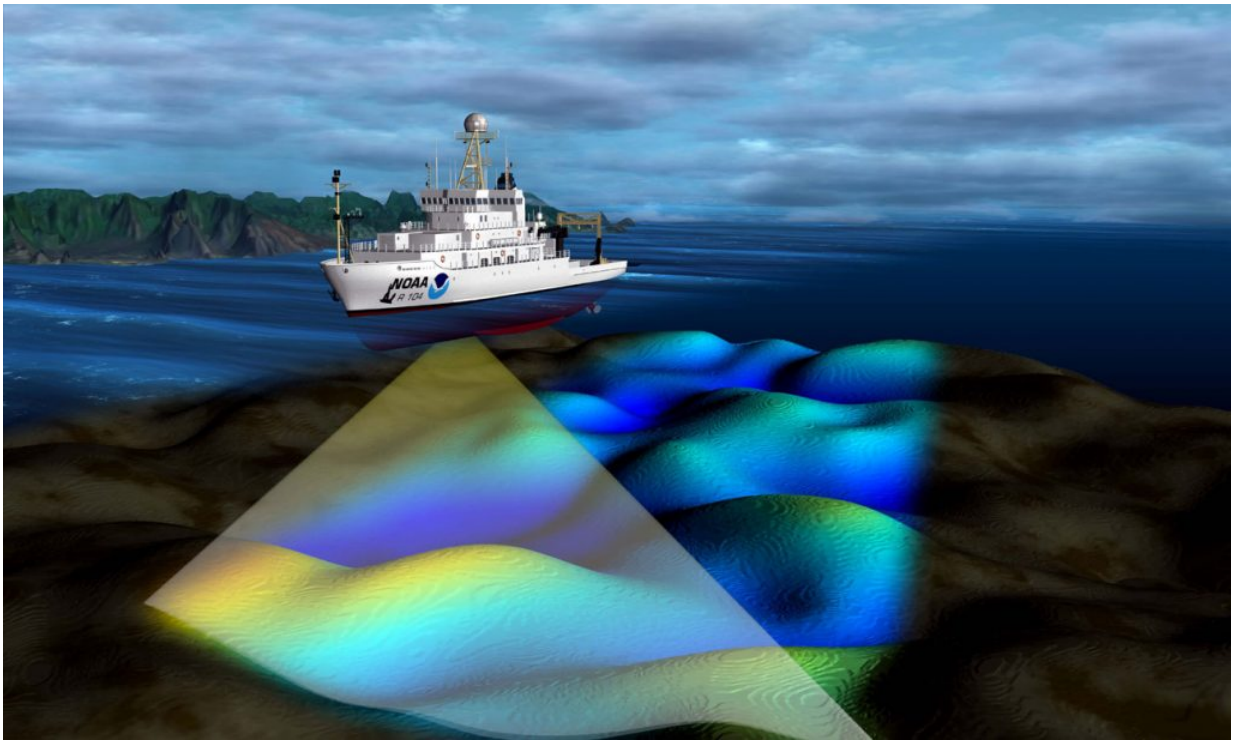


Figure 6.4.4 Multibeam sonar (NOAA).

Large-scale mapping of the ocean floor is also carried out by satellites (originally SEASAT, then GEOSAT, now the Jason satellites) which use radio waves to measure the height of the sea surface (radar altimetry). The sea surface is not flat; gravity causes it to be slightly higher over elevated features on the ocean floor, and slightly lower over trenches and other depressions. Satellites send out radio waves, and similar to an echosounder, can use the returning waves to detect differences in sea surface height down to 3-6 cm (Figure 6.4.5). These differences in sea surface heights allow us to determine the topography under the surface. Unlike the old lead line technology, where hundreds of soundings were necessary to map a small area, each taking an hour or more to complete, the current satellites can map over 90% of Earth's ice-free sea surface every 10 days!

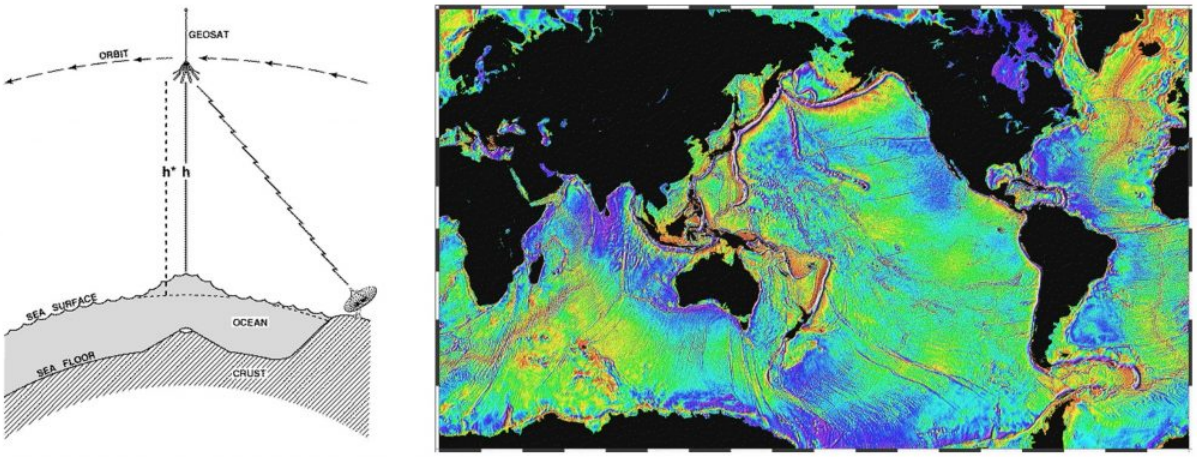


Figure 6.4.5 Radar altimetry (left) and a map of the seafloor produced by radar altimetry satellites (right) (NOAA).

Additional links for more information:

- NOAA page on side scan sonar: http://www.nauticalcharts.noaa.gov/hsd/SSS.html
- NOAA page on exploring the ocean using satellite altimetry data: https://www.ngdc.noaa.gov/mgg/bathym...d/explore.HTML

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CHAPTER OVERVIEW

7: Ocean Sediments

7.1: Classifying Sediments
7.2: Lithogenous Sediments
7.3: Biogenous Sediments
7.4: Hydrogenous Sediments
7.5: Sediment Distribution

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7.1: Classifying Sediments

The term “sediment” refers to the tiny particles of rocks and other materials that sink to the ocean floor and eventually settle and accumulate on the bottom. All regions of the seafloor contain some form of sediment, although there are many different types of sediments from a variety of sources, and the amount of accumulated sediment can vary greatly from place to place. Globally, ocean sediments average about 1 km thick, but they can exceed 15 km thick in areas of high accumulation (Figure 7.1.1). These areas include regions near the mouths of rivers where there is high sediment discharge, and passive margins near the continents where the seafloor has had millions of years for sediment to accumulate. On the other hand, sediments are sparse along divergent plate boundaries where new oceanic crust is being formed, as the crust is too new for significant accumulation, and in the central oceans that are far away from any significant sediment sources.

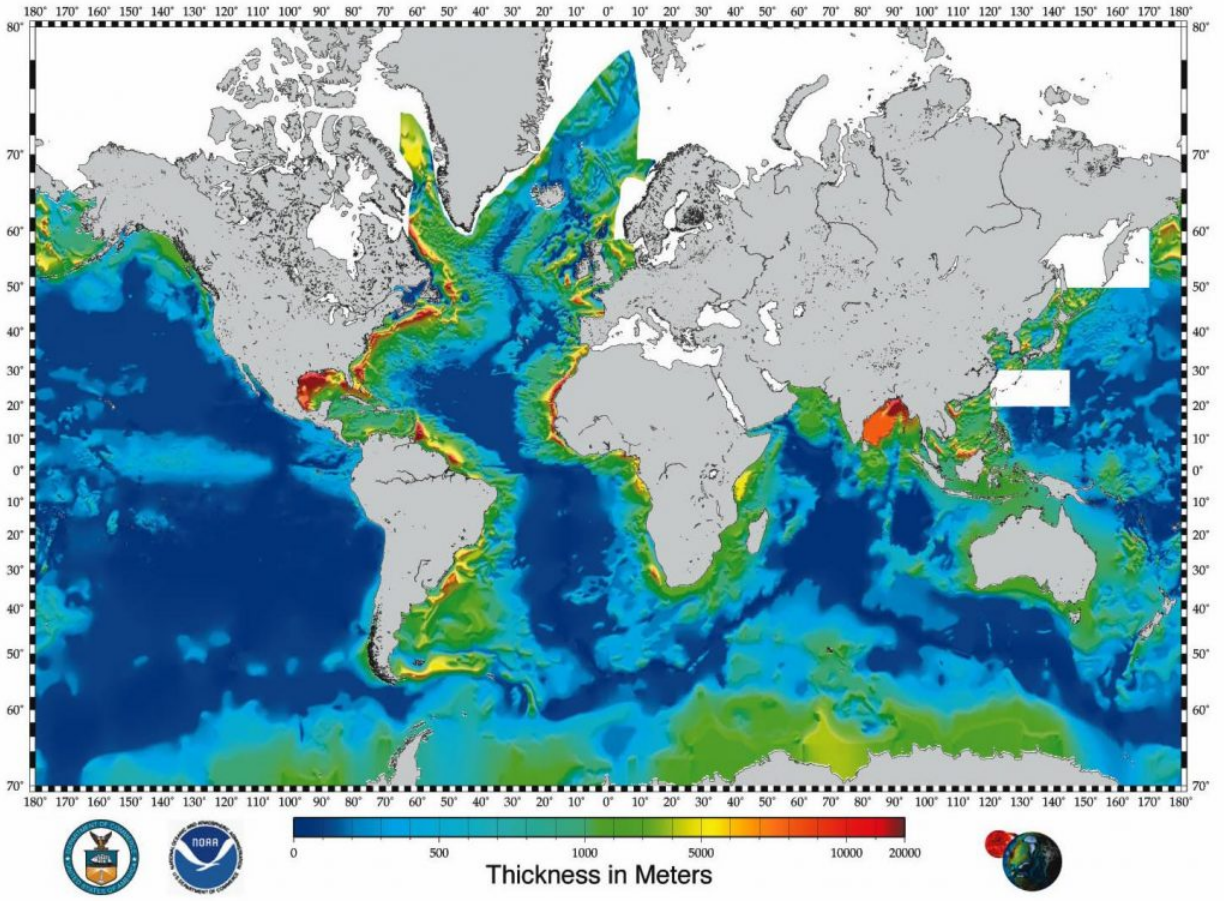


Figure 7.1.1 Total sediment thickness of the world ocean (By Divins, D.L., NGDC Total Sediment Thickness of the World’s Oceans & Marginal Seas, http://www.ngdc.noaa.gov/mgg/sedthick/sedthick.html (http://www.ngdc.noaa.gov/mgg/image/sedthick9.jpg) [Public domain], via Wikimedia Commons).

As time passes over millions of years, these sediments can become **lithified** or turned into sedimentary rock. It has been estimated that over half of the exposed rock on the continents is sedimentary rock originally deposited in ancient oceans and uplifted by plate tectonics. Many tall mountains, including Mt. Everest, are composed of rock formations that contain fossils of marine creatures. These rocks were originally formed as ocean sediments which were then lithified and pushed upwards during the process of mountain formation.

Sediment Classification

There are a number of ways that we can classify ocean sediments, and some of the most common distinctions are based on the sediment texture, the sediment composition, and the sediment’s origin.

Texture

Sediment texture can be examined through several variables. The first is **grain size**. Sediments are classified by particle size, ranging from the finest clays (diameter <0.004 mm) to the largest boulders (> 256 mm)(Figure 7.1.2). Among other things, grain size represents the conditions under which the sediment was deposited. High energy conditions, such as strong currents or waves, usually results in the deposition of only the larger particles as the finer ones will be carried away. Lower energy conditions will allow the smaller particles to settle out and form finer sediments.

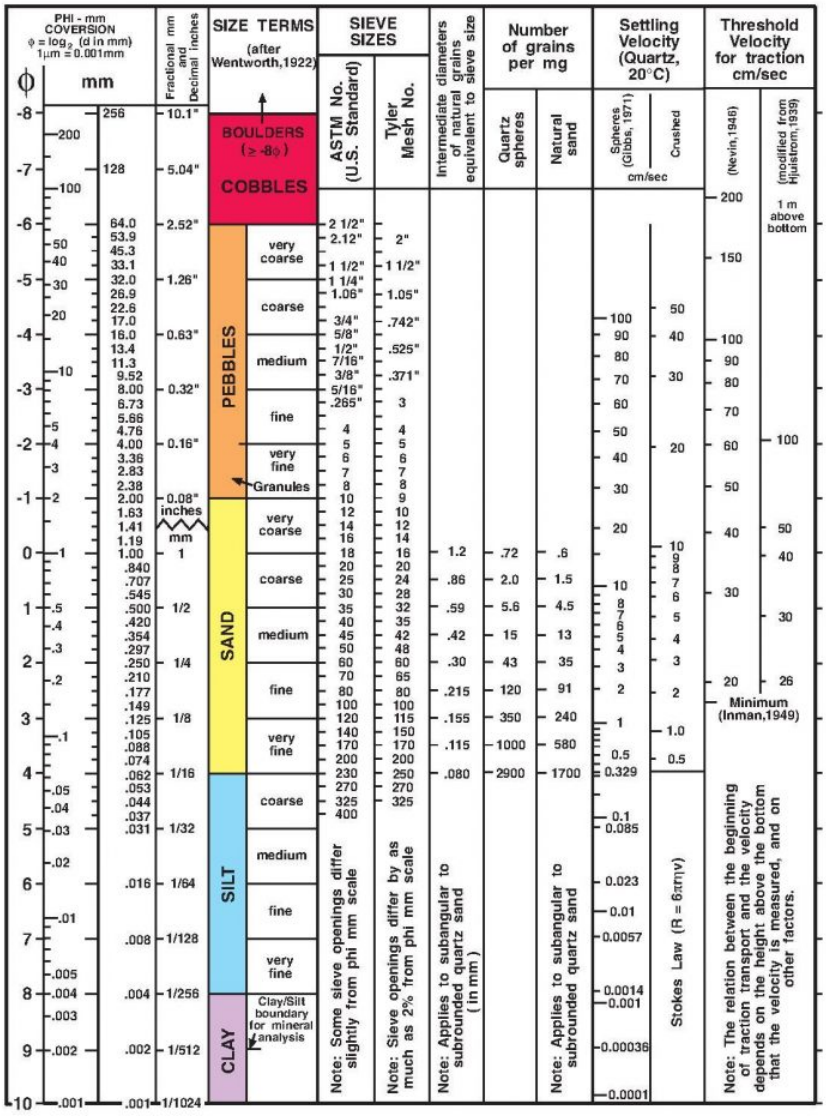


Figure 7.1.2 Wentworth grain size chart for classifying sediments (By Jeffress Williams, Matthew A. Arsenault, Brian J. Buczkowski, Jane A. Reid, James G. Flocks, Mark A. Kulp, Shea Penland, and Chris J. Jenkins, USGS [Public domain], via Wikimedia Commons).

Sorting is another way to categorize sediments. Sorting refers to how uniform the particles are in terms of size (Figure 7.1.3). If all of the particles are of a similar size, such as in beach sand, the sediment is well-sorted. If the particles are of very different sizes, the sediment is poorly sorted, such as in glacial deposits.

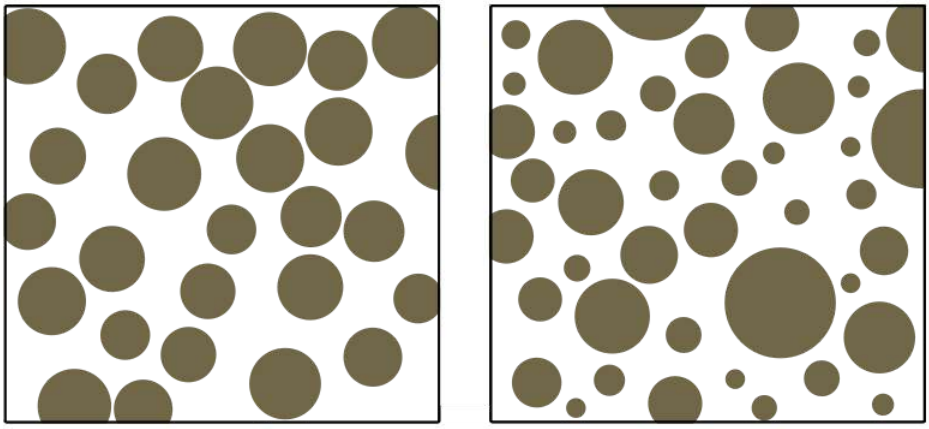


Figure 7.1.3 Well-sorted sediments (left) have particles that are all of a similar size. Poorly sorted sediments (right) consist of particles of a wide range of sizes (Woudloper (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

Finally, sediment texture can be described based on its **maturity**, or how long the particles have been transported by water or other vectors. Maturity can be reflected by the degree of rounding of the particles, the amount of sorting, and the composition of the sediment. In the case of **rounding**, the more mature the sediment, the rounder the particles, as a result of the particles being abraded over time. A high degree of sorting indicates maturity, because over time the smaller particles will be washed away, and a given amount of energy will move particles of a similar size over the same distance. Lastly, the older and more mature a sediment the higher the quartz content, at least in sediments derived from rock particles. Quartz is a common mineral in terrestrial rocks, and it is very hard and resistant to abrasion. Over time, particles made from other materials are worn away, leaving only quartz behind. Beach sand is a very mature sediment; it is composed primarily of quartz, and the particles are rounded and of similar size (well-sorted).

Sources of sediments

Sediments are also classified based on their source of origin. There are four main categories for the origin of marine sediments:

- **Lithogenous** sediments are derived from preexisting rock. They are also called **terrigenous** sediments since most of it comes from the land masses and makes its way into the ocean.
- **Biogenous** sediments are composed of the remains of marine organisms.
- **Hydrogenous** sediments are formed when materials that are dissolved in water precipitate out and form solid particles.
- **Cosmogenous** sediments are derived from extraterrestrial sources.

The next few sections will address each of these sediment types in more detail.

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7.2: Lithogenous Sediments

Lithogenous or terrigenous sediment is primarily composed of small fragments of preexisting rocks that have made their way into the ocean. These sediments can contain the entire range of particle sizes, from microscopic clays to large boulders, and they are found almost everywhere on the ocean floor. Lithogenous sediments are created on land through the process of weathering, where rocks and minerals are broken down into smaller particles through the action of wind, rain, water flow, temperature- or ice-induced cracking, and other erosive processes. These small eroded particles are then transported to the oceans through a variety of mechanisms:

- **Streams and rivers**: Various forms of runoff deposit large amounts of sediment into the oceans, mostly in the form of finer-grained particles (Figure 7.2.1). About 90% of the lithogenous sediment in the oceans is though to have come from river discharge, particularly from Asia. Most of this sediment, especially the larger particles, will be deposited and remain fairly close to the coastline, however, smaller clay particles may remain suspended in the water column for long periods of time and may be transported great distances from the source.



Figure 7.2.1 River discharge in the Yukon Delta, Alaska. The pale color demonstrates the large amounts of sediment released into the ocean via the rivers (By Jesse Allen and Robert Simmon (NASA Earth Observatory) [Public domain], via Wikimedia Commons).

- **Wind**: Wind-borne (**aeolian**) transport can take small particles of sand and dust and move them thousands of kilometers from the source. These small particles can fall into the ocean when the wind dies down, or can serve as the nuclei around which raindrops or snowflakes form. Aeolian transport is particularly important near desert areas (Figure 7.2.2).

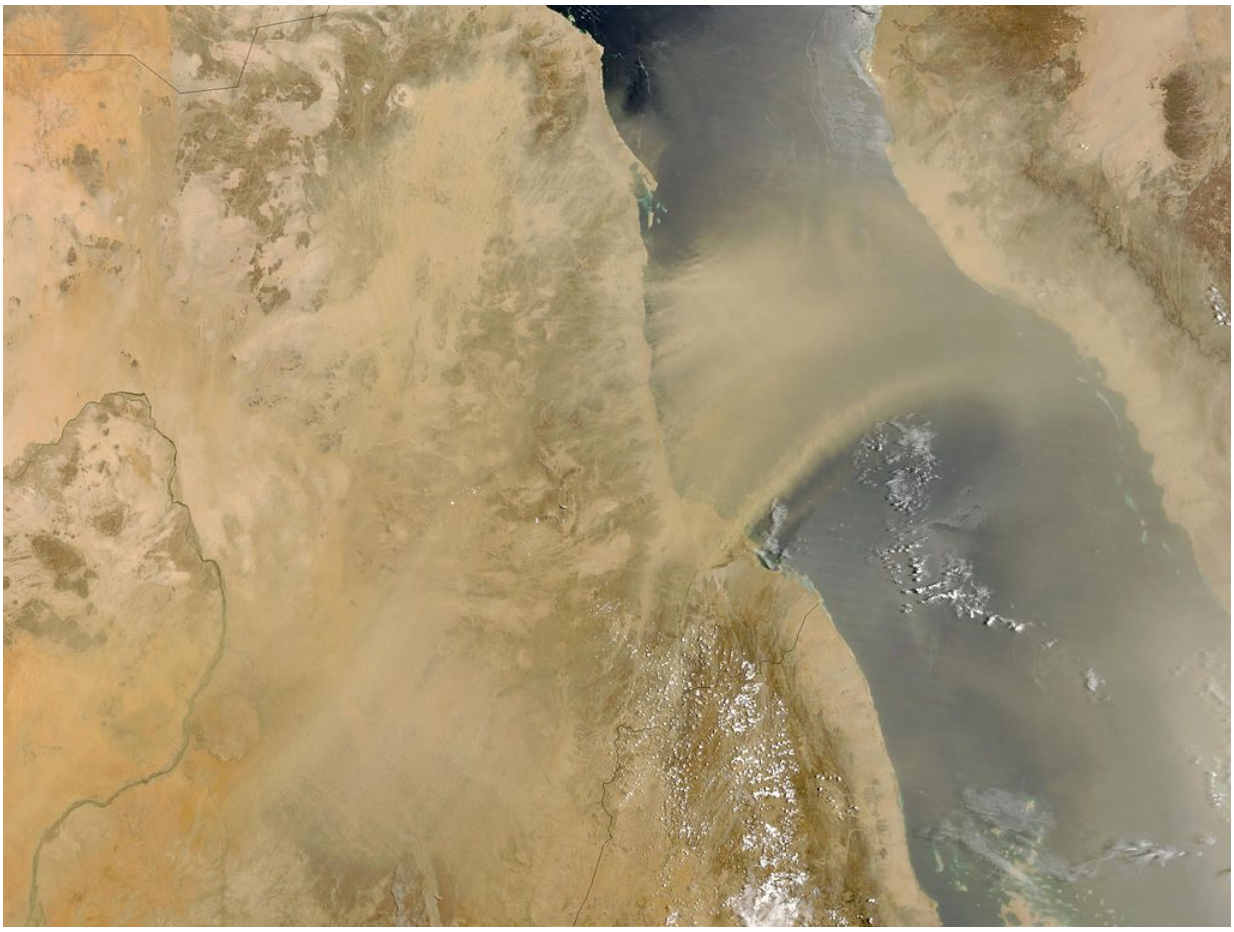


Figure 7.2.2 A plume of wind-borne particles from Sudan (left) blow over the Red Sea (By NASA (http://visibleearth.nasa.gov/view_rec.php?id=5645) [Public domain], via Wikimedia Commons).

- **Glaciers and ice rafting**: As glaciers grind their way over land, they pick up lots of soil and rock particles, including very large boulders, that get carried by the ice. When the glacier meets the ocean and begins to break apart or melt, these particles get deposited (Figure 7.2.3). Most of the deposition will happen close to where the glacier meets the water, but a small amount of material is also transported longer distances by rafting, where larger pieces of ice drift far from the glacier before releasing their sediment.



Figure 7.2.3 When glaciers reach the sea they can break apart, depositing their sediments into the ocean, including very large pieces of rock (By Ianaré Sévi (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

- **Gravity**: Landslides, mudslides, avalanches, and other gravity-driven events can deposit large amounts of material into the ocean when they happen close to shore.

- **Waves**: Wave action along a coastline will erode rocks and will pull loose particles from beaches and shorelines into the water.
- **Volcanoes**: Volcanic eruptions emit vast amounts of ash and other debris into the atmosphere, where it can then be transported by wind to eventually get deposited in the oceans (Figure 7.2.4).



Figure 7.2.4 Eruption of the Mayon Volcano, Philippines, in 1984. Much of the material spewed from a volcanic eruption may eventually make its way into the oceans (By C.G. Newhall [Public domain], via Wikimedia Commons).

- **Gastroliths**: An interesting, although relatively minor avenue for the transport of lithogenous sediments to the ocean is in the form of gastroliths. Gastrolith means “stomach stones” and comes from the fact that many animals, including seabirds, pinnipeds, and some crocodiles will deliberately swallow stones in one area and regurgitate them in another. Often these stones swallowed on land will be regurgitated at sea. Why swallow stones? Possible explanations include using the stones to help grind up food in the stomach, to act as ballast to aid in buoyancy regulation, or to fill the stomach and reduce feelings of hunger during fasting periods on shore.

Most of these processes deposit lithogenous sediment fairly close to shore. The sediment particles can then be transported farther away by waves and currents, where they may eventually escape the continental shelf and reach the deep ocean floor.

Composition

Lithogenous sediments usually reflect the composition of whatever materials they were derived from, so they are dominated by the major minerals that make up most terrestrial rock. This includes quartz, feldspar, clay minerals, iron oxides, and terrestrial organic matter. Quartz (silicon dioxide, the main component of glass) is one of the most common minerals found in nearly all rocks, and it is very resistant to abrasion, so it is a dominant component of lithogenous sediments, including sand.

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7.3: Biogenous Sediments

Biogenous sediments come from the remains of living organisms that settle out as sediment when the organisms die. It is the “hard parts” of the organisms that contribute to the sediments; things like shells, teeth or skeletal elements, as these parts are usually mineralized and are more resistant to decomposition than the fleshy “soft parts” that rapidly deteriorate after death.

Macroscopic sediments contain large remains, such as skeletons, teeth, or shells of larger organisms. This type of sediment is fairly rare over most of the ocean, as large organisms don’t die in enough of a concentrated abundance to allow these remains to accumulate. One exception is around coral reefs; here there is a great abundance of organisms that leave behind their remains, in particular the fragments of the stony skeletons of corals that make up a large percentage of tropical sand.

Microscopic sediment consists of the hard parts of microscopic organisms, particularly their shells, or **tests**. Although very small, these organisms are highly abundant and as they die by the billions every day their tests sink to the bottom to create biogenous sediments. Sediments composed of microscopic tests are far more abundant than sediments from macroscopic particles, and because of their small size they create fine-grained, mushy sediment layers. If the sediment layer consists of at least 30% microscopic biogenous material, it is classified as a biogenous **ooze**. The remainder of the sediment is often made up of clay.

The primary sources of microscopic biogenous sediments are unicellular algaes and protozoans (single-celled amoeba-like creatures) that secrete tests of either calcium carbonate ($CaCO_3$) or silica (SiO_2). Silica tests come from two main groups, the **diatoms** (algae) and the **radiolarians** (protozoans) (Figure 7.3.1).

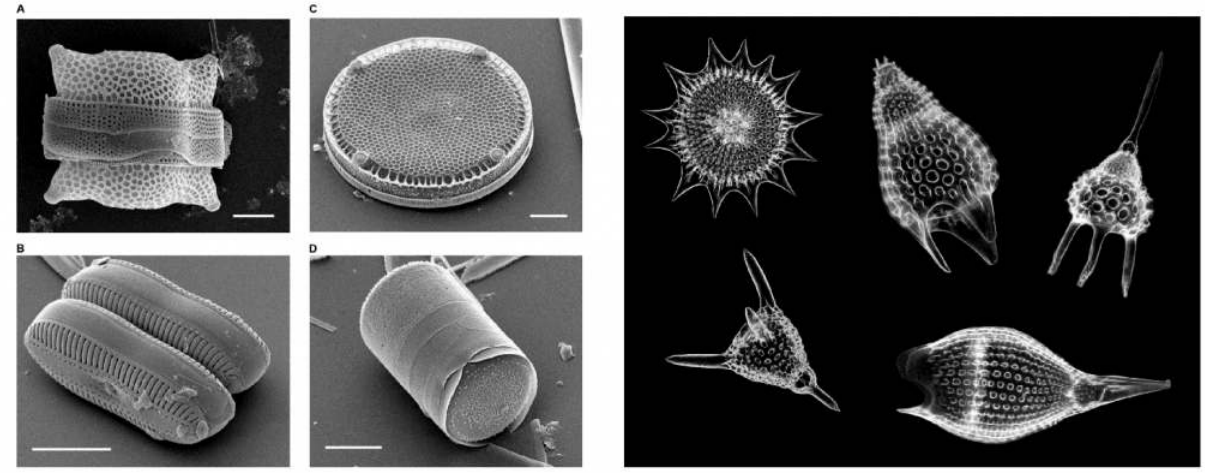


Figure 7.3.1 Various diatom (left) and radiolarian (right) tests (Diatom images courtesy of Mary Ann Tiffany, San Diego State University [CC BY 2.5], via Wikimedia Commons; radiolarian images by Andreas Drews, https://pxhere.com/en/photo/239774, [CC by 2.0]).

Diatoms are important members of the phytoplankton, the small, drifting algal photosynthesizers. A diatom consists of a single algal cell surrounded by an elaborate silica shell that it secretes for itself. Diatoms come in a range of shapes, from elongated, pennate forms, to round, or centric shapes that often have two halves, like a Petri dish (Figure 7.3.1 *left*). In areas where diatoms are abundant, the underlying sediment is rich in silica diatom tests, and is called **diatomaceous earth** (see box below).

What use are diatoms?

Diatoms are a vital piece of the global ecosystem for their role in oceanic primary production and the creation of much of the oxygen that organisms breathe. But diatoms are also important for many industrial and agricultural applications. Because of the very fine grain size, and the lattice-like structure of the diatom tests, diatomaceous earth has been used as a filtering agent in things like swimming pool filters and beer brewing. The microscopic tests have been added as an abrasive to toothpaste, facial cleansers and household cleaning agents. Alfred Nobel used diatomaceous earth to stabilize nitroglycerine in the production of dynamite. Diatomaceous earth also displays insecticide properties by stimulating dehydration in insects. It is marketed for this purpose in agriculture, as well as for household use to combat ants, cockroaches, and bedbugs. “Food grade” diatomaceous earth has also

entered the market, with proponents touting a range of health benefits arising from its consumption. That's a pretty impressive range of uses from a microscopic algae!

Radiolarians are planktonic protozoans (making them part of the zooplankton), that like diatoms, secrete a silica test. The test surrounds the cell and can include an array of small openings through which the radiolarian can extend an amoeba-like "arm" or pseudopod (Figure 7.3.1 *right*). Radiolarian tests often display a number of rays protruding from their shells which aid in buoyancy. Oozes that are dominated by diatom or radiolarian tests are called **siliceous oozes**.

Like the siliceous sediments, the calcium carbonate, or calcareous sediments are also produced from the tests of microscopic algae and protozoans; in this case the **coccolithophores** and **foraminiferans**. Coccolithophores are single-celled planktonic algae about 100 times smaller than diatoms. Their tests are composed of a number of interlocking $CaCO_3$ plates (coccoliths) that form a sphere surrounding the cell (Figure 7.3.2 *left*). When coccolithophores die the individual plates sink out and form an ooze. Over time, the coccolithophore ooze lithifies to becomes chalk. The famous White Cliffs of Dover in England are composed of coccolithophore-rich ooze that turned into chalk deposits (Figure 7.3.2 *right*).

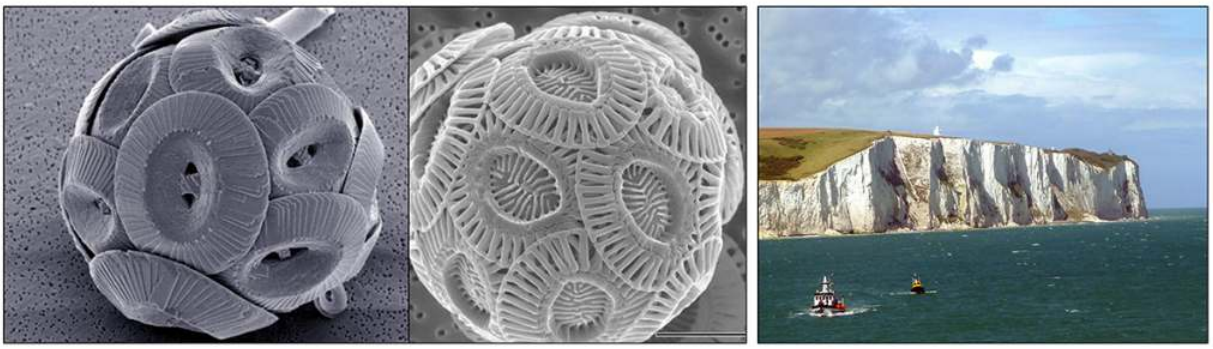


Figure 7.3.2 (**Left**) coccolithophore tests (left; By Richard Lampitt, Jeremy Young, The Natural History Museum, London (http://planktonnet.awi.de/); center; by Alison R. Taylor (University of North Carolina Wilmington Microscopy Facility) (PLoS Biology, June 2011, Cover ([1])) [Both images CC BY 2.5], via Wikimedia Commons). (**Right**); the White Cliffs of Dover (Immanuel Giel (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

Foraminiferans (also referred to as "forams") are protozoans whose tests are often chambered, similar to the shells of snails. As the organism grows, is secretes new, larger chambers in which to reside. Most foraminiferans are benthic, living on or in the sediment, but there are some planktonic species living higher in the water column. When coccolithophores and foraminiferans die, they form **calcareous oozes**.



Figure 7.3.3 Foraminifera tests collected from a beach in Myanmar (By Psammophile [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 3.0], via Wikimedia Commons).

Older calcareous sediment layers contain the remains of another type of organism, the **discoasters**; single-celled algae related to the coccolithophores that also produced calcium carbonate tests. Discoaster tests were star-shaped, and reached sizes of 5-40 µm across (Figure 7.3.4). Discoasters went extinct approximately 2 million years ago, but their tests remain in deep, tropical sediments that predate their extinction.

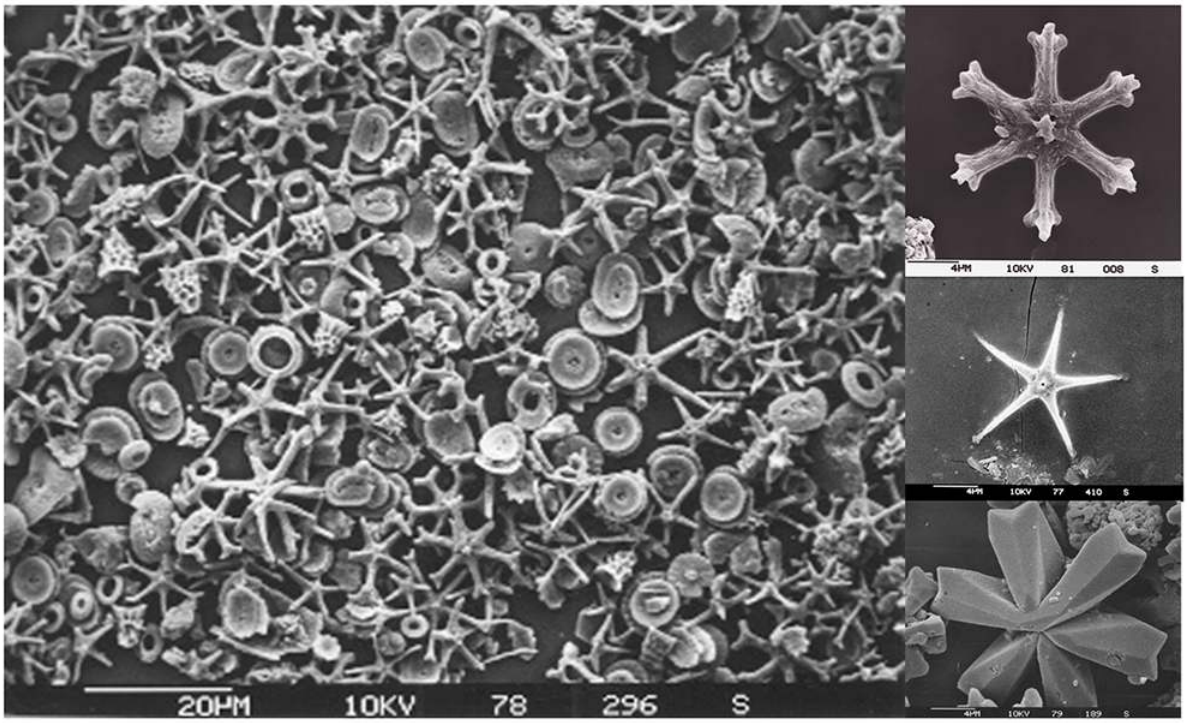


Figure 7.3.4 Discoaster tests. Left: discoster tests with assorted coccoliths. Top right *Discoaster surculus;* center right: *Discoaster pentaradiatus;* bottom right: *Discoaster surculus* (All images by Hannes Grobe (Own work) [CC BY 3.0], via Wikimedia Commons).

Because of their small size, these tests sink very slowly; a single microscopic test may take about 10-50 years to sink to the bottom! Given that slow descent, a current of only 1 cm/sec could carry the test as much as 15,000 km away from its point of origin before it reaches the bottom. Yet despite this, we find that the sediments in a particular location are well-matched to the types of organisms and degree of productivity that occurs in the water overhead. This means that the sediment particles must be sinking to the bottom at a much faster rate, so that they accumulate below their point of origin before the currents can disperse them. What is the mechanism for this increased sinking rate? Apparently most of the tests do not sink as individual particles; about 99% of them are first consumed by some other organism, and are then aggregated and expelled as large fecal pellets, which sink much more quickly and reach the ocean floor in only 10-15 days. This does not give the particles as much time to disperse, and the sediment below will reflect the production occurring near the surface. The increased rate of sinking through this mechanism is called the "fecal express."

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7.4: Hydrogenous Sediments

Seawater contains many different dissolved substances. Occasionally chemical reactions occur that cause these substances to precipitate out as solid particles, which then accumulate as **hydrogenous sediment**. These reactions are usually triggered by a change in conditions, such as a change in temperature, pressure, or pH, which reduces the amount of a substance that can remain in a dissolved state. There is not a lot of hydrogenous sediment in the ocean compared to lithogenous or biogenous sediments, but there are some interesting forms.

In **hydrothermal vent** systems, seawater percolates into the seafloor, where it becomes superheated by magma before being expelled by the vent. This superheated water contains many dissolved substances, and when it encounters the cold seawater after leaving the vent, these particles precipitate out, mostly as metal sulfides. These particles make up the “smoke” that flows from a vent, and may eventually settle on the bottom as hydrogenous sediment (Figure 7.4.1).

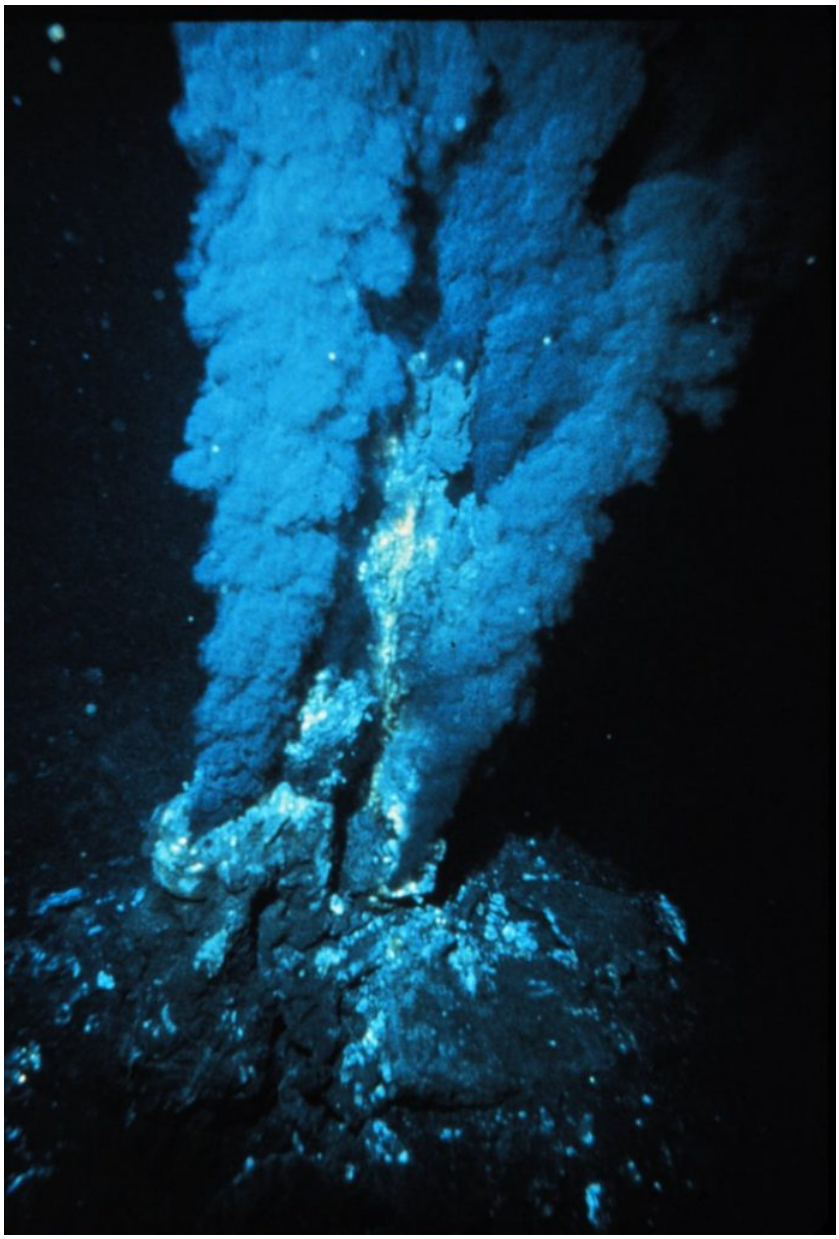


Figure 7.4.1 A “black smoker” hydrothermal vent. The “smoke” consists of dissolved particles that precipitate into solids when exposed to colder water (NOAA, http://www.photolib.noaa.gov/htmls/nur04506.htm).

Manganese nodules are rounded lumps of manganese and other metals that form on the seafloor, generally ranging between 3-10 cm in diameter, although they may sometimes reach up to 30 cm (Figure 7.4.2). The nodules form in a manner similar to pearls; there is a central object around which concentric layers are slowly deposited, causing the nodule to grow over time. The composition of the nodules can vary somewhat depending on their location and the conditions of their formation, but they are usually dominated by manganese- and iron oxides. They may also contain smaller amounts of other metals such as copper, nickel and cobalt. The precipitation of manganese nodules is one of the slowest geological processes known; they grow on the order of a few millimeters per million years. For that reason, they only form in areas where there are low rates of lithogenous or biogenous sediment accumulation, because any other sediment deposition would quickly cover the nodules and prevent further nodule growth. Therefore, manganese nodules are usually limited to areas in the central ocean, far from significant lithogenous or biogenous inputs, where they can sometimes accumulate in large numbers on the seafloor (Figure 7.4.2 *right*). Because the nodules contain a number of commercially valuable metals, there has been significant interest in mining the nodules over the last several decades, although most of the efforts have thus far remained at the exploratory stage. A number of factors have prevented large-scale extraction of nodules, including the high costs of deep sea mining operations, political issues over mining rights, and environmental concerns surrounding the extraction of these non-renewable resources.

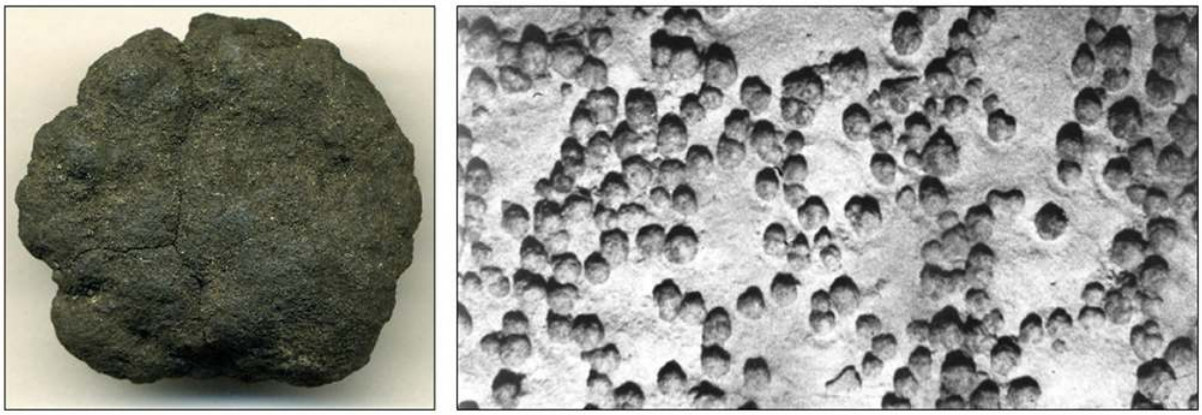


Figure 7.4.2 (Left) manganese nodule from the subtropical eastern Pacific Ocean. The nodule is 4.1 cm in diameter (By James St. John, https://www.flickr.com/photos/jsjgeo...n/photostream/ [CC-BY 2.0]). (Right) field of manganese nodules on the seafloor (By United States Geological Survey [Public domain], via Wikimedia Commons).

Evaporites are hydrogenous sediments that form when seawater evaporates, leaving the dissolved materials to precipitate into solids, particularly halite (salt, NaCl). In fact, the evaporation of seawater is the oldest form of salt production for human use, and is still carried out today. Large deposits of halite evaporites exist in a number of places, including under the Mediterranean Sea. Beginning around 6 million years ago, tectonic processes closed off the Mediterranean Sea from the Atlantic, and the warm climate evaporated so much water that the Mediterranean was almost completely dried out, leaving large deposits of salt in its place (an event known as the Messinian Salinity Crisis). Eventually the Mediterranean re-flooded about 5.3 million years ago, and the halite deposits were covered by other sediments, but they still remain beneath the seafloor.



Figure 7.4.3 Salt farmers harvesting salt left behind from the evaporation of seawater, Pak Thale, Ban Laem, Phetchaburi, Thailand (By JJ Harrison (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

Oolites are small, rounded grains formed from concentric layers of precipitation of material around a suspended particle. They are usually composed of calcium carbonate, but they may also from from phosphates and other materials. Accumulation of oolites results in oolitic sand, which is found in its greatest abundance in the Bahamas (Figure 7.4.4).



Figure 7.4.4 Oolites from a beach on Joulter's Cay, The Bahamas (By Wilson44691 (Own work) [Public domain], via Wikimedia Commons).

Methane hydrates are another type of hydrogenous deposit with a potential industrial application. All terrestrial erosion products include a small proportion of organic matter derived mostly from terrestrial plants. Tiny fragments of this material plus other organic matter from marine plants and animals accumulate in terrigenous sediments, especially within a few hundred kilometers of shore. As the sediments pile up, the deeper parts start to warm up (from geothermal heat), and bacteria get to work breaking down the contained organic matter. Because this is happening in the absence of oxygen (a.k.a. anaerobic conditions), the by-product of this metabolism is the gas methane (CH_4). Methane released by the bacteria slowly bubbles upward through the sediment toward the seafloor. At water depths of 500 m to 1,000 m, and at the low temperatures typical of the seafloor (close to 4°C), water and methane combine to create a substance known as methane hydrate. Within a few meters to hundreds of meters of the seafloor, the temperature is low enough for methane hydrate to be stable and hydrates accumulate within the sediment (Figure 7.4.5 *left*). Methane hydrate is flammable because when it is heated, the methane is released as a gas (Figure 7.4.5 *right*). The methane within seafloor sediments represents an enormous reservoir of fossil fuel energy. Although energy corporations and governments are anxious to develop ways to produce and sell this methane, anyone that understands the climate-change implications of its extraction and use can see that this would be folly.



Figure 7.4.5 (Left): Methane hydrate within muddy sea-floor sediment from an area offshore from Oregon (By Wusel007 (Own work) [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 3.0], via Wikimedia Commons). (Right): Methane hydrate on fire (USGS, http://www.usgs.gov/blogs/features/f.../New-Image.jpg).

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7.5: Sediment Distribution

Now that we have an understanding of the types of sediments found in the ocean, we can turn our attention to the processes that cause different types of sediments to dominate in different locations. Sediment accumulation will depend on the the amount of material coming from the source, the distance from the source, the amount of time that sediment has had to accumulate, how well the sediments are preserved, and the amounts of other types of sediments that are also being added to the system.

Rates of sediment accumulation are relatively slow throughout most of the ocean, in many cases taking thousands of years for any significant deposits to form. Lithogenous sediment accumulates the fastest, on the order of 1 m or more per thousand years for coarser particles. However, sedimentation rates near the mouths of large rivers with high discharge can be orders of magnitude higher. Biogenous oozes accumulate at a rate of about 1 cm per thousand years, while small clay particles are deposited in the deep ocean at around 1 mm per thousand years. Manganese nodules have an incredibly slow rate of accumulation, gaining 0.001 mm per thousand years.

Marine sediments are thickest near the continental margins (refer to figure 7.5.1) where they can be over 10 km thick. This is because the crust near passive continental margins is often very old, allowing for a long period of accumulation, and because there is a large amount of terrigenous sediment input coming from the continents. Near mid-ocean ridge systems where new oceanic crust is being formed, sediments are thinner, as they have had less time to accumulate on the younger crust. As you move away from the ridge spreading center the sediments get progressively thicker, increasing by approximately 100-200 m of sediment for every 1000 km distance from the ridge axis. With a seafloor spreading rate of about 20-40 km/million years, this represents a sediment accumulation rate of approximately 100-200 m every 25-50 million years.

Figure 7.5.1 shows the distribution of the major types of sediment on the ocean floor. Cosmogenous sediments could potentially end up in any part of the ocean, but they accumulate in such small abundances that they are overwhelmed by other sediment types and thus are not dominant in any location. Similarly, hydrogenous sediments can have high concentrations in specific locations, but these regions are very small on a global scale. So we will mostly ignore cosmogenous and hydrogenous sediments in the discussion of global sediment patterns.

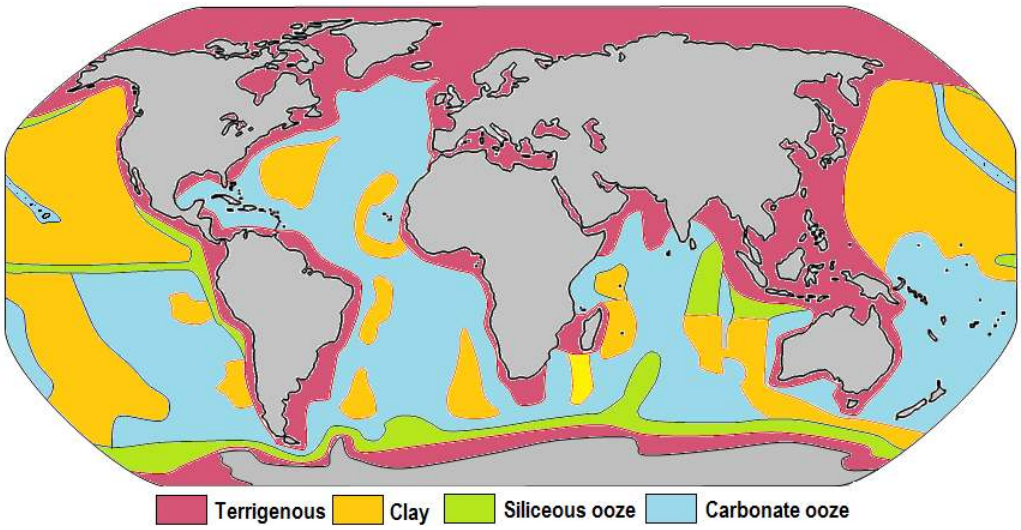


Figure 7.5.1 The distribution of sediment types on the seafloor. Within each colored area, the type of material shown is what dominates, although other materials are also likely to be present (Steven Earle, “Physical Geology”).

Coarse lithogenous/terrigenous sediments are dominant near the continental margins as runoff, river discharge, and other processes deposit vast amounts of these materials on the continental shelf. Much of this sediment remains on or near the shelf, while turbidity currents can transport material down the continental slope to the deep ocean floor. Lithogenous sediment is also common at the poles where thick ice cover can limit primary production, and glacial breakup deposits sediments along the ice edge. Coarse lithogenous sediments are less common in the central ocean, as these areas are too far from the sources for these sediments to accumulate. Very small clay particles are the exception, and as described below, they can accumulate in areas that other lithogenous sediment will not reach.

The distribution of biogenous sediments depends on their rates of production, dissolution, and dilution by other sediments. Coastal areas display very high primary production, so we might expect to see abundant biogenous deposits in these regions. However, recall that sediment must be >30% biogenous to be considered a biogenous ooze, and even in productive coastal areas there is so much lithogenous input that it swamps the biogenous materials, and that 30% threshold is not reached. So coastal areas remain dominated by lithogenous sediment, and biogenous sediments will be more abundant in pelagic environments where there is little lithogenous input.

In order for biogenous sediments to accumulate their rate of production must be greater than the rate at which the tests dissolve. Silica is undersaturated throughout the ocean and will dissolve in seawater, but it dissolves more readily in warmer water and lower pressures; in other words, it dissolves faster near the surface than in deep water. Silica sediments will therefore only accumulate in cooler regions of high productivity where they accumulate faster than they dissolve. This includes upwelling regions near the equator and at high latitudes where there are abundant nutrients and cooler water. Oozes formed near the equatorial regions are usually dominated by radiolarians, while diatoms are more common in the polar oozes. Once the silica tests have settled on the bottom and are covered by subsequent layers, they are no longer subject to dissolution and the sediment will accumulate. Approximately 15% of the seafloor is covered by siliceous oozes.

Biogenous calcium carbonate sediments also require production to exceed dissolution for sediments to accumulate, but the processes involved are a little different than for silica. Calcium carbonate dissolves more readily in more acidic water. Cold seawater contains more dissolved CO_2 and is slightly more acidic than warmer water. Therefore calcium carbonate tests are more likely to dissolve in colder, deeper, polar water than in warmer, tropical, surface water. At the poles the water is uniformly cold, so calcium carbonate readily dissolves at all depths, and carbonate sediments do not accumulate. In temperate and tropical regions calcium carbonate dissolves more readily as it sinks into deeper water. The depth at which calcium carbonate dissolves as fast as it accumulates is called the **calcium carbonate compensation depth**, or **calcite compensation depth**, or simply the **CCD**. The **lysocline** represents the depths where the rate of calcium carbonate dissolution increases dramatically (similar to the thermocline and halocline). At depths shallower than the CCD carbonate accumulation will exceed the rate of dissolution, and carbonate sediments will be deposited. In areas deeper than the CCD, the rate of dissolution will exceed production, and no carbonate sediments can accumulate (Figure 7.5.2). The CCD is usually found at depths of 4 – 4.5 km, although it is much shallower at the poles where the surface water is cold. Thus calcareous oozes will mostly be found in tropical or temperate waters less than about 4 km deep, such as along the mid-ocean ridge systems and atop seamounts and plateaus. The CCD is deeper in the Atlantic than in the Pacific since the Pacific contains more CO_2, making the water more acidic and calcium carbonate more soluble. This, along with the fact that the Pacific is deeper, means that the Atlantic contains more calcareous sediment than the Pacific. All told, about 48% of the seafloor is dominated by calcareous oozes.

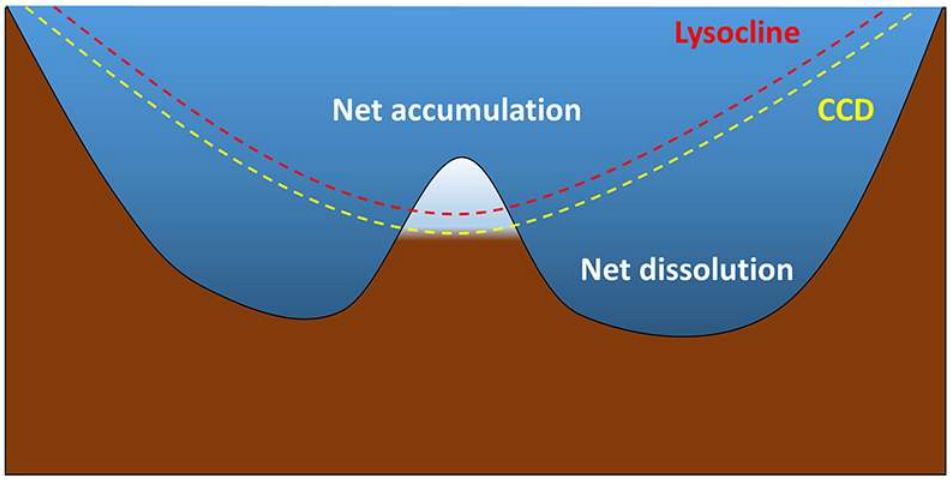


Figure 7.5.2 Calcareous sediment can only accumulate in depths shallower than the calcium carbonate compensation depth (CCD). Below the CCD, calcareous sediments dissolve and will not accumulate. The lysocline represents the depths where the rate of dissolution increases dramatically (PW).

Much of the rest of the deep ocean floor (about 38%) is dominated by abyssal clays. This is not so much a result of an abundance of clay formation, but rather the lack of any other types of sediment input. The clay particles are mostly of terrestrial origin, but

because they are so small they are easily dispersed by wind and currents, and can reach areas inaccessible to other sediment types. Clays dominate in the central North Pacific, for example. This area is too far from land for coarse lithogenous sediment to reach, it is not productive enough for biogenous tests to accumulate, and it is too deep for calcareous materials to reach the bottom before dissolving. Because clay particles accumulate so slowly, the clay-dominated deep ocean floor is often home to hydrogenous sediments like manganese nodules. If any other type of sediment was produced here it would accumulate much more quickly and would bury the nodules before they had a chance to grow.

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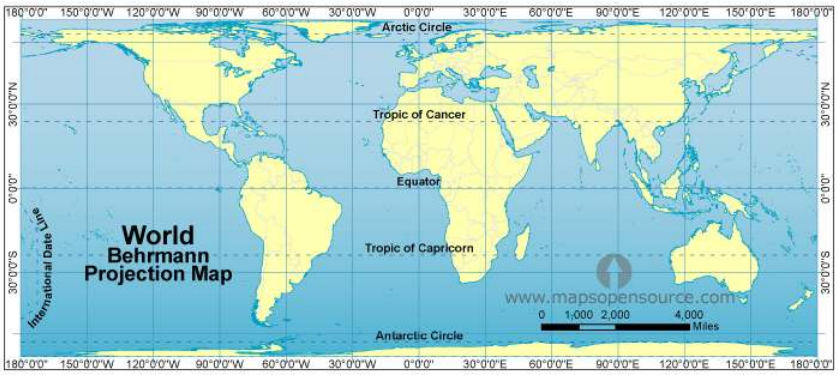
CHAPTER OVERVIEW

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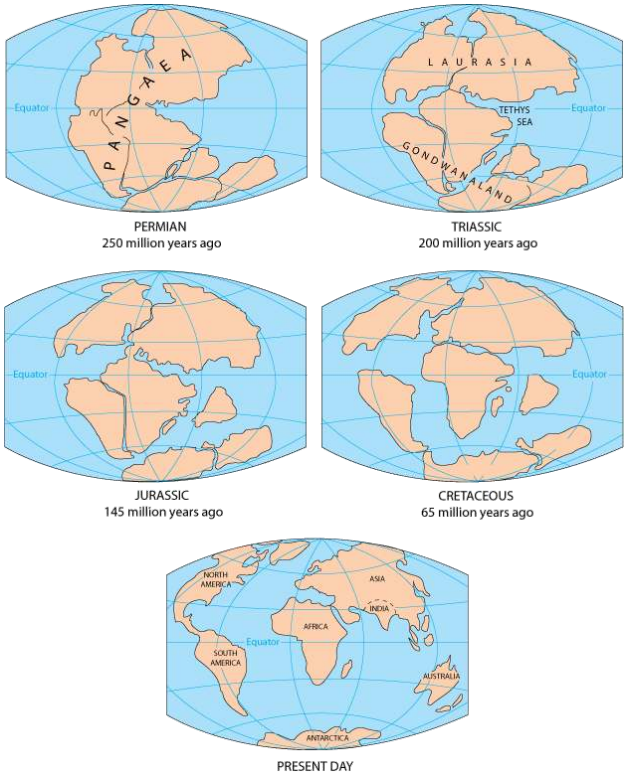
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8.1: Continental Drift



When you view a map of the earth, what do you notice about the shape of the continents? Many people notice that the continents seem like they could fit together, especially when you look at the coasts of the Americas and the coasts of Europe and Africa. This can be explained by the Theory of Plate Tectonics. In simple terms, the Theory of Plate Tectonics states that the earth is made up of plates that move in relation to one another.

The Theory of Plate Tectonics began as an idea called continental drift. Continental drift was first conceived by scholars and philosophers named Francis Bacon, George Buffon, and Alexander von Humboldt. However, it was Alfred Wagener, a meteorologist and geophysicist presented the complete explanation of continental drift in 1912. Continental drift stated that the continents once fit together but had drifted apart after millions of years. The continents, now far apart, showed similar characteristics in sediment, rock formation, and vegetation. This supported the idea that they were one landmass in the past. Wegener named the landmass Pangaea, which means "all the earth" in Greek.



According to the continental drift, the supercontinent Pangaea *began to break up about 225-200 million years ago, eventually fragmenting into the continents as we know them today.*

Under the Earth’s crust there are a series of plates that move and collide with each other causing landmasses on the Earth’s surface to shift and drift. Our continents will continue to move and in the future (millions of years) will be in completely different location’s on the earth. Alfred Wegener supported his explanation of continental drift with evidence of fossils, mountain ranges, and patterns of glaciation. While Wegener called the supercontinent Pangea, he called the super ocean Panthalassa, which means "all ocean" in Greek.

There were problems with Wegener's continental drift explanation, which were corrected when a more thorough explanation called Plate Tectonics was presented in the 1960s.

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8.2: Plates and Plate Motions

The idea of plate tectonics became widely accepted around 1965 as more and more geologists started thinking in these terms. By the end of 1967, Earth's surface had been mapped into a series of plates (Figure 8.2.1). The major plates are Eurasia, Pacific, India, Australia, North America, South America, Africa, and Antarctic. There are also numerous small plates (e.g., Juan de Fuca, Nazca, Scotia, Philippine, Caribbean), and many very small plates or sub-plates. For example the Juan de Fuca Plate is actually three separate plates (Gorda, Juan de Fuca, and Explorer) that all move in the same general direction but at slightly different rates.

The fact that the plates include both crustal material and lithospheric mantle material makes it possible for a single plate to be made up of both oceanic and continental crust. For example, the North American Plate includes most of North America, plus half of the northern Atlantic Ocean. Similarly the South American Plate extends across the western part of the southern Atlantic Ocean, while the European and African plates each include part of the eastern Atlantic Ocean. The Pacific Plate is almost entirely oceanic, but it does include the part of California west of the San Andreas Fault.

Rates of motions of the major plates range from less than 1 cm/year to over 10 cm/year (for comparison, human fingernails grow at around 6 cm/year). The Pacific Plate is the fastest at over 10 cm/year in some areas, followed by the Australian and Nazca Plates. The North American Plate is one of the slowest, averaging around 1 cm/year in the south up to almost 4 cm/year in the north. Plates move as rigid bodies, so it may seem surprising that the North American Plate can be moving at different rates in different places. The explanation is that plates move in a rotational manner. The North American Plate, for example, rotates counter-clockwise; the Eurasian Plate rotates clockwise.

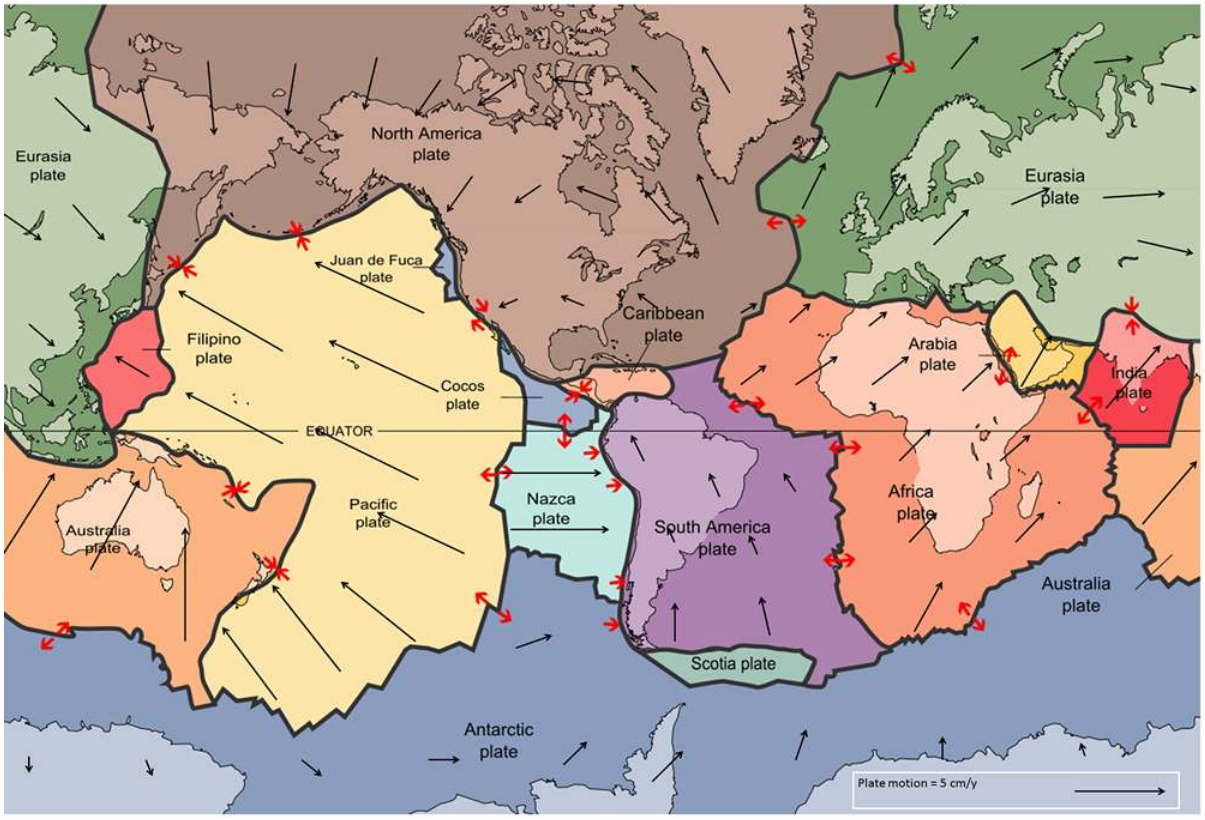


Figure 8.2.1 The major lithospheric plates of Earth. Arrows indicate direction of plate movement, and the length of the arrows represent the speed of plate motion (Steven Earle, "Physical Geology").

As originally described by Wegener in 1915, the present continents were once all part of the supercontinent Pangaea. More recent studies of continental match-ups and the magnetic ages of ocean-floor rocks have enabled us to reconstruct the history of the break-up of Pangaea.

Pangaea began to rift apart along a line between Africa and Asia and between North America and South America at around 200 Ma (Figure 8.2.2). During the same period, the Atlantic Ocean began to open up between northern Africa and North America, and India

broke away from Antarctica. At this stage, Pangaea was divided into Laurasia (now Europe, Asia and North America) and Gondwanaland (the southern continents; South America, Africa, India, Australia, and Antarctica). Between 200 and 150 Ma, rifting started between South America and Africa and between North America and Europe, and India separated from Antarctica and moved north toward Asia. By 80 Ma, Africa had separated from South America, and most of Europe had separated from North America. By 50 Ma, Australia had separated from Antarctica, and shortly after that, India collided with Asia.

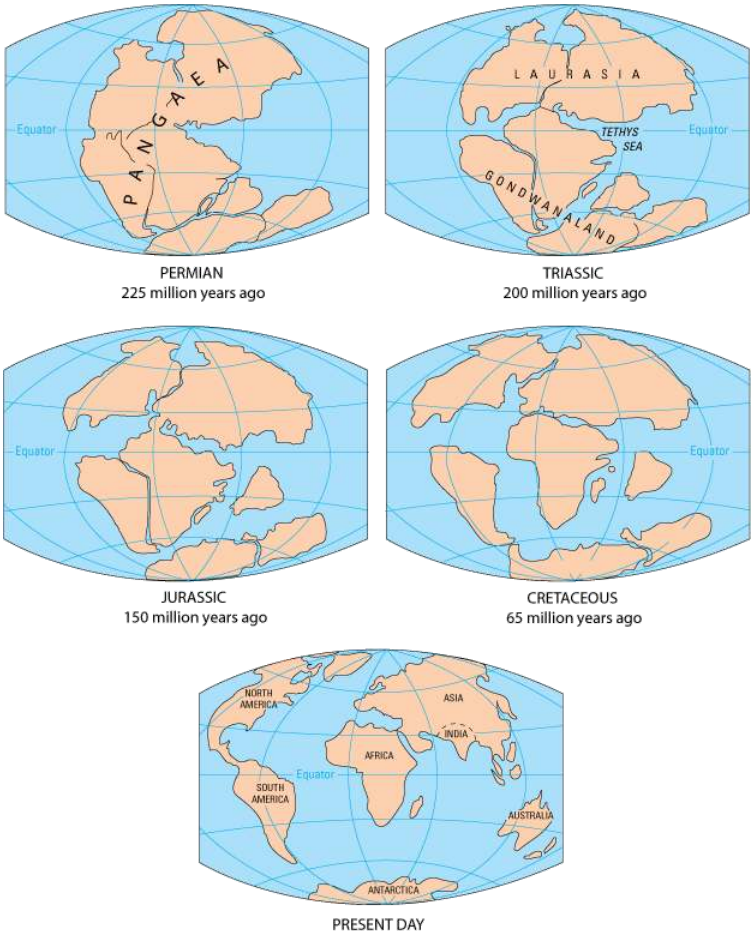


Figure 8.2.2 Movement of the continents over the past 225 million years (USGS, https://pubs.usgs.gov/gip/dynamic/historical.html).

Within the past few million years, rifting has taken place in the Gulf of Aden and the Red Sea, and also within the Gulf of California. Incipient rifting has begun along the Great Rift Valley of eastern Africa, extending from Ethiopia and Djibouti on the Gulf of Aden (Red Sea) all the way south to Malawi.

Over the next 50 million years, it is likely that there will be full development of the east African rift and creation of new ocean floor. Eventually Africa will split apart. There will also be continued northerly movement of Australia and Indonesia. The western part of California (including Los Angeles and part of San Francisco) will split away from the rest of North America, and eventually sail right by the west coast of Vancouver Island, en route to Alaska. Because the oceanic crust formed by spreading on the mid-Atlantic ridge is not currently being subducted (except in the Caribbean), the Atlantic Ocean is slowly getting bigger, and the Pacific Ocean is getting smaller. If this continues without changing for another couple hundred million years, we will be back to where we started, with one supercontinent.

Pangaea, which existed from about 350 to 200 Ma, was not the first supercontinent. In 1966, Tuzo Wilson proposed that there has been a continuous series of cycles of continental rifting and collision; that is, break-up of supercontinents, drifting, collision, and formation of other supercontinents. Pangaea was preceded by Pannotia (600 to 540 Ma), by Rodinia (1,100 to 750 Ma), and by other supercontinents before that.

With all of these plates constantly on the move, they inevitably end up interacting with each other at their plate boundaries. Plates can interact in three ways: they can move apart (divergent boundary), they can move towards each other (convergent boundary), or they can slide past each other (transform boundary). The following sections will examine each of these types of plate boundaries, and the geological features they create.

Additional links for more information:

- An interactive animation of plate motion over the past 550 million years: http://barabus.tru.ca/geol1031/plates.html

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8.3: Types of Plate Boundaries

Categorization of plate boundaries is based off of how two plates move relative to each other. There are essentially three types of plate boundaries: **divergent**, **convergent**, and **transform**.



In the case of **divergent plate boundaries**, two of earth's plates move away from each other. We can call these spreading centers, and new ocean crust is created here. An example of a divergent plate boundary is the Mid-Atlantic Ridge. The Mid-Atlantic Ridge is a long underwater mountain range that results from the creation of new oceanic crust. As new crust forms, rocks not only pile up to the form the mountain chain, but the Atlantic Ocean grows larger. Because the movement at the divergent boundary, shallow earthquakes can occur here.

When two plates move towards each other, the boundary is known as a **convergent boundary**. As previously mentioned, depending on the type of crust involved in this collision, different geographic features are formed. When two oceanic plates converge, the denser plate (usually the older one) will end up sinking below the less dense plate, leading to the formation of an oceanic **subduction zone**. Whenever a subduction zone is formed, the subducted plate will end up being partially melted by the earth's internal magma. This melted rock rises, uplifting the crust, eventually developing into a volcano. When this occurs in an ocean, the result is a series of volcanic islands. An example of this can be found in the Marianas Islands, which sit along the Marianas Trench. Trenches are created at convergence zones. Subduction zones are the reason why oceanic crust older than 200 million years old cannot be found. Old, dense crust tends to be subducted back into the earth. When an oceanic plate converges with a continental plate, the oceanic crust will always subduct under the continental crust; this is because oceanic crust is naturally denser. This creates volcanic mountains and trenches along the coast of the continent. An oceanic-continental convergent boundary exists along the Nazca plate and South American Plate. This convergent boundary results in the formation of the Peru-Chile Trench and the Andes Mountains. When two continental crusted plates converge, they eventually collide and end up producing mountains; this was how the Himalayas were created. Neither continental crust will subduct underneath one another because of their similar densities. Convergent boundaries are commonly associated with larger, deeper earthquakes.



The last type of plate boundary is the **transform boundary**, which is where two plates slide past one another. Unlike the other two types of plate boundaries in which new seafloor is created at divergent boundaries and where old seafloor is subducted at convergent boundaries, transform plate boundaries neither create nor destroy crust. Transform faults are commonly known for creating earthquakes, which can be quite large. One example of a transform fault boundary would be the San Andreas fault.



https://www.youtube.com/watch?v=dXDYoCqwSbM

We will explore plate boundaries in more detail in the next sections.

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8.4: Divergent Plate Boundaries

Divergent boundaries are spreading boundaries, where new oceanic crust is created to fill in the space as the plates move apart. Most divergent boundaries are located along mid-ocean oceanic ridges (although some are on land). The **mid-ocean ridge** system is a giant undersea mountain range, and is the largest geological feature on Earth; at 65,000 km long and about 1000 km wide, it covers 23% of Earth's surface (Figure 8.4.1). Because the new crust formed at the plate boundary is warmer than the surrounding crust, it has a lower density so it sits higher on the mantle, creating the mountain chain. Running down the middle of the mid-ocean ridge is a **rift valley** 25-50 km wide and 1 km deep. Although oceanic spreading ridges appear to be curved features on Earth's surface, in fact the ridges are composed of a series of straight-line segments, offset at intervals by faults perpendicular to the ridge, called **transform faults**. These transform faults make the mid-ocean ridge system look like a giant zipper on the seafloor (Figure 8.4.2). Movements along transform faults between two adjacent ridge segments are responsible for many earthquakes.

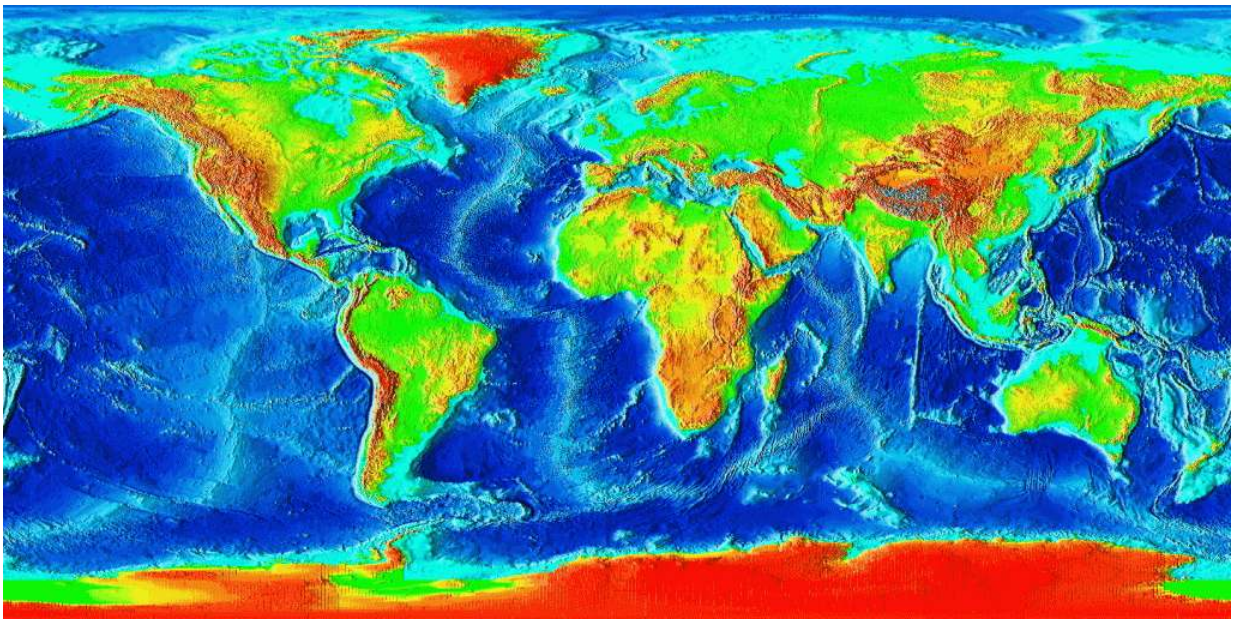


Figure 8.4.1 Ocean floor topography. The mid-ocean ridge system can be seen as the light blue chain of mountains running throughout the oceans (http://www.ngdc.noaa.gov/mgg/image/mggd.gif).

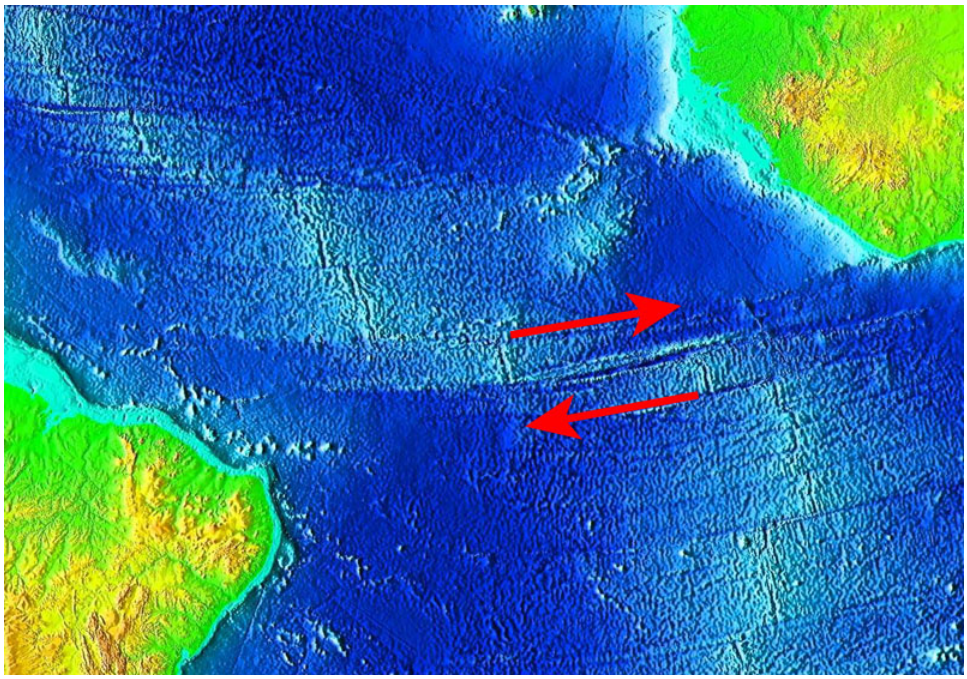


Figure 8.4.2 Closeup of the mid-Atlantic ridge system, showing transform faults perpendicular to the ridge axis. Arrows indicate the direction of plate motion on either side of the fault (USGS, Public domain, via Wikimedia Commons).

The crustal material created at a spreading boundary is always oceanic in character; in other words, it is igneous rock (e.g., basalt or gabbro, rich in ferromagnesian minerals), forming from magma derived from partial melting of the mantle caused by decompression as hot mantle rock from depth is moved toward the surface (Figure 8.4.3). The triangular zone of partial melting near the ridge crest is approximately 60 km thick and the proportion of magma is about 10% of the rock volume, thus producing crust that is about 6 km thick. This magma oozes out onto the seafloor to form pillow basalts, breccias (fragmented basaltic rock), and flows, interbedded in some cases with limestone or chert. Over time, the igneous rock of the oceanic crust gets covered with layers of sediment, which eventually become sedimentary rock.

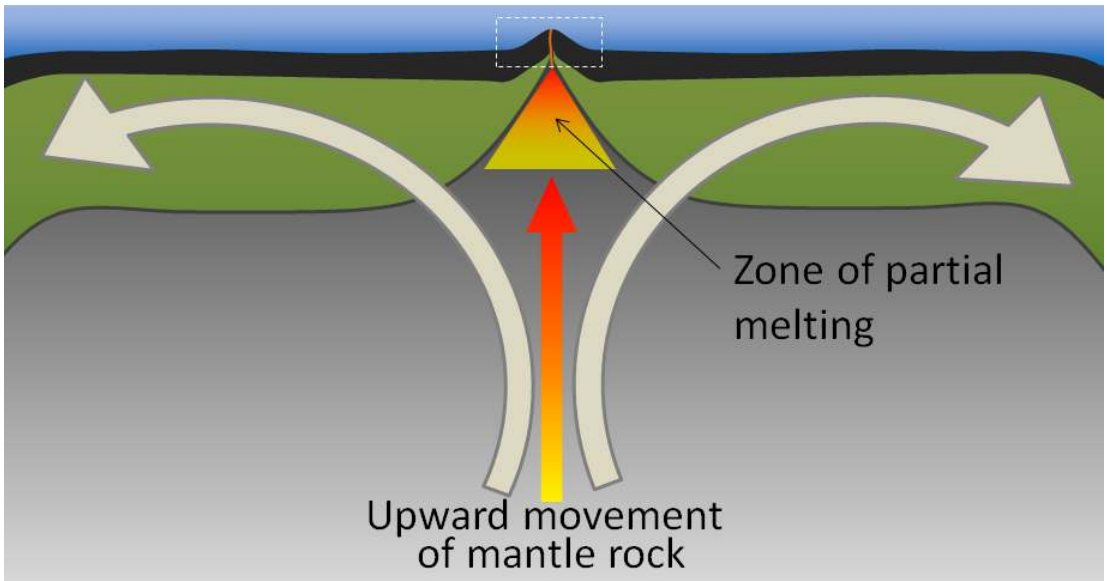


Figure 8.4.3 Mechanism for divergent plate boundaries. The region in the outlined rectangle represent the mid-ocean ridge (Steven Earle, “Physical Geology”).

Spreading is hypothesized to start within a continental area with up-warping or doming of crust related to an underlying mantle plume or series of mantle plumes. The buoyancy of the mantle plume material creates a dome within the crust, causing it to

fracture. When a series of mantle plumes exists beneath a large continent, the resulting rifts may align and lead to the formation of a rift valley (such as the present-day Great Rift Valley in eastern Africa). It is suggested that this type of valley eventually develops into a linear sea (such as the present-day Red Sea), and finally into an ocean (such as the Atlantic). It is likely that as many as 20 mantle plumes, many of which still exist, were responsible for the initiation of the rifting of Pangaea along what is now the mid-Atlantic ridge.

There are multiple lines of evidence demonstrating that new oceanic crust is forming at these seafloor spreading centers:

1. Age of the crust:

Comparing the ages of the oceanic crust near a mid-ocean ridge shows that the crust is youngest right at the spreading center, and gets progressively older as you move away from the divergent boundary in either direction, aging approximately 1 million years for every 20-40 km from the ridge. Furthermore, the pattern of crust age is fairly symmetrical on either side of the ridge (Figure 8.4.4).

The oldest oceanic crust is around 280 Ma in the eastern Mediterranean, and the oldest parts of the open ocean are around 180 Ma on either side of the north Atlantic. It may be surprising, considering that parts of the continental crust are close to 4,000 Ma old, that the oldest seafloor is less than 300 Ma. Of course, the reason for this is that all seafloor older than that has been either subducted or pushed up to become part of the continental crust. As one would expect, the oceanic crust is very young near the spreading ridges (Figure 8.4.4), and there are obvious differences in the rate of sea-floor spreading along different ridges. The ridges in the Pacific and southeastern Indian Oceans have wide age bands, indicating rapid spreading (approaching 10 cm/year on each side in some areas), while those in the Atlantic and western Indian Oceans are spreading much more slowly (less than 2 cm/year on each side in some areas).

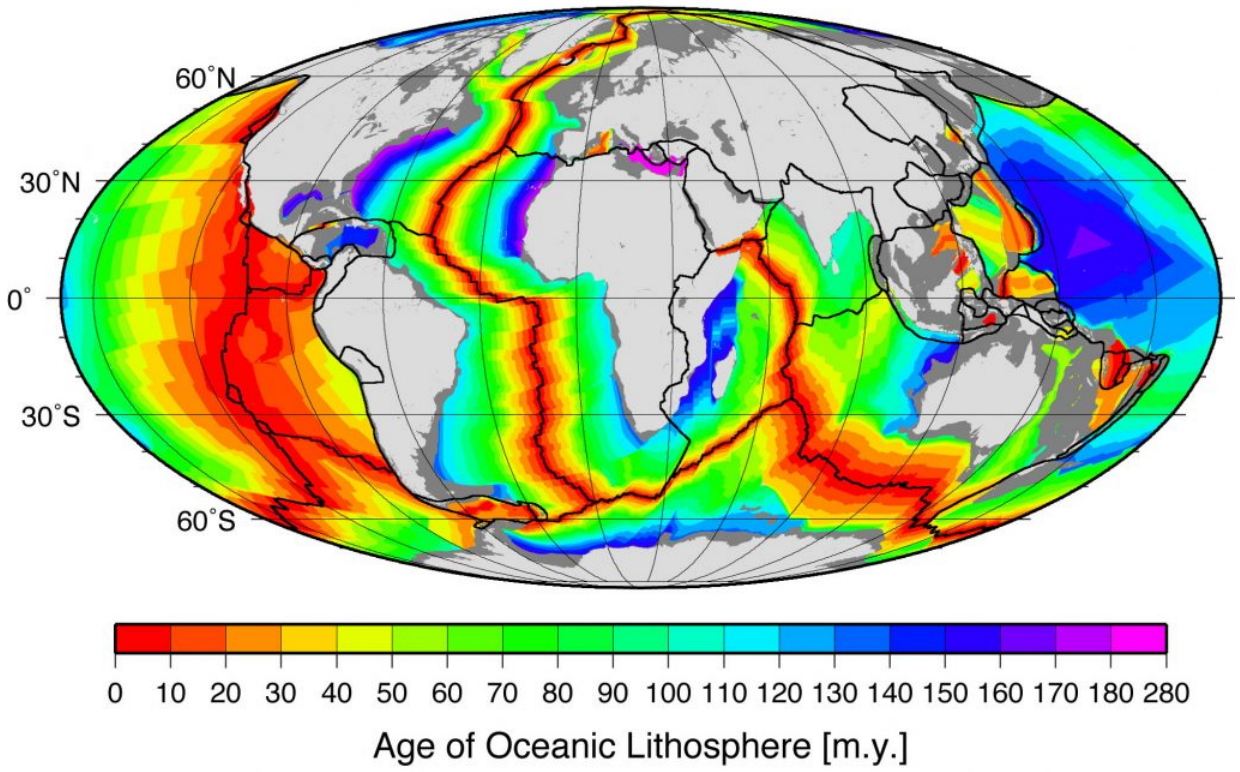


Figure 8.4.4 Age of the oceanic crust (http://www.ngdc.noaa.gov/mgg/ocean_a...eanic_lith.jpg).

2. Sediment thickness:

With the development of seismic reflection sounding (similar to echo sounding described previously) it became possible to *see through* the seafloor sediments and map the bedrock topography and crustal thickness. Hence sediment thicknesses could be mapped, and it was soon discovered that although the sediments were up to several thousands of meters thick near the continents, they were relatively thin — or even non-existent — in the ocean ridge areas (Figure 8.4.5). This makes sense when combined with the data on the age of the oceanic crust; the farther from the spreading center the older the crust, the longer it has had to accumulate

sediment, and the thicker the sediment layer. Additionally, the bottom layers of sediment are older the farther you get from the ridge, indicating that they were deposited on the crust long ago when the crust was first formed at the ridge.

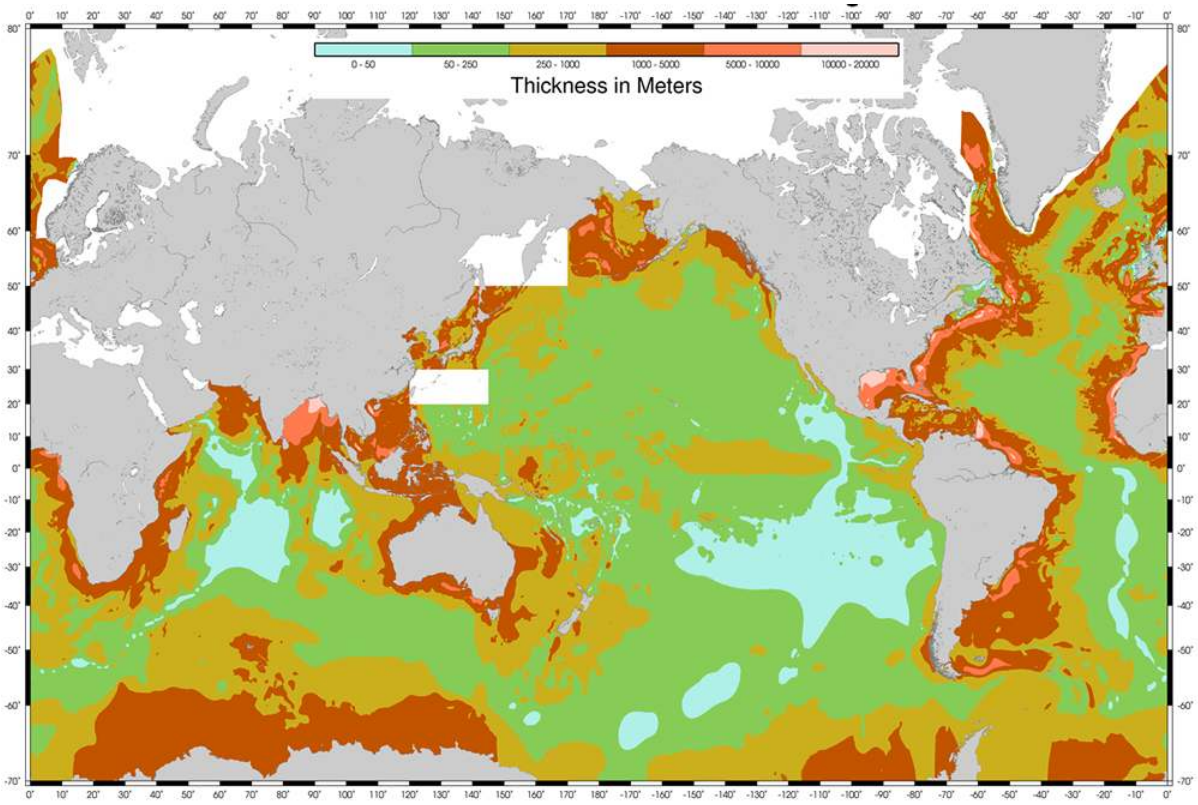


Figure 8.4.5 Seafloor sediment thickness (Modified from https://www.ngdc.noaa.gov/mgg/sedthick/).

3. Heat flow:

Measurements of rates of heat flow through the ocean floor revealed that the rates are higher than average (about 8x higher) along the ridges, and lower than average in the trench areas (about 1/20th of the average). The areas of high heat flow are correlated with upward convection of hot mantle material as new crust is formed, and the areas of low heat flow are correlated with downward convection at subduction zones.

4. Magnetic reversals:

We saw that rocks could retain magnetic information that they acquired when they were formed. However, Earth's magnetic field is not stable over geological time. For reasons that are not completely understood, the magnetic field decays periodically and then becomes re-established. When it does re-establish, it may be oriented the way it was before the decay, or it may be oriented with the reversed polarity. During periods of reversed polarity, a compass would point south instead of north. Over the past 250 Ma, there have a few hundred magnetic field reversals, and their timing has been anything but regular. The shortest ones that geologists have been able to define lasted only a few thousand years, and the longest one was more than 30 million years, during the Cretaceous (Figure 8.4.6). The present "normal" event has persisted for about 780,000 years.

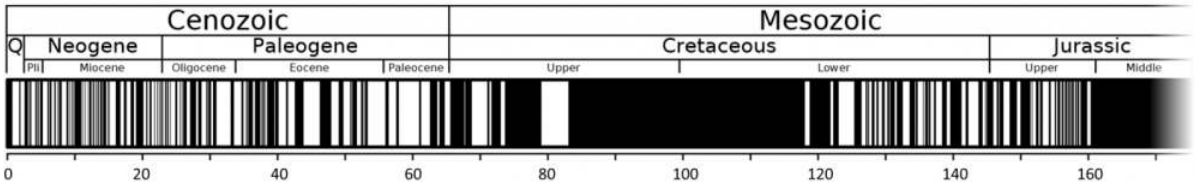


Figure 8.4.6 Magnetic field reversal chronology for the past 170 Ma (Steven Earle after: http://upload.wikimedia.org/wikipedi...y_0-169_Ma.svg).



Figure 8.4.7 Pattern of magnetic anomalies in oceanic crust in the Pacific northwest (Steven Earle, "Physical Geology").

Beginning in the 1950s, scientists started using magnetometer readings when studying ocean floor topography. The first comprehensive magnetic data set was compiled in 1958 for an area off the coast of British Columbia and Washington State. This survey revealed a mysterious pattern of alternating stripes of low and high magnetic intensity in sea-floor rocks (Figure 8.4.7). Subsequent studies elsewhere in the ocean also observed these magnetic anomalies, and most importantly, the fact that the magnetic patterns are symmetrical with respect to ocean ridges. In the 1960s, in what would become known as the Vine-Matthews-Morley (VMM) hypothesis, it was proposed that the patterns associated with ridges were related to the magnetic reversals, and that oceanic crust created from cooling basalt during a *normal* event would have polarity aligned with the present magnetic field, and thus would produce a positive anomaly (a black stripe on the sea-floor magnetic map), whereas oceanic crust created during a *reversed* event would have polarity opposite to the present field and thus would produce a negative magnetic anomaly (a white stripe). The widths of the anomalies varied according to the spreading rates characteristic of the different ridges. This process is illustrated in Figure 8.4.8. New crust is formed (panel a) and takes on the existing normal magnetic polarity. Over time, as the plates continue to diverge, the magnetic polarity reverses, and new crust formed at the ridge now takes on the reversed polarity (white stripes in Figure 8.4.8). In panel b, the poles have reverted to normal, so once again the new crust shows normal polarity before moving away from the ridge. Eventually, this creates a series of parallel, alternating bands of reversals, symmetrical around the spreading center (panel c).

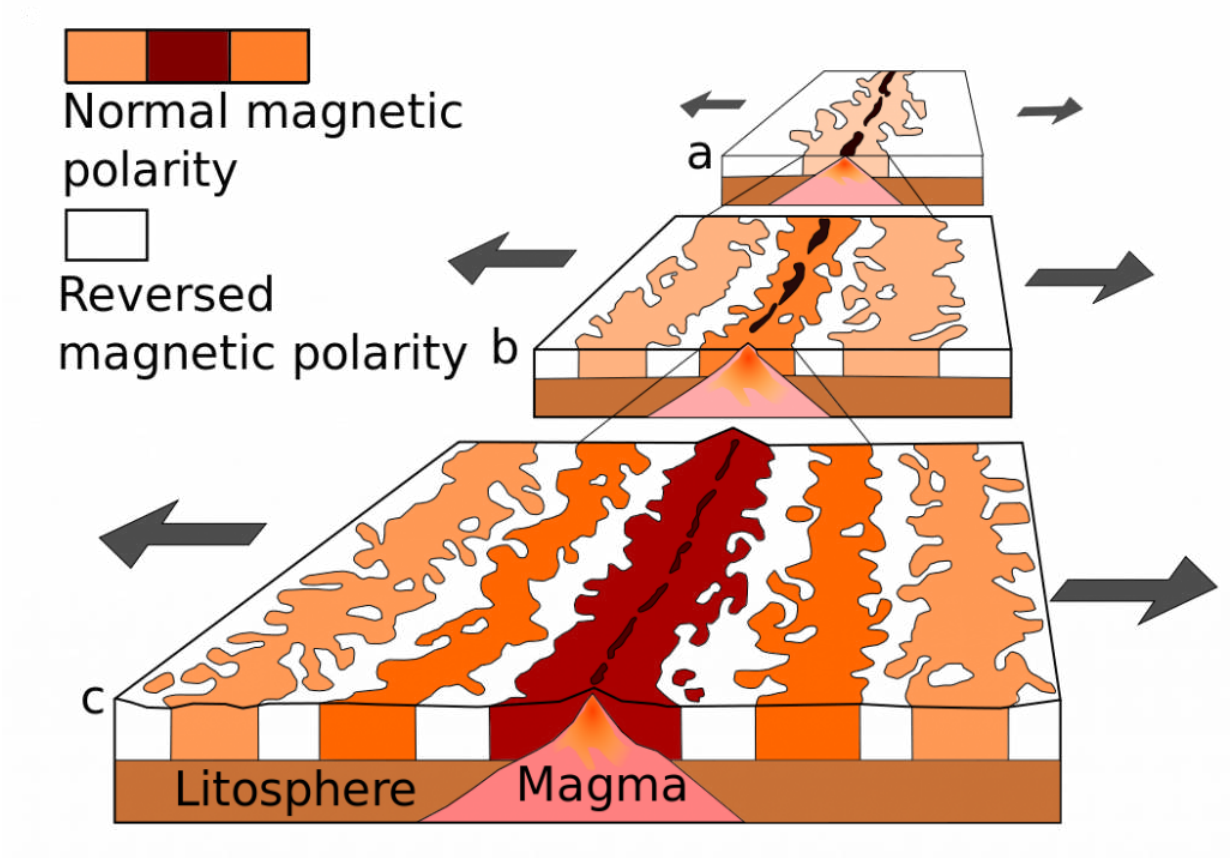


Figure 8.4.8 Formation of alternating patterns of magnetic polarity along a mid-ocean ridge (Steven Earle, "Physical Geology").

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8.5: Convergent Plate Boundaries

Convergent boundaries, where two plates are moving toward each other, are of three types, depending on the type of crust present on either side of the boundary — oceanic or continental. The types are ocean-ocean, ocean-continent, and continent-continent.

At an ocean-ocean convergent boundary, one of the plates (oceanic crust and lithospheric mantle) is pushed, or subducted**,** under the other (Figure 8.5.1). Often it is the older and colder plate that is denser and subducts beneath the younger and warmer plate. There is commonly an ocean trench along the boundary as the crust bends downwards. The subducted lithosphere descends into the hot mantle at a relatively shallow angle close to the subduction zone, but at steeper angles farther down (up to about 45°). The significant volume of water within the subducting material is released as the subducting crust is heated. It mixes with the overlying mantle, and the addition of water to the hot mantle lowers the crust's melting point and leads to the formation of magma (flux melting). The magma, which is lighter than the surrounding mantle material, rises through the mantle and the overlying oceanic crust to the ocean floor where it creates a chain of volcanic islands known as an **island arc**. A mature island arc develops into a chain of relatively large islands (such as Japan or Indonesia) as more and more volcanic material is extruded and sedimentary rocks accumulate around the islands. Earthquakes occur relatively deep below the seafloor, where the subducting crust moves against the overriding crust.

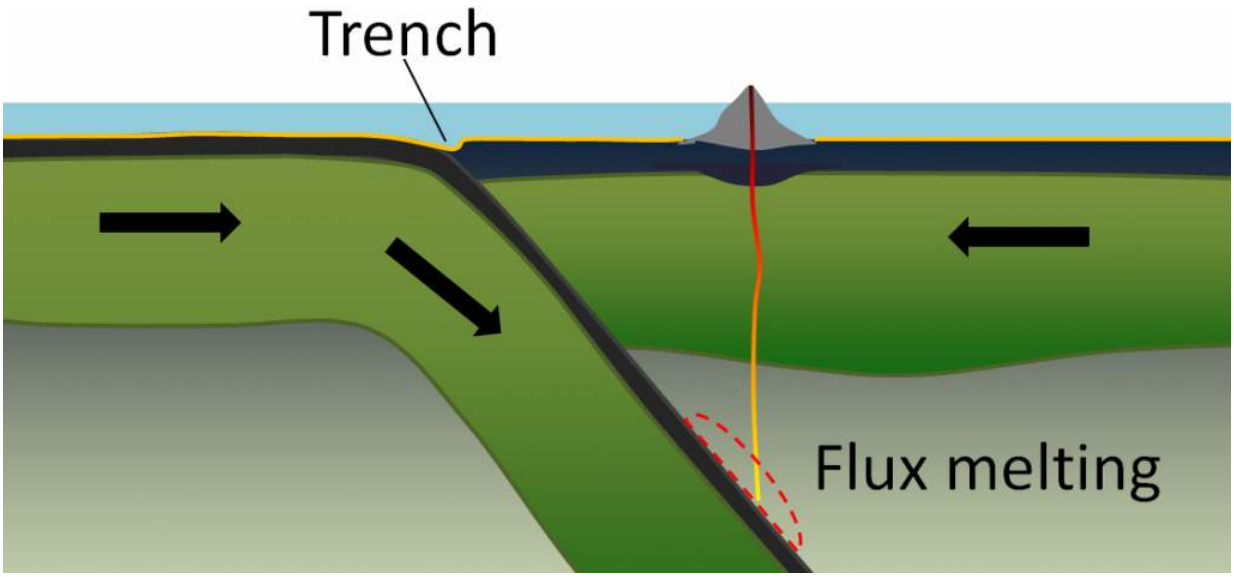


Figure 8.5.1 A trench and volcanic island formed from an ocean-ocean convergent zone (Steven Earle, "Physical Geology").

Examples of ocean-ocean convergent zones are subduction of the Pacific Plate south of Alaska (creating the Aleutian Islands) and under the Philippine Plate, where it creates the Marianas Trench, the deepest part of the ocean.

At an ocean-continent convergent boundary, the denser oceanic plate is pushed under the less dense continental plate in the same manner as at an ocean-ocean boundary. Sediment that has accumulated on the seafloor is thrust up into an accretionary wedge, and compression leads to thrusting within the continental plate (Figure 8.5.2). The magma produced adjacent to the subduction zone rises to the base of the continental crust and leads to partial melting of the crustal rock. The resulting magma ascends through the crust, producing a mountain chain with many volcanoes. As with an ocean-ocean boundary, the subducting crust can produce a deep trench running parallel to the coastline.

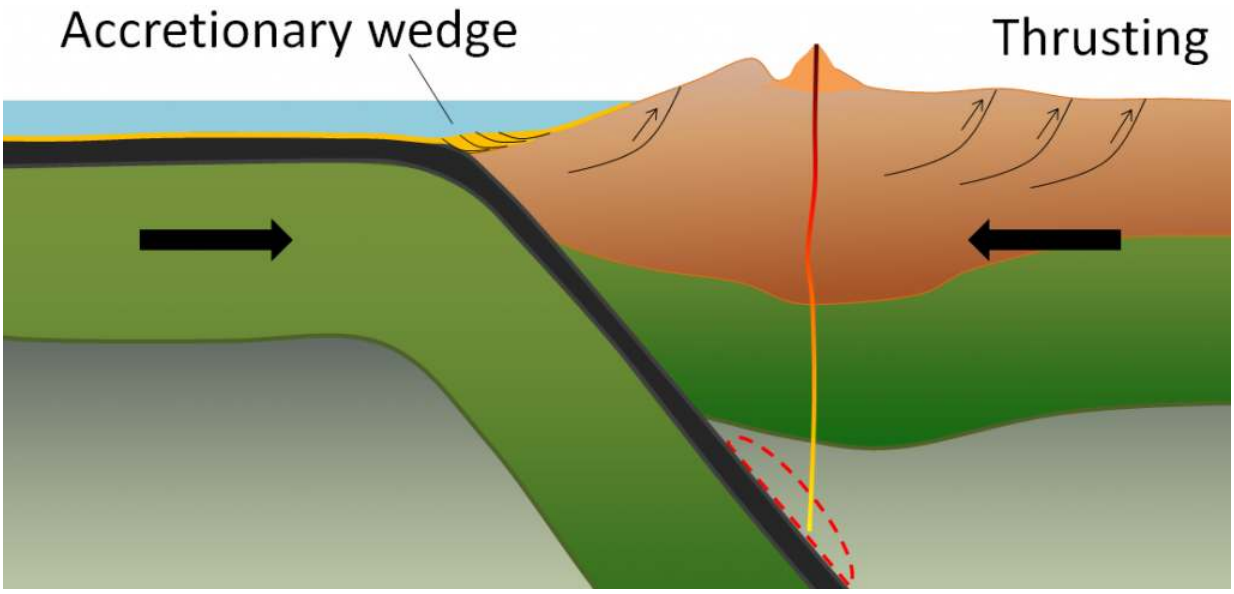


Figure 8.5.2 A trench and volcanic mountains formed from an ocean-continent convergent zone (Steven Earle, “Physical Geology”).

Examples of ocean-continent convergent boundaries are subduction of the Nazca Plate under South America (which has created the Andes Mountains and the Peru Trench) and subduction of the Juan de Fuca Plate under North America (creating the Cascade Range).

A continent-continent collision occurs when a continent or large island that has been moved along with subducting oceanic crust collides with another continent (Figure 8.5.3). The colliding continental material will not be subducted because it is too light (i.e., because it is composed largely of light continental rocks), but the root of the oceanic plate will eventually break off and sink into the mantle. There is tremendous deformation of the pre-existing continental rocks, forcing the material upwards and creating mountains.

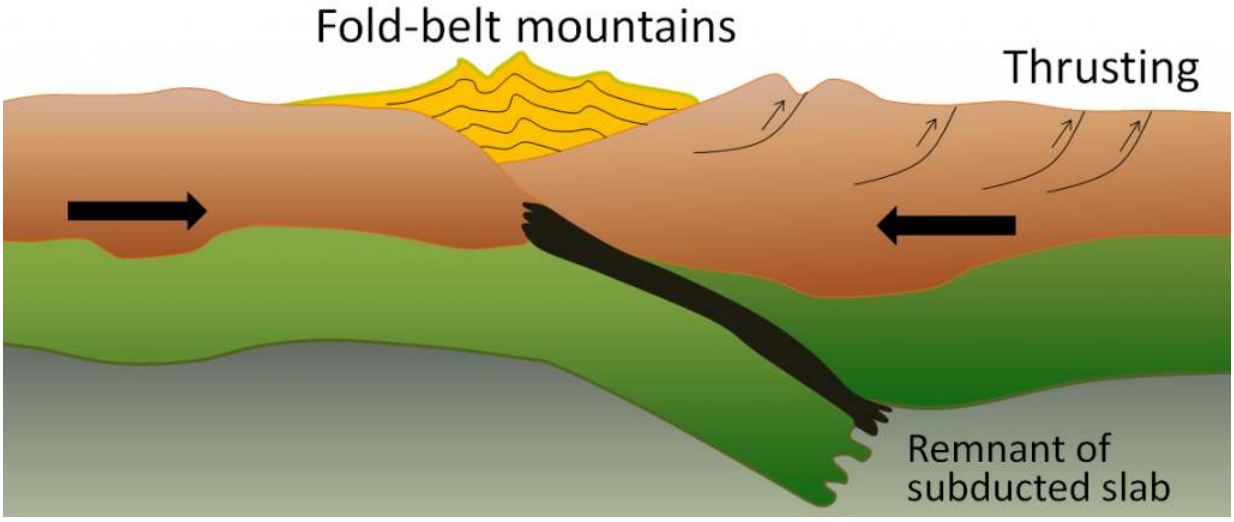


Figure 8.5.3 Mountains formed from a continent-continent convergent zone (Steven Earle, “Physical Geology”).

Examples of continent-continent convergent boundaries are the collision of the India Plate with the Eurasian Plate, creating the Himalaya Mountains, and the collision of the African Plate with the Eurasian Plate, creating the series of ranges extending from the Alps in Europe to the Zagros Mountains in Iran. The Rocky Mountains in North America are also a result of continent-continent collisions.

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8.6: Transform Plate Boundaries

Transform boundaries exist where one plate slides past another without production or destruction of crustal material. Most transform faults connect segments of mid-ocean ridges and are thus ocean-ocean plate boundaries. Some transform faults connect continental parts of plates. An example is the San Andreas Fault, which connects the southern end of the Juan de Fuca Ridge with the northern end of the East Pacific Rise (ridge) in the Gulf of California (Figure 8.6.1). The part of California west of the San Andreas Fault and all of Baja California are on the Pacific Plate. Transform faults do not just connect divergent boundaries. For example, the Queen Charlotte Fault connects the north end of the Juan de Fuca Ridge, starting at the north end of Vancouver Island, to the Aleutian subduction zone.

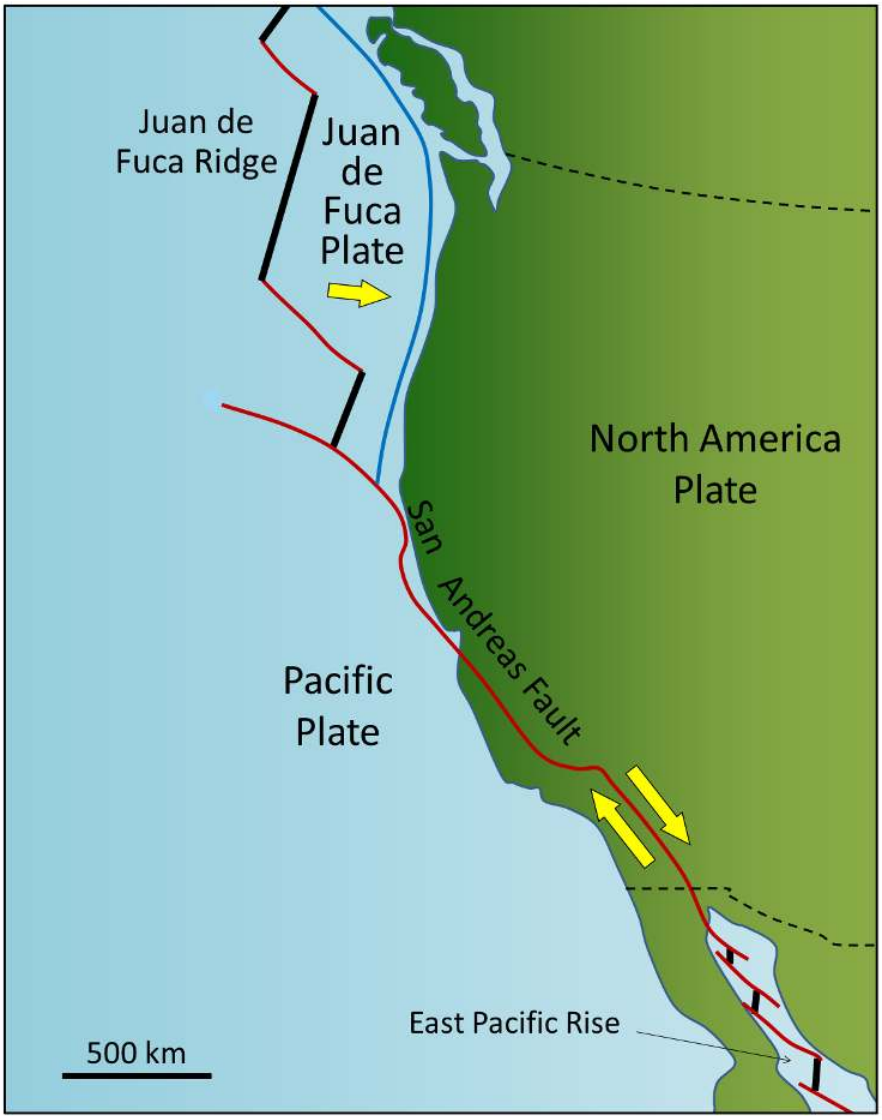


Figure 8.6.1 Transform faults along the U.S. west coast (Steven Earle, “Physical Geology”).

As we will see in the next section, earthquakes are common along transform faults, as the two plates slide past each other.

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8.7: Seamounts and Hot Spots

The ocean floor is dotted with **seamounts**, some isolated and some in chains. Seamounts are underwater volcanoes, and most are much younger than the oceanic crust on which they formed. If a seamount gets large enough to break the ocean surface, it becomes a volcanic island. Some seamounts are formed from magma rising at a divergent boundary, and as the plates move apart, the seamounts move with them, which can result in a seamount chain. Other seamounts form from the rising magma at an ocean-ocean subduction zone; these include the Aleutians, extending from Alaska to Russia, and the Lesser Antilles in the eastern part of the Caribbean. Sometimes the crust on which an island or seamount sits will subside, taking the seamount with it. As this happens, the top of the seamount can become eroded flat, and these flat-topped seamounts are then called **tablemounts** or **guyots**.

However, some seamounts are formed far away from plate boundaries, in places where we would not usually expect much volcanic activity. Some seamounts and ocean islands are formed above a **mantle plume** or **hot spot** — a place where hot mantle material rises in a stationary and semi-permanent plume, and affects the overlying crust. Mantle plumes are thought to rise at approximately 10 times the rate of mantle convection. The ascending column may be on the order of kilometers to tens of kilometers across, but near the surface it spreads out to create a mushroom-style head that is several tens to over 100 kilometers across. Near the base of the lithosphere (the rigid part of the mantle), the mantle plume (and possibly some of the surrounding mantle material) partially melts to form magma that rises to feed volcanoes.

A great example of seamounts created from a hot spot includes the Hawaiian and Emperor Seamount island chains in the Pacific Ocean (Figure 8.7.1). The oldest of the Hawaiian/Emperor seamounts is dated at around 80 Ma, and it is situated on oceanic crust aged around 90 to 100 Ma. The volcanic rock making up these islands gets progressively younger toward the southeast, culminating with the island of Hawaii itself, which consists of rock that is almost all younger than 1 Ma. It appears that a stationary plume of hot upwelling mantle material is the source of the Hawaiian volcanism, and that the ocean crust of the Pacific Plate is moving toward the northwest over this hot spot. A seamount will be formed through volcanic activity over the hot spot, then the plate will move and displace the seamount before the hot spot produces the next seamount, and so on. In this way, over time, the seamounts are formed in chains. Near the Midway Islands, the chain takes a pronounced change in direction, from northwest-southeast for the Hawaiian Islands to nearly north-south for the Emperor Seamounts. This change is widely ascribed to a change in direction of the Pacific Plate moving over the stationary mantle plume, but it is also possible that the Hawaiian mantle plume has not actually been stationary throughout its history, and in fact moved at least 2,000 km south over the period between 81 and 45 Ma.

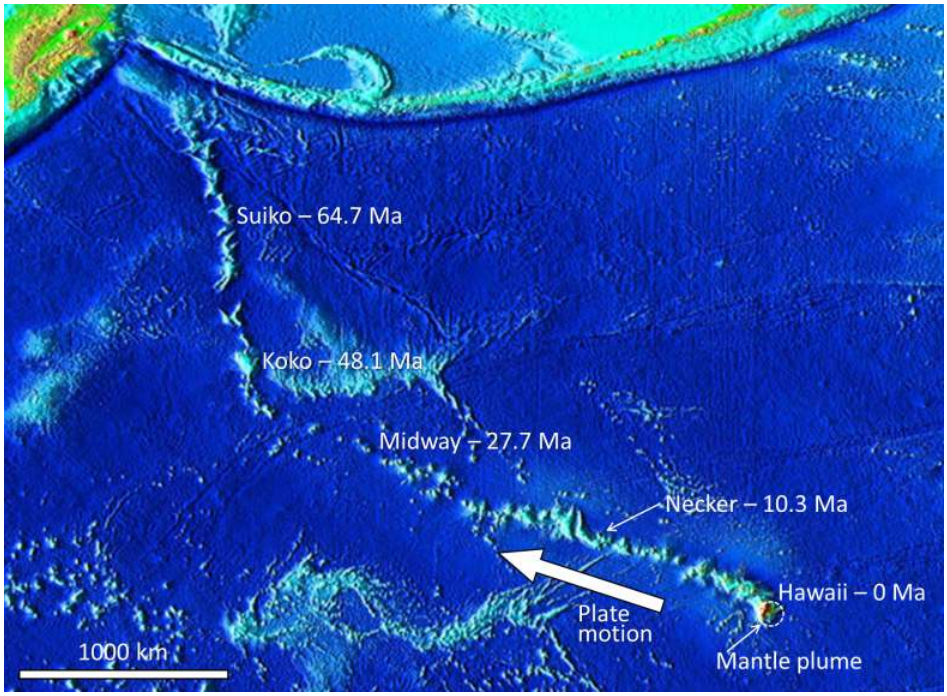


Figure 8.7.1 The Hawaiian Islands/Emperor Seamount chain, with ages of selected structures. This chain has formed as the Pacific Plate moved northwest over a hot spot (Steven Earle, “Physical Geology”).

Since most mantle plumes are beneath the oceans, the early stages of volcanism typically take place on the seafloor. Over time, very large islands may form like those in Hawaii. In fact, if you measure it from its base on the seafloor to its summit, Mauna Loa on the island of Hawaii is the largest mountain on Earth, rising 9700 m (in comparison, the elevation of the summit of Mt. Everest is 8848 m). While the island of Hawaii is the youngest in the chain, there is actually a new volcano named Loihi, that is still submerged at a depth of 980 m SE of Hawaii, and may one day become a new Hawaiian island when it emerges 10,000 – 100,000 years from now.

There is evidence of many such mantle plumes around the world. Most are within the ocean basins, including places like Hawaii, Iceland, and the Galapagos Islands, but some are under continents. One example is the Yellowstone hot spot in the west-central United States, and another is the one responsible for the Anahim Volcanic Belt in central British Columbia. It is evident that mantle plumes are very long-lived phenomena, lasting for at least tens of millions of years, possibly for hundreds of millions of years in some cases.

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8.8: Paleomagnetic Evidence for Plate Tectonics

Although Alfred Wegener would not live to see it, his ideas about continental drift would gradually gain acceptance within the scientific community after other scientists adjusted and molded it to form the Theory of Plate Tectonics. With the continued research within the scientific community more evidence began to accumulate. Some of the most important evidence came from the study of **paleomagnetism**, or changes in Earth's magnetic field over millions of years.

Earth's magnetic field is defined by the North and South Poles that align generally with the axis of rotation (Figure 8.8.1). The lines of magnetic force flow into Earth in the Northern Hemisphere and out of Earth in the Southern Hemisphere. Because of the shape of the field lines, the magnetic force trends at different angles to the surface in different locations (red arrows of Figure 8.8.1). At the North and South Poles, the force is vertical. Anywhere on the equator the force is horizontal, and everywhere in between, the magnetic force is at some intermediate angle to the surface.

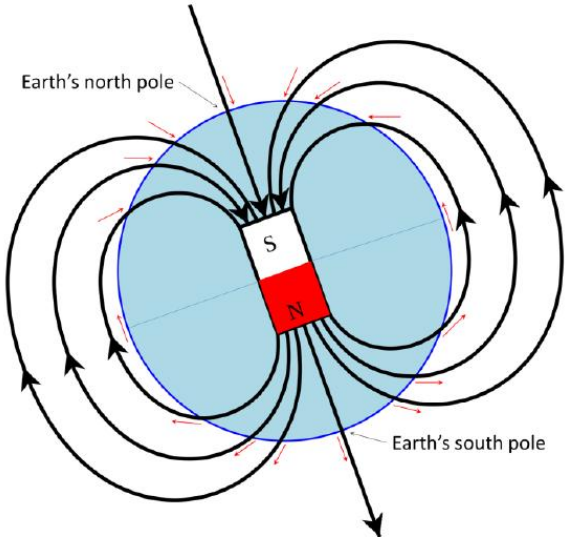


Figure 8.8.1 Depiction of Earth's magnetic field as a bar magnet coinciding with the core. The south pole of such a magnet points to Earth's North Pole. The red arrows represent the orientation of the magnetic field at various locations on Earth's surface (Steven Earle after: http://upload.wikimedia.org/wikipedia/commons/ 1/17/Earths_Magnetic_Field_ Confusion.svg).

In its fluid form, the minerals that make up magma are free to move in any direction and take on any orientation. But as the magma cools and solidifies, movement ceases and the mineral orientation and position become fixed. As the mineral magnetite (Fe_3O_4) crystallizes from magma, it becomes magnetized with an orientation parallel to that of Earth's magnetic field at that time, similar to the way a compass needle aligns with the magnetic field to point north. This magnetic record in the rock is called **remnant magnetism**. Rocks like basalt, which cool from a high temperature and commonly have relatively high levels of magnetite, are particularly susceptible to being magnetized in this way, but even sediments and sedimentary rocks, as long as they have small amounts of magnetite, will take on remnant magnetism because the magnetite grains gradually become reoriented following deposition. By studying both the horizontal and vertical components of the remnant magnetism, one can tell not only the direction to magnetic north at the time of the rock's formation, but also the latitude where the rock formed relative to magnetic north.

In the early 1950s, a group of geologists from Cambridge University, including Keith Runcorn, Edward Irving and several others, started looking at the remnant magnetism of Phanerozoic British and European volcanic rocks, and collecting paleomagnetic data. They found that rocks of different ages sampled from generally the same area showed quite different apparent magnetic pole positions (green line, Figure 8.8.2). They initially assumed that this meant that Earth's magnetic field had, over time, departed significantly from its present position, which is close to the rotational pole.

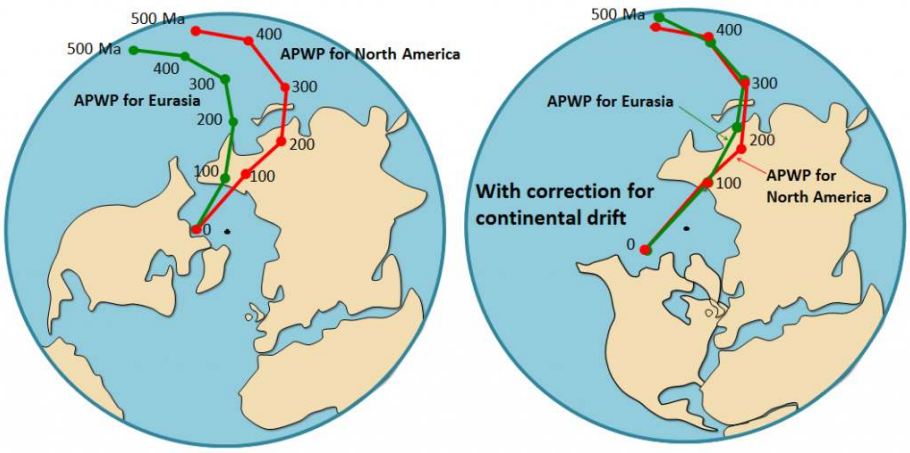


Figure 8.8.2 Polar wandering curves. Curves from Eurasia and North America seem to show that the north magnetic pole was located in two places simultaneously throughout history (left). However, if the continents are rearranged into Pangaea, the two curves overlap, showing that it is the continents than have moved, not the pole (right) (Steven Earle, "Physical Geology").

The curve defined by the paleomagnetic data was called a **polar wandering path** because Runcorn and his colleagues initially thought that their data represented actual movement of the magnetic poles (since geophysical models of the time suggested that the magnetic poles did not need to be aligned with the rotational poles). We now know that the magnetic data define movement of continents, and *not* of the magnetic poles, so we call it an ***apparent* polar wandering path** (APWP). Runcorn and colleagues soon extended their work to North America, and this also showed *apparent* polar wandering, but the results were not consistent with those from Europe (Figure 8.8.2). For example, the 200 Ma pole for North America placed somewhere in China, while the 200 Ma pole for Europe placed in the Pacific Ocean. Since there could only have been one pole position at 200 Ma, this evidence strongly supported the idea that North America and Europe had moved relative to each other since 200 Ma. Subsequent paleomagnetic work showed that South America, Africa, India, and Australia also have unique polar wandering curves. Rearranging the continents based on their positions in Pangaea caused these wandering curves to overlap, showing that the continents had moved over time.

Additional evidence for movement of the continents came from analysis of **magnetic dip**. Recall from Figure 8.8.1 that the angle of the magnetic field changes as a function of latitude, with the field directed vertically downwards at the north pole, upwards at the south pole, and horizontal at the equator. Every latitude between the equator and the poles will have a corresponding angle between horizontal and vertical (red arrows, Figure 8.8.1). By looking at the dip angle in rocks, we can determine the latitude at which those rocks were formed. Combining that with the age of the rocks, we can trace the movements of the continents over time. For example, at around 500 Ma, what we now call Europe was south of the equator, and so European rocks formed then would have acquired an upward-pointing magnetic field orientation (Figure 8.8.3). Between then and now, Europe gradually moved north, and the rocks forming at various times acquired steeper and steeper *downward-pointing* magnetic orientations.

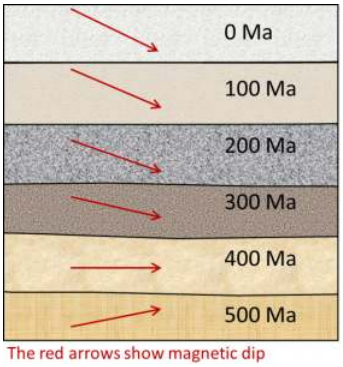


Figure 8.8.3 Hypothetical magnetic dip angles from layers of rock. This rock would have been south of the equator 500 Ma, at the equator 400 Ma, and since then has been moving further north (Steven Earle, "Physical Geology").

This paleomagnetic work of the 1950s was the first new evidence in favor of continental drift, and it led a number of geologists to start thinking that the idea might have some merit.

*"Physical Geology" by Steven Earle used under a CC-BY 4.0 international license. Download this book for free at http://open.bccampus.ca

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CHAPTER OVERVIEW

9: Chemical Oceanography

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9.3: Classifying Matter According to Its State—Solid, Liquid, and Gas
9.4: Cutting Aluminum until you get Atoms
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9.1: In Your Room

We will begin our discussion of chemical oceanography with some basic chemistry.

Matter is any substance that has mass and takes up space. Matter includes atoms and anything made up of atoms, but not other energy phenomena or waves such as light or sound. While this simple definition is easily applied, the way people view matter is often broken down into two characteristic length scales: the macroscopic and the microscopic.



Figure 9.1.1: A typical American university and college dormitory room in 2002 (CC BY-SA 3.0; Raul654).

The macroscopic scale is the length scale on which objects or phenomena are large enough to be visible almost practically with the naked eye, without magnifying optical instruments. Everything that one can see, touch, and handle in the dorm room of Figure 9.1.1 is within the macroscopic scale. To describe each of these objects, only a few macroscopic properties are required. However, each of these items can be decomposed into smaller microscopic scale properties.

The microscopic scale is the scale of objects and events smaller than those that can easily be seen by the naked eye, requiring a lens or microscope to see them clearly. All the everyday objects that we can bump into, touch or squeeze are ultimately composed of atoms. This ordinary atomic matter is in turn made up of interacting subatomic particles—usually a nucleus of protons and neutrons, and a cloud of orbiting electrons. Because of this, a large number of variables are needed to describe such a system which complicates the characterization.

Matter vs. MAss

Matter should not be confused with mass, as the two are not the same in modern physics. Matter is a physical substance of which systems may be composed, while mass is not a substance, but rather a quantitative property of matter and other substances or systems.

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9.2: What is Matter?

Learning Objectives

- Define matter and explain how it is composed of building blocks known as "atoms".

We are all familiar with matter. The definition of **Matter** is anything that has mass and volume (takes up space). For most common objects that we deal with every day, it is fairly simple to demonstrate that they have mass and take up space. You might be able to imagine, however, the difficulty for people several hundred years ago to demonstrate that air had mass and volume. Air (and all other gases) are invisible to the eye, have very small masses compared to equal amounts of solids and liquids, and are quite easy to compress (change volume). Without sensitive equipment, it would have been difficult to convince people that gases are matter. Today, we can measure the mass of a small balloon when it is deflated and then blow it up, tie it off, and measure its mass again to detect the additional mass due to the air inside. The mass of air, under room conditions, that occupies a one quart jar is approximately 0.0002 pounds. This small amount of mass would have been difficult to measure in times before balances were designed to accurately measure very small masses. Later, scientists were able to compress gases into such a small volume that the gases turned into liquids, which made it clear that gases are matter.



Figure 9.2.1: Everything from an ant, to a truck, to the earth, and even the entire galaxy is composed of matter. Images used with permission from Wikipedia (CC_SA-BY-3.0; credit High Contrast).

Even though the universe consists of "things" as wildly different as ants and galaxies, the matter that makes up all of these "things" is composed of a very limited number of building blocks. These building blocks are known as **atoms**, and so far, scientists have discovered or created a grand total of 118 different types of atoms. Scientists have given a name to each different type of atom. A substance that is composed of only one type of atom is called an element. At this point, what should amaze you is that all forms of matter in our universe are made with only 118 different building blocks. In some ways, it's sort of like cooking a gourmet, five-course meal using only three ingredients! How is it possible? To answer that question, you have to understand the ways in which different elements are put together to form matter.

The most important method that nature uses to organize atoms into matter is the formation of **molecules**. Molecules are groups of two or more atoms that have been bonded together. There are millions of different ways to bond atoms together, which means that there are millions of different possible molecules. Each of these molecules has its own set of chemical properties, and it's these properties with which chemists are most concerned. You will learn a lot more about atoms and molecules, including how they were discovered, in a later part of the textbook.

Summary

All matter has mass and occupies space. All physical objects are made of matter. Matter itself is composed of tiny building blocks known as "atoms". There are only 118 different types of atoms known to man. Frequently, atoms are bonded together to form "molecules".

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9.3: Classifying Matter According to Its State—Solid, Liquid, and Gas

Learning Objectives

- To describe the solid, liquid and gas phases.

Water can take many forms. At low temperatures (below $0^{\circ}C$), it is a solid. When at "normal" temperatures (between $0^{\circ}C$ and $100^{\circ}C$), it is a liquid. While at temperatures above $100^{\circ}C$, water is a gas (steam). The state that water is in depends upon the temperature. Each state has its own unique set of physical properties. Matter typically exists in one of three states: **solid**, **liquid**, or **gas**.

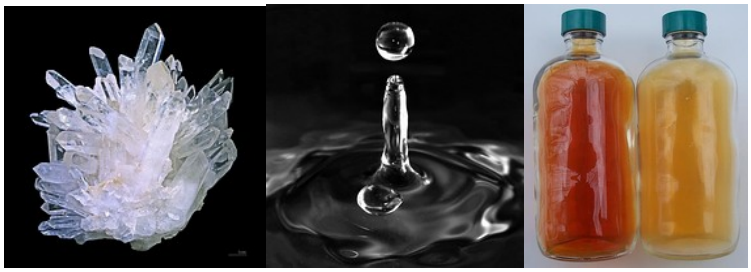


Figure $9.3.1$: Matter is usually classified into three classical states, with plasma sometimes added as a fourth state. From left to right: quartz (solid), water (liquid), nitrogen dioxide (gas).

The state that a given substance exhibits is also a physical property. Some substances exist as gases at room temperature (oxygen and carbon dioxide), while others, like water and mercury metal, exist as liquids. Most metals exist as solids at room temperature. All substances can exist in any of these three states. Figure $9.3.2$ shows the differences among solids, liquids, and gases at the molecular level. A solid has definite volume and shape, a liquid has a definite volume but no definite shape, and a gas has neither a definite volume nor shape.

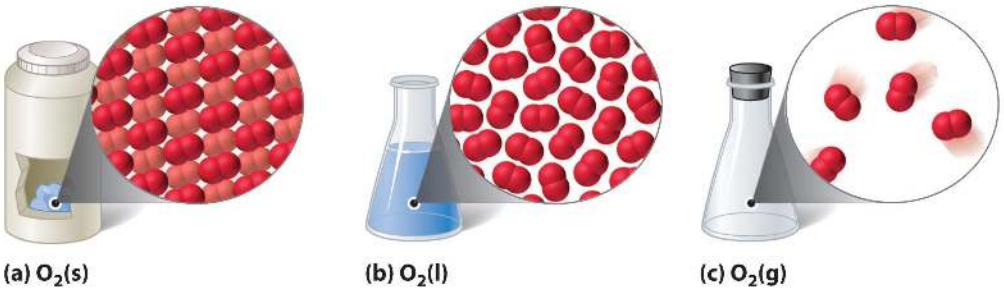
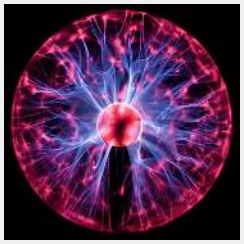


Figure $9.3.2$: A Representation of the Solid, Liquid, and Gas States. (a) Solid O_2 has a fixed volume and shape, and the molecules are packed tightly together. (b) Liquid O_2 conforms to the shape of its container but has a fixed volume; it contains relatively densely packed molecules. (c) Gaseous O_2 fills its container completely—regardless of the container's size or shape—and consists of widely separated molecules.

Plasma: A Fourth State of Matter

Technically speaking, a fourth state of matter called plasma exists, but it does not naturally occur on earth, so we will omit it from our study here.



A plasma globe operating in a darkened room. (CC BY-SA 3.0; Chocolateoak).

Solids

In the solid state, the individual particles of a substance are in fixed positions with respect to each other because there is not enough thermal energy to overcome the intermolecular interactions between the particles. As a result, solids have a definite shape and volume. Most solids are hard, but some (like waxes) are relatively soft. Many solids composed of ions can also be quite brittle.

Solids are defined by the following characteristics:

- Definite shape (rigid)
- Definite volume
- Particles vibrate around fixed axes

If we were to cool liquid mercury to its freezing point of -39°C, and under the right pressure conditions, we would notice all of the liquid particles would go into the solid state. Mercury can be solidified when its temperature is brought to its freezing point. However, when returned to room temperature conditions, mercury does not exist in solid state for long, and returns back to its more common liquid form.

Solids usually have their constituent particles arranged in a regular, three-dimensional array of alternating positive and negative ions called a **crystal**. The effect of this regular arrangement of particles is sometimes visible macroscopically, as shown in Figure 9.3.3. Some solids, especially those composed of large molecules, cannot easily organize their particles in such regular crystals and exist as amorphous (literally, “without form”) solids. Glass is one example of an amorphous solid.

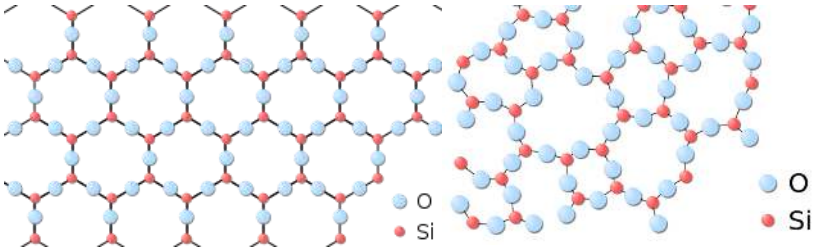


Figure 9.3.3: (left) The periodic crystalline lattice structure of quartz SiO_2 in two-dimensions. (right) The random network structure of glassy SiO_2 in two-dimensions. Note that, as in the crystal, each Silicon atom is bonded to 4 oxygen atoms, where the fourth oxygen atom is obscured from view in this plane. Images used with permission (public domain).

Liquids

If the particles of a substance have enough energy to partially overcome intermolecular interactions, then the particles can move about each other while remaining in contact. This describes the liquid state. In a liquid, the particles are still in close contact, so liquids have a definite volume. However, because the particles can move about each other rather freely, a liquid has no definite shape and takes a shape dictated by its container.

Liquids have the following characteristics:

- No definite shape (takes the shape of its container).
- Has definite volume.
- Particles are free to move over each other, but are still attracted to each other.

A familiar liquid is mercury metal. Mercury is an anomaly. It is the only metal we know of that is liquid at room temperature. Mercury also has an ability to stick to itself (surface tension)—a property that all liquids exhibit. Mercury has a relatively high surface tension, which makes it very unique. Here you see mercury in its common liquid form.



Video 9.3.1*: Mercury boiling to become a gas.*

If we heat liquid mercury to its boiling point of 357°C under the right pressure conditions, we would notice all particles in the liquid state go into the gas state.

Gases

If the particles of a substance have enough energy to completely overcome intermolecular interactions, then the particles can separate from each other and move about randomly in space. This describes the gas state, which we will consider in more detail elsewhere. Like liquids, gases have no definite shape, but unlike solids and liquids, gases have no definite volume either. The change from solid to liquid usually does not significantly change the volume of a substance. However, the change from a liquid to a gas significantly increases the volume of a substance, by a factor of 1,000 or more. Gases have the following characteristics:

- No definite shape (takes the shape of its container)
- No definite volume
- Particles move in random motion with little or no attraction to each other
- Highly compressible

Table 9.3.1: Characteristics of the Three States of Matter

Characteristics	Solids	Liquids	Gases
shape	definite	indefinite	indefinite
volume	definite	definite	indefinite
relative intermolecular interaction strength	strong	moderate	weak
relative particle positions	in contact and fixed in place	in contact but not fixed	not in contact, random positions

Example 9.3.1

What state or states of matter does each statement, describe?

a. This state has a definite volume, but no definite shape.
b. This state has no definite volume.
c. This state allows the individual particles to move about while remaining in contact.

Solution

a. This statement describes the liquid state.
b. This statement describes the gas state.
c. This statement describes the liquid state.

Exercise 9.3.1

What state or states of matter does each statement describe?

a. This state has individual particles in a fixed position with regard to each other.
b. This state has individual particles far apart from each other in space.
c. This state has a definite shape.

Answer a:

solid

Answer b:

gas

Answer c:

solid

Summary

- Three states of matter exist—solid, liquid, and gas.
- Solids have a definite shape and volume.
- Liquids have a definite volume, but take the shape of the container.
- Gases have no definite shape or volume.

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9.4: Cutting Aluminum until you get Atoms

Take some aluminum foil. Cut it in half. Now there are two smaller pieces of aluminum foil. Cut one of the pieces in half again. Cut one of those smaller pieces in half again. Continue cutting, making smaller and smaller pieces of aluminum foil. It should be obvious that the pieces are still aluminum foil; they are just becoming smaller and smaller. But how far can this exercise be taken, at least in theory? Can one continue cutting the aluminum foil into halves forever, making smaller and smaller pieces? Or is there some limit, some absolute smallest piece of aluminum foil? Thought experiments like this—and the conclusions based on them—were debated as far back as the fifth century BC by Democritus and other ancient Greek philosophers (Figure 9.4.1).

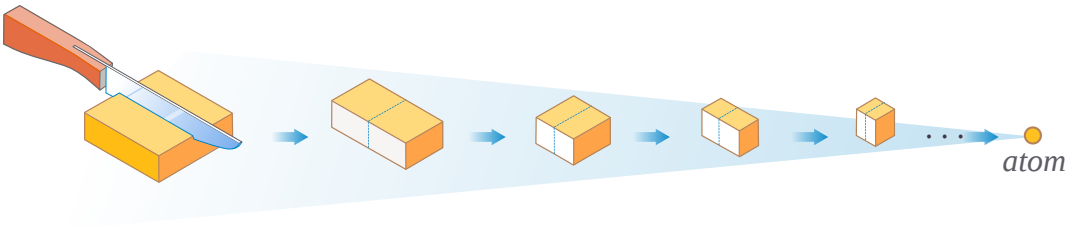


Figure 9.4.1: Democritus argued that matter, like an aluminum block, cannot be repeatedly cut in half perpetually. At some point, a limit is reached before the substance can no longer be called "aluminum"; this is the atomic limit. (CC BY-NC; Ümit Kaya via LibreTexts)

Most elements in their pure form exist as individual atoms. For example, a macroscopic chunk of iron metal is composed, microscopically, of individual atoms. Some elements, however, exist as groups of atoms called molecules. Several important elements exist as two-atom combinations and are called diatomic molecules. In representing a diatomic molecule, we use the symbol of the element and include the subscript 2 to indicate that two atoms of that element are joined together. The elements that exist as diatomic molecules are hydrogen (H_2), oxygen (O_2), nitrogen (N_2), fluorine (F_2), chlorine (Cl_2), bromine (Br_2), and iodine (I_2).

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9.5: The Structure of Atoms

Learning Objectives

- Describe the three main subatomic particles.
- State how the subatomic particles are arranged in atoms.

There have been several minor but important modifications to Dalton's atomic theory. For one thing, Dalton considered atoms to be indivisible. We know now that atoms not only can be divided but also are composed of three different kinds of particle, *subatomic particles*, with their own properties that are different from the chemical properties of atoms.

Subatomic Particles

The first subatomic particle was identified in 1897 and called the electron. It is an extremely tiny particle, with a mass of about 9.109×10^{-31} kg. Experiments with magnetic fields showed that the electron has a negative electrical charge. By 1920, experimental evidence indicated the existence of a second particle. A proton has the same amount of charge as an electron, but its charge is positive, not negative. Another major difference between a proton and an electron is mass. Although still incredibly small, the mass of a proton is 1.673×10^{-27} kg, which is almost 2,000 times greater than the mass of an electron. Because opposite charges attract each other (while like charges repel each other), protons attract electrons (and vice versa).

Finally, additional experiments pointed to the existence of a third particle. Evidence produced in 1932 established the existence of the neutron, a particle with about the same mass as a proton but with no electrical charge, it is *neutral*. We understand now that all atoms can be broken down into subatomic particles: protons, neutrons, and electrons. Table 9.5.1 lists some of their important characteristics and the symbols used to represent each particle.

Table 9.5.1: Properties of the Subatomic Particles

Particle	Symbol	Mass (kg)	Relative Mass (proton = 1)	Relative Charge
proton	p^+	1.673×10^{-27}	1	+1
neutron	n^0	1.675×10^{-27}	1	0
electron	e^-	9.109×10^{-31}	0.00055	−1

The Nucleus

How are these subatomic particles arranged? Between 1909 and 1911, Ernest Rutherford, a Cambridge physicist, and his associates Hans Geiger and Ernest Marsden performed experiments that provided strong evidence concerning the internal structure of an atom. They took a very thin metal foil, such as gold or platinum, and aimed a beam of positively charged particles (called alpha particles, which are combinations of two protons and two neutrons) from a radioactive source toward the foil. Surrounding the foil was a detector—either a scintillator (a material that glows when hit by such particles) or some unexposed film (which is exposed where the particles hit it). The detector allowed the scientists to determine the distribution of the alpha particles after they interacted with the foil. Figure 9.5.1 shows a diagram of the experimental setup.

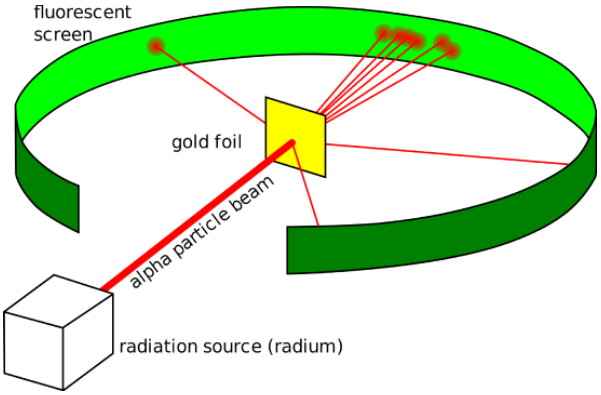


Figure 9.5.1: The Geiger-Marsden Experimental Setup. Experiments using this setup were used to investigate the structure of atoms. In this experiment, most of the particles traveled straight through the foil, but some alpha particles were deflected off to one side. Some were even deflected back toward the source. This was unexpected. Rutherford once said, "It was almost as incredible as if you fired a 15-inch shell at a piece of tissue paper and it came back and hit you." (CC SA-BY 3.0; Kurzon).

Rutherford proposed the following model to explain these experimental results. Protons and neutrons are concentrated in a central region he called the nucleus (plural, *nuclei*) of the atom. Electrons are outside the nucleus and orbit about it because they are attracted to the positive charge in the nucleus. Most of the mass of an atom is in the nucleus, while the orbiting electrons account for an atom's size. As a result, an atom consists largely of empty space. Rutherford called his description the "planetary model" of the atom. Figure 9.5.2 shows how this model explains the experimental results.

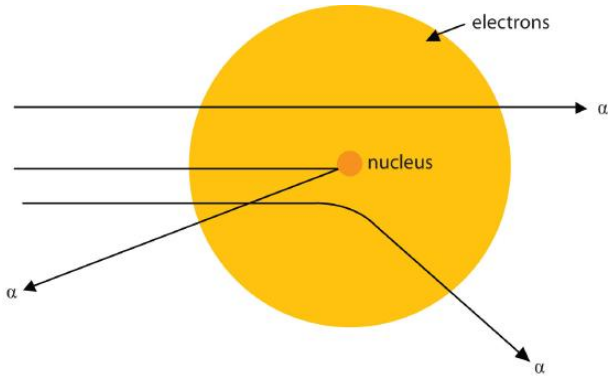


Figure 9.5.2: Rutherford's Metal Foil Experiments. Rutherford explained the results of the metal foil experiments by proposing that most of the mass and the positive charge of an atom are located in its nucleus, while the relatively low-mass electrons orbit about the nucleus. Most alpha particles go straight through the empty space, a few particles are deflected, and fewer still ricochet back toward the source. The nucleus is much smaller proportionately than depicted here.

> *The planetary model of the atom replaced the plum pudding model, which had electrons floating around aimlessly like plums in a "pudding" of positive charge.*

Rutherford's model is essentially the same model that we use today to describe atoms but with one important modification. The planetary model suggests that electrons occupy certain specific, circular orbits about the nucleus. We know now that this model is overly simplistic. A better description is that electrons form fuzzy clouds around nuclei. Figure 9.5.3 shows a more modern version of our understanding of atomic structure.

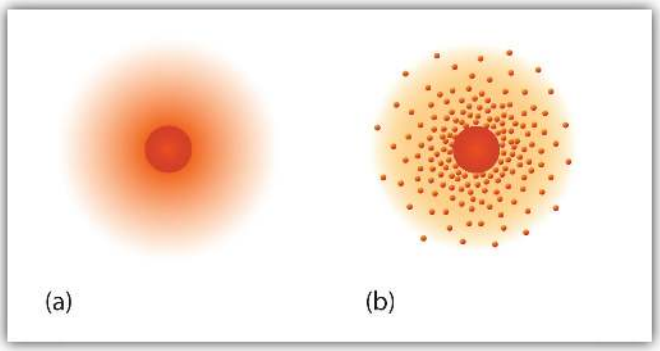


Figure 9.5.3: A Modern Depiction of Atomic Structure. A more modern understanding of atoms, reflected in these representations of the electron in a hydrogen atom, is that electrons occupy regions of space about the nucleus; they are not in discrete orbits like planets around the sun. (a) The darker the color, the higher the probability that an electron will be at that point. (b) In a two-dimensional cross section of the electron in a hydrogen atom, the more crowded the dots, the higher the probability that an electron will be at that point. In both (a) and (b), the nucleus is in the center of the diagram.

Concept Review Exercises

1. What are the charges and the relative masses of the three subatomic particles?
2. Describe the structure of an atom in terms of its protons, neutrons, and electrons.

Answers

1. proton: +1, large; neutron: 0, large; electron: −1, small
2. Protons and neutrons are located in a central nucleus, while electrons orbit about the nucleus.

Key Takeaways

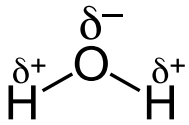
- Atoms are composed of three main subatomic particles: protons, neutrons, and electrons.
- Protons and neutrons are grouped together in the nucleus of an atom, while electrons orbit about the nucleus.

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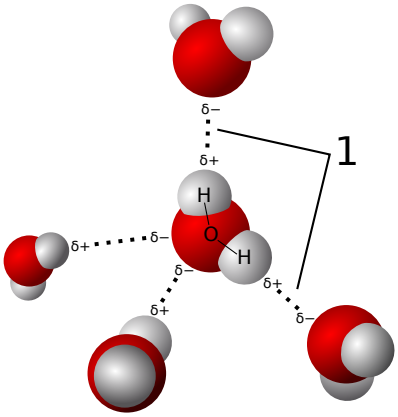
9.6: Unique properties of water

Now, let's look at the main chemical discussed in oceanography- WATER!



Chemical Properties of Water

Water is made up of two hydrogen atoms (represented by H in the image above) and one oxygen atom (represented by the O in the image above). It is unique in that it is dipolar. This means that the molecule has a positive charge on one side and a negative charge on the other. The side of the water molecule where the hydrogen atoms are located ends up being positively charged, and the side of the molecule where the oxygen atom is located is negatively charged. This is because electrons are not equally distributed in the water molecule. The electrons are pulled toward the oxygen atom, resulting in its negative charge. Each hydrogen atom in water is covalently bonded to the oxygen via a shared pair of electrons. Oxygen also has two unshared pairs of electrons, thus there are four pairs of electrons surrounding the oxygen atom, two pairs involved in covalent bonds with hydrogen, and two unshared pairs on the opposite side of the oxygen atom.



One water molecule can connect with another molecule via a hydrogen bond between a hydrogen atom of one molecule and an oxygen atom of the other. Because it has a negative side and a positive side, water molecules can easily bond to other polar or charged molecules. This can be seen by the many constituents dissolved in seawater. Their ability to dissolve is due to the polar nature of the water molecule.

Water is a polar molecule means that there is an uneven distribution of electron density. Water has a partial negative charge near the oxygen atom due the unshared pairs of electrons,and partial positive charges near the hydrogen atoms. An electrostatic attraction between the partial positive charge near the hydrogen atoms and the partial negative charge near the oxygen results in the formation of a hydrogen bond. And the ability of ions and other molecules to dissolve in water is due to polarity.

Online resource: watereducation.utah.gov/water...es/default.asp

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9.7: Properties of Water

The most obvious feature of the oceans is that they contain water. Water is so ubiquitous that it may not seem like a very interesting substance, but it has many unique properties that impact global oceanographic and climatological processes. Many of these processes are due to **hydrogen bonds** forming between water molecules.

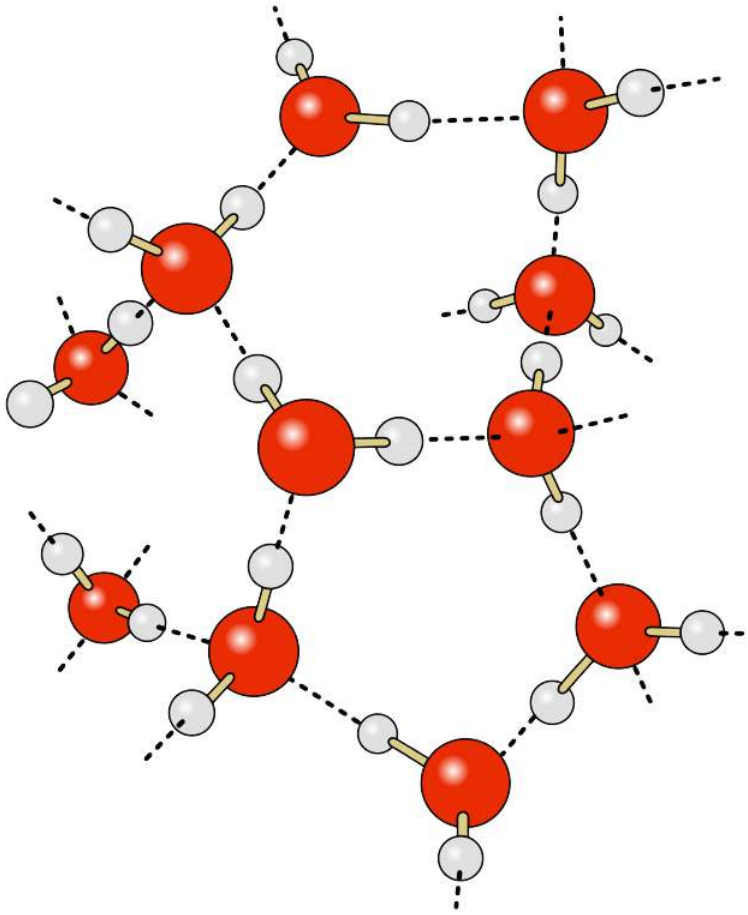


Figure 9.7.1 Hydrogen bonds (dashed lines) between water molecules. Oxygen atoms are shown in red, hydrogen atoms in white (Public domain, via Wikimedia Commons).

The water molecule consists of two hydrogen atoms and one oxygen atom. The electrons responsible for the bonds between the atoms are not distributed equally throughout the molecule, so that the hydrogen ends of water molecules have a slight positive charge, and the oxygen end has a slight negative charge, making water a **polar molecule**. The negative oxygen side of the molecule forms an attraction to the positive hydrogen end of a neighboring molecule. This rather weak force of attraction is called a hydrogen bond (Figure 9.7.1). If not for hydrogen bonds, water would vaporize at -68° C, meaning liquid water (and thus life) could not exist on Earth. These hydrogen bonds are responsible for some of water's unique properties:

1. Water is the only substance to naturally exist in a solid, liquid, and gaseous form under the normal range of temperatures and pressures found on Earth. This is due to water's relatively high freezing and vaporizing points (see below).

2. Water has a **high heat capacity**, which is the amount of heat that must be added to raise its temperature. Specific heat is the heat required to raise the temperature of 1 g of a substance by 1° C. Water has the highest specific heat of any liquid except ammonia (Table 5.1.1).

[table id=3 /]

Water is therefore one of the most difficult liquids to heat or cool; it can absorb large amounts of heat without increasing its temperature. Remember that temperature reflects the average kinetic energy of the molecules within a substance; the more vigorous the motion, the higher the temperature. In water, the molecules are held together by hydrogen bonds, and these bonds must be

overcome to allow the molecules to move freely. When heat is added to water the energy must first go to breaking the hydrogen bonds before the temperature can begin to rise. Therefore, much of the added heat is absorbed by breaking H bonds, not by increasing the temperature, giving water a high heat capacity.

Hydrogen bonds also give water a high latent heat; the heat required to undergo a phase change from solid to liquid, or liquid to gas. The **latent heat of fusion** is the heat required to go from solid to liquid; 80 cal/g in the case of ice melting to water. Ice is a solid because hydrogen bonds hold the water molecules into a solid crystal lattice (see below). As ice is heated, the temperature rises up to 0° C. At that point, any additional heat goes to melting the ice by breaking the hydrogen bonds, not to increasing the temperature. So as long as ice is present, the water temperature will not increase. This is why your drink will remain cold as long as it contains ice; any heat absorbed goes to melting the ice, not to warming the drink.

When all of the ice is melted, additional heat will increase the temperature of the water 1° C for each calorie of heat added, until it reaches 100° C. At that point, any additional heat goes to overcoming the hydrogen bonds and turning the liquid water into water vapor, rather than increasing the water temperature. The heat required to evaporate liquid water into water vapor is the **latent heat of vaporization** which has a value of 540 cal/g (Figure 9.7.2).

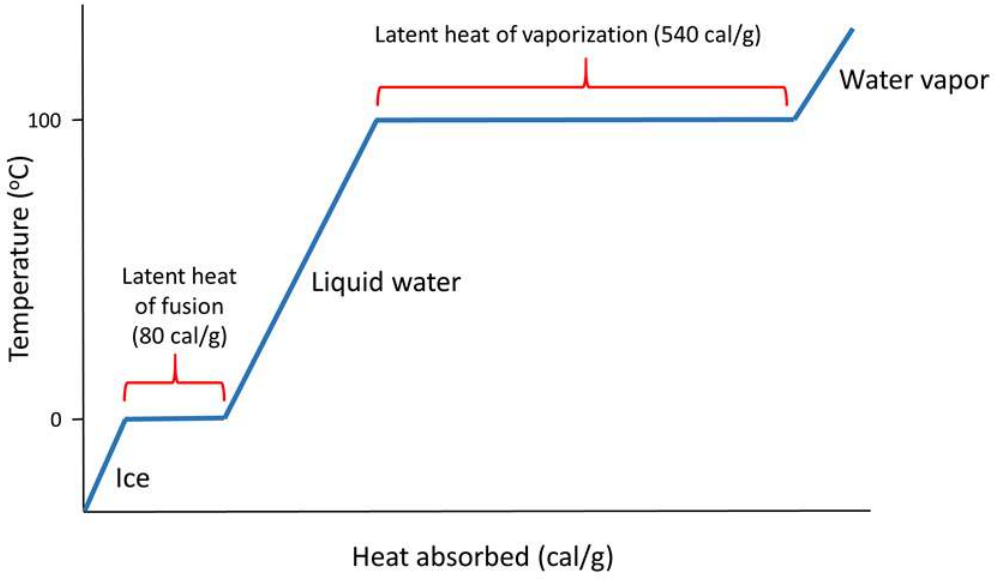


Figure 9.7.2 Latent heat required for phase changes in water. Latent heat of fusion is the heat required to melt ice (80 cal/g), and latent heat of vaporization is the heat required to turn liquid water into water vapor (540 cal/g) (PW).

The high heat capacity of water helps regulate global climate, as the oceans slowly absorb and release heat, preventing rapid swings in temperature (see section 8.1). It also means that aquatic organisms aren't as subjected to the same rapid temperature changes as terrestrial organisms. A deep ocean organism may not experience more than a 0.5^o C change in temperature over its entire life, while a terrestrial species may encounter changes of more than 20° C in a single day!

3. Water dissolves more substances than any other liquid; it is a **"universal solvent"**, which is why so many substances are dissolved in the ocean. Water is especially good at dissolving ionic salts; molecules made from oppositely charged ions such as NaCl (Na^+ and Cl^-). In water, the charged ions attract the polar water molecules. The ions get surrounded by a layer of water molecules, weakening the bond between the ions by up to 80 times. With the bonds weakened between ions, the substance dissolves (Figure 9.7.3).

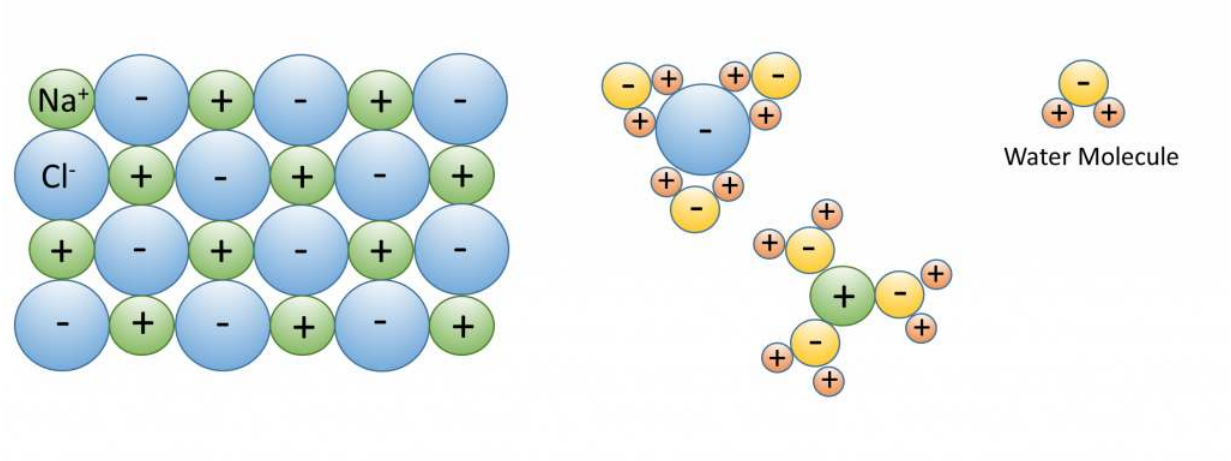


Figure 9.7.3 Attraction between polar water molecules and charged ions (such as in NaCl) is greater than the attraction between the charged ions, causing the ions to dissociate and the salt to dissolve (PW).

4. The **solid phase is less dense than the liquid phase**. In other words, ice floats. Most substances are denser in the solid form than in the liquid form, as their molecules are more closely packed together as a solid. Water is an exception: the density of fresh water is 1.0 g/cm^3, while the density of ice is 0.92 g/cm^3, and once again, this is due to the action of hydrogen bonds.

As water temperature cools the molecules slow down, eventually slowing enough that hydrogen bonds can form and hold the water molecules in a crystal lattice. The molecules in the lattice are spaced farther apart than the molecules in liquid water, which makes ice less dense than liquid water (Figure 9.7.4). This is familiar to anyone who has ever left a full water bottle in the freezer, only to have it burst as the water freezes and expands.

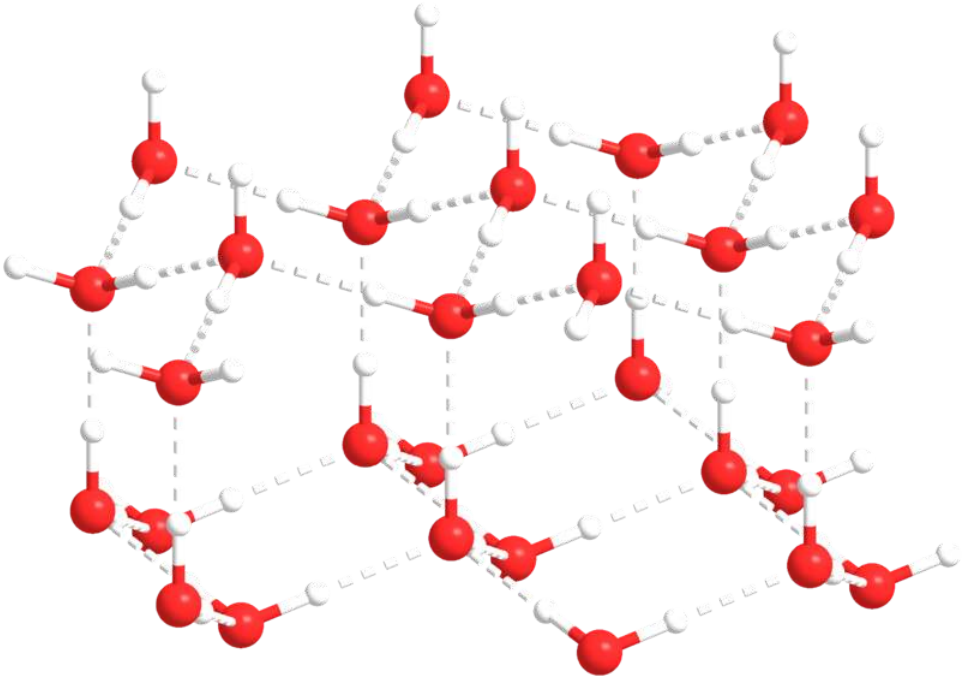


Figure 9.7.4 Crystal lattice structure of ice, showing water molecules held together by hydrogen bonds (By Adam001d (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

But the relationship between temperature and water density is not a simple linear one. As water cools, its density increases as expected, as the water molecules slow down and get closer together. However, fresh water reaches its maximum density at a temperature of 4° C, and as it cools beyond that point its density declines as the hydrogen bonds begin to form and the intermolecular spacing increases (Figure 9.7.5 inset). The density continues to decline until the temperature reaches 0° C and ice crystals form, reducing the density dramatically (Figure 9.7.5).

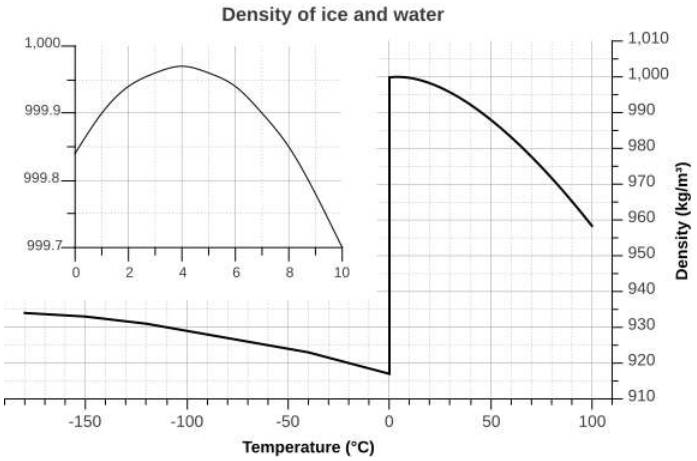


Figure 9.7.5 As temperature declines, the density of water increases until it reaches maximum density at 4° C (inset). Density then declines slightly down to 0° C, where it declines dramatically as ice crystals form (Klaus-Dieter Keller, [CC BY-SA 3.0], via Wikimedia Commons).

There are a number of important implications to ice being less dense than water. Ice floating on the surface of the ocean helps regulate ocean temperatures, and therefore global climate, by influencing the amount of sunlight that is reflected rather than absorbed (see section 8.1). On a smaller scale, surface ice can prevent lakes and ponds from freezing solid during the winter. As fresh surface water cools, the water gets denser, and sinks to the bottom. The new surface water then cools and sinks, and the process is repeated in what is referred to as **overturning**, with denser water sinking and less dense water moving to the surface only to be cooled and sink itself. In this way, the entire body of water is cooled somewhat evenly. This process continues until the surface water cools below 4° C. Below 4° C, the water becomes *less* dense as it cools, so it no longer sinks. Instead, it remains as the surface, getting colder and less dense, until it freezes at 0° C. Once fresh water freezes, the ice floats and insulates the rest of the water beneath it, reducing further cooling. The densest bottom water is still at 4° C, so it does not freeze, allowing the bottom of a lake or pond to remain unfrozen (which is good news for the animals living there) no matter how cold it gets outside.

The dissolved salts in seawater inhibit the formation of the crystal lattice, and therefore make it harder for ice to form. So seawater has a freezing point of about -2° C (depending on salinity), and freezes before a temperature of maximum density is reached. Thus seawater will continue to sink as it gets colder, until it finally freezes.

5. Water has a very **high surface tension**, the highest of any liquid except mercury (Table 5.1.2). Water molecules are attracted to each other by hydrogen bonds. For molecules not at the water surface, they are surrounded by other water molecules in all directions, so the attractive forces are evenly distributed in all directions. But for molecules at the surface there are few adjacent molecules above them, only below, so all of the attractive forces are directed inwards, away from the surface (Figure 9.7.6). This inwards force is what causes water droplets to take on a spherical shape, and water to bead up on a surface, as the spherical shape provides the minimum possible surface area. These attractive forces also cause the surface of the water to act like an elastic “skin” which allows things like insects to sit on the water’s surface without sinking.

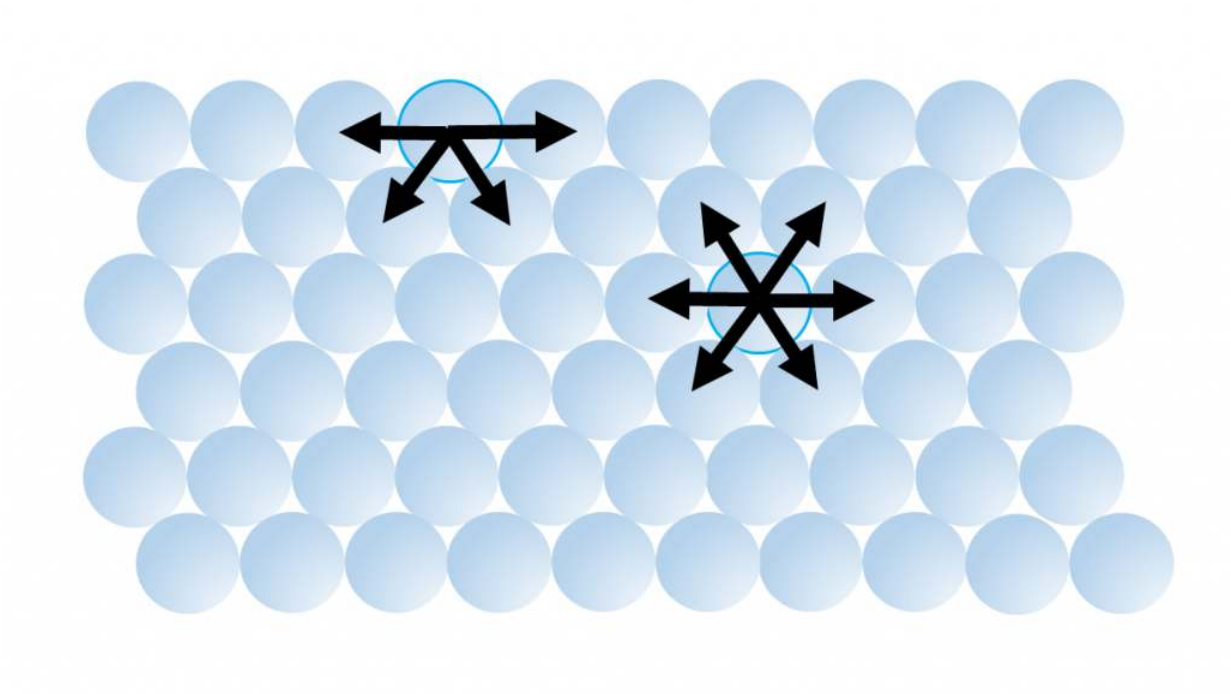


Figure 9.7.6 The net attractive force between molecules at the surface is inwards, leading to surface tension. For molecules in the center, the force is equal in all directions (PW).

[table id=8 /]

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9.8: Salinity Patterns

All of the salts and ions that dissolve in seawater contribute to its overall **salinity**. Salinity of seawater is usually expressed as the grams of salt per kilogram (1000 g) of seawater. On average, about 35 g of salt is present in each 1 kg of seawater, so we say that the average salinity of the ocean salinity is 35 parts per thousand (ppt). Note that 35 ppt is equivalent to 3.5% (parts per hundred). Some sources now use practical salinity units (PSU) to express salinity values, where 1 PSU = 1 ppt. The units are not included, so we can refer simply to a salinity of 35.

Many different substances are dissolved in the ocean, but six ions comprise about 99.4% of all the dissolved ions in seawater. These six **major ions** are (Table 5.3.1):

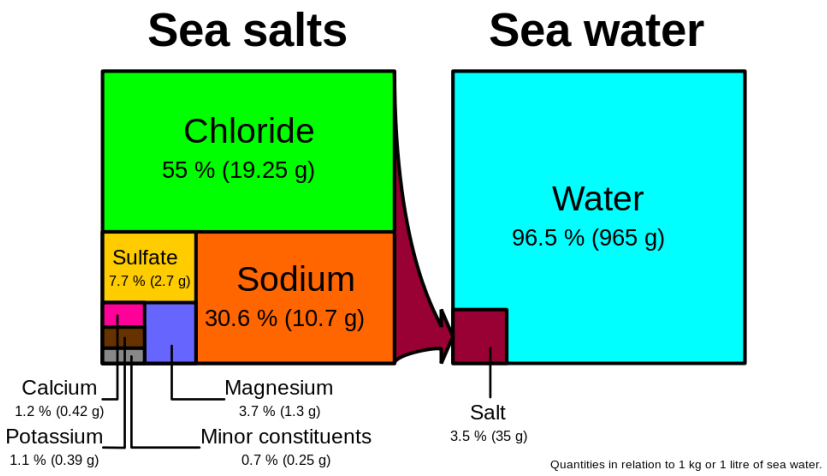


Figure 9.8.1 The relative proportions of ions in seawater. (By derivative work: Tcncv (talk) Sea_salt-e_hg.svg: Hannes Grobe, Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany; SVG version by Stefan Majewsky (Sea_salt-e_hg.svg) [CC BY-SA 2.5], via Wikimedia Commons)

Chloride and sodium, the components of table salt (sodium chloride NaCl), make up over 85% of the ions in the ocean, which is why seawater tastes salty (Figure 9.8.1). In addition to the major constituents, there are numerous minor constituents; radionucleotides, organic compounds, metals etc. These minor constituents are found in concentrations of ppm (parts per million) or ppb (parts per billion), unlike the major ions that are far more abundant (ppt) (Table 5.3.2). To put this into perspective, 1 ppm = 1 mg/kg, or the equivalent of 1 teaspoon of sugar dissolved in 14,000 cans of soda. 1 ppb = 1 µg/kg, or the equivalent of 1 teaspoon of a substance dissolved in five Olympic-sized swimming pools! These minor constituents represent numerous substances, but together they make up less than 1% of the ions in the seawater. Some of these may be important as minerals and trace elements vital to living organisms, but they don't have much impact on the overall salinity. But given the vast size of the oceans, even materials found in trace abundance can represent fairly large reservoirs. For example gold is a trace element in seawater, found in concentrations of parts per trillion, yet if you could extract all of the gold in just one km^3 of seawater, it would be worth about $20 million!

Because the six major ions in seawater comprise over 99% of the total salinity, changes in abundance of the minor constituents have little effect on overall salinity. Furthermore, the **rule of constant proportions** states that even though the absolute salinity of ocean water might differ in different places, the relative *proportions* of the six *major ions* within that water are always constant. For example, no matter the total salinity of a seawater sample, 55% of the total salinity will be due to chloride, 30% due to sodium, and so on. Since the proportion of these major ions does not change, we call these **conservative ions**.

Given these constant proportions, in order to calculate total salinity you can simply measure the concentration of just one of the major ions and use that value to calculate the rest. Traditionally chloride has been the ion measured because it is the most abundant, and thus the simplest to measure accurately. Multiplying the concentration of chloride by 1.8 gives the total salinity. For example, looking at Figure 9.8.1, 19.25 g/kg (ppt) chloride x 1.8 = 35 ppt. Today, for rapid measurements of salinity, electrical conductivity is often used rather than determining chloride concentrations (see box below).

Measuring Salinity

There are a number of methods available for measuring the salinity of water. The most precise measurements utilize direct chemical analysis of the seawater in a lab setting, but there are a number of ways to get immediate salinity measurements in the field. For a quick estimate of salinity, a hand-held **refractometer** can be used (right).



Handheld refractometer (Photo by CEphoto, Uwe Aranas / CC-BY-SA-3.0 via Wikimedia Commons)

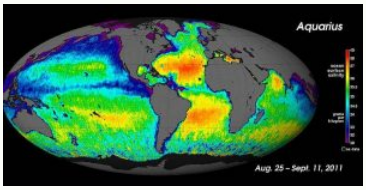
This instrument measures the degree of bending, or refraction, of light rays as they pass through a fluid. The greater the amount of dissolved salts in the sample, the greater the degree of light refraction. The observer traps a drop of water on the blue screen, and looks through the eyepiece. The dividing line between the blue and white sections of the scale (inset) can be used to read the salinity.

For more accurate measurements, most oceanographers use an instrument that measures electrical **conductivity**. An electrical current is passed between two electrodes immersed in water, and the higher the salinity, the more readily the current will be conducted (the ions in seawater conduct electrical currents). Conductivity probes are often bundled into an instrument called a **CTD**, which stands for Conductivity, Temperature, and Depth, which are the most commonly-measured parameters. Modern CTDs can be outfitted with an array of probes measuring parameters like light, turbidity (water clarity), dissolved gases etc. CTDs can be large instruments (below), but small hand-held salinity probes are also widely available.



Left: The inner workings of a CTD. An array of different probes are attached. Right: A CTD in a protective cage ready to be deployed at sea. (Left: Hannes Grobe [CC BY 3.0 (https://creativecommons.org/licenses/by/3.0)]; Right: © Hans Hillewaert (CC BY-SA 4.0) via Wikimedia Commons)

For large-scale salinity measurements, oceanographers can use satellites, such as the Aquarius satellite, which was able to measure surface salinity differences as small as 0.2 PSU as it mapped the ocean surface every seven days (below).



(NASA, Public Domain)

It is important to be aware that while the rule of constant proportions applies to most of the ocean, there may be certain coastal areas where lots of river discharge may alter these proportions slightly. Furthermore, it is important to remember that the rule of constant proportions only applies to the *major* ions. The proportions of the minor ions may fluctuate, but remember that they make a very minor contribution to overall salinity. Because the concentrations of the minor ions are not constant, these are referred to as **non-conservative ions**.

Why are the major ions found in constant proportions? There is constant input of ions from river runoff and other processes, usually in very different proportions from what is found in the ocean. So why don't the proportions in the oceans change? Most of the ions discharged by rivers have fairly low residence times (see section 5.2) compared to ions in seawater, usually because they are used in biological processes. These low residence times do not allow the ions to accumulate and alter salinity. Also, the mixing

time of the world ocean is around 1000 years, which is very short compared to the residence times of the major ions, which may be tens of millions of years long. So during the residence time of a single ion the ocean has mixed numerous times, and the major ions have become evenly distributed throughout the ocean.

Figure 9.8.4 Salinity profiles from two hypothetical sites in the open ocean, one from high latitude and one from low latitude (PW).

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9.9: Oxygen

Ions are not the only materials that are dissolved in seawater. The oceans also contain dissolved gases that are very important to living organisms, particularly oxygen (O_2), carbon dioxide (CO_2), and nitrogen (N_2). Oxygen is required for respiration in marine plants, algae, and phytoplankton (the primary producers) and animals. Carbon dioxide is utilized by the primary producers to power photosynthesis, a byproduct of which is oxygen. Nitrogen gas dissolved in the ocean is fixed by bacteria and converted into the forms required for primary production, such as nitrate and nitrite.

All of these gases are found in the atmosphere, and can enter the ocean by dissolving into the water at the ocean's surface. But the amount of each gas in air is very different from the amount found in the ocean (Table 5.4.1).

[table id=6 /]

The amount of each gas that can dissolve in the ocean depends on the solubility and saturation of the gas in water. **Solubility** refers to the amount of a dissolved gas that the water can hold under a particular set of conditions, which are usually defined as 0° C and 1 atmosphere of pressure. The solubility of a gas increases with increasing pressure, decreased temperature, and decreased salinity. **Saturation** refers to the amount of gas currently dissolved in the water, relative to the maximum possible content. If the water is undersaturated, more gas can dissolve. If the water is saturated or supersaturated, gas may be released. Most atmospheric gases are saturated in the ocean, but O_2 and CO_2 are not saturated because they are rapidly used by living organisms.

Oxygen

Typical oceanic dissolved oxygen profiles are shown in Figure 9.9.1. The shape of the profile is determined by the various processes that add or remove oxygen from the water at different depths.

Oxygen content is highest at the surface for two main reasons; this is where oxygen dissolves into the ocean from the atmosphere, and the surface water is where oxygen is produced by phytoplankton through photosynthesis. Respiration is also occurring in the surface waters, but the rate of photosynthetic oxygen production is greater than the rate of removal through respiration. It should be noted that even though dissolved oxygen is highest at the surface, there is still far less oxygen in the water than is found in the air. Well-oxygenated surface water may only contain around 8 mg O_2/l, while the air contains 210 mg O_2/l.

As depth increases, dissolved oxygen declines, reaching a minimum between a few hundred meters and 1000 m deep, the aptly-named **oxygen minimum layer**. At these depths and below, the water is too far removed from the surface for any atmospheric exchange, and there is not enough light to support photosynthesis, so there is little if any oxygen added to the water. At the same time, oxygen is removed from the water through the respiration of deep water organisms, and the decomposition of organic material by bacteria as it sinks to depth.

Below the oxygen minimum layer there is often an increase in dissolved oxygen at the greatest depths (Figures 5.4.1, 5.4.2). This bottom water is usually colder than the surface water and is under enormous pressure; as stated above, lower temperatures and higher pressure increase the solubility of dissolved gases. But there is another reason that bottom water contains more oxygen than mid-water depths that has to do with the way water circulates throughout the deep ocean (see section 9.8). In polar regions, the cold surface water absorbs lots of oxygen. This cold, oxygen-rich water sinks to the bottom due to its high density, taking the oxygen with it. The oxygen-rich bottom water will then spend the next thousand years or so moving over the seafloor throughout the major ocean basins. This deep water circulation is the source of oxygen for bottom-dwelling (benthic) organisms. The oxygen-rich bottom water forms in the polar regions of the Atlantic, and slowly makes its way to the Pacific, with oxygen being removed for respiration along the way. This is why dissolved oxygen levels in Pacific deep water are generally lower than in the Atlantic (Figure 9.9.1).

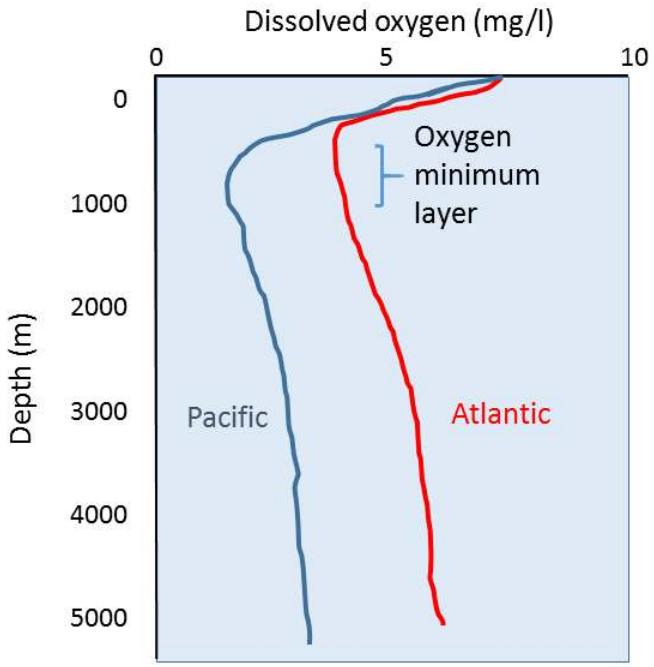


Figure 9.9.1 Representative dissolved oxygen profiles for the Pacific and Atlantic oceans (PW).

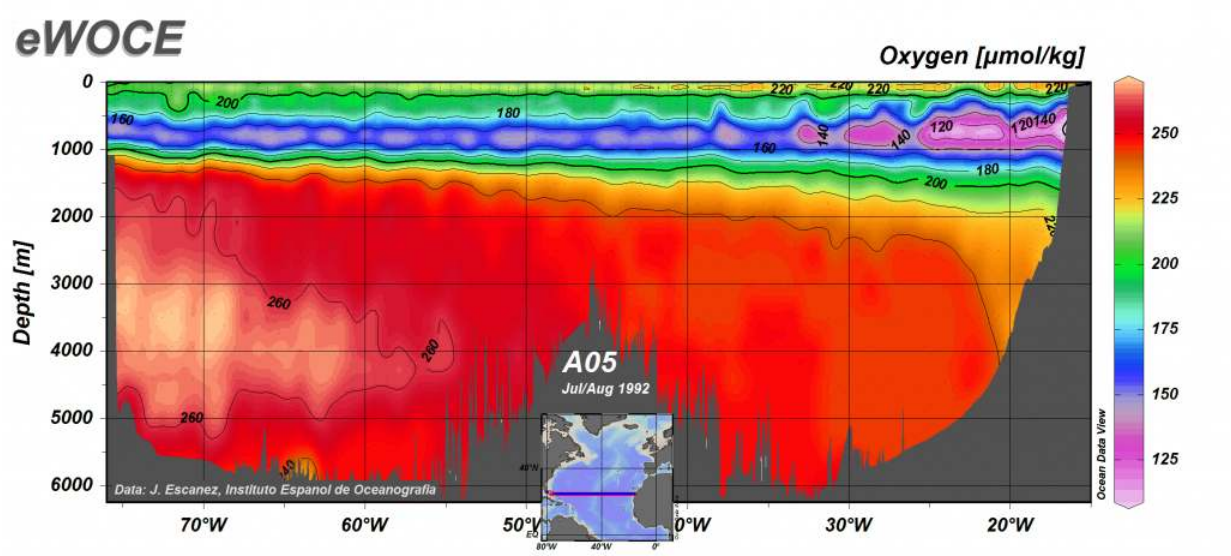


Figure 9.9.2 Dissolved oxygen profile from a transect across the Atlantic Ocean from Florida to the coast of Africa (inset). The oxygen minimum layer is visible between 500-1000 m (eWOCE, http://www.ewoce.org/gallery/eWOCE_T....html#Atlantic).

Areas where dissolved oxygen levels are too low to support most life are referred to as **hypoxic** zones (they are experiencing **hypoxia**, or low oxygen). Hypoxia is usually defined as oxygen levels below 2 mg/L. **Anoxic** zones (anoxia = without oxygen) show more severe forms of hypoxia, with oxygen below 0.5 mg/L. Some parts of the oceans may experience seasonal or temporary periods of hypoxia, while in other areas these conditions may last much longer. These hypoxic conditions often lead to mass die-offs of marine organisms who struggle to survive without sufficient oxygen.

Additional links for more information:

- NOAA site on oceanic hypoxic zones: http://oceanservice.noaa.gov/hazards/hypoxia/

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9.10: Carbon Dioxide and pH

Oxygen and carbon dioxide are involved in the same biological processes in the ocean, but in opposite ways; photosynthesis consumes CO_2 and produces O_2, while respiration and decomposition consume O_2 and produce CO_2. Therefore it should not be surprising that oceanic CO_2 profiles are essentially the opposite of dissolved oxygen profiles (Figure 9.10.1). At the surface, photosynthesis consumes CO_2 so CO_2 levels remain relatively low. In addition, organisms that utilize carbonate in their shells are common near the surface, further reducing the amount of dissolved CO_2.

In deeper water, CO_2 concentration increases as respiration exceeds photosynthesis, and decomposition of organic matter adds additional CO_2 to the water. As with oxygen, there is often more CO_2 at depth because cold bottom water holds more dissolved gases, and high pressures increase solubility. Deep water in the Pacific contains more CO_2 than the Atlantic as the Pacific water is older and has accumulated more CO_2 from the respiration of benthic organisms.

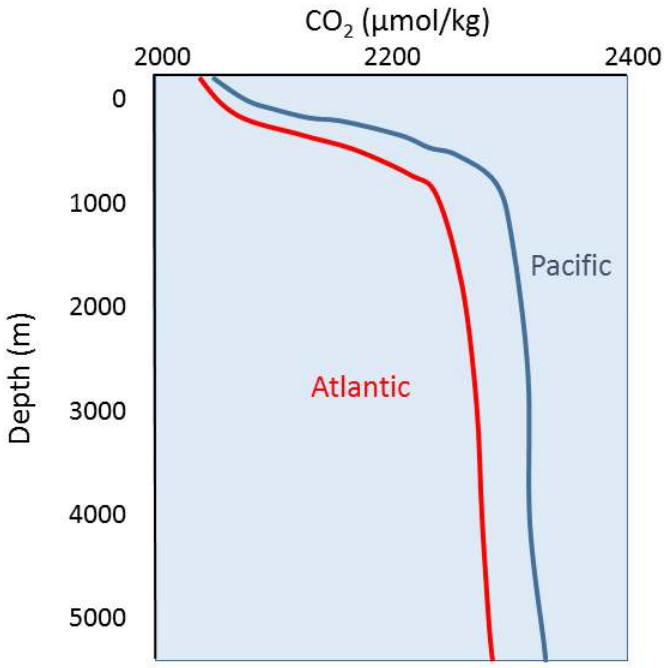


Figure 9.10.1 Representative carbon dioxide profiles for the Pacific and Atlantic oceans (PW).

But the behavior of carbon dioxide in the ocean is more complex than the figure above would suggest. When CO_2 gas dissolves in the ocean, it interacts with the water to produce a number of different compounds according to the reaction below:

$$CO_2 + H_2O \leftrightarrow H_2CO_3 \leftrightarrow H^+ + HCO_3^- \leftrightarrow 2H^+ + CO_3^{2-}$$

CO_2 reacts with water to produce carbonic acid (H_2CO_3), which then dissociates into bicarbonate (HCO_3^-) and hydrogen ions (H^+). The bicarbonate ions can further dissociate into carbonate (CO_3^{2-}) and additional hydrogen ions (Figure 9.10.2).

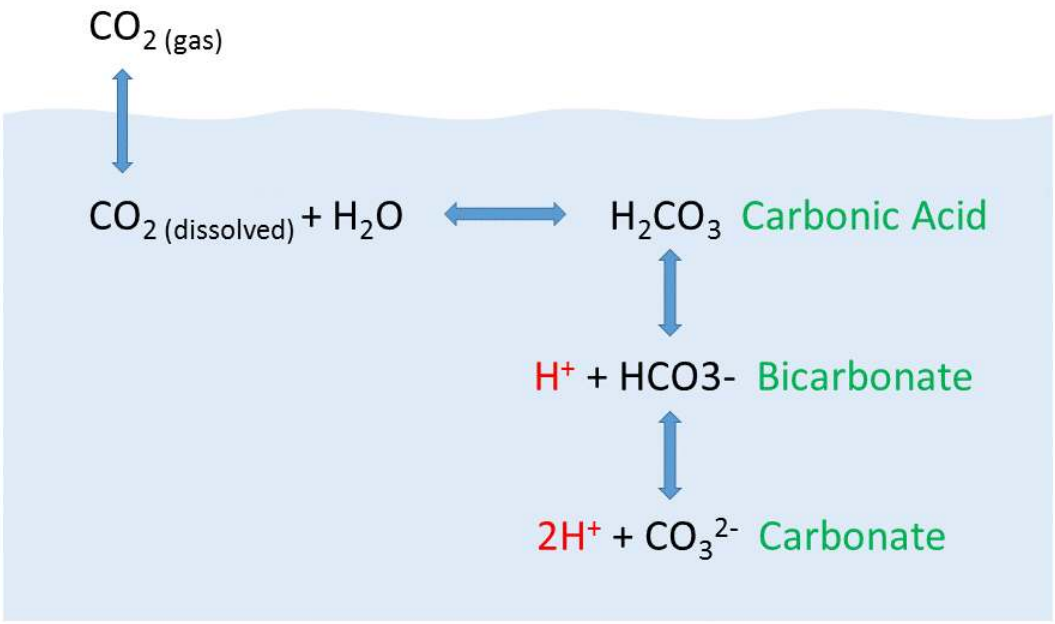


Figure 9.10.2 The fate of dissolved carbon dioxide in the oceans. Most of the carbon ends up in the form of bicarbonate (PW).

Most of the CO_2 dissolving or produced in the ocean is quickly converted to bicarbonate. Bicarbonate accounts for about 92% of the CO_2 dissolved in the ocean, and carbonate represents around 7%, so only about 1% remains as CO_2, and little gets absorbed back into the air. The rapid conversion of CO_2 into other forms prevents it from reaching equilibrium with the atmosphere, and in this way, water can hold 50-60 times as much CO_2 and its derivatives as the air.

CO_2 and pH

The equation above also illustrates carbon dioxide's role as a buffer, regulating the pH of the ocean. Recall that pH reflects the acidity or basicity of a solution. The pH scale runs from 0-14, with 0 indicating a very strong acid, and 14 representing highly basic conditions. A solution with a pH of 7 is considered neutral, as is the case for pure water. The pH value is calculated as the negative logarithm of the hydrogen ion concentration according to the equation:

$$pH = -\log_{10}[H^+]$$

Therefore, a high concentration of H^+ ions leads to a low pH and acidic condition, while a low H^+ concentration indicates a high pH and basic conditions. It should also be noted that pH is described on a logarithmic scale, so every one point change on the pH scale actually represents an order of magnitude (10 x) change in solution strength. So a pH of 6 is 10 times more acidic than a pH of 7, and a pH of 5 is 100 times (10 x 10) more acidic than a pH of 7.

Carbon dioxide and the other carbon compounds listed above play an important role in buffering the pH of the ocean. Currently, the average pH for the global ocean is about 8.1, meaning seawater is slightly basic. Because most of the inorganic carbon dissolved in the ocean exists in the form of bicarbonate, bicarbonate can respond to disturbances in pH by releasing or incorporating hydrogen ions into the various carbon compounds. If pH rises (low $[H^+]$), bicarbonate may dissociate into carbonate, and release more H^+ ions, thus lowering pH. Conversely, if pH gets too low (high $[H^+]$), bicarbonate and carbonate may incorporate some of those H^+ ions and produce bicarbonate, carbonic acid, or CO_2 to remove H^+ ions and raise the pH. By shuttling H^+ ions back and forth between the various compounds in this equation, the pH of the ocean is regulated and conditions remain favorable for life.

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9.11: Nitrogen and Nutrients

Nitrogen is the most abundant gas in the atmosphere, and like the other atmospheric gases it dissolves into the surface layers of the ocean. But most marine organisms cannot directly utilize dissolved nitrogen in the form in which it exists in air (N_2), so it must first be converted into other nitrogenous products. These conversions are undertaken by marine bacteria, who take the dissolved N_2 and convert it into ammonium (NH_4^+), nitrite (NO_2^{2-}) or nitrate (NO_3^-). These bacteria also take the ammonium excreted by marine organisms and the byproducts of decomposition and ultimately convert them into nitrate. Nitrate is the main nitrogenous compound utilized by primary producers in the ocean; it is a major **nutrient** required for photosynthesis. Note that in this context, a nutrient refers to a chemical needed to support photosynthesis and primary production. It does not refer to the nutritional needs of consumer organisms. Since nutrients are rapidly used in biological processes, they are non-conservative, and their concentrations vary regionally and seasonally. They are generally found in low quantities in the oceans because they are quickly consumed by primary producers.

Since nitrate is one of the most important nutrients, for now we will focus only on nitrate as we discuss general nutrient patterns in the ocean. Other important nutrients, such as phosphate and silica, show similar patterns to nitrate, and will be discussed in the section on primary production (chapter 7).

A representative nutrient (nitrate) profile is shown below (Figure 9.11.1). Nutrient concentrations are low at the surface, because that is where the primary producers are located; the nutrients are rapidly consumed and they do not have the chance to accumulate. Nutrient levels increase at depth, as they are no longer being consumed by producers, and they are being regenerated through the decomposition of organic material by bacteria.

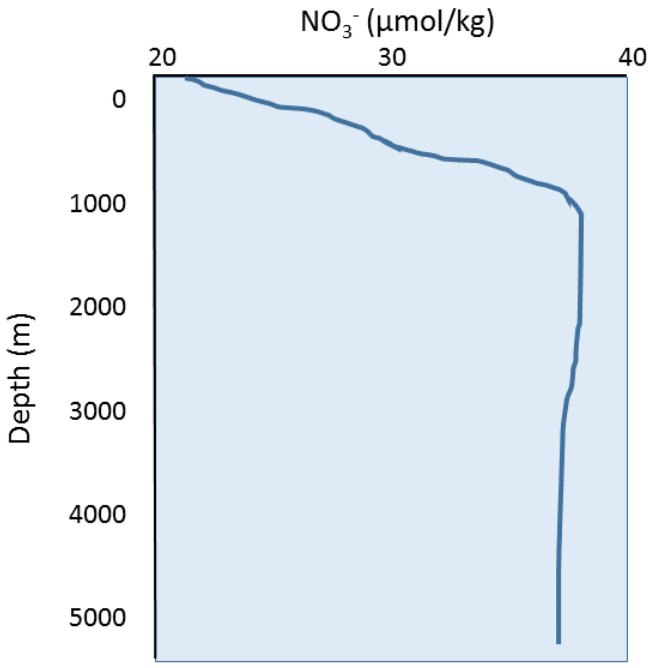


Figure 9.11.1 Representative nutrient (nitrate) profile for the open ocean (PW).

Comparisons of nutrient and dissolved oxygen profiles between the Pacific and Atlantic Oceans reveal some interesting differences (Figure 9.11.2). In general, the Atlantic has more dissolved oxygen but lower nutrient concentrations than the Pacific. Water masses form in the North Atlantic that are very cold and dense, so the water sinks to the bottom. This water will then spend the next thousand years or more moving along the seafloor from the Atlantic, to the Indian, and finally into the Pacific Ocean (see section 9.8). This water is initially oxygen-rich surface water, and as it sinks it brings oxygen to the deep seafloor. As the bottom water moves across the ocean basins, oxygen is removed through respiration and decomposition, and by the time it arrives in the Pacific it has been depleted of much of its oxygen. At the same time, decomposition of sinking organic matter adds nutrients to the deep

water as it moves through the oceans, so nutrients accumulate and the Pacific water becomes nutrient-rich. Comparison of the ratios of oxygen to nutrients in the deep water can therefore provide an indication of the age of the water, i.e. how much time has passed since it initially sank from the surface in the North Atlantic. Water with a high oxygen and low nutrient content is relatively young, while older water will have less oxygen but higher nutrient concentrations.

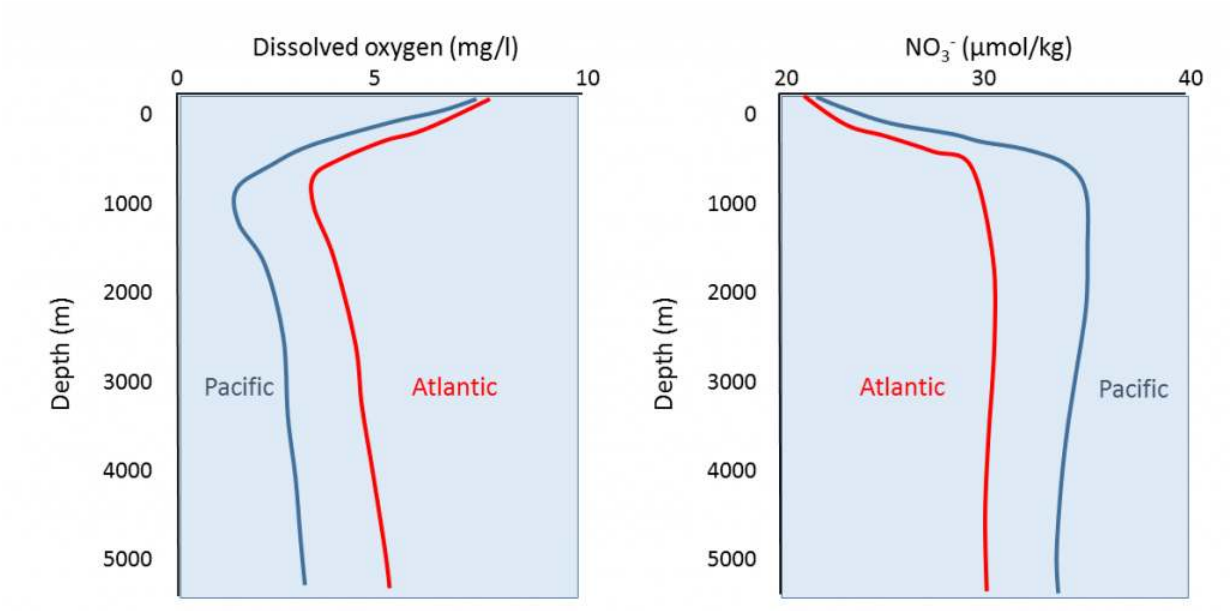


Figure 9.11.2 Dissolved oxygen (left) and nutrient (right) profiles for the Pacific and Atlantic Oceans. As water circulates from the Atlantic to the Pacific, oxygen is consumed while nutrients accumulate (PW).

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CHAPTER OVERVIEW

10: Atmospheric Circulation

10.1: How is Earth Heated

10.2: Winds and the Coriolis Effect

10.3: Winds and Climate

10.4: Hurricanes

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10.1: How is Earth Heated

Differential Heating of Earth's Surface

If the Earth was a flat surface facing the sun, every part of that surface would receive the same amount of incoming solar radiation. However, because the Earth is a sphere, sunlight is not equally distributed over the Earth's surface, so different regions of Earth will be heated to different degrees. This differential heating of Earth's surface occurs for a number of reasons. First, because of the curvature of Earth, sunlight only falls perpendicularly to the surface at the center of the sphere (equatorial regions). At any other point on Earth, the angle between the surface and the incoming solar radiation is less than 90°. Because of this, the same amount of incoming solar radiation will be concentrated in a smaller area at the equator, but will be spread over a much larger area at the poles (Figure 10.1.3). Thus the tropics receive more intense sunlight and a greater amount of heating per unit of area than the polar regions.

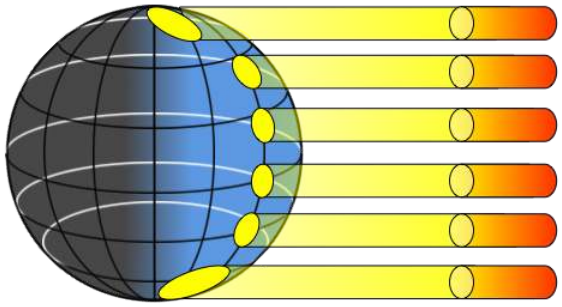


Figure 10.1.3 Because of the curvature of the Earth, the same amount of sunlight will be spread out over a larger area at the poles compared to the equator. The equator therefore receives more intense sunlight, and a greater amount of heat per unit of area (By Thebiologyprimer (Own work) [CC0], via Wikimedia Commons).

The angle at which sunlight strikes the Earth contributes to differential heating of the surface in an additional way. At the poles, because of the angle at which the solar energy strikes the surface, more of the light will glance off of the surface and the atmosphere and be reflected back into space. At the equator, the direct angle with which light reaches the surface results in more of the energy being absorbed rather than reflected. Finally, the poles reflect more solar energy than other parts of the Earth because the poles have a higher **albedo**. The albedo refers to reflectivity of a surface. Lighter surfaces are more reflective than darker surfaces (which absorb more energy), and therefore have a higher albedo. At the poles, the ice, snow and cloud cover create a much higher albedo, and the poles reflect more and absorb less solar energy than the lower latitudes. Through all of these mechanisms, the poles absorb much less solar radiation than equatorial regions, which is why the poles are cold and the tropics are very warm.

But there is an interesting twist to this global distribution of heat. The tropical regions actually receive more radiant heat than they emit, and the poles emit more heat than they receive (Figure 10.1.4). We should therefore expect that the tropics will be getting continually warmer, while the poles become increasingly cold. Yet this is not the case; so what is happening? Rather than the heat remaining isolated near the equator, about 20% of the heat from the tropics is transported to the poles before it is emitted. This large scale transport of energy moderates the climates at both extremes. The mechanisms for this heat transfer are ocean and atmospheric circulation, the topic of the next section.

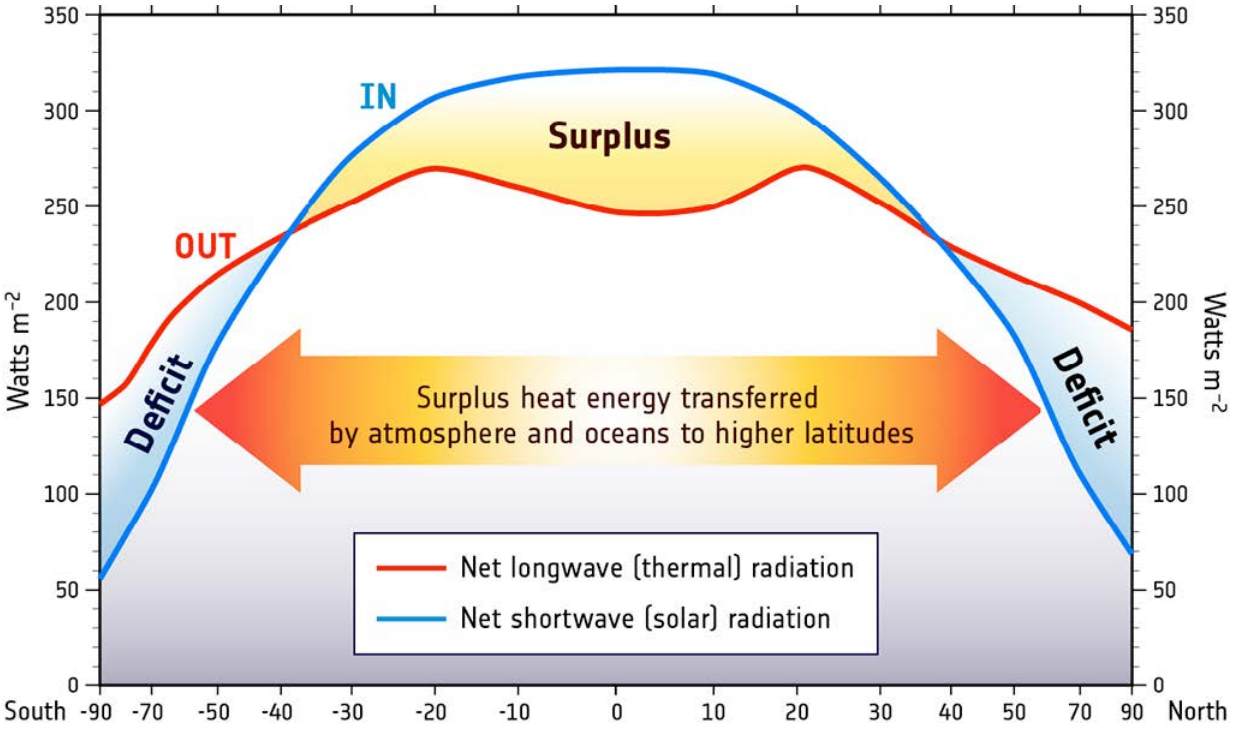


Figure 10.1.4 The balance between heat gain and heat loss as a function of latitude. Excess heat received near the equator is transferred towards the poles (National Oceanography Centre (NOC). Creative Commons 3.0 unported license).

The idea of differential heating of the Earth's surface is fundamental to understanding a wide range of oceanographic and atmospheric processes. This differential heating leads to atmospheric convection, which creates winds, which blow over the water and create waves and surface currents, and these currents influence nutrient distribution, which promotes primary production, which then supports the rest of the ocean ecosystem. So there's a lot riding on the simple fact that more light reaches the tropics than the poles!

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10.2: Winds and the Coriolis Effect

Differential heating of the Earth's surface results in equatorial regions receiving more heat than the poles. As air is warmed at the equator it becomes less dense and rises, while at the poles the cold air is denser and sinks. If the Earth was non-rotating, the warm air rising at the equator would reach the upper atmosphere and begin moving horizontally towards the poles. As the air reached the poles it would cool and sink, and would move over the surface of Earth back towards the equator. This would result in one large atmospheric convection cell in each hemisphere (Figure 10.2.1), with air rising at the equator and sinking at the poles, and the movement of air over the Earth's surface creating the winds. On this non-rotating Earth, the prevailing winds would thus blow from the poles towards the equator in both hemispheres (Figure 10.2.1).

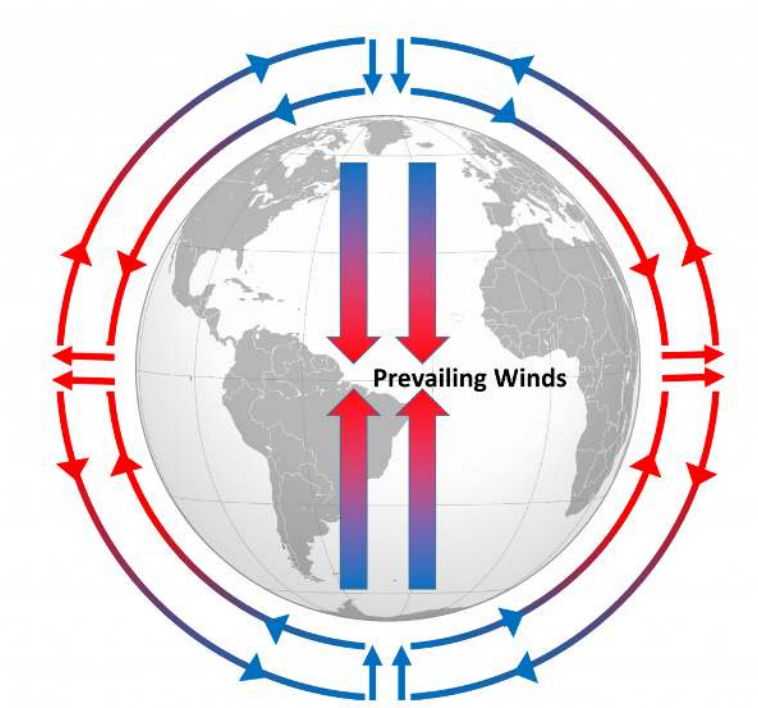


Figure 10.2.1 Hypothetical atmospheric convection cells on a non-rotating Earth. Air rises at the equator and sinks at the poles, creating a single convection cell in each hemisphere. The prevailing winds moving over the Earth's surface blow from the poles towards the equator in both hemispheres (Modified by PW from globe image by Location_of_Cape_Verde_in_the_globe.svg: Eddo derivative work: Luan fala! [CC BY-SA 3.0], via Wikimedia Commons).

The non-rotating situation in Figure 10.2.1 is of course only hypothetical, and in reality the Earth's rotation makes this atmospheric circulation a bit more complex. The paths of the winds on a rotating Earth are deflected by the **Coriolis Effect**. The Coriolis Effect is a result of the fact that different latitudes on Earth rotate at different speeds. This is because every point on Earth must make a complete rotation in 24 hours, but some points must travel farther, and therefore faster, to complete the rotation in the same amount of time. In 24 hours a point on the equator must complete a rotation distance equal to the circumference of the Earth, which is about 40,000 km. A point right on the poles covers no distance in that time; it just turns in a circle. So the speed of rotation at the equator is about 1600 km/hr, while at the poles the speed is 0 km/hr. Latitudes in between rotate at intermediate speeds; approximately 1400 km/hr at 30° and 800 km/hr at 60°. As objects move over the surface of the Earth they encounter regions of varying speed, which causes their path to be deflected by the Coriolis Effect.

To explain the Coriolis Effect, imagine a cannon positioned at the equator and facing north. Even though the cannon appears stationary to someone on Earth, it is in fact moving east at about 1600 km/hr due to Earth's rotation. When the cannon fires the projectile travels north towards its target; but it also continues to move to the east at 1600 km/hr, the speed it had while it was still in the cannon. As the shell moves over higher latitudes, its momentum carries it eastward faster than the speed at which the ground beneath it is rotating. For example, by 30° latitude the shell is moving east at 1600 km/hr while the ground is moving east at only 1400 km/hr. Therefore, the shell gets "ahead" of its target, and will land to the east of its intended destination. From the point of view of the cannon, the path of the projectile appears to have been deflected to the right (red arrow, Figure 10.2.2). Similarly, a

cannon located at 60° and facing the equator will be moving east at 800 km/hr. When its shell is fired towards the equator, the shell will be moving east at 800 km/hr, but as it approaches the equator it will be moving over land that is traveling east *faster* than the projectile. So the projectile gets "behind" its target, and will land to the west of its destination. But from the point of view of the cannon facing the equator, the path of the shell still appears to have been deflected to the right (green arrow, Figure 10.2.2). Therefore, in the Northern Hemisphere, the apparent Coriolis deflection will *always* be to the **right**.

In the Southern Hemisphere the situation is reversed (Figure 10.2.2). Objects moving towards the equator from the south pole are moving from low speed to high speed, so are left behind and their path is deflected to the left. Movement from the equator towards the south pole also leads to deflection to the left. In the Southern Hemisphere, the Coriolis deflection is always to the **left** from the point of origin.

The magnitude of the Coriolis deflection is related to the difference in rotation speed between the start and end points. Between the poles and 60° latitude, the difference in rotation speed is 800 km/hr. Between the equator and 30° latitude, the difference is only 200 km/hr (Figure 10.2.2). Therefore the strength of the Coriolis Effect is stronger near the poles, and weaker at the equator.

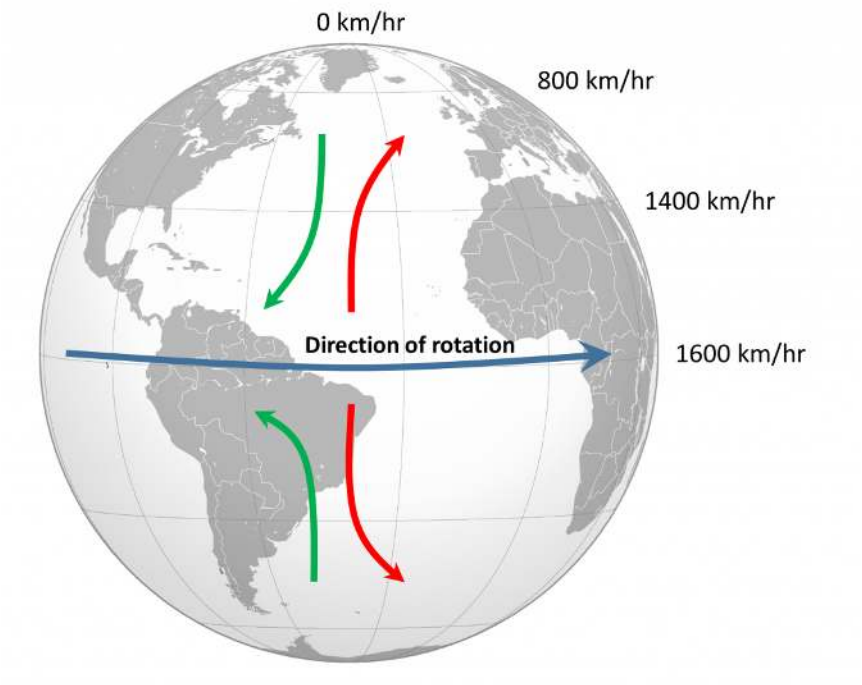


Figure 10.2.2 The Coriolis Effect. Objects moving from the equator towards the poles (red arrows) move into a region of slower rotational speed and their paths are deflected "ahead" of their point of origin. Movement from high latitudes to low latitudes (green arrows) goes from a region of low speed to a region of higher rotation speed, and there is deflection "behind" their point of origin. In the Northern Hemisphere this deflection is always to the right from the point of origin, and in the Southern Hemisphere the deflection is always to the left (Modified by PW from globe image by Location_of_Cape_Verde_in_the_globe.svg: Eddo derivative work: Luan fala! [CC BY-SA 3.0], via Wikimedia Commons).

Because of the rotation of the Earth and the Coriolis Effect, rather than a single atmospheric convection cell in each hemisphere, there are three major cells per hemisphere. Warm air rising at the equator cools as it moves through the upper atmosphere, and it descends at around 30° latitude. The convection cells created by rising air at the equator and sinking air at 30° are referred to as **Hadley Cells**, of which there is one in each hemisphere. The cold air that descends at the poles moves over the Earth's surface towards the equator, and by about 60° latitude it begins to rise, creating a **Polar Cell** between 60° and 90°. Between 30° and 60° lie the **Ferrel Cells**, composed of sinking air at 30° and rising air at 60° (Figure 10.2.3). With three convection cells in each hemisphere that rotate in alternate directions, the surface winds no longer always blow from the poles towards the equator as in the non-rotating Earth in Figure 10.2.1. Instead, surface winds in both hemispheres blow towards the equator between 90° and 60° latitude, and between 0° and 30° latitude. Between 30° and 60° latitude, the surface winds blow towards the poles (Figure 10.2.3).

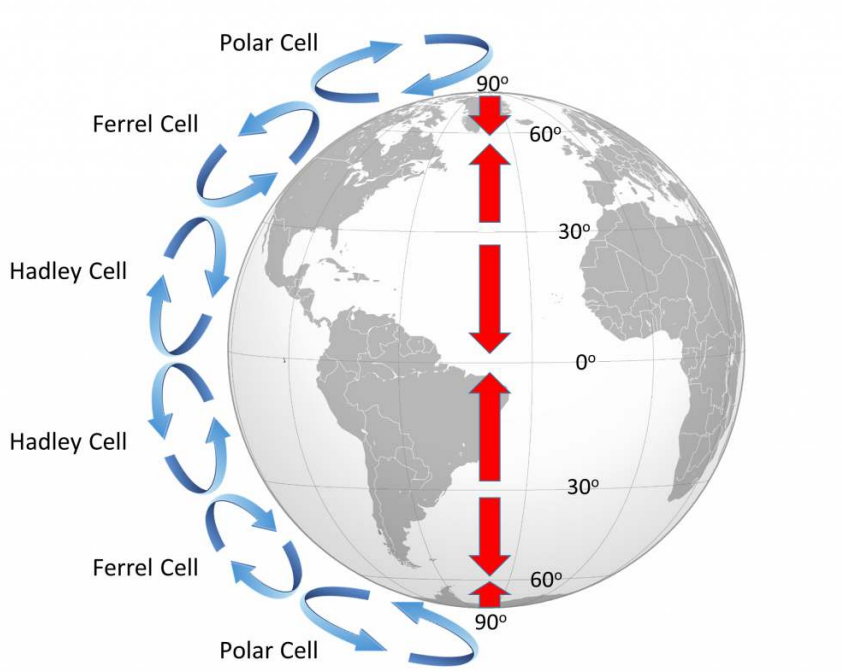


Figure 10.2.3 On a rotating Earth, there are three atmospheric convection cells in each hemisphere, leading to alternating bands of surface winds (red arrows) (Modified by PW from globe image by Location_of_Cape_Verde_in_the_globe.svg: Eddo derivative work: Luan fala! [CC BY-SA 3.0], via Wikimedia Commons).

The surface winds created by the atmospheric convection cells are also influenced by the Coriolis Effect as they change latitudes. The Coriolis Effect deflects the path of the winds to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Adding this deflection leads to the pattern of prevailing winds illustrated in Figure 10.2.4. Between the equator and 30° latitude are the **trade winds**; the northeast trade winds in the Northern Hemisphere and the southeast trade winds in the Southern Hemisphere (note that winds are named based on the direction from which they originate, not where they are going). The **westerlies** are the dominant winds between 30° and 60° in both hemispheres, and the **polar easterlies** are found between 60° and the poles.

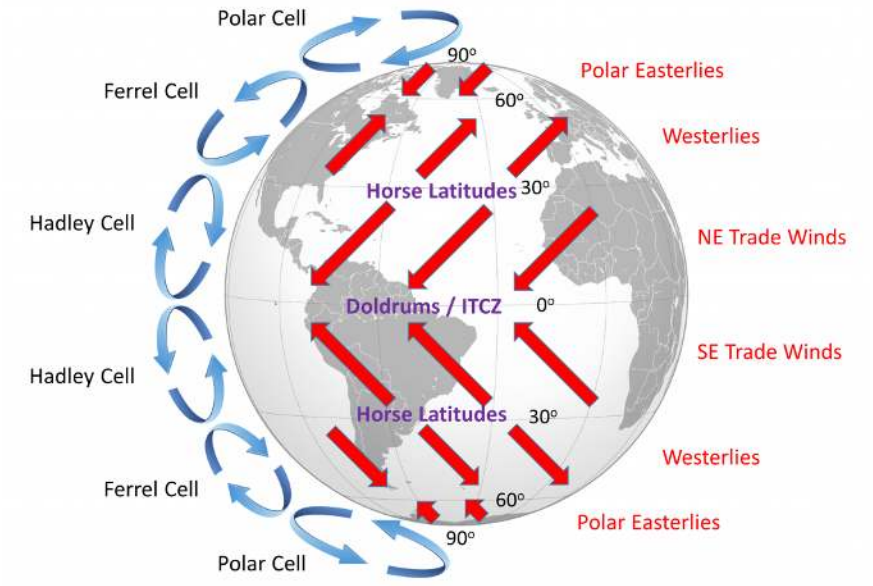


Figure 10.2.4 The prevailing wind patterns of Earth (Modified by PW from globe image by Location_of_Cape_Verde_in_the_globe.svg: Eddo derivative work: Luan fala! [CC BY-SA 3.0], via Wikimedia Commons).

In between these wind bands lie regions of high and low pressure. High pressure zones occur where air is descending, while low pressure zones indicate rising air. Along the equator the rising air creates a low pressure region called the **doldrums**, or the **Intertropical Convergence Zone** (ITCZ)(convergence zone because this is where the trade winds converge). At 30° latitude there are high pressure zones of descending air known as the **horse latitudes**, or the subtropical highs. Finally, at 60° lies another low pressure region called the **polar front**. It should be noted that these high and low pressure zones are not fixed in place; their latitude fluctuates depending on the season, and these fluctuations have important implications for regional climates.

Doldrums? Horse latitudes? Trade winds?

These may seem like some odd names for these atmospheric phenomena, but many of them can be traced back to maritime traditions and lore.

The **doldrums** refer to regions of low pressure around the equator. In these areas, air is rising rather than moving horizontally, so these regions commonly encounter very light winds. The lack of wind could leave sailing ships becalmed for days or weeks at a time, which was not good for the morale of the ship's crew.

Like the doldrums the **horse latitudes** are also areas with light winds, this time due to descending air, which could leave ships becalmed. One explanation for the term "horse latitudes" is that when these ships became stranded they ran the risk of running out of food or water. To conserve these resources, sailors would throw their dead or dying horses overboard, hence the "horse latitudes." Another explanation is that many sailors received part of their pay before a voyage, and often spent it before departing. This meant that they would spend the first part of the voyage working without pay and in debt, a period called the "dead horse" time, which might last for a few months. When they started earning their pay once again, they had a "dead horse" ceremony and threw a pretend horse overboard. The timing of this ceremony often coincided with reaching the horse latitudes, leading to the association of the ceremony with the location. A third explanation is that a ship was referred to as "horsed" when winds were weak and the ship instead had to rely on ocean currents to move them. This could be a common occurrence in the high pressure zones around 30° latitude, so they were referred to as the horse latitudes.

The term **trade winds** may have originally derived from the terms for "track" or "path", but the term may have become more common during European exploration and commercialization of the New World. Mariners sailing from Europe to the New World could sail south until they reached the trade winds, which would then propel their ships across the Atlantic to the Caribbean. To return to Europe, ships could sail to the northeast until they entered the westerlies, which would then steer them back to Europe.

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10.3: Winds and Climate

In the previous section we learned that rising air creates low pressure systems, and sinking air creates high pressure. In addition to their role in creating the surface winds, these high and low pressure systems also influence other climatic phenomena. Along the equator air is rising as it is warmed by solar radiation. Warm air contains more water vapor than cold air, which is why we experience humidity during the summer and not during the winter. The water content of air roughly doubles with every 10° C increase in temperature. So the air rising at the equator is warm and full of water vapor; as it rises into the upper atmosphere it cools, and the cool air can no longer hold as much water vapor, so the water condenses and forms rain. Therefore, low pressure systems are associated with precipitation, and we see wet habitats like tropical rainforests near the equator (Figure 10.3.1).

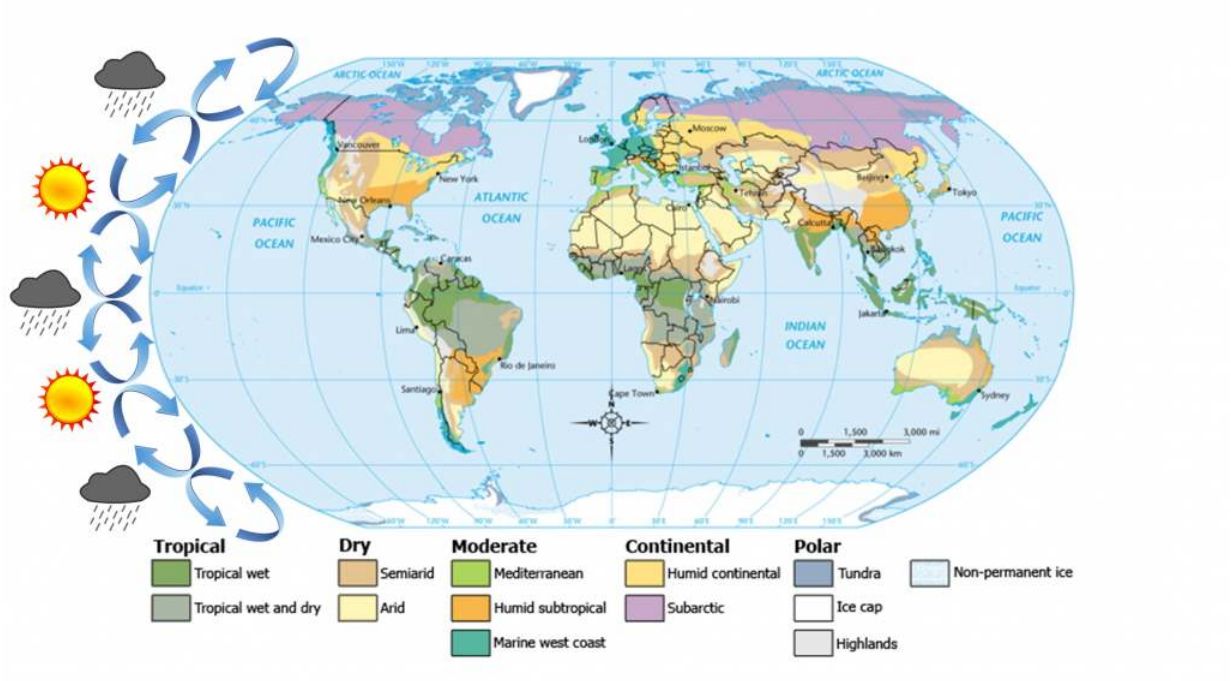


Figure 10.3.1 Major global climatic regions in relation to atmospheric convection cells. Rising air and low pressure creates rain and wet environments at 0° and 60° latitudes, while high pressure, sinking air creates drier conditions at 30° and 90° latitudes. (Modified by PW; Map by Waitak at en.wikipedia Later version(s) were uploaded by Splette at en.wikipedia; Sun by Inductiveload (Own work Based on File:Nuvola_apps_kweather.svg); Raincloud by Calusarul (Own work); all [CC BY-SA 3.0], via Wikimedia Commons).

After rising and producing rain near the equator, the air masses move towards 30° latitude and sink back towards Earth as part of the Hadley convection cells. This air has lost most of its moisture after producing the equatorial rains, so the sinking air is dry, resulting in arid climates near 30° latitude in both hemispheres. Many of the major desert regions on Earth are located near 30° latitude, including much of Australia, the Middle East, and the Sahara Desert of Africa (Figure 10.3.1). The air also becomes compressed and heats up as it sinks, absorbing any moisture from the clouds and creating clear skies. Thus high pressure systems are associated with dry weather and clear skies. This cycle of high and low pressure regions continues with the Ferrel and Polar convection cells, leading to rain and the boreal forests at 60° latitude in the Northern Hemisphere (there are no corresponding large land masses at these latitudes in the Southern Hemisphere). At the poles, descending, dry air produces little precipitation, leading to the polar desert climate.

The elevation of the land also plays a role in precipitation and climactic characteristics. As moist air moves over land and encounters mountains it rises, expands, and cools because of the declining pressure and temperature. The cool air holds less water vapor, so condensation occurs and rain falls on the windward side of the mountains. As the air passes over the mountains to the leeward side, it is now dry air, and as it sinks the pressure increases, it heats back up, any moisture revaporizes, and it creates dry, deserts regions behind the mountains (Figure 10.3.2). This phenomenon is referred to as a **rain shadow**, and can be found in areas such as the Tibetan Plateau and Gobi Desert behind the Himalayas, Death Valley behind the Sierra Nevada mountains, and the dry San Joaquin Valley in California.

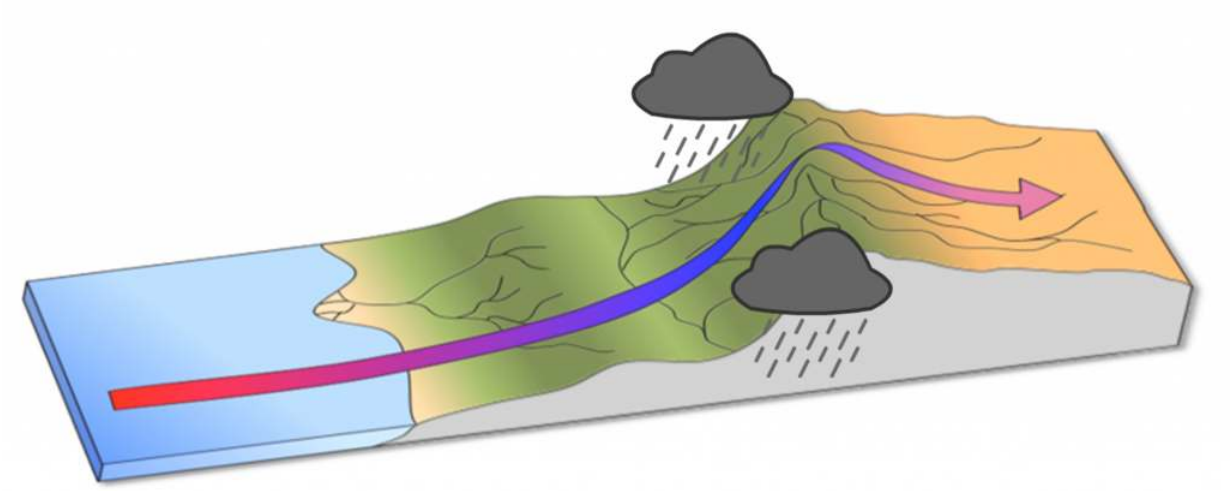


Figure 10.3.2 A rain shadow. Air rising over mountains cools and condenses and forms rain, leaving dry descending air and arid conditions on the other side of the mountain (Modified by PW from Thebiologyprimer, Public domain via Wikimedia Commons).

Rising and falling air are also responsible for more localized, short-term wind patterns in coastal areas. Due to the high heat capacity of water, land heats up and cools down about five times faster than water. During the day the sun heats up the land faster than it heats the water, setting up a convection cell of warmer rising air over the land and sinking cooler air over the water. This creates winds blowing from the water towards the land during the day and early evening; a **sea breeze** (Figure 10.3.3). The opposite occurs at night, when the land cools more quickly than the ocean. Now the ocean is warmer than the land, so air rises over the water and sinks over the land, creating a convection cell where winds blow from land towards the water. This is a **land breeze**, which blows at night and into the early morning (Figure 10.3.3).

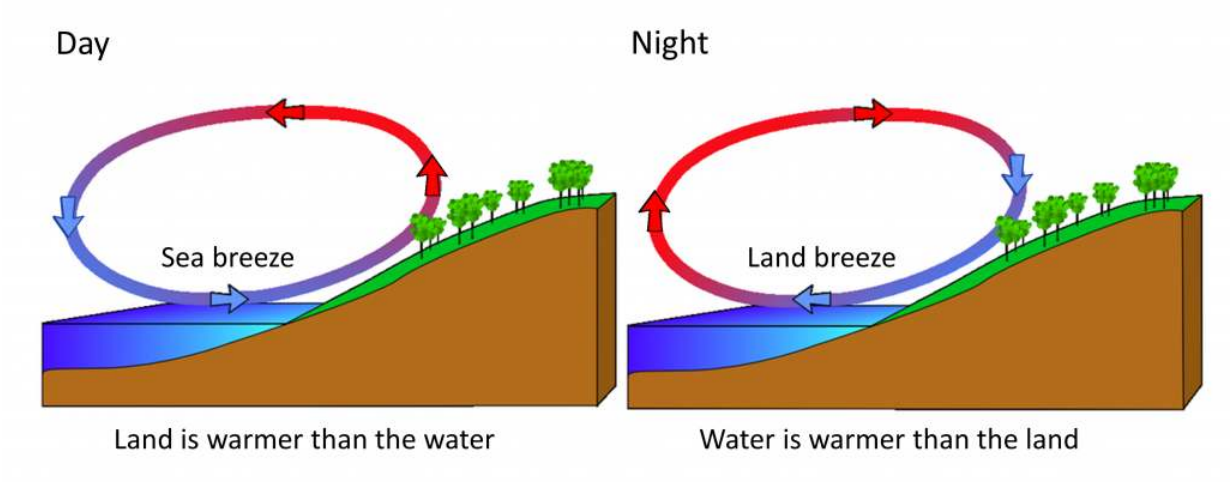


Figure 10.3.3 Land heats and cools faster than the ocean, so during the day the land is warmer than the water leading to rising air over land and a sea breeze. At night, the ocean is warmer than the land, creating a land breeze (Modified by PW derivative work: Ingwik (Diagrama de formacion de la brisa-breeze.png) [CC-BY-SA-3.0 or GFDL (http://www.gnu.org/copyleft/fdl.html)], via Wikimedia Commons).

The same phenomenon leads to seasonal climatic changes in many areas. During the winter the lower pressure is over the warmer ocean, and the high pressure is over the colder land, so winds blow from land to sea. In summer the land is warmer than the ocean, causing low pressure over the land and winds to blow from the ocean towards the land. The winds blowing from the ocean contain a lot of water vapor, and as the moist air passes over land and rises, it cools and condenses causing seasonal rains, such as the summer **monsoons** of southeast Asia (Figure 10.3.4).

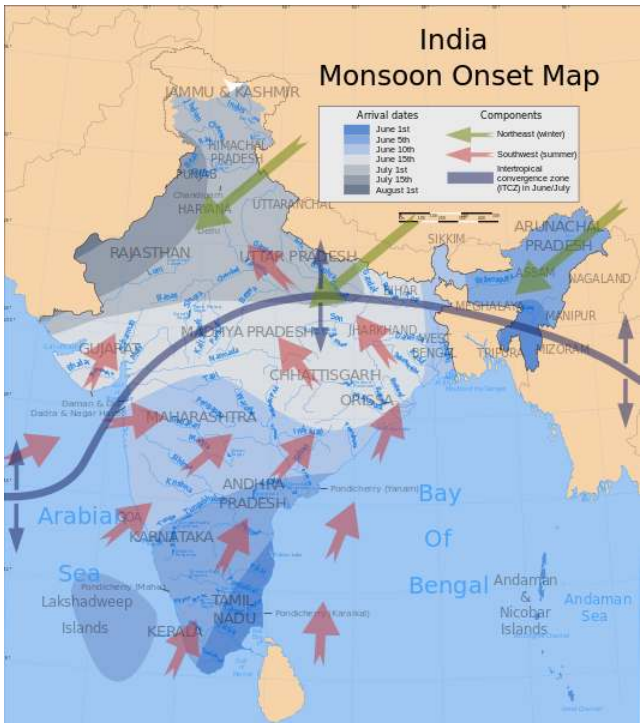


Figure 10.3.4 Seasonal wind patterns and monsoons over India. In summer, moist air from the ocean moves over the continent and rises, creating rain and the summer monsoons (pink arrows). In winter, winds are blowing from land to the sea, leading to the dry season (green arrows) (By Saravask, based on work by Planemad and Nichalp [CC BY-SA 3.0], via Wikimedia Commons).

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10.4: Hurricanes

The most dramatic examples of low pressure systems leading to storms and rain are hurricanes, cyclones, and typhoons. All three of these terms describe the same atmospheric processes and the same types of storms; it's just that different terminology is used in different parts of the world. In the Atlantic and Northeast Pacific, the storms are called hurricanes, in the Indian and South Pacific Oceans they are referred to as cyclones, and they are called typhoons in the Northwest Pacific.

Hurricanes begin as low pressure systems formed over warm, tropical water. They only form in tropical regions because they need the heat from the warm water to fuel the storm. The warm, moist air rises, cools, and condenses, forming rain, and the condensation releases more latent heat into the atmosphere. This heat causes even more air to rise and condense, further fueling the storm.

As the air rises towards the center of the storm, more warm tropical air rushes in to replace it, causing very strong winds. But the air does not move directly towards the center of the storm. Because of the large size of hurricanes, the air rushing towards the center will be deflected by the Coriolis Effect, causing the entire storm to rotate. In the Northern Hemisphere that deflection is to the right, causing Northern Hemisphere hurricanes to rotate counterclockwise. In the Southern Hemisphere, the winds are deflected to the left, leading to a clockwise rotation (Figure 10.4.1).

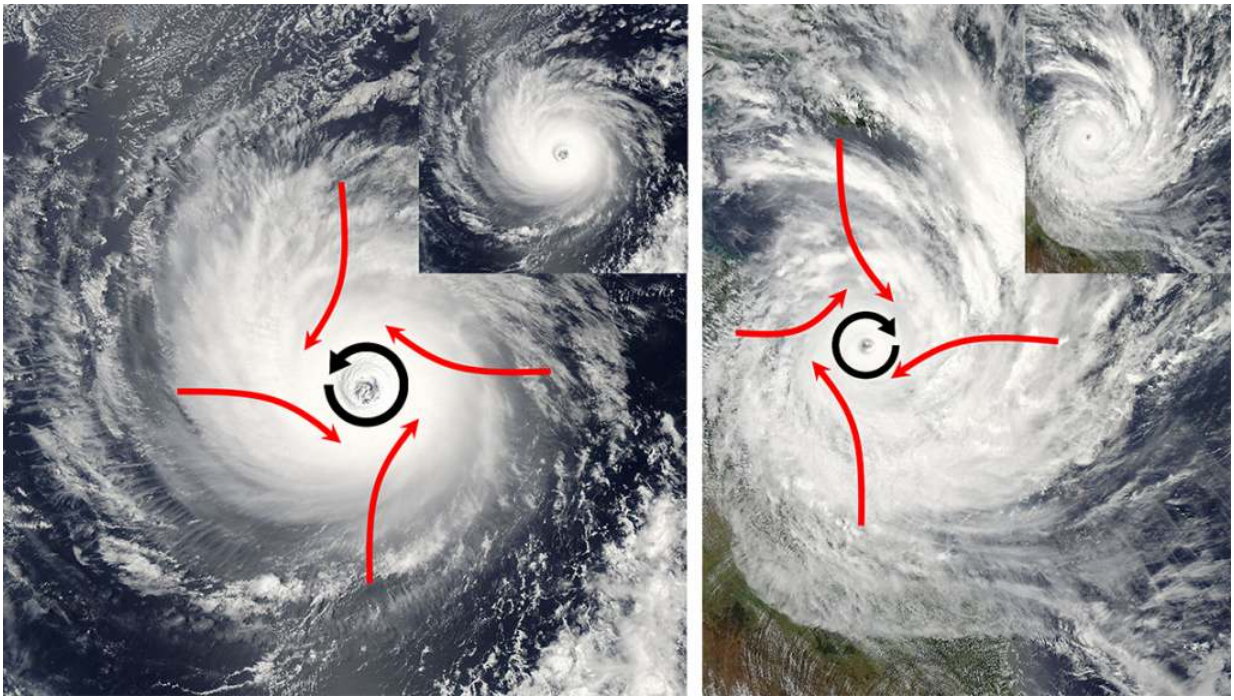


Figure 10.4.1 Hurricanes in the Northern Hemisphere rotate counterclockwise (left, Hurricane Daniel, 2006), as air rushes towards the center and is deflected to the right by the Coriolis Effect. In the Southern Hemisphere, hurricanes rotate clockwise as the Coriolis deflection is to the left (right, Cyclone Yasi, 2011) (Modified by PW; Daniel image by NASA image courtesy Jeff Schmaltz, MODIS Land Rapid Response Team at NASA GSFC; Yasi image by NASA; MODIS [Public domain], via Wikimedia Commons).

The violent winds characteristic of hurricanes are the result of the spiraling air that is moving towards the center of the storm, and once its winds exceed 74 mph the storm officially becomes a hurricane. At the very center of the hurricane, the pressure is so low that cool, dry air from the upper atmosphere get sucked downwards, leading to a central region of calm, clear skies; the hurricane's eye (Figure 10.4.2).

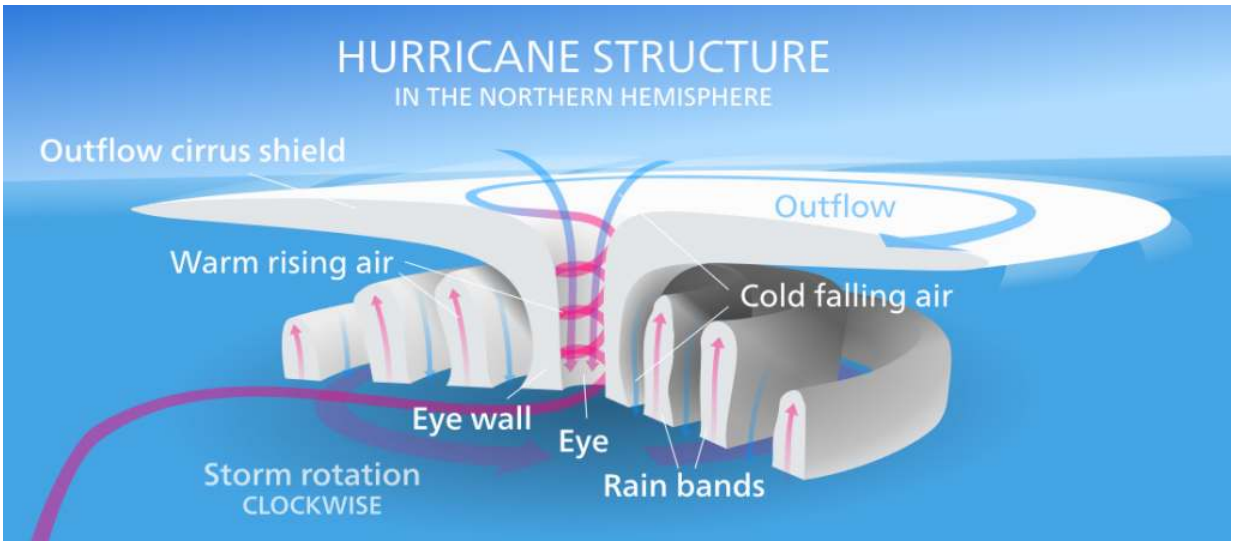


Figure 10.4.2 Hurricane structure. Air rising in the center of the hurricane is replaced by warm air moving inwards, and the Coriolis Effect deflects the winds, causing the storm to rotate. In the eye, the extreme low pressure causes cool, dry air to sink, creating calm, clear conditions within the eye (By Kelvinsong (Own work) [CC BY 3.0], via Wikimedia Commons).

Hurricanes in the North Atlantic form as tropical storms over the warm water off of the African coast, and are moved east to west by the trade winds (Figure 10.4.3). As the storms move west over the tropical ocean, their energy increases until they reach hurricane status. As they approach the Caribbean, the Coriolis Effect deflects their path to the right, causing them to move towards the north (Figure 10.4.3). Eventually hurricanes might make landfall, causing extensive damage to coastal areas through the high winds, rain, and flooding. However, hurricanes often die out fairly soon after reaching land. When a storm moves over land it becomes cut off from the warm moist ocean air that has sustained it. Without that fuel source, the storm loses power and begins to dissipate.

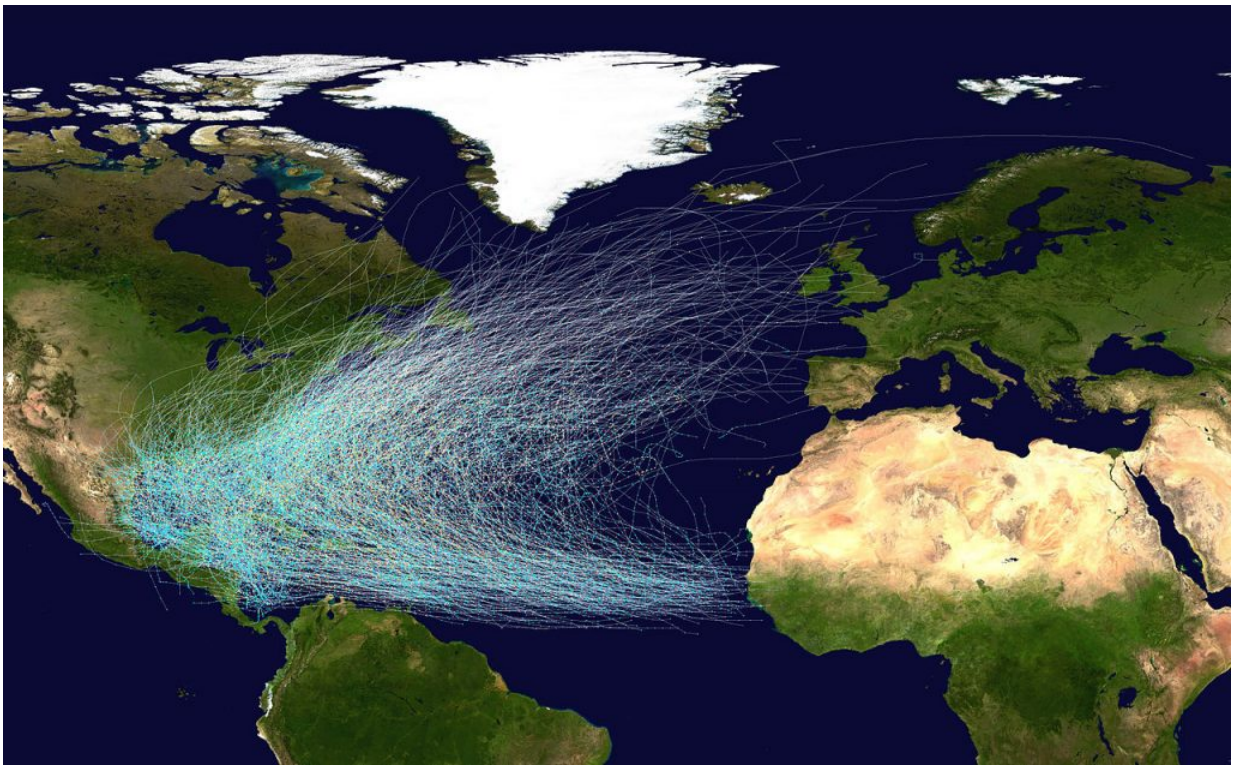


Figure 10.4.3 Hurricane tracks in the North Atlantic from 1980-2005. Hurricanes begin near the coast of Africa and are blown westward by the trade winds. Coriolis deflection causes them to take a northward path as they approach the Caribbean (By Nilfanion [Public domain], via Wikimedia Commons).

A similar pattern occurs in the Pacific and in the Southern Hemisphere. The trade winds move the storms from east to west, and they are deflected as they approach the coasts; to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (Figure 10.4.4).

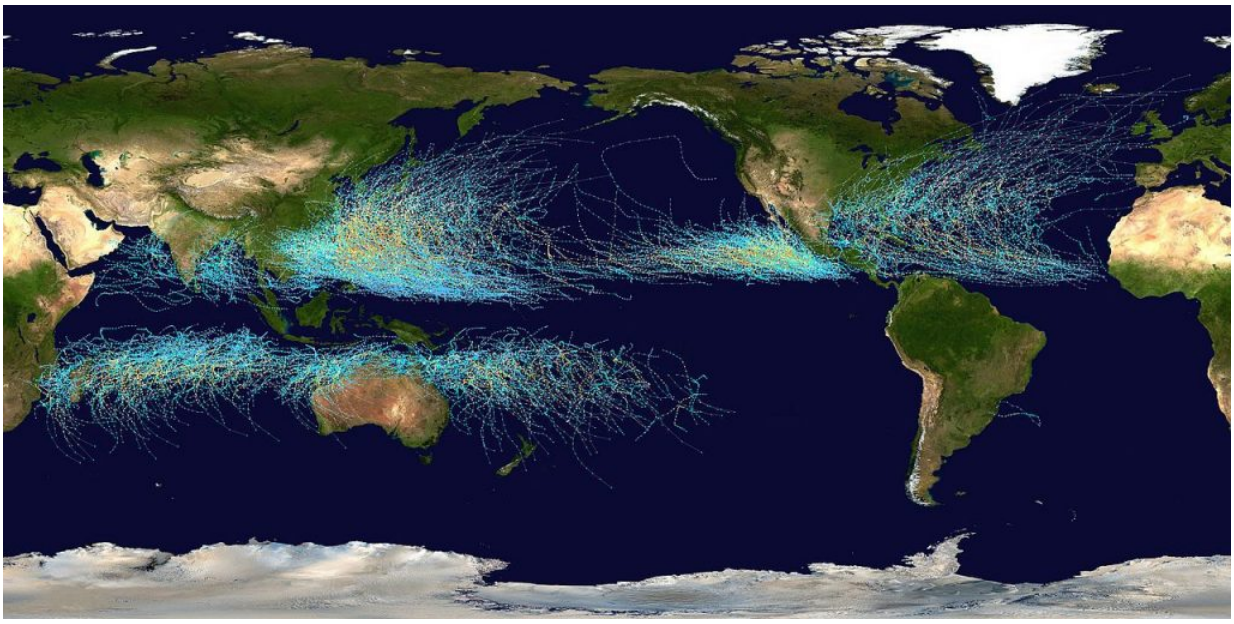


Figure 10.4.4 Global hurricane/cyclone tracks 1985-2005. Hurricanes move west via the trade winds, and their paths are deflected away from the equator in both hemispheres as they approach the continents (Background image: NASA this version: Nilfanion [Public domain], via Wikimedia Commons).

While the very high winds and intense rain of hurricanes can cause significant damage, in many cases it is the **storm surge** that leads to the most death and destruction. The storm surge is a "hill" of water that forms on the ocean surface below a hurricane. The surge is the result of two processes; a small hill is produced due to the extreme low pressure in the eye of a hurricane, which pulls water upwards towards the eye, creating a pressure surge. A larger surge is produced by the winds blowing and piling up water in the direction the storm is traveling (Figure 10.4.5). As the hurricane makes landfall, the effect of the storm surge is equivalent to a very large and sudden rise in sea level as the surge moves over the land, causing extensive flooding.

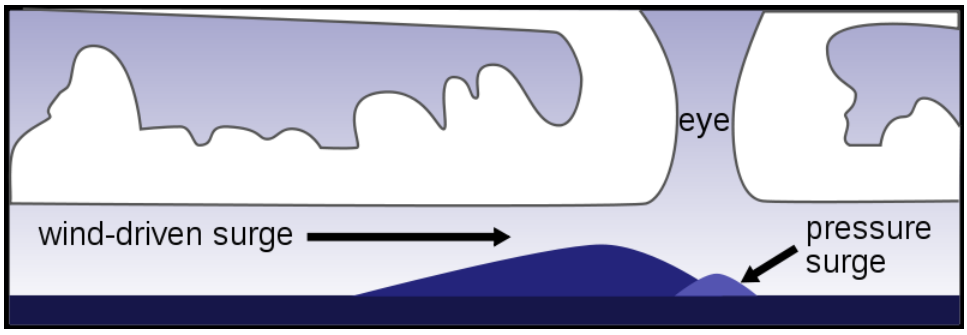


Figure 10.4.5 Storm surges are created by hurricanes and move with the storm, causing a rapid rise in sea level when they reach shore. Pressure surges are due to the low pressure within the hurricane's eye, wind-driven surges are a product of high winds piling up water (By Howcheng. Original graphic by Robert Simmon, NASA GSFC. [Public domain], via Wikimedia Commons).

In 1970 the Bhola Cyclone struck Bangladesh with a 40 ft. storm surge, leading to the death of about 500,000 people, the deadliest hurricane in history. The east coast of the United States was hit by the New England Hurricane of 1938, which had a 16 ft. storm surge and left almost 700 people dead.

Preventing Storm Surge Damage

In response to the hurricane-related tragedies like those listed above, may cities have built hurricane barriers designed to reduce the flooding and damage associated with storm surges. Downtown Providence, Rhode Island, USA, was submerged under 13 feet of water during the Great New England hurricane of 1938, and was flooded again following Hurricane Carol in 1954. In the 1960s the

Fox Point Hurricane Barrier was constructed at the mouth of the Providence River. It consists of a high wall with three “doors” that are left open under normal conditions, but can be closed during a hurricane to prevent a storm surge of up to 20.5 feet from inundating the city (Figure 10.4.6, left). A related concept is seen in the storm surge barrier on the Hollandse IJssel river in the Netherlands, where the barrier is lowered to prevent flooding (Figure 10.4.6, right).



Figure 10.4.6 The Fox Point Hurricane Barrier in Providence, RI, USA (left) and the surge barrier on the Hollandse IJssel river in southern Holland. (Fox Point by Marcbela (Marc N. Belanger) (Own work) [Public domain], via Wikimedia Commons; Dutch barrier by Mark Voorendt (Own work) [CC BY-SA 4.0], via Wikimedia Commons).

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CHAPTER OVERVIEW

11: Ocean Circulation

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11.1: Density

Density refers to the amount of mass per unit volume, such as grams per cubic centimeter (g/cm^3). The density of fresh water is 1 g/cm^3 at 4° C (see section 5.1), but the addition of salts and other dissolved substances increases surface seawater density to between 1.02 and 1.03 g/cm^3. The density of seawater can be increased by reducing its temperature, increasing its salinity, or increasing the pressure. Pressure has the least impact on density as water is fairly incompressible, so pressure effects are not very significant except at extreme depths. However, if not for the slight compression of water due to pressure, sea level would be approximately 50 m higher than it is today! That leaves temperature and salinity as the primary factors determining density, and of these, temperature has the greatest impact (Figure 11.1.1).

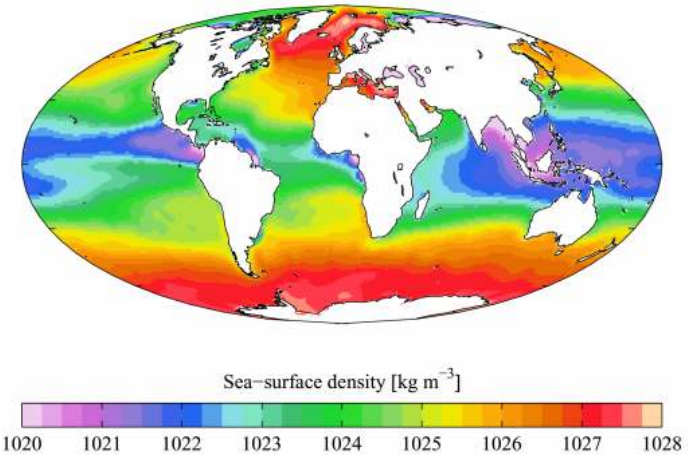


Figure 11.1.1 Global sea surface density. Colder polar regions display higher densities than warmer tropical zones (By Plumbago (Own work) [CC BY-SA 3.0, or GFDL (http://www.gnu.org/copyleft/fdl.html)], via Wikimedia Commons).

Since temperature has the greatest effect on density, density profiles are usually mirror images of temperature profiles (Figure 11.1.2). Density is lowest at the surface, where the water is the warmest. As depth increases, there is a region of rapidly increasing density with increasing depth, which is called the **pycnocline**. The pycnocline coincides with the thermocline, as it is the sudden decrease in temperature that leads to the increase in density. Below the pycnocline, density may be fairly constant (as is temperature), or it may continue to increase slightly towards the bottom.

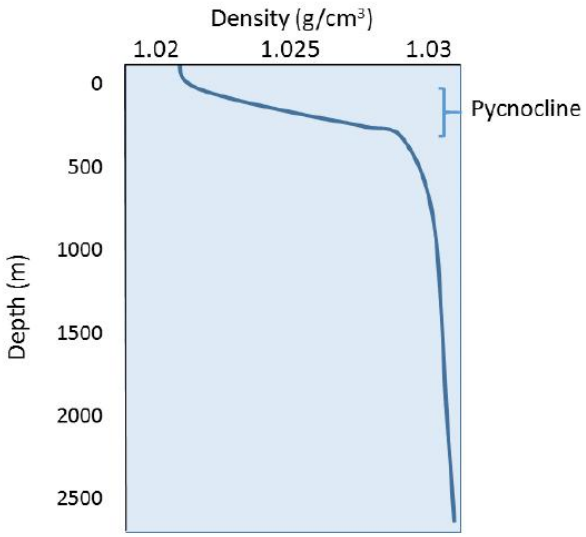


Figure 11.1.2 Representative density profile for the open ocean at mid-latitudes. The warm surface water causes a decrease in surface density (PW).

The profile above represents a stable state, or a high degree of stratification, where the warm, low density layer sits atop the colder, denser layer. If denser water happened to form at the surface, the water masses would be unstable, and the denser water would sink

to the bottom, to be replaced by less dense water at the surface. This vertical movement of water masses based on density (as determined by temperature and salinity) is referred to as thermohaline circulation, which is the topic of section 9.8. By creating a stratified water column, the thermocline and pycnocline together create a barrier that prevents mixing between the warmer, less dense surface water and the colder, denser bottom water. In this way, nutrient-rich deep water may be prevented from coming to the surface to support primary production.

As with temperature, there are also latitudinal differences in density. In the tropics the surface water is warm and low density, and there is a pronounced thermocline separating it from the colder, denser deep water. As stated above, this stratification prevents nutrient-rich water from reaching the surface and as a result tropical regions often have low productivity. In the high latitudes the water is uniformly cold at all depths, so there is little density stratification. The lack of a pycnocline (or a thermocline) allows cold, nutrient-rich deep water to more easily mix with the surface water, leading to higher primary production in polar regions.

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11.2: Salinity Patterns

Because salinity contributes to water density, and, thus, thermohaline circulation, it is important to discuss variations in salinity across oceans and, also, with depth.

Variations in Salinity

Total salinity in the open ocean averages 33-37 ppt, but it can vary significantly in different locations. But since the major ion proportions are constant, the regional salinity differences must be due more to water input and removal rather than the addition or removal of ions. Fresh water input comes through processes like precipitation, runoff from land, and melting ice. Fresh water removal primarily comes from evaporation and freezing (when seawater freezes, the resulting ice is mostly fresh water and the salts are excluded, making the remaining water even saltier). So differences in rates of precipitation, evaporation, river discharge, and ice formation play a significant role in regional salinity variations. For example, the Baltic Sea has a very low surface salinity of around 10 ppt, because it is a mostly enclosed body of water with lots of river input. Conversely, the Red Sea is very salty (around 40 ppt), due to the lack of precipitation and the hot environment which leads to high levels of evaporation.

One of the saltiest large bodies of water on Earth is the Dead Sea, between Israel and Jordan. Salinity in the Dead Sea is around 330 ppt, which is almost ten times saltier than the ocean. This extremely high salinity is a result of the hot, arid conditions in the Middle East that lead to high rates of evaporation. In addition, in the 1950s the flow from the Jordan River was diverted away from the Dead Sea, so there is no longer significant fresh water input. With no input and high evaporation, the water level in the Dead Sea is receding at a rate of about 1 m per year. The high salinity makes the water very dense, which creates buoyant forces that allow people to easily float at the surface. But the high salinity also means that the water is too salty for most living organisms, so only microbes are able to call it home; hence the name the Dead Sea. But as salty as the Dead Sea may be, it is not the saltiest body of water on Earth. That distinction currently belongs to Gaet'ale Pond in Ethiopia, with a salinity of 433 ppt!

Latitudinal Variations

While local conditions are important for determining salinity patterns in any single location, there are some global patterns that bear further investigation. Temperature is highest at the equator, and lowest near the poles, so we would expect higher rates of evaporation, and therefore higher salinity, in equatorial regions (Figure 11.2.2). This is generally the case, but in the figure below salinity right along the equator seems to be a little lower than at slightly higher latitudes. This is because equatorial regions also get a high volume of rain on a regular basis, which dilutes the surface water along the equator. So the higher salinities are found at subtropical, warm latitudes with high evaporation and less precipitation. At the poles there is little evaporation, which, coupled with ice and snow melting, produces a relatively low surface salinity. The image below shows high salinity in the Mediterranean Sea; this is located in a warm region with high evaporation, and the sea is largely isolated from mixing with the rest of the North Atlantic water, leading to high salinity. Lower salinities, such as those around southeast Asia, are the result of precipitation and high volumes of river input.

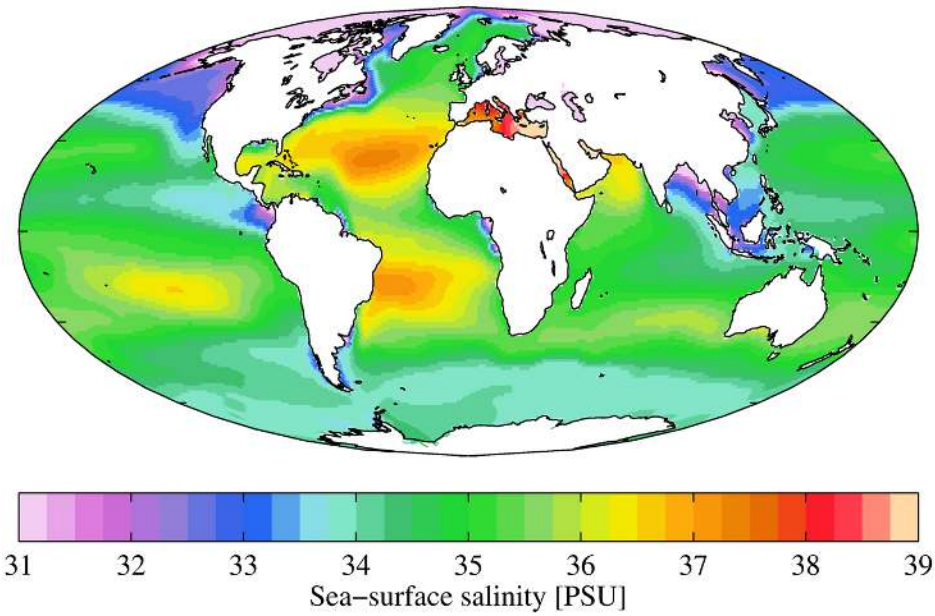


Figure 11.2.2 Annual mean global sea surface salinity (Data from World Ocean Atlas 2009. By Plumbago – Own work, [CC BY-SA 3.0], via Wikimedia Commons).

Figure 11.2.3 shows the mean global differences between evaporation and precipitation (evaporation – precipitation). Green colors represent areas where precipitation exceeds evaporation, while brown regions are where evaporation is greater than precipitations. Note the correlation between precipitation, evaporation, and surface salinity as seen in Figure 11.2.2.

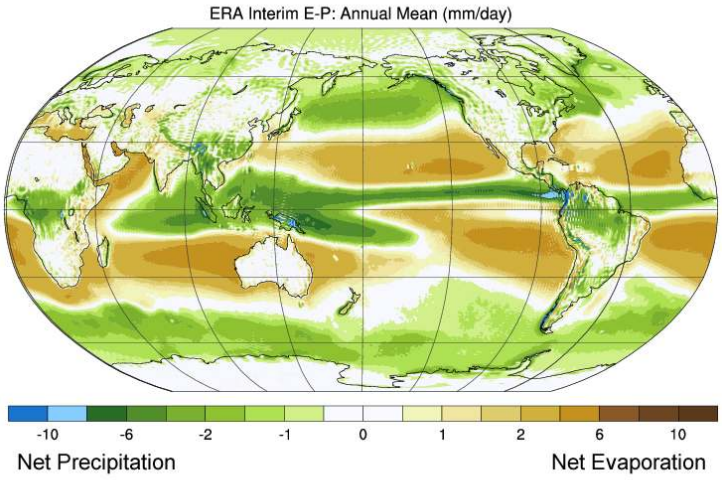


Figure 11.2.3 Annual mean global differences in evaporation and precipitation (J.Fasullo. National Center for Atmospheric Research Staff (Eds). "The Climate Data Guide: ERA-Interim: derived components." Retrieved from https://climatedataguide.ucar.edu/cl...ved-components).

Vertical Variation

In addition to geographical variation in salinity, there are also changes in salinity with depth. As we have seen, most differences in salinity are due to variations in evaporation, precipitation, runoff, and ice cover. All of these process occur at the ocean surface, not at depth, so the most pronounced differences in salinity should be found in surface waters. Salinity in deeper water remains relatively uniform, as it is unaffected by these surface processes. Some representative salinity profiles are shown in Figure 11.2.4. At the surface, the top 200 m or so show relatively uniform salinity in what is called the **mixed layer**. Winds, waves, and surface

currents stir up the surface water, causing a great deal of mixing in this layer and fairly uniform salinity conditions. Below the mixed layer is an area of rapid salinity change over a small change in depth. This zone of rapid change is called the **halocline,** and it represents a transition between the mixed layer and the deep ocean. Below the halocline, salinity may show little variation down to the seafloor, as this region is far removed from the surface processes that impact salinity. In the figure below, note the low surface salinity at high latitudes, and higher surface salinity at low latitudes as discussed above. Yet despite the surface differences, salinity at depth in both locations may be very similar.

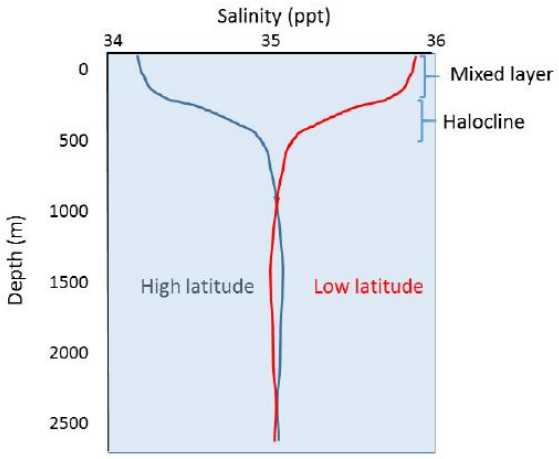


Figure 11.2.4 Salinity profiles from two hypothetical sites in the open ocean, one from high latitude and one from low latitude (PW).

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11.3: Thermohaline Circulation

This page is a draft and is under active development.

Circulation of water throughout the world's oceans occurs by one of two major modes: surface currents or thermohaline circulation. The former resides in the upper few hundred meters of the oceans and are predominantly controlled by winds. The latter is regulated by density. The main drivers of density in the ocean are temperature (*thermo-*) and salinity (*-haline*). Together, these two physical parameters dictate particular ocean water mass densities, the driving force behind large-scale, deep circulation. Cold, salty, dense water sinks and pushes other water masses around the ocean on a very slow scale. In general, as a result, thermohaline circulation causes warm water to flow poleward near the surface; in turn, this warm water cools and becomes denser. The water mass then sinks and begins flowing south toward the equator.

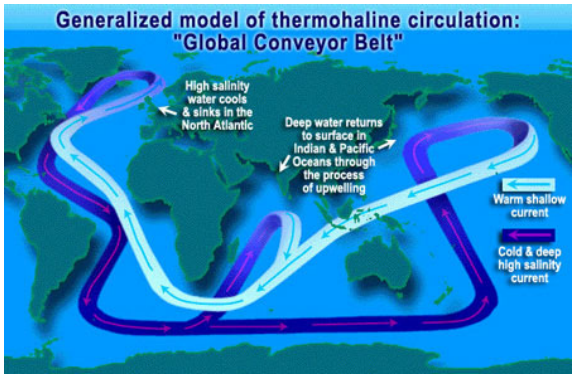


photo courtesy of http://nsidc.org/cryosphere/seaice/e...l_climate.html

Deepwater Formation

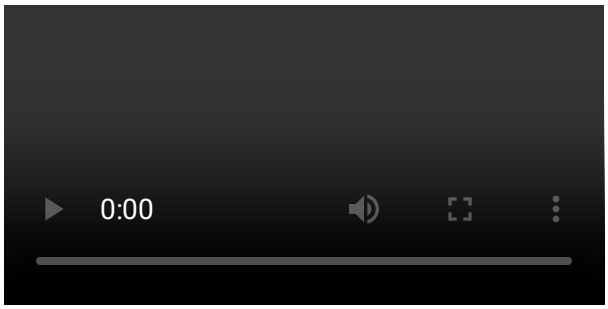
Where does it all begin? Thermohaline circulation is driven by the formation of North Atlantic Deep-Water (NADW), and Antarctic Bottom Water (AABW). Collectively these terms are referred to as deep water. Because the Earth is a sphere, the sun's energy warms the globe unequally, with most of the energy concentrated along the equator and the least amount of energy reaching the poles. The Gulf Stream facilitates the movement of warm equatorial waters through the tropics and subtropics north to the North Atlantic. Along this transit, the water becomes more saline due to evaporation. As the salty mass proceeds northward toward the sub-Arctic regions, it also cools. It can cool enough for sea ice to form, which leaves salt behind, causing the water to become even saltier and denser. Consequently, water jettisoned to the Norwegian Sea and the Arctic Ocean by the Gulf Stream becomes dense enough to sink out, and thus NADW is formed.

Kitchen Oceanography Experiment: Ocean Overturning Circulation [Link]

Try this easy-to-do at-home experiment shared by Dr. Mirjam S. Glessmer. It can potentially help you better understand how the temperature and salinity differences of water lead to North Atlantic Deep-Water circulation!

Transit Through the Ocean

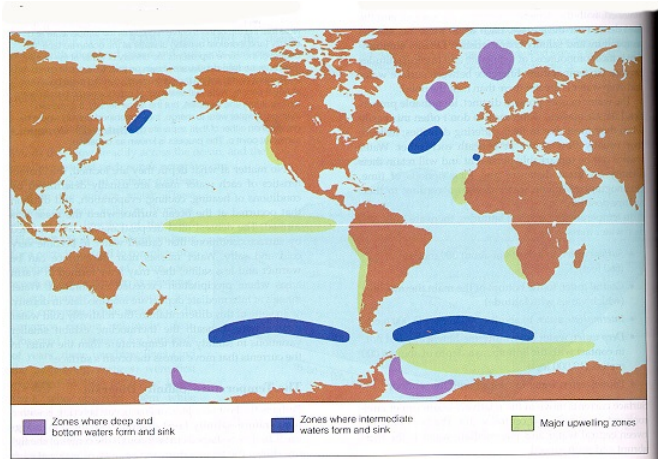
"This animation first depicts thermohaline surface flows over surface density, and illustrates the sinking of water in the dense ocean near Iceland and Greenland. The surface of the ocean then fades away and the animation pulls back to show the global thermohaline circulation," (Olsen et al 2013).



Thermohaline Surface Flow
http://svs.gsfc.nasa.gov/cgi-bin/details.cgi?aid=3658

Locations of Deep Water Formation

Deepwater is formed in several places around the world. As discussed above, most NADW forms in the North Atlantic, specifically in the Norwegian, Labrador, and Greenland Seas were the Gulf Current cools (also known as Atlantic Meridional Overturning Circulation). However, some NADW forms in the warmer waters of the Mediterranean. This water may not be as cold as polar sources but gains its density from its extremely salinity. Deepwater production sites in the Mediterranean include the Gulf of Lions, the Southern Adriatic Pit, and the Cretan Sea. Antarctic bottom water is formed during the production of large amounts of sea ice off the coast of Antarctica. The ice formation increases the salinity of the frigid surrounding waters, producing some of the densest seawater in the world. As its name implies, this water sinks right to the seafloor, displacing even NADW. This phenomenon occurs mostly in the Weddell and Ross Seas. Because of the proximity of these sites to large ice sheets, global warming could have strong effects on the rates of deep water formation. Freshwater runoff from melting glaciers can dilute surface salinity, decreasing density and stopping water from sinking. This in turn could slow global thermohaline circulation and have profound effects on how heat cycles on our planet.



Map illustrating upwelling zones and sinking zones

Image courtesy of education.nationalgeographic....r-belt/?ar_a=1

Further Resources

Flight Center Scientific Visualization Studio: Thermohaline Circulation - http://svs.gsfc.nasa.gov/cgi-bin/details.cgi?aid=3884

See Toggweiler and Key (2001) for an excellent introduction to thermohaline circulation www.researchgate.net/publication/242569649_Thermohaline_Circulation

NOAA: Thermohaline Circulation: http://oceanservice.noaa.gov/educati...conveyor1.html

References

1. www.nationalgeographic.org/encyclopedia/ocean-conveyor-belt/
2. https://oceanservice.noaa.gov/education/tutorial_currents/05conveyor1.html
3. https://pangea.stanford.edu/courses/EESS146Bweb/Lecture%2013.pdf
4. https://www.nature.com/articles/990069
5. http://adsabs.harvard.edu/abs/2018EGUGA..2019128S

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11.4: Surface Gyres

In the previous chapter the major wind patterns on Earth were derived. It is these prevailing winds that blow across the water surface to create the major ocean surface currents. However, only about 2% of the wind energy is actually transferred to the water, so a 50 knot wind only creates a 1 knot current. Furthermore, wind-driven surface currents only affect the top 100-200m of water, meaning surface currents only involve about 10% of the world's ocean water. In section 9.8 we will examine deep, thermohaline circulation, which impacts around 90% of the ocean water.

Surface currents generally move in the same direction as the winds that created them. However, because of Coriolis deflection, the surface currents are offset approximately 45° relative to the wind direction; 45° to the right in the Northern Hemisphere, and 45° to the left in the Southern Hemisphere. This creates a general circulation pattern where in both hemispheres, surface currents flow east to west between the equator and 30° latitude, west to east between 30° and 60°, and east to west between 60° and the poles (Figure 11.4.1).

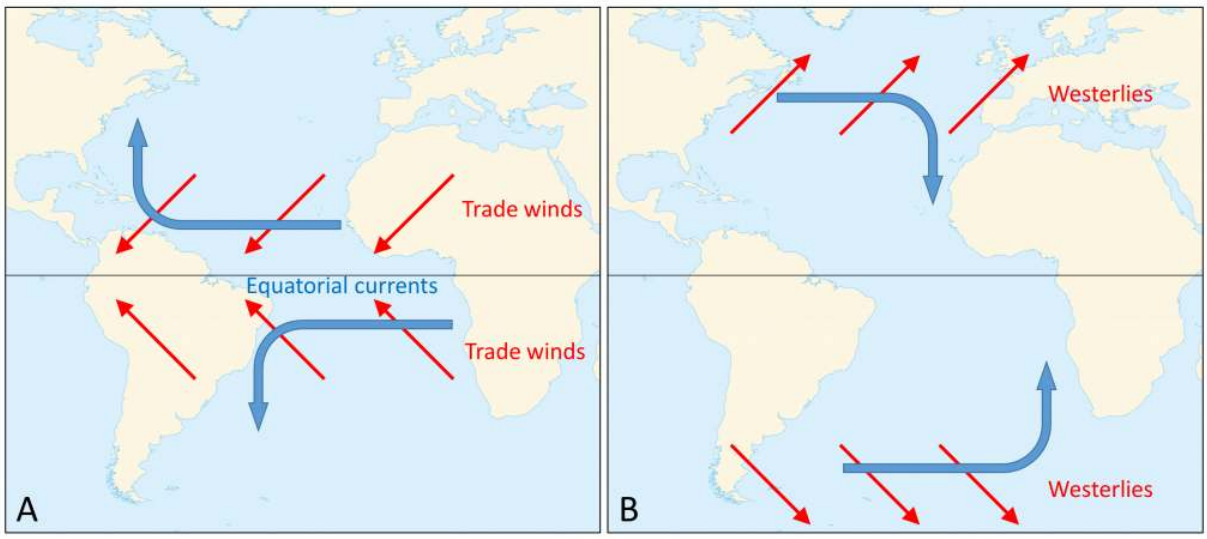


Figure 11.4.1 Generalized surface currents in the Atlantic Ocean. A) Surface water moving at 45° relative to the trade winds create the westward flowing equatorial currents. B) Between 30-60° latitude, the westerlies form eastward flowing surface currents (PW, map by Catrin (Own work, Using GMT) [CC BY-SA 3.0], via Wikimedia Commons).

The trade winds create the equatorial currents that flow east to west along the equator; the North Equatorial and South Equatorial currents. If there were no continents, these surface currents would travel all the way around the Earth, parallel to the equator. However, the presence of the continents prevents this unimpeded flow. When these equatorial currents reach the continents, they are diverted and deflected away from the equator by the Coriolis Effect; deflection to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. These currents then become western boundary currents; currents that run along the western side of the ocean basin (i.e. the east coasts of the continents). Since these currents come from the equator, they are warm water currents, bringing warm water to the higher latitudes and distributing heat throughout the ocean.

At the same time, between 30-60° latitude the westerlies move surface water towards the east. The Coriolis Effect and the presence of the continents deflect the currents towards the equator, creating eastern boundary currents (on the eastern side of the ocean basins). These currents come from high latitude areas, so they deliver cold water to the lower latitudes. Together, these currents combine to create large-scale circular patterns of surface circulation called **gyres**. In the Northern Hemisphere the gyres rotate to the right (clockwise), while in the Southern Hemisphere the gyres rotate to the left (counterclockwise).

There are five major gyres in the oceans; the North Atlantic, South Atlantic, North Pacific, South Pacific, and Indian (Figure 11.4.2). The North Pacific gyre is composed of the North Equatorial Current on its southern boundary, which turns into the Kuroshio Current (a.k.a. the Japan Current) bringing warm water north towards Japan. The Kuroshio flows into the North Pacific Current which moves east towards North America, where it becomes the California Current to complete the gyre. The North Atlantic gyre is formed by the North Equatorial Current flowing into the Gulf Stream along the east coast of the United States. The

Gulf Stream merges into the North Atlantic Current to move water towards Europe, which then becomes the Canary Current as it moves south to join the North Equatorial Current.

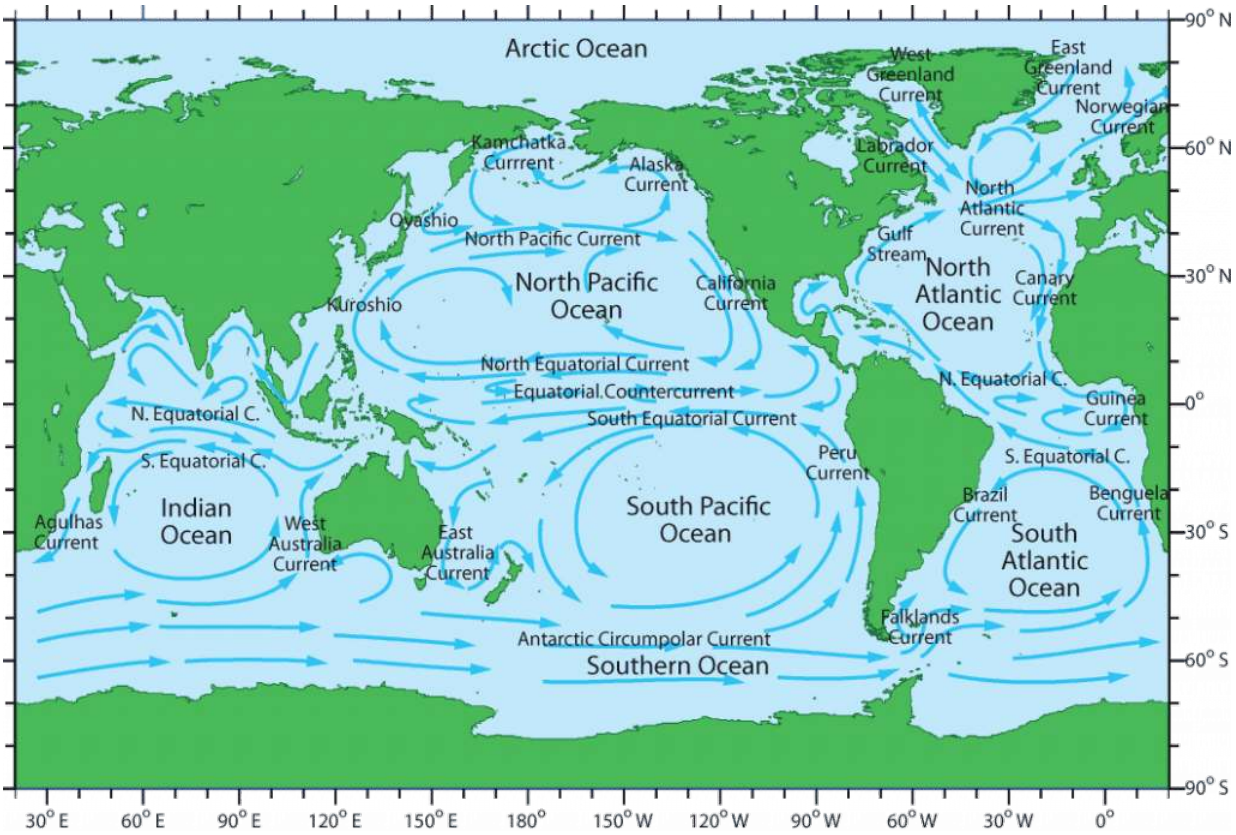


Figure 11.4.2 The major global surface currents (http://www.seos-project.eu/modules/o...s-c02-p01.html, Source: NOC; CC BY-NC-SA 2.0).

Near Antarctica the circulation is somewhat different. Because there is little in the way of continental land masses between 50-60º south, the surface current created by the westerly winds can make its way completely around the Earth, creating the Antarctic Circumpolar Current (ACC) or West Wind Drift (WWD) that flows from west to east (Figure 11.4.2). The Antarctic Circumpolar Current is the only current that connects all of the major ocean basins, and in terms of the amount of water that it transports, it is the largest surface current on Earth. Above 60º latitude the prevailing winds are the polar easterlies, which create a current flowing from east to west along the edge of the Antarctic continent, the East Wind Drift or the Antarctic Coastal Current.

The Antarctic Circumpolar Current creates the southern boundary for all of the Southern Hemisphere gyres. In the South Pacific gyre the ACC becomes the Peru Current (also known as the Humboldt Current) moving up the west coast of South America, before joining the South Equatorial Current. The South Equatorial Current flows southwards as the East Australia Current, before completing the gyre with the ACC. The South Atlantic gyre is composed of the South Equatorial Current, the Brazil Current, the ACC, and the Benguela Current. Finally, the currents making up the Indian gyre are the ACC, the West Australia Current, the South Equatorial Current, and the Agulhas Current.

Not all of the equatorial water that is moved westward by the trade winds and reaches the continents gets transported to higher latitudes in the gyres, because the Coriolis Effect is weakest along the equator. Instead, some of the water piles up along the western edge of the ocean, and then flows eastward due to gravity, creating narrow Equatorial Countercurrents between the North and South Equatorial Currents (Figure 11.4.2). Some of this water also moves east as equatorial undercurrents that flow at depths between 50-200 m, underneath the Equatorial Currents. These undercurrents are called the Lomonosov Current in the Atlantic, and the Cromwell Current in the Pacific.

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https://geo.libretexts.org/@go/page/13023

11.5: The Gulf Stream

The primary surface current along the east coast of the United States is the Gulf Stream, which was first mapped by Benjamin Franklin in the 18th century (Figure 11.5.1). As a strong, fast current, it reduced the sailing time for ships traveling from the United States back to Europe, so sailors would use thermometers to locate its warm water and stay within the current.

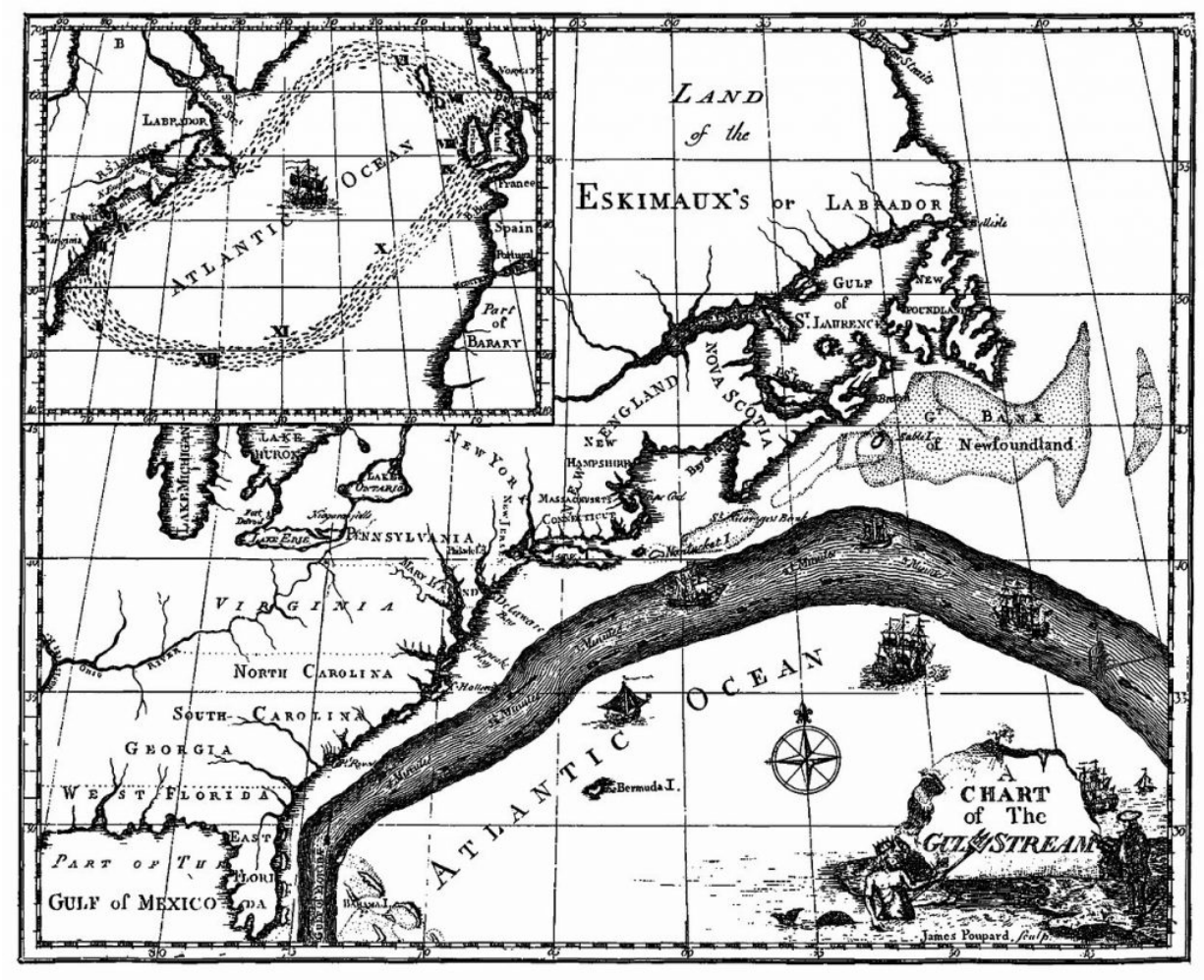


Figure 11.5.1 Benjamin Franklin's original map of the Gulf Stream (Public domain, via Wikimedia Commons).

The Gulf Stream is formed from the convergence of the North Atlantic Equatorial Current bringing tropical water from the east, and the Florida Current that brings warm water from the Gulf of Mexico. The Gulf Stream takes this warm water and transports it northwards along the U.S. east coast (Figure 11.5.2). As a western boundary current, the Gulf Stream experiences western intensification (section 9.4), making the current narrow (50-100 km wide), deep (to depths of 1.5 km) and fast. With an average speed of 6.4 km/hr, and a maximum speed of about 9 km/hr, it is the fastest current in the world ocean. It also transports huge amounts of water, more than 100 times greater than the combined flow of all of the rivers on Earth.

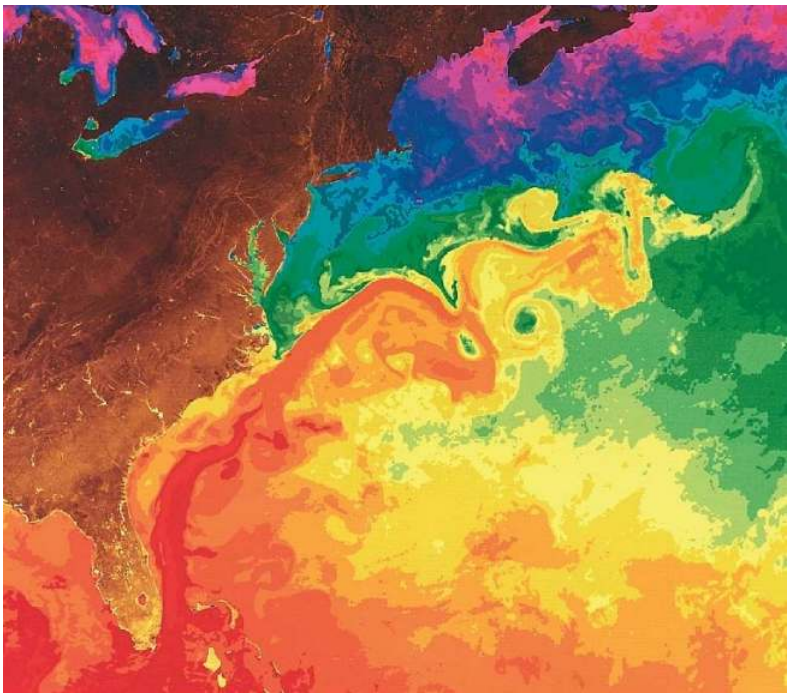


Figure 11.5.2 Sea surface temperature map illustrating the Gulf Stream. Warmer water is shown in red, colder water in blue and violet. Meanders and eddies are visible where the current moves towards the northeast (NASA, Public Domain via Wikimedia Commons).

As the Gulf Stream approaches Canada, the current becomes wider and slower as the flow dissipates and it encounters the cold Labrador Current moving in from the north. At this point, the current begins to meander, or change from a fast, straight flow to a slower, looping current (Figure 11.5.2). Often these meanders loop so much that they pinch off and form large rotating water masses called **rings** or **eddies**, that separate from the Gulf Stream. If an eddy pinches off from the north side of the Gulf Stream, it entraps a mass of warm water and moves it north into the surrounding cold water of the North Atlantic. These **warm core rings** are shallow, bowl-shaped water masses about 1 km deep, and about 100 km across, that rotate clockwise as they carry warm water in to the North Atlantic (Figure 11.5.3). If the meanders pinch off at the southern boundary of the Gulf Stream, they form **cold core rings** that rotate counterclockwise and move to the south. Cold core rings are cone-shaped water masses extending down to over 3.5 km deep, and may be over 500 km wide at the surface.

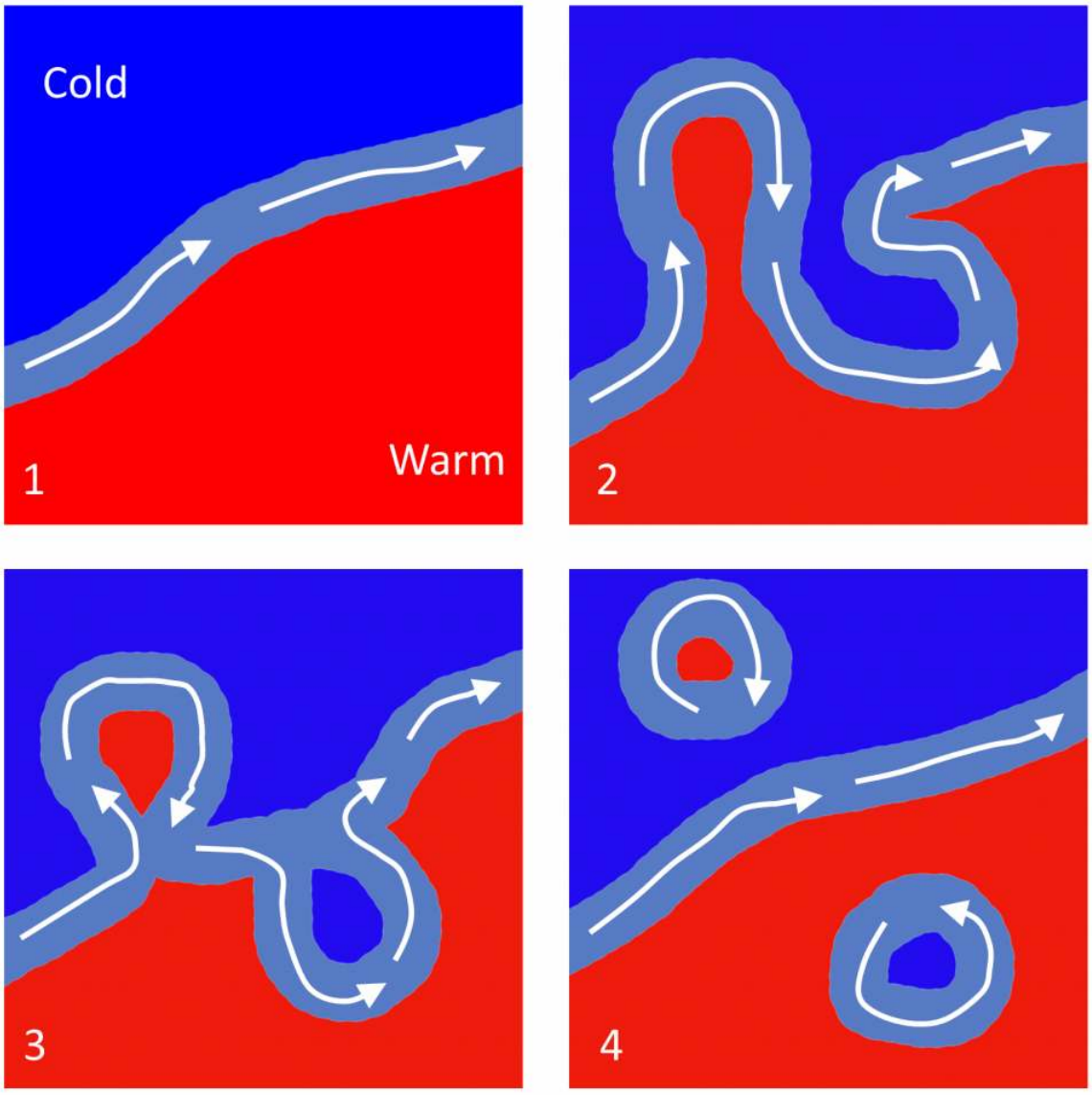


Figure 11.5.3 Formation of warm and cold core rings from meanders in the Gulf Stream. As the Gulf Stream flows to the northeast (1), it starts to meander as it slows, forming warm or cold water extensions on either side of the current (2). If the meanders pinch off the extensions, they trap pockets of warm or cold water (3), that can separate from the Gulf Stream and travel north or south. Warm core rings rotate clockwise, while cold core rings rotate counterclockwise (4) (PW).

After the Gulf Stream meets the cold Labrador Current, it joins the North Atlantic Current, which transports the warm water towards Europe, where it moderates the European climate. It is estimated that Northern Europe is up to 9º C warmer than expected because of the Gulf Stream, and the warm water helps to keep many northern European ports ice-free in the winter.

In the east, the Gulf Stream merges into the Sargasso Sea, which is the area of the ocean within the rotation center of the North Atlantic gyre. The Sargasso Sea gets its name from the large floating mats of the marine algae *Sargassum* that are abundant on the surface (Figure 11.5.4). These *Sargassum* mats may play an important role in the early life stages of sea turtles, who may live and feed within the algae for many years before reaching adulthood.



Figure 11.5.4 Map of the Sargasso Sea (left), and closeup photo of Sargassum algae (right) (Map by USFWS; photo by Bogdan Giuşcă; Public Domain via Wikimedia Commons).

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11.6: The Ekman Spiral and Geostrophic Flow

Ekman spiral

Winds blowing over the ocean are ultimately what create the surface currents. However, not all of the water moved by the surface currents is transported in the same direction. The Coriolis Effect causes the surface water to move in a direction about 45 degrees offset from the wind direction, with the deflection to the right of the wind in the Northern Hemisphere and to the left in the Southern Hemisphere. The frictional movement of the topmost layer of water sets in motion the layer directly underneath it, which then sets in motion the next layer under that, and so on as the water gets deeper. Some energy is lost in each transition, so each successive layer of water will not move as far as the layer above it; in other words, there is decreasing energy with increasing depth. But at the same time, the Coriolis Effect deflects each layer relative to the layer above it (again, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere). The movement of the successive layers therefore creates a spiraling pattern of water motion called the **Ekman spiral**, which usually penetrates to about 100 m deep before the motion ceases. If the magnitudes and directions of the movements of each layer are added together, the result is that the net movement of the upper 100 m of the water column is 90° relative to the original wind direction (90° to the right of the wind in the Northern Hemisphere, and 90° to the left in the Southern Hemisphere). This net water movement is called **Ekman transport** (Figure 11.6.1).

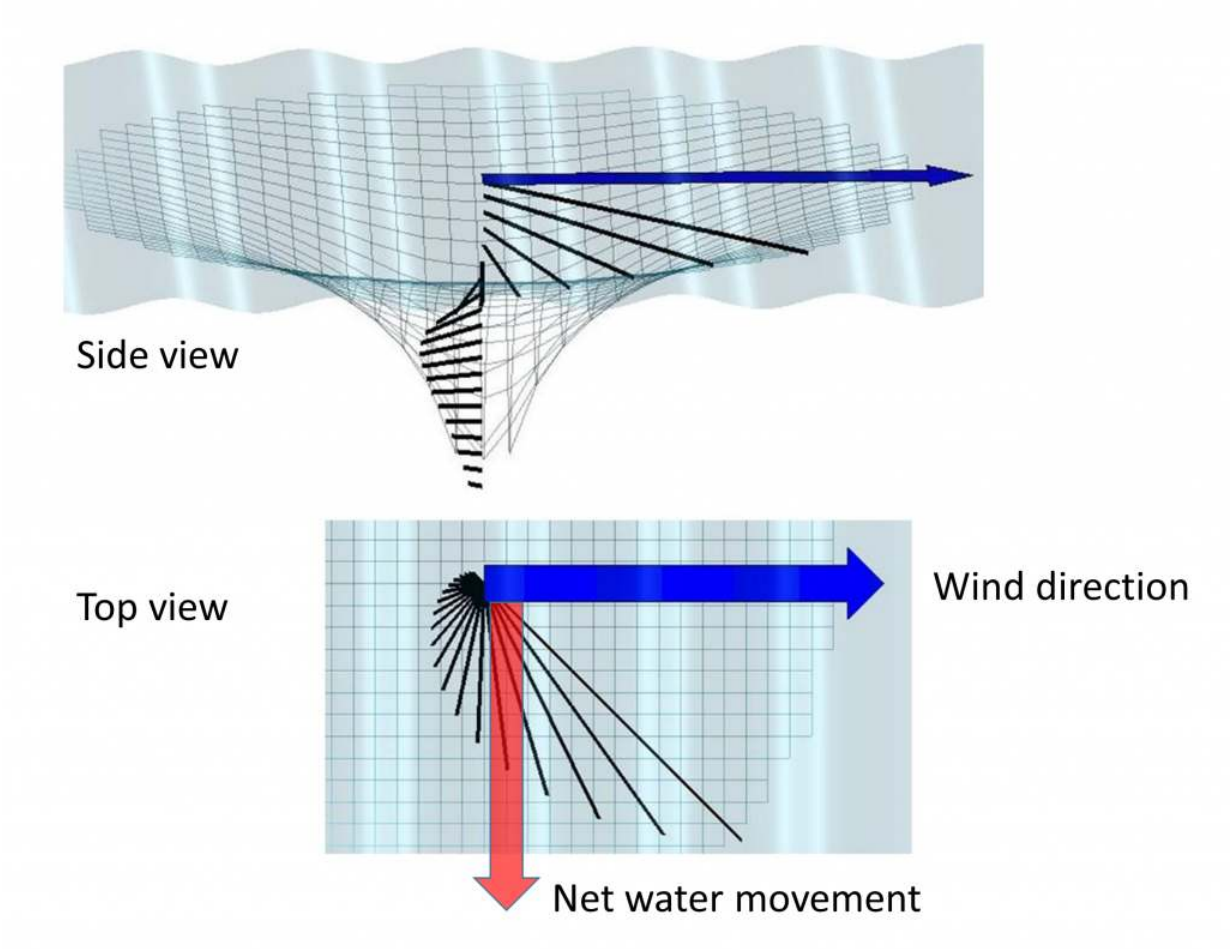


Figure 11.6.1 The Ekman spiral, shown for the Northern Hemisphere. Wind blowing over the water (blue arrow) creates a surface current 45° offset from the wind. Each successive layer of water is moved and deflected by the layer above, creating a spiraling pattern of water movement that diminishes with depth. The net movement of the water within the spiral is 90° relative to the wind direction (red arrow) (Modified by PW from Ekman spiral By Schlusenbach (unbekannt) [Public domain], via Wikimedia Commons).

Geostrophic flow

Gyre rotation is dependent on the wind and the Coriolis Effect impacting the surface currents (see section 9.1). But the rotation is also affected by movement below the surface due to Ekman transport. In the Northern Hemisphere, as the gyre rotates clockwise the net movement of the Ekman transport is 90° to the right of the wind; in other words, towards the center of the gyre. The Ekman transport piles up water in the center of the gyre, making the water level higher in the gyre center than on the edges of the gyres. This pile of water then has a tendency to flow back "downhill" due to gravity. As the water flows "downhill" away from the gyre center, it is deflected to the right by the Coriolis force. This results in a clockwise current around the central "hill" called **geostrophic flow**, which moves in the same direction as the rotating gyre. Water is thus pushed into the "hill" by Ekman transport, and away from the "hill" by gravity, with both flows modified by the Coriolis Effect to create the rotation. As with the gyres, geostrophic flow is clockwise in the Northern Hemisphere, and counterclockwise in the Southern Hemisphere.

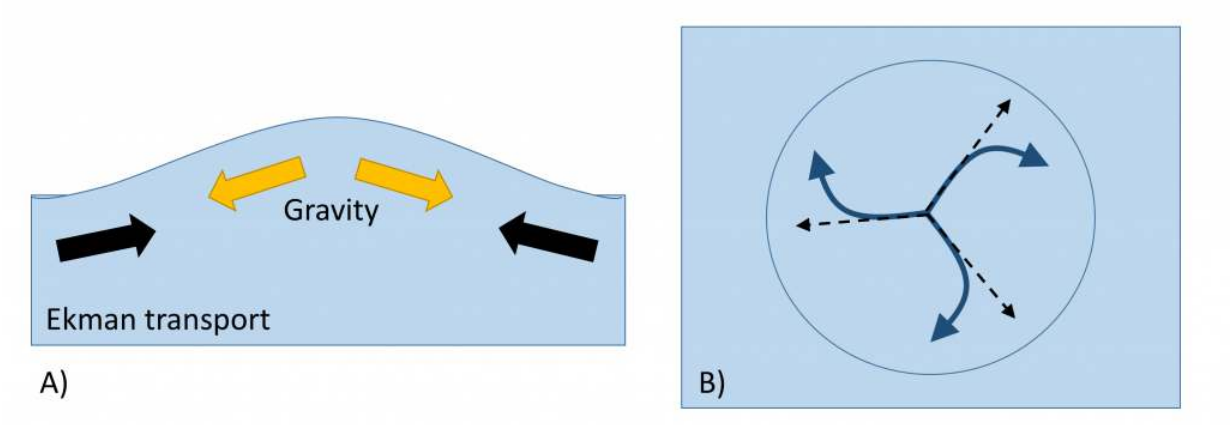


Figure 11.6.2 Geostrophic flow in the Northern Hemisphere. A) Ekman transport moves water into the middle of the gyre, where it "piles up." Gravity causes the water to flow back "downhill." B) Viewed from above, as the water in the center flows "downhill" (dotted arrows) the Coriolis force deflects the movement to the right (solid arrows), causing the system to rotate clockwise (PW).

Most major surface currents are a combination of wind-driven and geostrophic currents. Since winds can be variable, geostrophic flow ensure that the gyre currents keep moving at a fairly constant rate even when the wind dies down. The larger the area, and the higher the slope, the longer the geostrophic flow will continue to move and power the gyre after the wind subsides.

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11.7: Western Intensification

In both hemispheres, the currents making up the western side of the gyre are much more intense than the currents on the eastern side. In other words, the currents off of the east coast of the continents are more intense than currents off of the west coast of the continents. This phenomenon is known as western intensification, and once again it is due to the Coriolis Effect.

As discussed previously, the Coriolis Effect is a result of the fact that different latitudes of the Earth are rotating at different speeds, and the apparent path taken by an object is deflected as it moves between areas of different rotation speeds. The greater the change in rotation speed, the stronger the Coriolis force. At the poles, the speed of rotation is 0 km/hr. The speed increases to about 800 km/hr at 60° latitude, 1400 km/hr at 30° latitude, and 1600 km/hr at the equator. Therefore there is an 800 km/hr difference between 60° and 90° latitude, while there is only a 200 km/hr difference between the equator and 30°. Thus the speed of Earth's rotation changes more quickly with latitude near the poles than at the equator, making the Coriolis force strongest near the poles and weakest at the equator.

The high latitude surface currents of the major gyres experience a strong Coriolis force due to their proximity to the poles. As the currents move eastward, the strong Coriolis force begins to deflect the currents towards the equator relatively early. The currents on the eastern side of the gyre are therefore spread out over a wide area as they move towards the equator (Figure 11.7.1). Near the equator, the westward flowing currents experience a much weaker Coriolis force, so their deflection does not happen until the current is all the way over to the western side of the ocean basin. These western currents must therefore move through a much narrower area (Figure 11.7.1). This imbalance means that the center of rotation of the gyre is not in the center of the ocean basins, but it closer to the western side of the gyre.

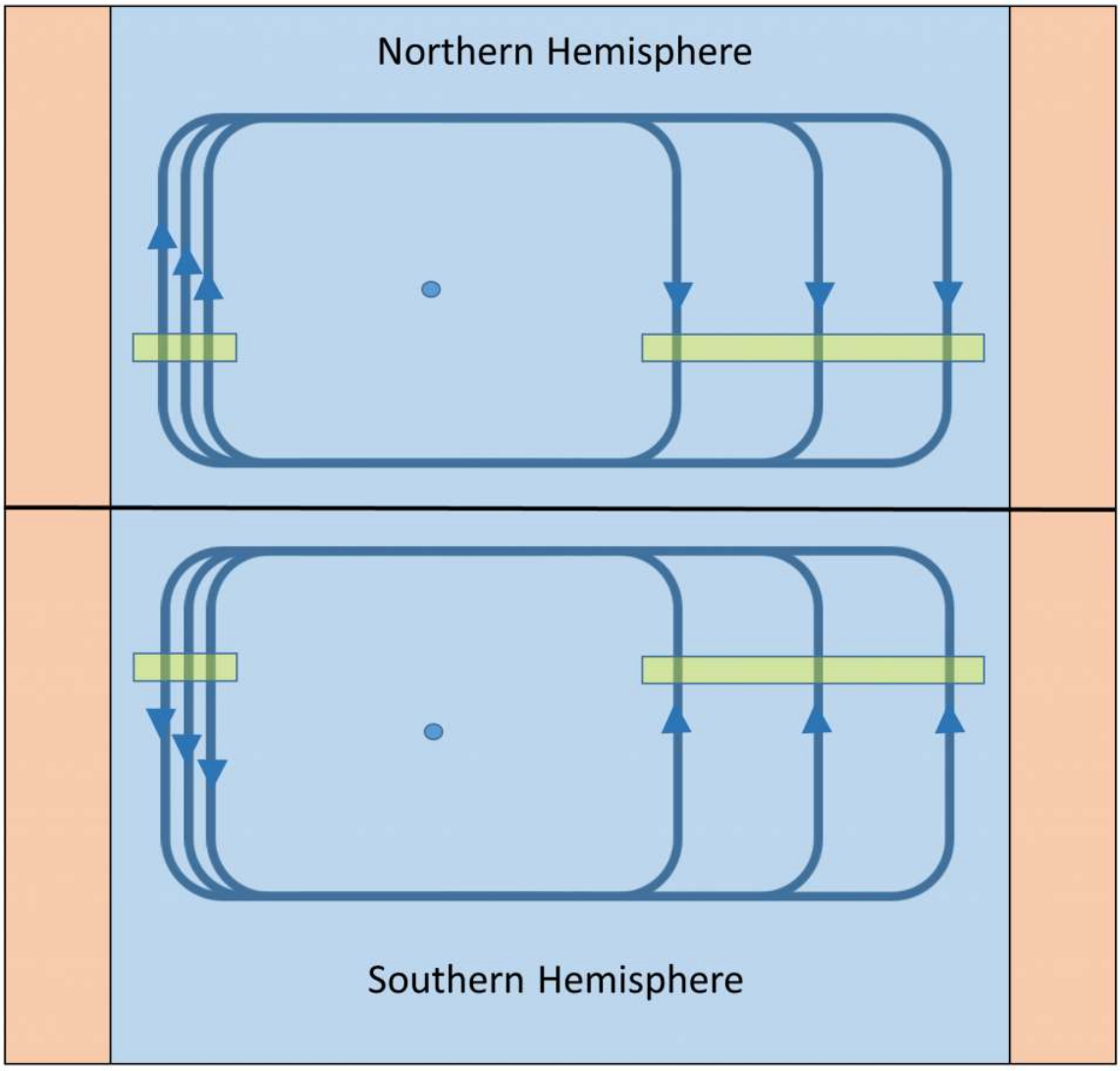


Figure 11.7.1 Western intensification. In both hemispheres, currents on the western side of the gyres travel through a much narrower area than the currents on the eastern side (yellow rectangles). To move the same volume of water through each side, western boundary currents are faster, deeper, and narrower than eastern boundary currents. The center of rotation of the gyre is also closer to the western side of the gyre (blue dots) (PW).

The same volume of water must pass through both the east and west sides of the gyre. In the western gyre currents, that volume is passing through a narrower area, so the current must travel faster in order to transport the same amount of water in the same amount of time. On the eastern side of the gyre the current is much wider, so the flow is slower. A simple analogy is the water flowing from a garden hose. You can make the water flow from the hose much faster and more strongly by covering part of the opening with your thumb. The same amount of water is exiting the hose whether the opening is covered or uncovered, but to get that water through the covered opening the flow has to be much faster and stronger. In the same way, western boundary currents are not only faster, but also deeper than eastern boundary currents, as they move the same volume through a narrower space. For example, the Kuroshio Current in the western Pacific is around 15 times faster, 20 times narrower, and 5 times deeper than the California Current in the eastern Pacific.

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11.8: Currents, Upwelling and Downwelling

The movement of surface currents also plays a role in the vertical movements of deeper water, mixing the upper water column. **Upwelling** is the process that brings deeper water to the surface, and its major significance is that it brings nutrient-rich deep water to the nutrient-deprived surface, stimulating primary production (see section 7.3). **Downwelling** is where surface water is forced downwards, where it may deliver oxygen to deeper water. Downwelling leads to reduced productivity, as it extends the depth of the nutrient-limited layer.

Upwelling occurs where surface currents are diverging, or moving away from each other. As the surface waters diverge, deeper water must be brought to the surface to replace it, creating upwelling zones. The upwelled water is cold and rich in nutrients, leading to high productivity. Many of the most productive regions on Earth are found in upwelling zones. In the equatorial Pacific, the trade winds blow the North and South Equatorial Currents towards the west, while Ekman transport causes the upper layers to move to the north and south in their respective hemispheres. This creates a divergence zone, and a region of upwelling and high productivity (Figure 11.8.1).

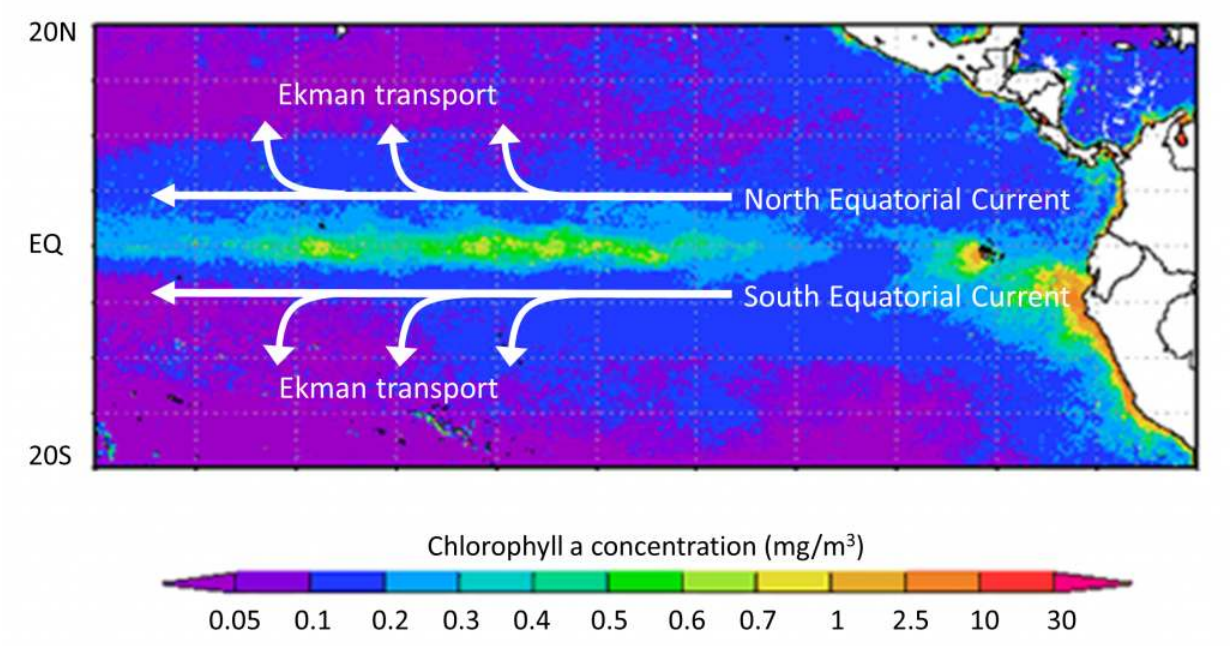


Figure 11.8.1 Equatorial upwelling and increased productivity as a result of divergence between the north and south equatorial currents (Modified by PW from image by NASA [Public domain], via Wikimedia Commons).

A similar process occurs near the Antarctic continent, creating one of the most productive regions on Earth, the Antarctic divergence. In this case, the West Wind Drift (Antarctic Circumpolar Current) is flowing parallel to, but in the opposite direction of the East Wind Drift. With both currents occurring in the Southern Hemisphere, Ekman transport will be to the left, so the eastward-flowing West Wind Drift water will be transported to the north, and the westward-flowing East Wind Drift water will be transported to the south, creating a highly productive divergence zone (Figure 11.8.2).

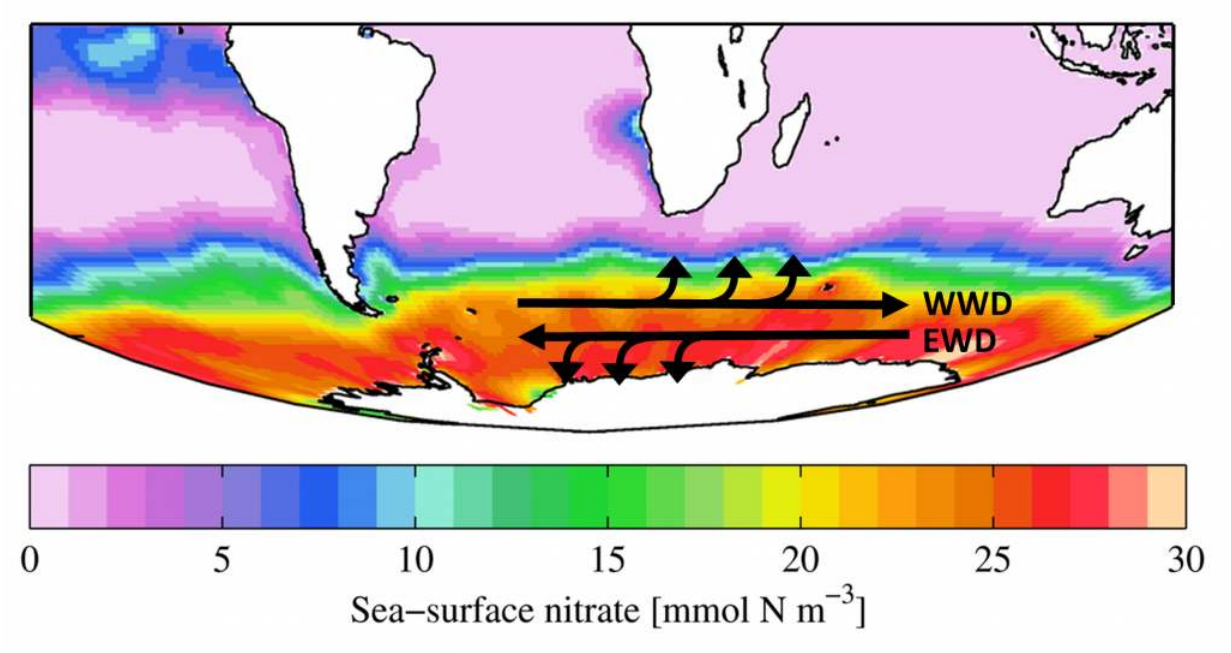


Figure 11.8.2 High nutrient levels in the Antarctic divergence zone, as a result of the diverging West Wind Drift and East Wind Drift currents creating strong upwelling (Modified by PW from Plumbago (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

Downwelling occurs where surface currents converge. The converging water has nowhere to go but down, so the surface water sinks. Since surface water is usually low in nutrients, downwelling leads to low productivity zones. An example of a downwelling region is off of the Labrador coast in Canada, where the Gulf Stream, Labrador, and East Greenland Currents converge.

Coastal Upwelling

Upwelling and downwelling also occur along coasts, when winds move water towards or away from the coastline. Surface water moving away from land leads to upwelling, while downwelling occurs when surface water moves towards the land. Historically, some of the most productive commercial fishing grounds have been associated with coastal upwelling. Along the coast of California, the local prevailing winds blow towards the south. Ekman transport moves the surface layer 90° to the right of the wind, meaning the net Ekman transport is in an offshore direction. The water displaced near the coast is replaced by cold, nutrient-rich deeper water that is brought to the surface through upwelling, leading to high productivity (Figure 11.8.3).

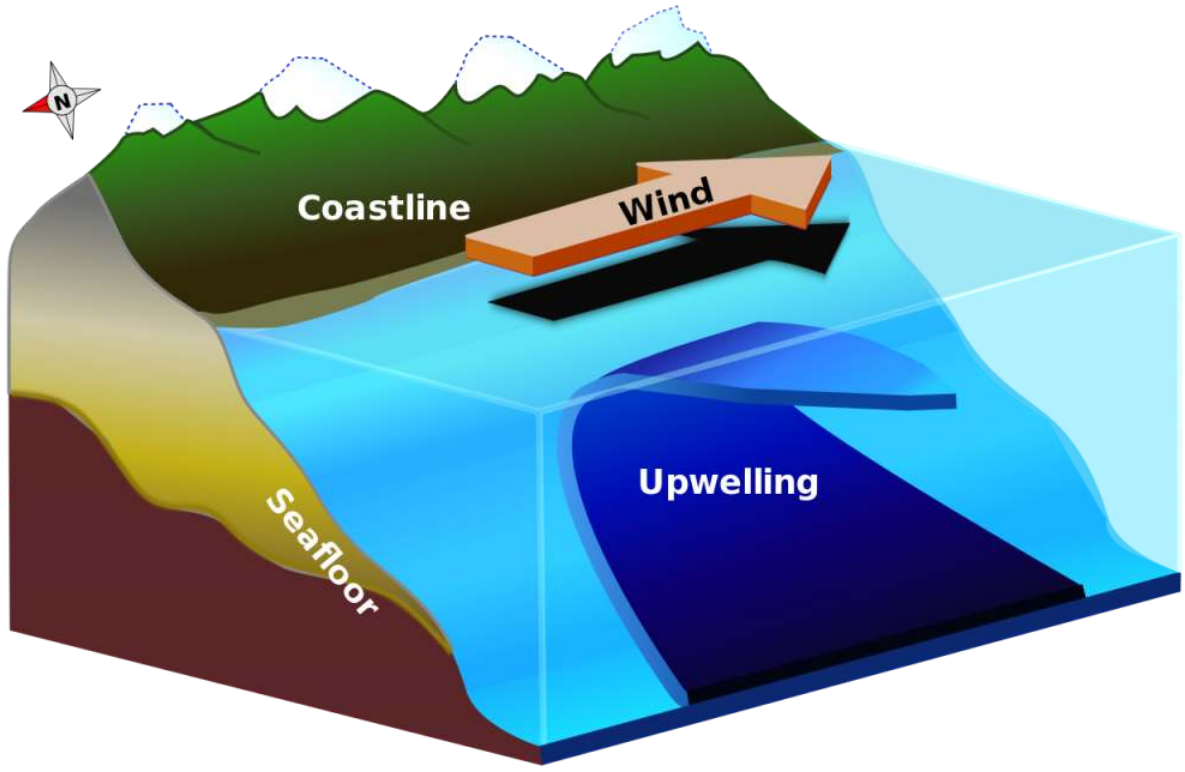


Figure 11.8.3 Coastal upwelling. As wind blows along a coastline, Ekman transport moves the surface layer in a direction 90° to the wind. In this figure, it is to the right of the wind, indicating a Northern Hemisphere location. As surface water moves offshore it it replaced by upwelled deeper water (By Lichtspiel [Public domain], via Wikimedia Commons).

The same process happens off of the coast of Peru, which for a long time had the world's largest commercial fishery. Winds along the Peruvian coast blow towards the north, and since Peru is in the Southern Hemisphere, the Ekman transport is 90° to the left of the wind, which causes the surface water to move offshore and leads to upwelling and productivity. In any coastal upwelling location, if the winds reverse, surface water moves towards the shore and downwelling is the result.

Upwelling can also occur due to geological features of the ocean floor. For example, as deep water currents encounter seamounts or other raised features, the water is forced upwards, bringing nutrient-rich water to the surface. This helps explain why productivity is often high in the water over seamounts.

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11.9: El Niño and La Niña

As we saw in the previous section, coastal upwelling off of Peru makes that region one of the world's most productive fishing grounds. But every so often, the conditions in the region are very different. Every few years, the cold, nutrient-rich water is replaced by unusually warm water that is low in nutrients, leading to a decline in fish populations. In addition, the normally dry areas receive lots of rain. Because this phenomenon occurs in the northern winter close to Christmas, it is called **El Niño** (the child). More formally, the event is referred to as **El Niño-Southern Oscillation (ENSO)**. The Southern Oscillation portion refers to the fluctuating atmospheric conditions that lead to the localized ocean warming of El Niño. While the exact reasons for the oscillation events are unclear, it is easier to understand how they lead to an El Niño.

Under normal conditions in the equatorial Pacific, the trade winds blow towards the west, moving large amounts of warm surface water towards the western Pacific around Southeast Asia. As the surface water moves west, it is replaced by cold, nutrient-rich deep water through upwelling (Figure 11.9.1). The coastal upwelling leads to a shallow thermocline in the eastern Pacific. In terms of atmospheric conditions, the trade winds are part of a convection cell called the Walker Cell. There is low pressure over the western Pacific, leading to rising moist air and significant precipitation in the region. In the eastern Pacific near South America, there is high pressure, leading to drier conditions (Figure 11.9.1).

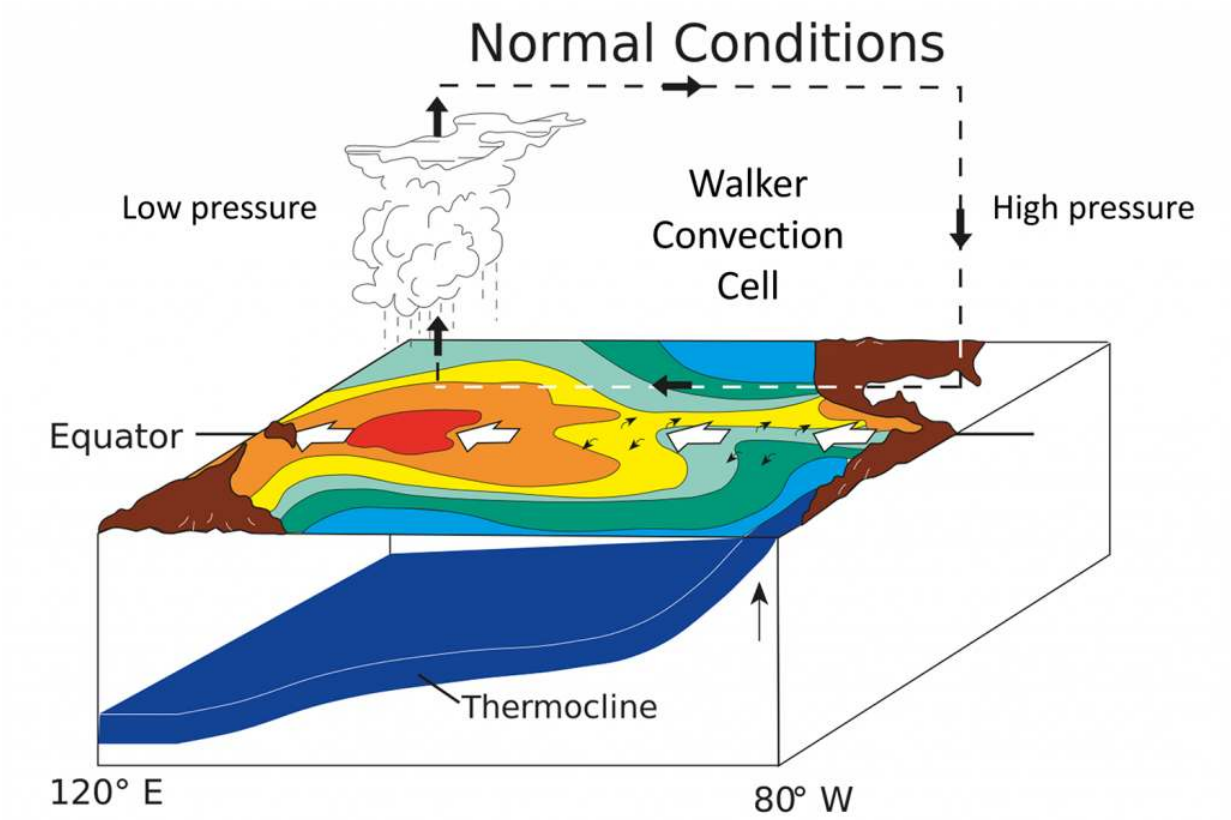


Figure 11.9.1 Normal conditions in the equatorial Pacific. Low pressure in the western Pacific and high pressure in the eastern Pacific cause the trade winds to move surface water to the west, leading to upwelling and a shallow thermocline near South America (Modified by PW from Fred the Oyster (Own work) [Public domain], via Wikimedia Commons).

During an El Niño-Southern Oscillation, the high pressure system over the eastern Pacific diminishes, so the trade winds are weakened, or in extreme cases will even reverse. When this happens, warm surface water begins to flow east across the Pacific towards South America (Figure 11.9.2), warming the coastal South American water by up to 8° C in strong ENSO years. This influx of low density warm water deepens the thermocline and prevents upwelling, which dramatically reduces productivity and can devastate populations of fish and other marine life.

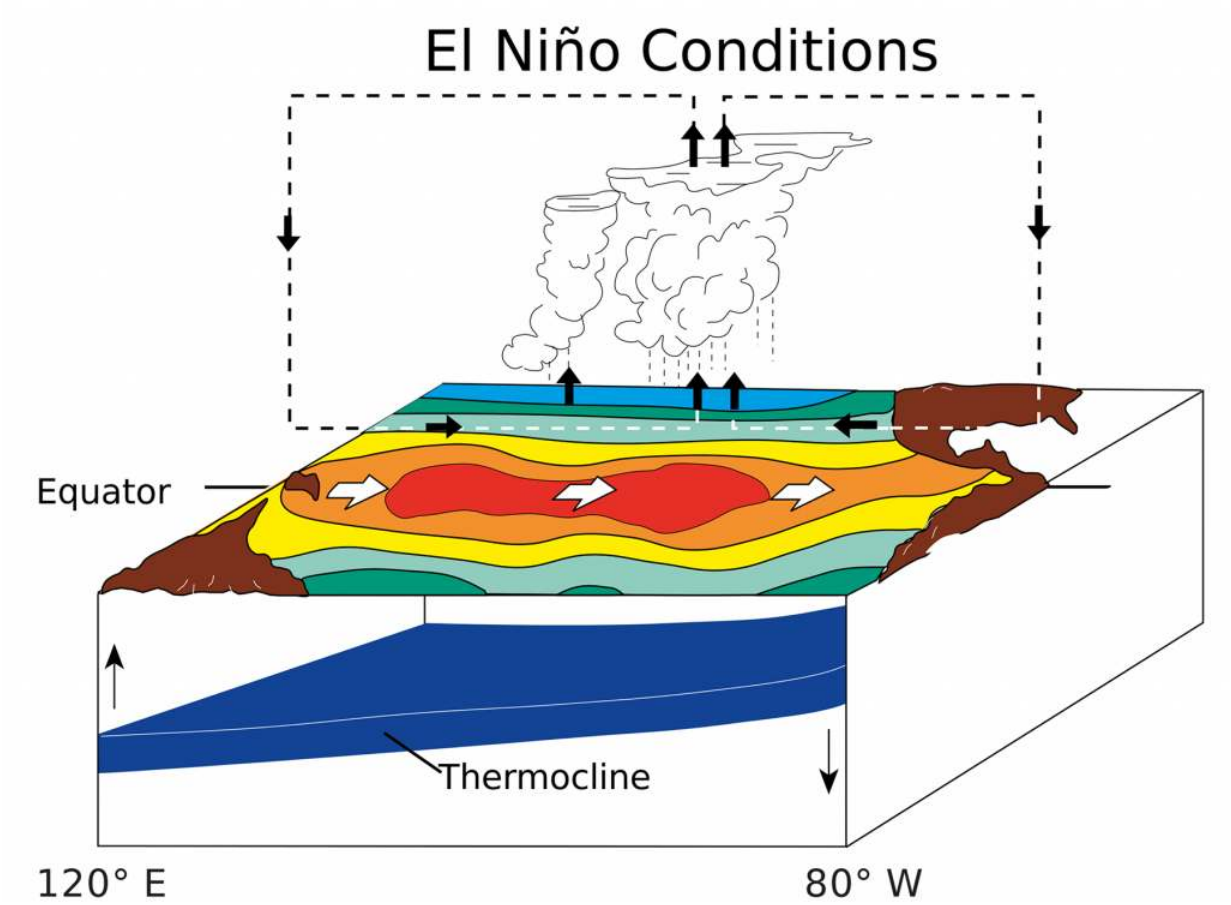


Figure 11.9.2 El Niño conditions in the equatorial Pacific. Weakening or reversal of the trade winds transport warm surface water eastward towards South America, disrupting coastal upwelling (Public domain, via Wikimedia Commons).

In the atmosphere, the low pressure system in the western Pacific is replaced by high pressure, bringing dry or even drought conditions to Southeast Asia and Australia. The low pressure system moves east across the Pacific, potentially reaching as far as South America in strong El Niño years. The low pressure over the eastern Pacific brings lots of rain and flooding to South America (Figure 11.9.2). But the effects of El Niño are not just limited to the Pacific; it can influence weather patterns throughout the globe (see box below).

Because the Southern Oscillation is a cyclic pattern, the eastern Pacific is not subject just to unusually warm conditions. There are also periods of abnormally cold water in the region known as **La Niña** events. During a La Niña the trade winds are unusually strong, leading to increased upwelling and transport of deep, cold water to the surface (Figure 11.9.3). The effects of a La Niña are essentially the opposite of an El Niño, bringing cooler and wetter conditions to the northwestern United States and Canada, while the southeastern US receives below-average precipitation. Monsoon seasons in Asia are drier during El Niños but wetter during La Niña events.

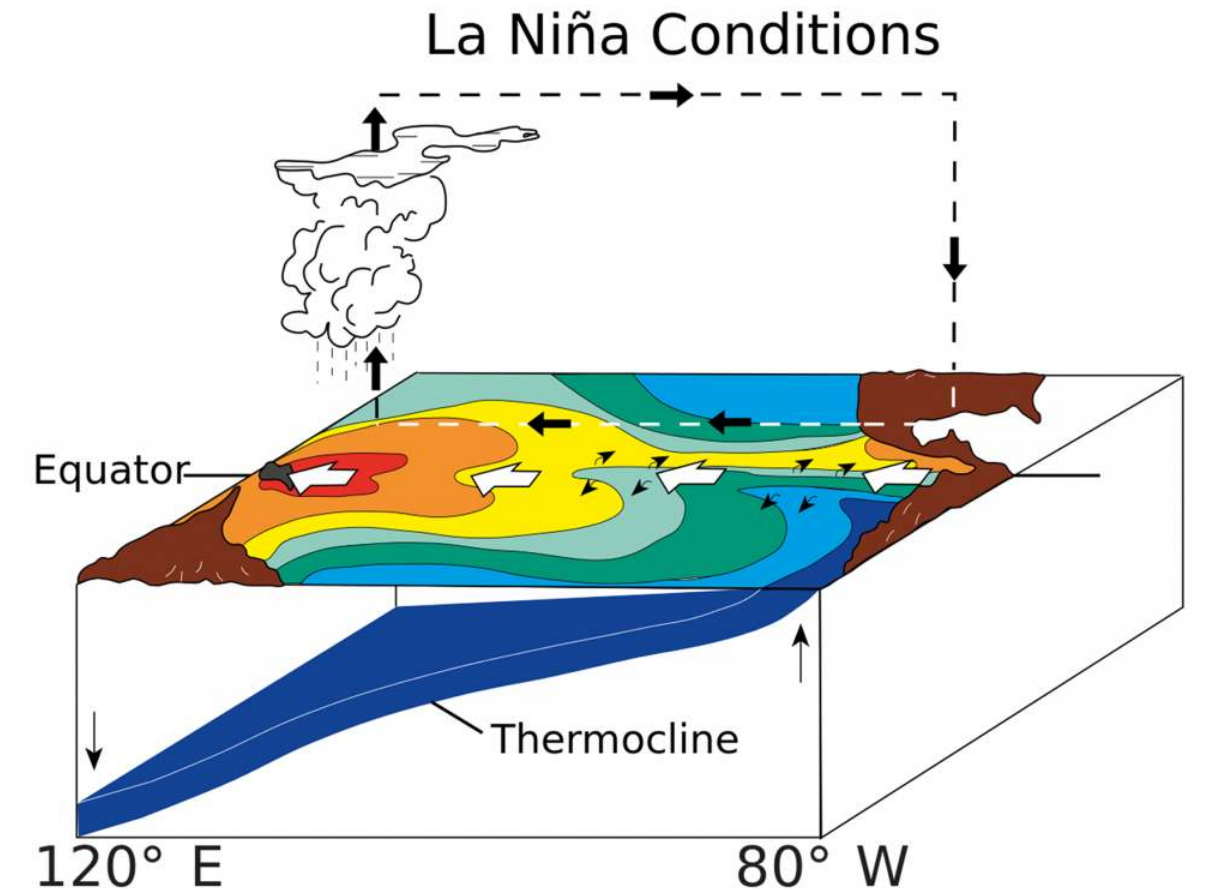


Figure 11.9.3 La Niña conditions. Stronger trade winds promote more intense upwelling in the eastern Pacific, leading to cooler than usual water temperatures (Public domain, via Wikimedia Commons).

El Niño and La Niña events alternate, although the presence of one does not always mean the other will automatically follow. El Niños occur roughly every 2-7 years, and each event may last from a few months to a year or more. Although we do not understand exactly why or when the ENSO events will occur, we can anticipate their arrival by monitoring a number of ocean and atmospheric phenomena that make up the Multivariate ENSO Index (MEI). Examination of the MEI over time demonstrates the cyclic nature of ENSO events (Figure 11.9.4).

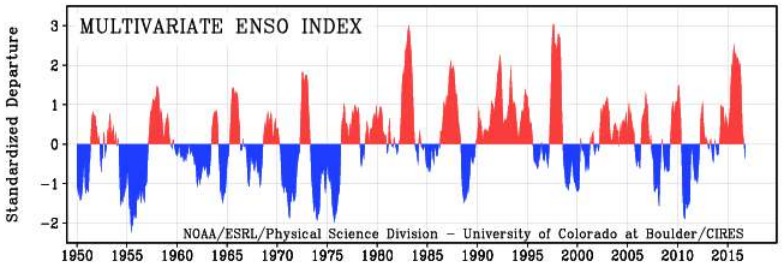


Figure 11.9.4 Multivariate ENSO Index over time. Positive (red) values indicate warmer than normal conditions, while negative (blue) values represent conditions that are cooler than average. The greater the deviation from zero, the stronger the event. Note the intense El Niños in 1983, 1997-1998, and 2015 (NOAA, https://www.esrl.noaa.gov/psd/enso/mei/).

Figure 11.9.5 shows a comparison of sea surface temperatures in the equatorial Pacific during normal, El Niño, and La Niña periods.

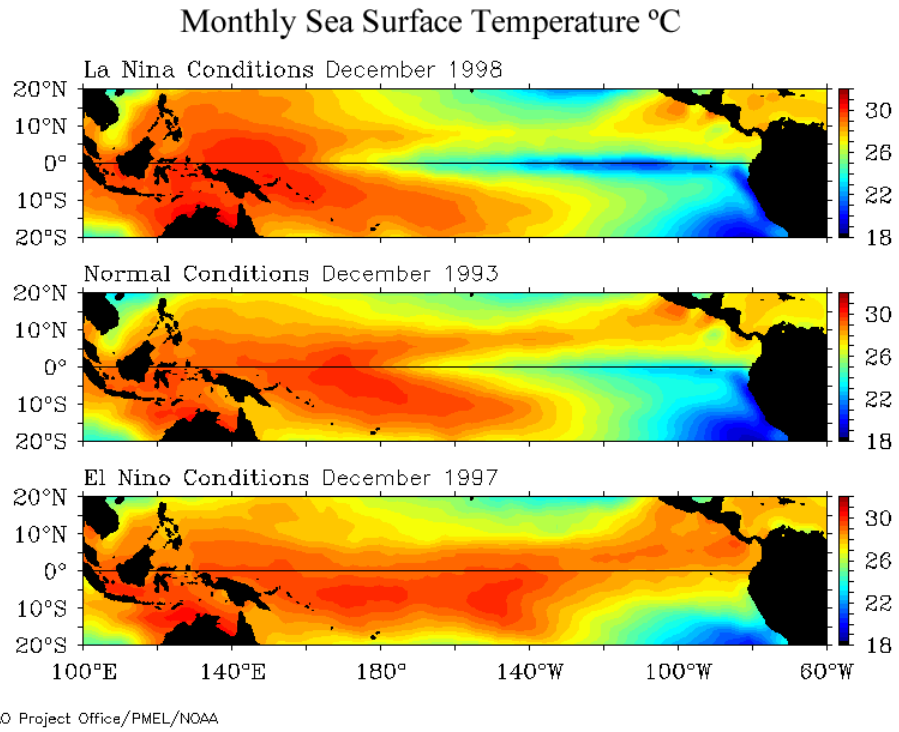


Figure 11.9.5 Comparison of mean December sea surface temperatures in the equatorial Pacific during La Niña (top), normal (middle), and El Niño (bottom) conditions (NOAA, http://www.pmel.noaa.gov/elnino/site...mal-elnino.gif).

Impacts of El Niño

The 2014-2016 El Niño was one of the strongest ENSO events on record (Figure 11.9.4). Some of the recorded global impacts of this El Niño included:

- Widespread droughts in the Philippines and many South Pacific island nations.
- Severe coral bleaching on the Great Barrier Reef in Australia.
- One of the most destructive bushfire seasons in Australia, in part due to low rainfall.
- High rainfall in the southeastern United States and parts of California, leading to flooding.
- Mild, low-precipitation winter in the New England region of the United States.
- Severe flooding in Peru and Argentina.
- Droughts in many portions of southern Africa.
- Nearly 100 million people worldwide suffered a lack of food or water from flooding and droughts.
- Peru suspended its second anchovy fishing season due to low biomass, and an anticipated 20% reduction in the yearly catch.

Additional links for more information

- What are the current ENSO conditions?: https://www.cpc.ncep.noaa.gov/products/precip/CWlink/MJO/enso.shtml

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CHAPTER OVERVIEW

12: Waves

12.1: Waves

12.1.1: Wave Basics

12.1.2: Waves at Sea

12.1.3: Waves on the Shore

12.1.4: Tsunamis

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SECTION OVERVIEW

12.1: Waves

Learning Objectives

After reading this chapter you should:

- know the parts of a basic wave
- know the terminology used to describe the motion of a wave (i.e. period, frequency, speed etc.)
- understand the circular motion of water particles involved in wave motion
- understand the difference between deep water waves and shallow water waves
- know what factors influence wave speed in deep and shallow waves
- know the three factors that determine the energy of wind-generated waves
- understand the concept of restoring force
- understand the difference between seas and swell
- understand the concepts of destructive, constructive and mixed interference
- understand why waves break as they approach shore
- know the differences in the different types of breakers, and how the bottom topography impacts breaker type
- understand why waves always approach parallel to shore, and why waves are larger off of points and smaller in bays
- understand what causes tsunamis, and how they behave in the ocean

What was the largest wave ever recorded? 50 feet? 100 feet? Not even close. That record belongs to a wave created in Lituya Bay, Alaska, on July 9, 1958. On that day, a magnitude 7.8 earthquake caused a massive rockslide that slid down a mountainside and into the headwaters of the bay. The rockslide created a splash wave that was high enough to flatten vegetation up to 1722 ft (525 m) above sea level! The wave then moved through the narrow bay towards the sea, destroying a number of fishing boats along the way. Miraculously, a father and son on one fishing boat were carried above the trees by the wave, and survived to tell the story. This is by far the largest wave, a megatsunami, ever reliably recorded. The waves we will discuss in this chapter may not be quite that dramatic, but it is still important to know how they form, how they are propagated, and what happens to them as they interact with the shore.



A view of Lituya Bay taken a few weeks after the 1958 megatsunami. The rockslide occurred in the mountains at the head of the bay, producing the wave that them moved through the bay towards the sea (D.J. Miller, United States Geological Survey, [Public domain], via Wikimedia Commons).

12.1.1: Wave Basics

12.1.2: Waves at Sea

12.1.3: Waves on the Shore

12.1.4: Tsunamis

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12.1.1: Wave Basics

Waves generally begin as a disturbance of some kind, and the energy of that disturbance gets propagated in the form of waves. We are most familiar with the kind of waves that break on shore, or rock a boat at sea, but there are many other types of waves that are important to oceanography:

- **Internal waves** form at the boundaries of water masses of different densities (i.e. at a pycnocline), and propagate at depth. These generally move more slowly than surface waves, and can be much larger, with heights exceeding 100 m. However, the height of the deep wave would be unnoticeable at the surface.
- **Tidal waves** are due to the movement of the tides. What we think of as tides are basically enormously long waves with a wavelength that may span half the globe (see section 11.1). Tidal waves are not related to tsunamis, so don't confuse the two.
- **Tsunamis** are large waves created as a result of earthquakes or other seismic disturbances. They are also called seismic sea waves (section 10.4).
- **Splash waves** are formed when something falls into the ocean and creates a splash. The giant wave in Lituya Bay that was described in the introduction to this chapter was a splash wave.
- **Atmospheric waves** form in the sky at the boundary between air masses of different densities. These often create ripple effects in the clouds (Figure 12.1.1.1).



Figure 12.1.1.1 Wake patterns in cloud cover over Possession Island, East Island, Ile aux Cochons, Ile de Pingouins. The ripple pattern is a result of internal waves in the atmosphere (NASA [Public domain], via Wikimedia Commons).

There are several components to a basic wave (Figure 12.1.1.2):

- **Still water level:** where the water surface would be if there were no waves present and the sea was completely calm.
- **Crest:** the highest point of the wave.
- **Trough:** the lowest point of the wave.
- **Wave height:** the distance between the crest and the trough.
- **Wavelength:** the distance between two identical points on successive waves, for example crest to crest, or trough to trough.
- **Wave steepness:** the ratio of wave height to length (H/L). If this ratio exceeds 1/7 (i.e. height exceeds 1/7 of the wavelength) the wave gets too steep, and will break.

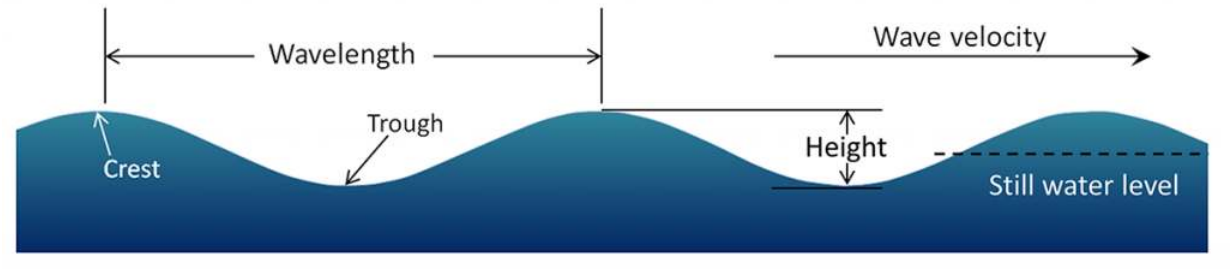


Figure 12.1.1.2 Components of a basic wave (Modified by PW from Steven Earle “Physical Geology”).

There are also a number of terms used to describe wave motion:

- **Period:** the time it takes for two successive crests to pass a given point.
- **Frequency:** the number of waves passing a point in a given amount of time, usually expressed as waves per second. This is the inverse of the period.
- **Speed**: how fast the wave travels, or the distance traveled per unit of time. This is also called celerity (c), where

$$c = \text{wavelength} \times \text{frequency}$$

Therefore, the longer the wavelength, the faster the wave.

Although waves can travel over great distances, the water itself shows little horizontal movement; it is the *energy* of the wave that is being transmitted, not the water. Instead, the water particles move in circular orbits, with the size of the orbit equal to the wave height (Figure 12.1.1.3). This orbital motion occurs because water waves contain components of both longitudinal (side to side) and transverse (up and down) waves, leading to circular motion. As a wave passes, water moves forwards and up over the wave crests, then down and backwards into the troughs, so there is little horizontal movement. This is evident if you have ever watched an object such as a seabird floating at the surface. The bird bobs up and down as the wave pass underneath it; it does not get carried horizontally by a single wave crest.

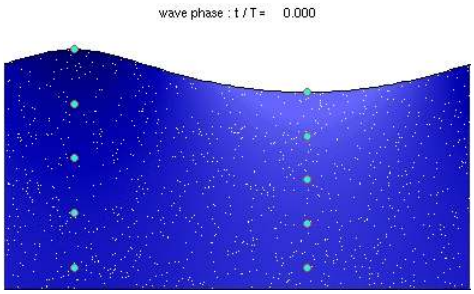


Figure 12.1.1.3 Animation showing the orbital motion of particles in a surface wave (By Kraaiennest (Own work) [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 4.0], via Wikimedia Commons).

The circular orbital motion declines with depth as the wave has less impact on deeper water and the diameter of the circles is reduced. Eventually at some depth there is no more circular movement and the water is unaffected by surface wave action. This depth is the **wave base** and is equivalent to half of the wavelength (Figure 12.1.1.4). Since most ocean waves have wavelengths of less than a few hundred meters, most of the deeper ocean is unaffected by surface waves, so even in the strongest storms marine life or submarines can avoid heavy waves by submerging below the wave base.

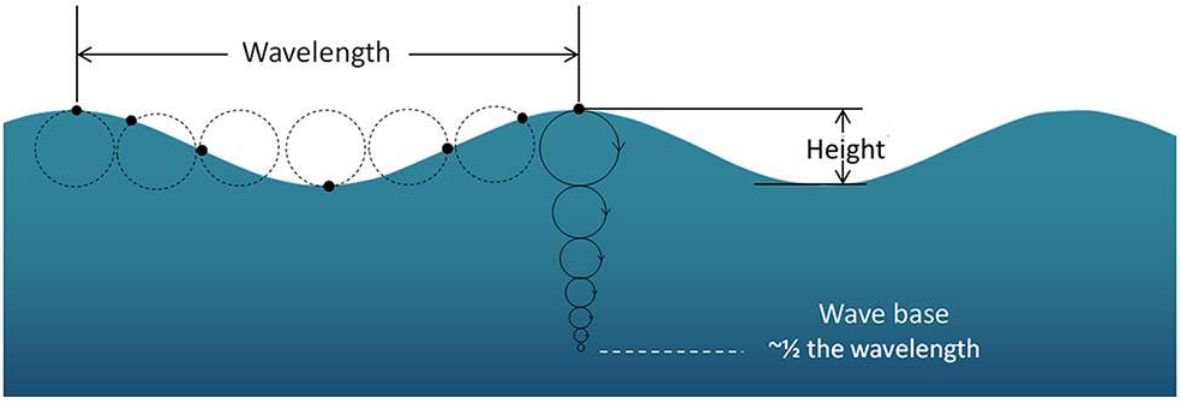
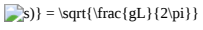
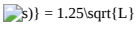


Figure 12.1.1.4 Orbital motion of water within a wave, extending down to the wave base at a depth of half of the wavelength (Modified by PW from Steven Earle, “Physical Geology”).

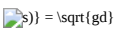
When the water below a wave is deeper than the wave base (deeper than half of the wavelength), those waves are called **deep water waves**. Most open ocean waves are deep water waves. Since the water is deeper than the wave base, deep water waves experience no interference from the bottom, so their speed only depends on the wavelength:



where g is gravity and L is wavelength in meters. Since g and π are constants, this can be simplified to:



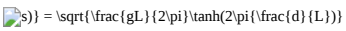
Shallow water waves occur when the depth is less than 1/20 of the wavelength. In these cases, the wave is said to “touch bottom” because the depth is shallower than the wave base so the orbital motion is affected by the seafloor. Due to the shallow depth, the orbits are flattened, and eventually the water movement becomes horizontal rather than circular just above the bottom. The speed of shallow water waves depends only on the depth:



where g is gravity and d is depth in meters. This can be simplified to:



Intermediate or **transitional waves** are found in depths between ½ and 1/20 of the wavelength. Their behavior is a bit more complex, as their speed is influenced by both wavelength and depth. The speed of an intermediate wave is calculated as:



which contains both depth and wavelength variables.

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12.1.2: Waves at Sea

Most ocean waves are generated by wind. Wind blowing across the water's surface creates little disturbances called **capillary waves**, or ripples that start from gentle breezes (Figure 12.1.2.1). Capillary waves have a rounded crest with a V-shaped trough, and wavelengths less than 1.7 cm. These small ripples give the wind something to "grip" onto to generate larger waves when the wind energy increases, and once the wavelength exceeds 1.7 cm the wave transitions from a capillary wave to a wind wave. As waves are produced, they are opposed by a **restoring force** that attempts to return the water to its calm, equilibrium condition. The restoring force of the small capillary waves is surface tension, but for larger wind-generated waves gravity becomes the restoring force.



Figure 12.1.2.1 Small capillary waves or ripples caused by winds blowing over the surface of calm water (By Blue Elf (Own work) [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 3.0], via Wikimedia Commons).

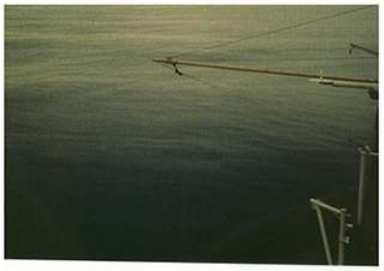
As the energy of the wind increases, so does the size, length and speed of the resulting waves. There are three important factors determining how much energy is transferred from wind to waves, and thus how large the waves will get:

- Wind **speed.**
- The **duration** of the wind, or how long the wind blows continuously over the water.
- The distance over which the wind blows across the water in the same direction, also known as the **fetch**.

Increasing any of these factors increases the energy of wind waves, and therefore their size and speed. But there is an upper limit to how large wind-generated waves can get. As wind energy increases, the waves receive more energy and they get both larger and steeper (recall from section 10.1 that wave steepness = height/wavelength). When the wave height exceeds 1/7 of the wavelength, the wave becomes unstable and collapses, forming whitecaps. Under strong wind conditions, the ocean surface becomes an irregular mixture of choppy, whitecapped wind-generated waves. The term **sea state** describes the size and extent of the wind-generated waves in a particular area. When the waves are at their maximum size for the existing wind speed, duration, and fetch, it is referred to as a fully developed sea. The sea state is often reported on the **Beaufort scale**, ranging from 0-12, where 0 means calm, windless and waveless conditions, while Beaufort 12 is a hurricane (see box below).

The Beaufort Scale

The Beaufort scale is used to describe the wind and sea state conditions on the ocean. It is an observational scale based on the judgement of the observer, rather than one dictated by accurate measurements of wave height. Beaufort 0 represents calm, flat conditions, while Beaufort 12 represents a hurricane.



BEAUFORT FORCE 0
WIND SPEED: LESS THAN 1 KNOT

SEA: SEA LIKE A MIRROR



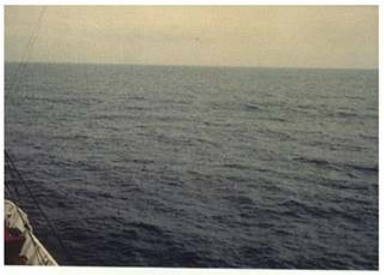
BEAUFORT FORCE 1
WIND SPEED: 1-3 KNOTS

SEA: WAVE HEIGHT .1M (.25FT), RIPPLES WITH THE APPEARANCE OF SCALES, BUT WITHOUT FOAM CRESTS



BEAUFORT FORCE 2
WIND SPEED: 4-6 KNOTS

SEA: WAVE HEIGHT .2-.3M (.5-1FT), SMALL WAVELETS, CRESTS HAVE A GLASSY APPEARANCE AND DO NOT BREAK



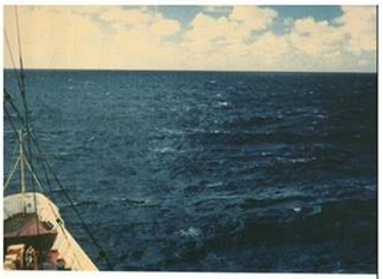
BEAUFORT FORCE 3
WIND SPEED: 7-10 KNOTS

SEA: WAVE HEIGHT .6-1M (2-3FT), LARGE WAVELETS, CRESTS BEGIN TO BREAK, ANY FOAM HAS GLASSY APPEARANCE, SCATTERED WHITECAPS



BEAUFORT FORCE 4
WIND SPEED: 11-16 KNOTS

SEA: WAVE HEIGHT 1-1.5M (3.5-5FT), SMALL WAVES BECOMING LONGER, FAIRLY FREQUENT WHITE HORSES



BEAUFORT FORCE 5
WIND SPEED: 17-21 KNOTS

SEA: WAVE HEIGHT 2-2.5M (6-8FT), MODERATE WAVES TAKING MORE PRONOUNCED LONG FORM, MANY WHITE HORSES, CHANCE OF SOME SPRAY



BEAUFORT FORCE 6
WIND SPEED: 22-27 KNOTS

SEA: WAVE HEIGHT 3-4M (9.5-13 FT), LARGER WAVES BEGIN TO FORM, SPRAY IS PRESENT, WHITE FOAM CRESTS ARE EVERYWHERE



BEAUFORT FORCE 7
WIND SPEED: 28-33 KNOTS

SEA: WAVE HEIGHT 4-5.5M (13.5-19 FT), SEA HEAPS UP, WHITE FOAM FROM BREAKING WAVES BEGINS TO BE BLOWN IN STREAKS ALONG THE WIND DIRECTION



BEAUFORT FORCE 8
WIND SPEED: 34-40 KNOTS

SEA: WAVE HEIGHT 5.5-7.5M (18-25FT), MODERATELY HIGH WAVES OF GREATER LENGTH, EDGES OF CREST BEGIN TO BREAK INTO THE SPINDRIFT, FOAM BLOWN IN WELL MARKED STREAKS ALONG WIND DIRECTION.

https://geo.libretexts.org/@go/page/13032



BEAUFORT FORCE 9
WIND SPEED: 41-47 KNOTS

SEA: WAVE HEIGHT 7-10M (23-32FT). HIGH WAVES, DENSE STREAKS OF FOAM ALONG DIRECTION OF THE WIND, WAVE CRESTS BEGIN TO TOPPLE, TUMBLE, AND ROLL OVER. SPRAY MAY AFFECT VISIBILITY.



BEAUFORT FORCE 10
WIND SPEED: 48-55 KNOTS

SEA: WAVE HEIGHT 9-12.5M (29-41FT). VERY HIGH WAVES WITH LONG OVERHANGING CRESTS. THE RESULTING FOAM, IN GREAT PATCHES, IS BLOWN IN DENSE WHITE STREAKS ALONG WIND DIRECTION. ON THE WHOLE, SEA SURFACE TAKES A WHITE APPEARANCE. TUMBLING OF THE SEA IS HEAVY AND SHOCK-LIKE, VISIBILITY AFFECTED.



BEAUFORT FORCE 11
WIND SPEED: 56-63 KNOTS

SEA: WAVE HEIGHT 11.5-16M (37-52FT). EXCEPTIONALLY HIGH WAVES, SMALL-MEDIUM SIZED SHIPS MAY BE LOST TO VIEW BEHIND THE WAVES. SEA COMPLETELY COVERED WITH LONG WHITE PATCHES OF FOAM LYING ALONG WIND DIRECTION. EVERYWHERE, THE EDGES OF WAVE CRESTS ARE BLOWN INTO FROTH.



BEAUFORT FORCE 12
WIND SPEED: 64 KNOTS

SEA: SEA COMPLETELY WHITE WITH DRIVING SPRAY, VISIBILITY VERY SERIOUSLY AFFECTED. THE AIR IS FILLED WITH FOAM AND SPRAY

(Images by United States National Weather Service (http://www.crh.noaa.gov/mkx/marinefcst.php) [Public domain], via Wikimedia Commons).

A fully developed sea often occurs under stormy conditions, where high winds create a chaotic, random pattern of waves and whitecaps of varying sizes. The waves will propagate outwards from the center of the storm, powered by the strong winds. However, as the storm subsides and the winds weaken, these irregular seas will sort themselves out into more ordered patterns. Recall that open ocean waves will usually be deep water waves, and their speed will depend on their wavelength (section 10.1). As the waves move away from the storm center, they sort themselves out based on speed, with longer wavelength waves traveling faster than shorter wavelength waves. This means that eventually all of the waves in a particular area will be traveling with the same wavelength, creating regular, long period waves called **swell** (Figure 12.1.2.2). We experience swell as the slow up and down or rocking motion we feel on a boat, or with the regular arrival of waves on shore. Swell can travel very long distances without losing much energy, so we can observe large swells arriving at the shore even where there is no local wind; the waves were produced by a storm far offshore, and were sorted into swell as they traveled towards the coast.

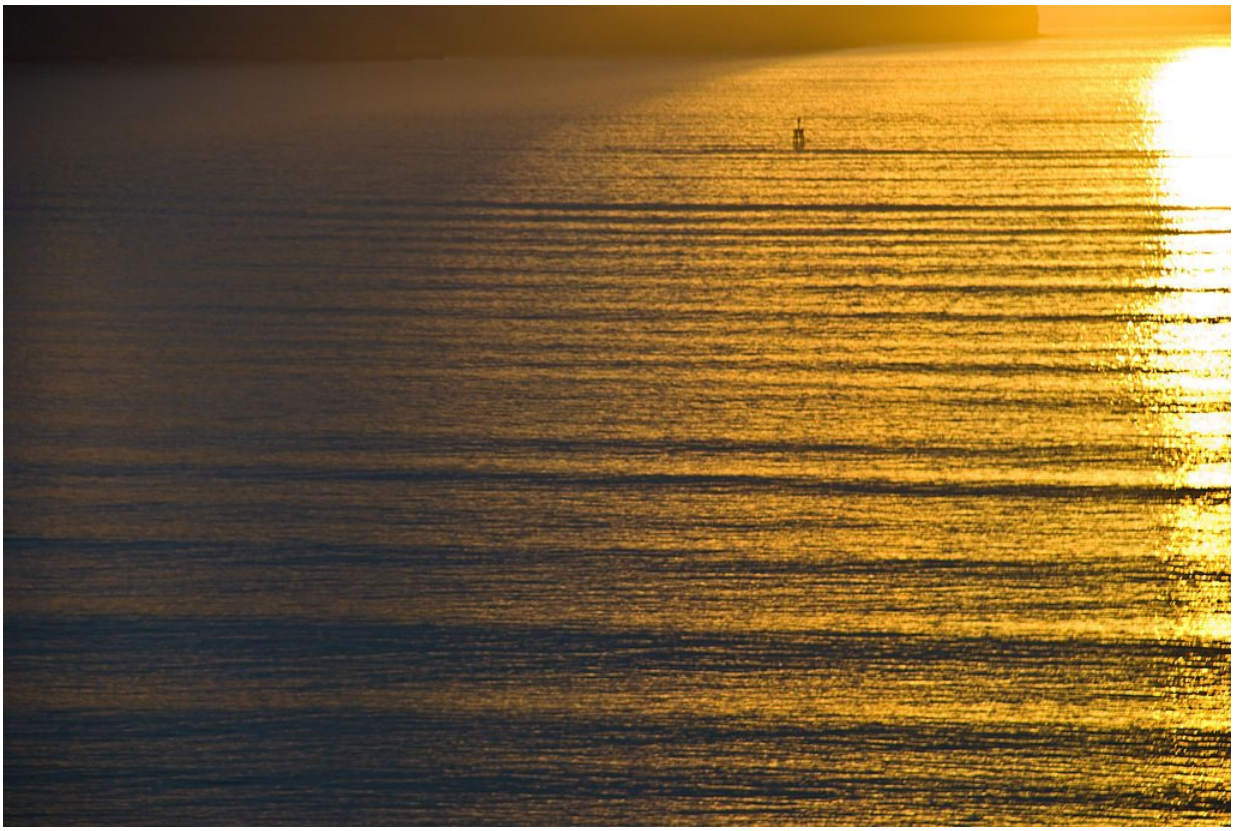


Figure 12.1.2.2 Ocean swell, the regular pattern of waves of equal wavelength (Phillip Capper [CC BY 2.0], via Wikimedia Commons).

Because swell travels such long distances, eventually swells coming from different directions will run into each other, and when they do they create interference patterns. The interference pattern is created by adding the features of the waves together, and the type of interference that is created depends on how the waves interact with each other (Figure 12.1.2.3). **Constructive interference** occurs when the two waves are completely in phase; the crest of one wave lines up exactly with the crest of the other wave, as do the troughs of the two waves. Adding the two crest together creates a crest that is higher than in either of the source waves, and adding the troughs creates a deeper trough than in the original waves. The result of constructive interference is therefore to create waves that are larger than the original source waves. In **destructive interference**, the waves interact completely out of phase, where the crest of one wave aligns with the trough of the other wave. In this case, the crest and the trough work to cancel each other out, creating a wave that is smaller than either of the source waves. In reality, it is rare to find perfect constructive or destructive interference as displayed in Figure 12.1.2.3. Most interference by swells at sea is **mixed interference**, which contains a mix of both constructive and destructive interference. The interacting swells do not have the same wavelength, so some points show constructive interference, and some points show destructive interference, to varying degrees. This results in an irregular pattern of both small and large waves, called **surf beat**.

It is important to point out that these interference patterns are only temporary disturbances, and do not affect the properties of the source waves. Moving swells interact and create interference where they meet, but each wave continues on unaffected after the swells pass each other.

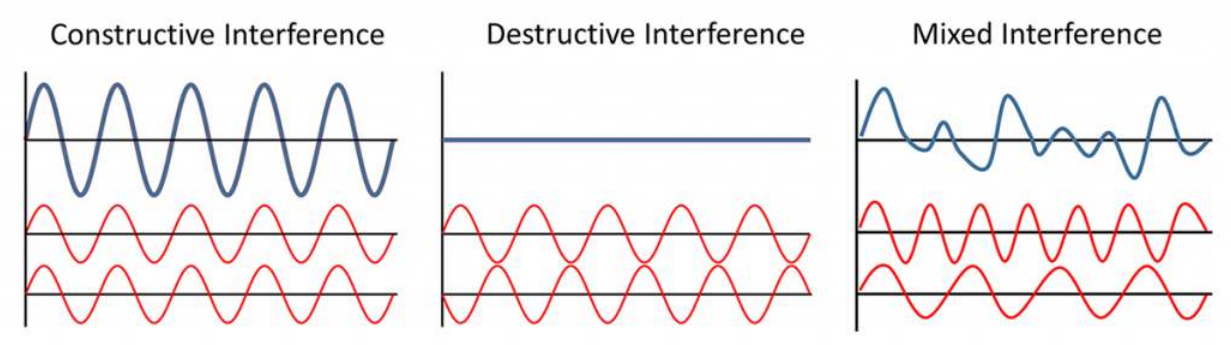


Figure 12.1.2.3 Wave interference patterns. In constructive interference the source waves (red) are completely in phase, and when added together produce waves that are larger than the original waves (blue). In destructive interference the source waves are out of phase, so they cancel each other out and produce waves that are smaller than the originals. In mixed interference, constructive and destructive interference occur at various point, creating an irregular wave pattern. (Modified by PW from original version: Haade; vectorization: Wjh31, Quibik (Vecorized from File:Interference of two waves.png) [CC BY-SA 3.0 or GFDL (http://www.gnu.org/copyleft/fdl.html)], via Wikimedia Commons).

About half of the waves in the open sea are less than 2 m high, and only 10-15% exceed 6 m. But the ocean can produce some extremely large waves. The largest wind wave reliably measured at sea occurred in the Pacific Ocean in 1935, and was measured by the navy tanker the USS Ramapo. Its crew measured a wave of 34 m or about 112 ft high! Occasionally constructive interference will produce waves that are exceptionally large, even when all of the surrounding waves are of normal height. These random, large waves are called **rogue waves** (Figure 12.1.2.4). A rogue wave is usually defined as a wave that is at least twice the size of the significant wave height, which is the average height of the highest one-third of waves in the region. It is not uncommon for rogue waves to reach heights of 20 m or more.

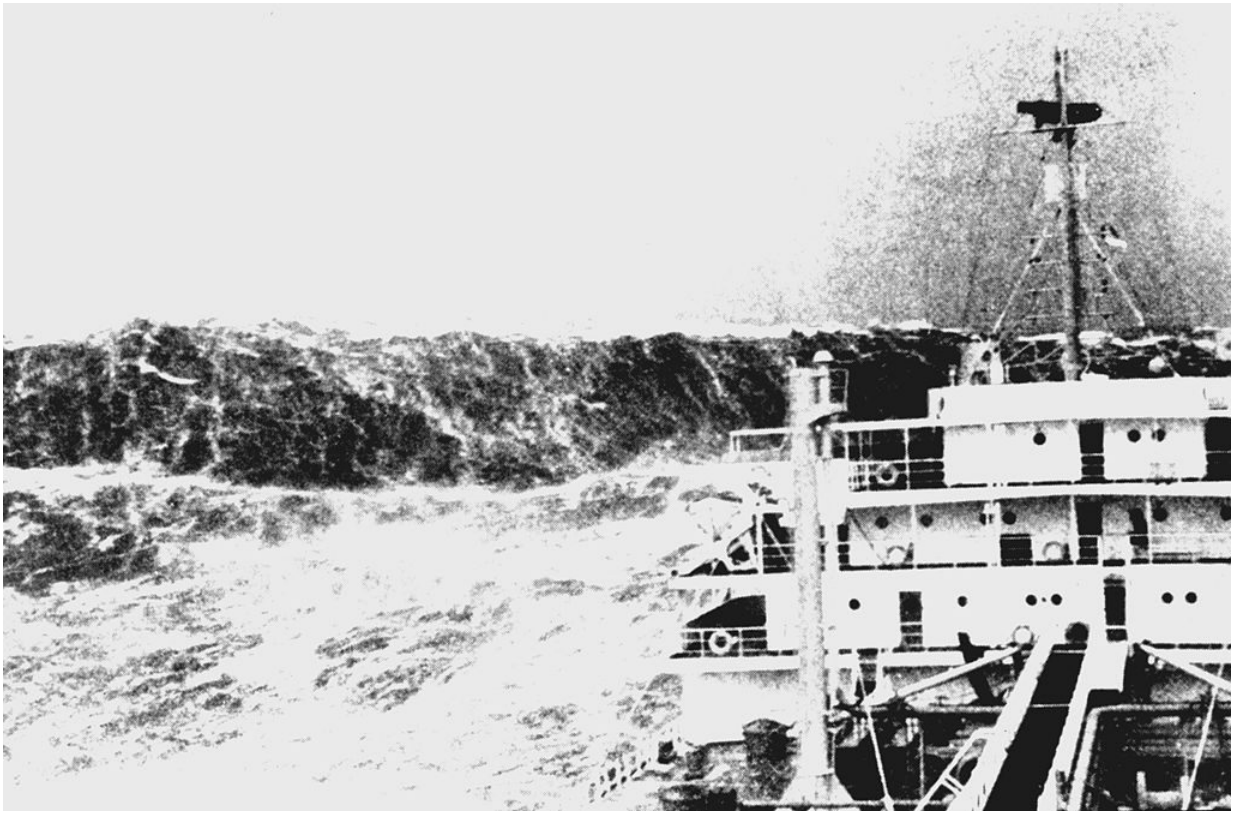


Figure 12.1.2.4 A rogue wave in the Bay of Biscay, off of the French coast, ca. 1940 (NOAA, [Public domain], via Wikimedia Commons).

Rogue waves are particularly common off of the southeast coast of South Africa, a region referred to as the “wild coast.” Here, Antarctic storm waves move north into the oncoming Agulhas Current, and the wave energy gets focused over a narrow area,

leading to constructive interference. This area may be responsible for sinking more ships than anywhere else on Earth. On average about 100 ships are lost every year across the globe, and many of these losses are probably due to rogue waves.

Waves in the Southern Ocean are generally fairly large (the red areas in Figure 12.1.2.2) because of the strong winds and the lack of landmasses, which provide the winds with a very long fetch, allowing them to blow unimpeded over the ocean for very long distances. These latitudes have been termed the “Roaring Forties”, “Furious Fifties”, and “Screaming Sixties” due to the high winds.

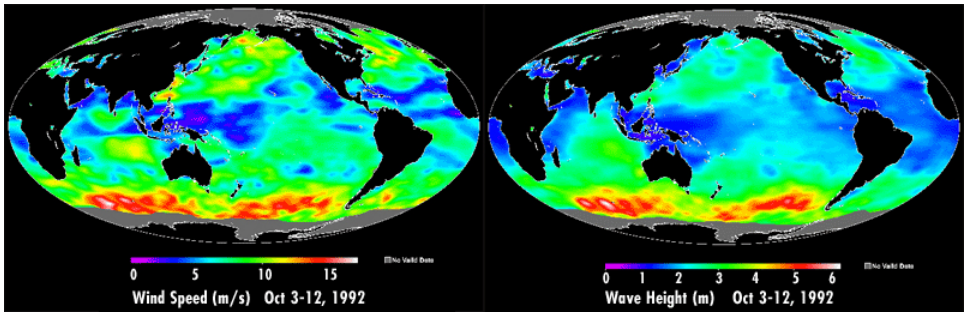


Figure 12.1.2.5 Wind speed and wave height data for a 9-day period in 1992. The Southern Ocean is notorious for its high winds and large waves (NASA, Public Domain via Wikimedia Commons).

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12.1.3: Waves on the Shore

Most of the waves discussed in the previous section referred to deep water waves in the open ocean. But what happens when these waves move towards shore and encounter shallow water? Remember that in deep water, a wave's speed depends on its wavelength, but in shallow water wave speed depends on the depth (section 10.1). When waves approach the shore they will "touch bottom" at a depth equal to half of their wavelength; in other words, when the water depth equals the depth of the wave base (Figure 12.1.3.1). At this point their behavior will begin to be influenced by the bottom.

When the wave touches the bottom, friction causes the wave to slow down. As one wave slows down, the one behind it catches up to it, thus decreasing the wavelength. However, the wave still contains the same amount of energy, so while the wavelength decreases, the wave height increases. Eventually the wave height exceeds 1/7 of the wavelength, and the wave becomes unstable and forms a **breaker**. Often breakers will start to curl forwards as they break. This is because the bottom of the wave begins to slow down before the top of the wave, as it is the first part to encounter the seafloor. So the crest of the wave gets "ahead" of the rest of the wave, but has no water underneath it to support it (Figure 12.1.3.1).

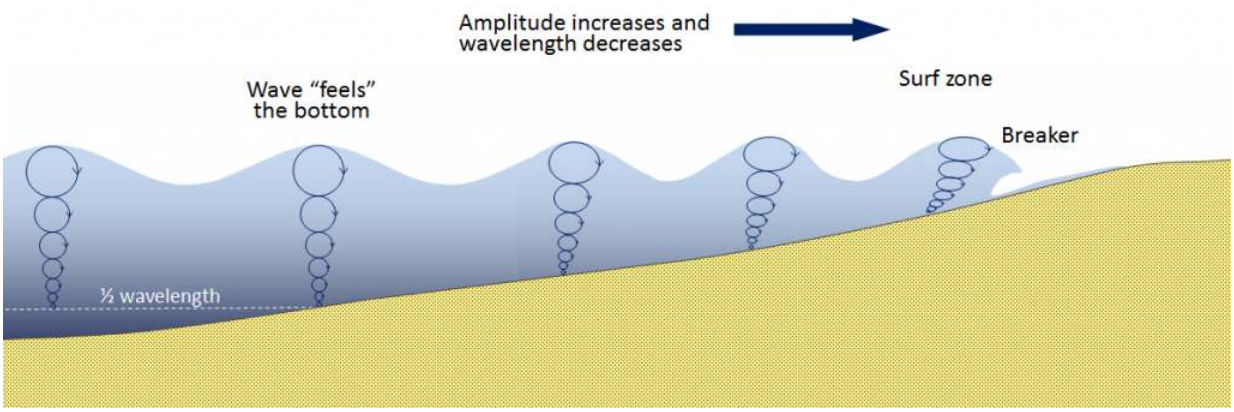


Figure 12.1.3.1 As waves approach shore they "touch bottom" when the depth equals half of the wavelength, and the wave begins to slow down. As is slows, the wavelength decreases and the wave height increases, until the wave breaks (Steven Earle "Physical Geology").

There are three main types of breakers: spilling, plunging, and surging. These are related to the steepness of the bottom, and how quickly the wave will slow down and its energy will get dissipated.

- **Spilling** breakers form on gently sloping or flatter beaches, where the energy of the wave is dissipated gradually. The wave slowly increases in height, then slowly collapses on itself (Figure 12.1.3.2). For surfers, these waves provide a longer ride, but they are less exciting.

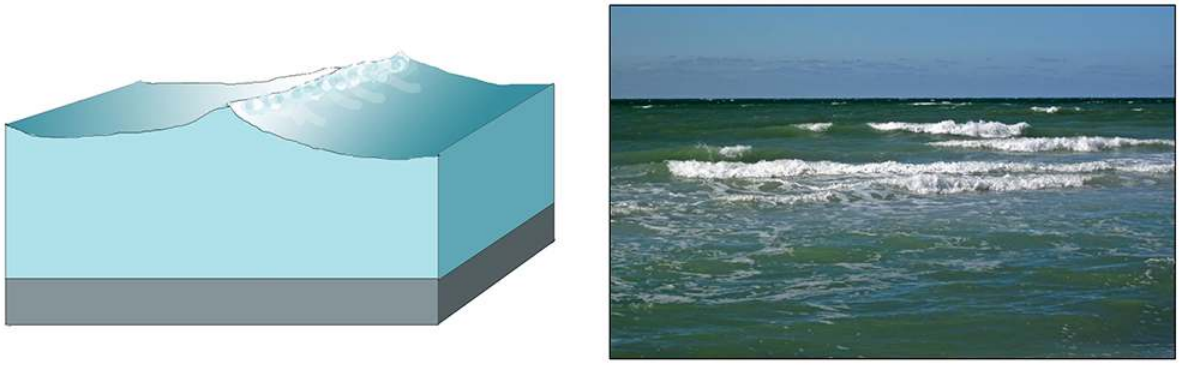


Figure 12.1.3.2 A spilling breaker. The gentle slope of the bottom causes the wave height to slowly increase until the wave collapses on itself (left: JR, right: James St. John, [CC-BY-2.0], https://www.flickr.com/photos/jsjgeology/23769708334).

- **Plunging** breakers form on more steeply-sloped shores, where there is a sudden slowing of the wave and the wave gets higher very quickly. The crest outruns the rest of the wave, curls forwards and breaks with a sudden loss of energy (Figure 12.1.3.3). These are the "pipeline" waves that surfers seek out.

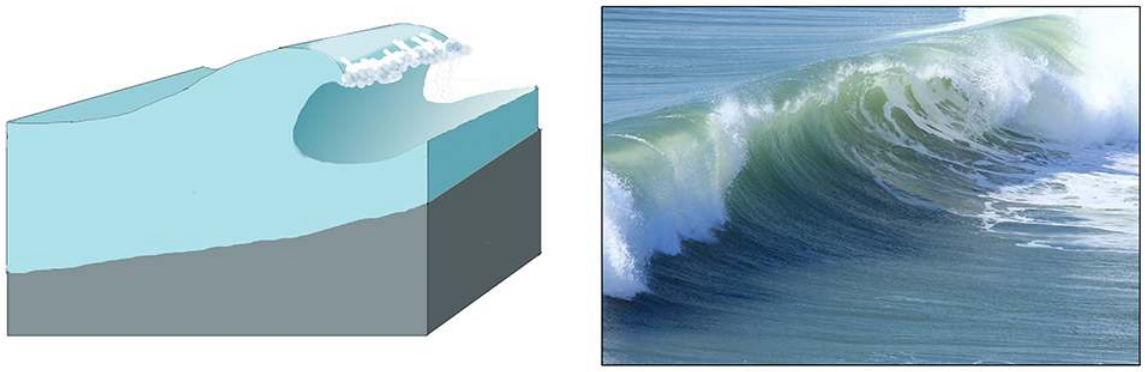


Figure 12.1.3.3 A plunging breaker. The steeper slope causes the wave height to increase more rapidly, with the crest of the wave outrunning the base of the wave, causing it to curl as it breaks (left: JR, right: Andrew Schmidt, Public Domain [CC-0], publicdomainpictures.net).

- **Surging** breakers form on the steepest shorelines. The wave energy is compressed very suddenly right at the shoreline, and the wave breaks right onto the beach (Figure 12.1.3.4). These waves give too short (and potentially painful) a ride for surfers to enjoy.

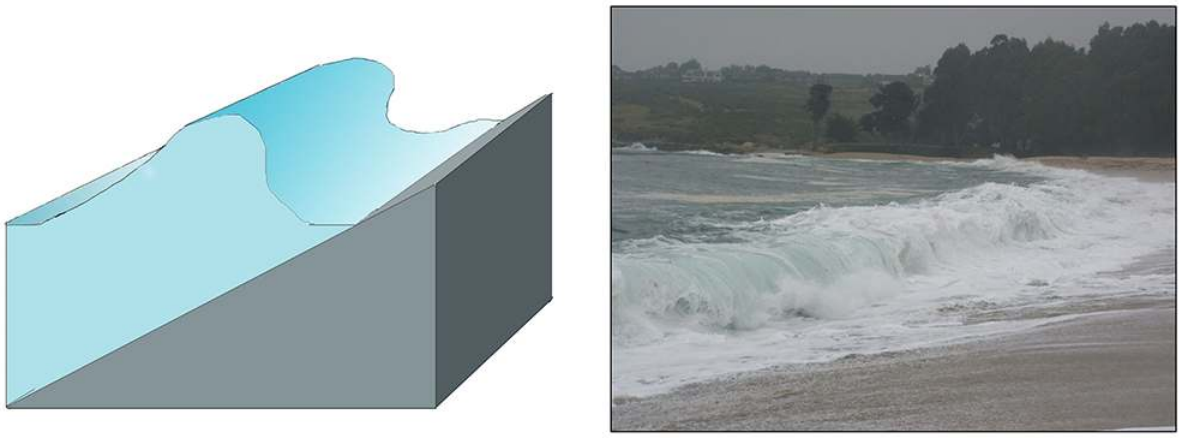


Figure 12.1.3.4 A surging breaker. The very steep slope causes the wave height to increase suddenly and break right on the beach (left: JR, right: Tewy, [CC-BY-SA-3.0], via Wikimedia Commons).

Wave Refraction

Swell can be generated anywhere in the ocean and therefore can arrive at a beach from almost any direction. But if you have ever stood at the shore you have probably noticed that the waves usually approach the shore somewhat parallel to the coast. This is due to wave refraction. If a wave front approaches shore at an angle, the end of the wave front closest to shore will touch bottom before the rest of the wave. This will cause that shallower part of the wave to slow down first, while the rest of the wave that is still in deeper water will continue on at its regular speed. As more and more of the wave front encounters shallower water and slows down, the wave font refracts and the waves tend to align themselves nearly parallel to the shoreline (they are refracted towards the region of slower speed). As we will see in section 13.2, the fact that the waves do not arrive perfectly parallel to the beach causes longshore currents and longshore transport that run parallel to the shore.

Refraction can also explain why waves tend to be larger off of points and headlands, and smaller in bays. A wave front approaching shore will touch the bottom off of the point before it touches bottom in a bay. Once again, the shallower part of the wave front will slow down, and cause the rest of the wave front to refract towards the slower region (the point). Now all of the initial wave energy is concentrated in a relatively small area off of the point, creating large, high energy waves (Figure 12.1.3.6). In the bay, the refraction has caused the wave fronts to refract away from each other, dispersing the wave energy, and leading to calmer water and smaller waves. This makes the large waves of a “point break” ideal for surfing, while water is calmer in a bay, which is where people would launch a boat. This difference in wave energy also explains why there is net erosion on points, while sand and sediments get deposited in bays (see section 13.3).

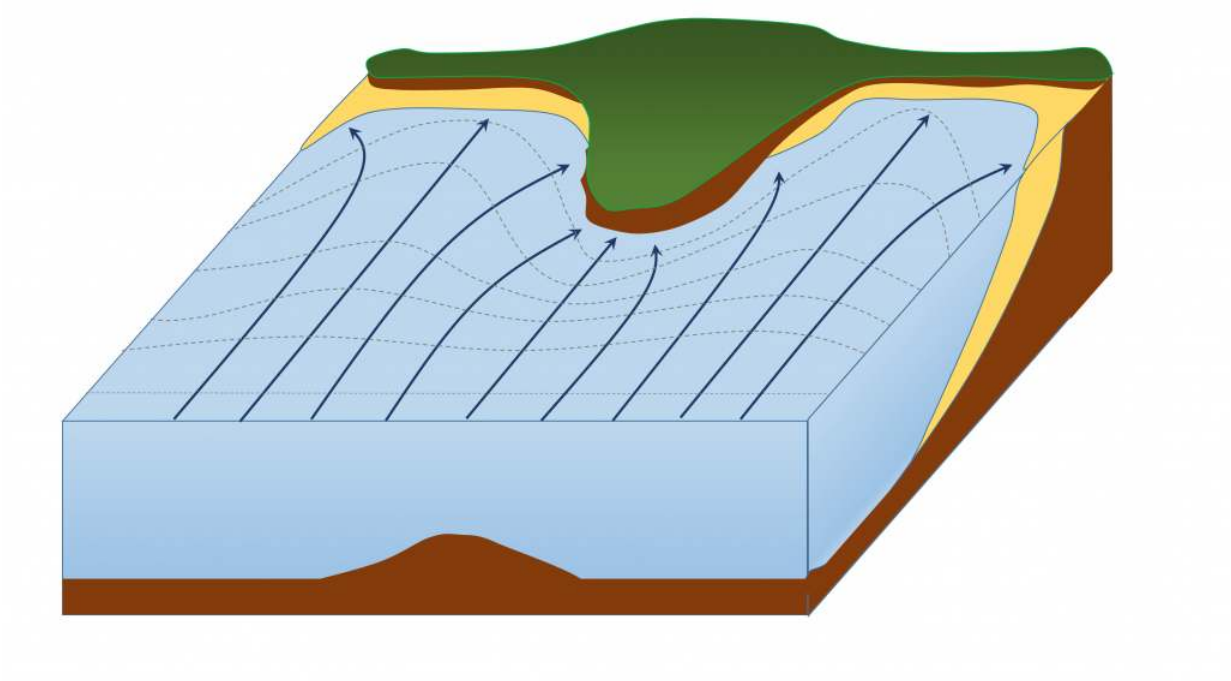


Figure 12.1.3.6 Waves approaching shore (blue lines) touch bottom sooner off of points and are refracted towards the points, concentrating their wave energy. Wave energy is spread out in bays, causing smaller waves. Dotted lines represent the bottom contours (PW).

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12.1.4: Tsunamis

Tsunamis loom large in popular culture, but there are a number of misconceptions about these large waves. First, tsunamis have nothing to do with the tides, so it is a misnomer to refer to them as “tidal waves.” There are actual tidal waves, but they are not related to tsunamis. Second, the giant, curling wave that is taller than skyscrapers and destroys cities in science fiction movies is also a fabrication, as tsunamis do not behave that way, as described below.

Tsunamis are large waves that are usually the result of seismic activity, such as the rising or falling of the seafloor due to earthquakes, although volcanic activity and landslides can also cause tsunamis in the form of splash waves. As the seafloor rises or falls, so does the water column above it, creating waves. Only vertical seismic disturbances cause tsunamis, not horizontal movements. These vertical seafloor movements are usually less than 10 m high, so the resulting wave will be of an equal or lesser height at sea. While the tsunamis have a relatively small height at the point of origin, they have very long wavelengths (100-200 km). Because of the long wavelength, they behave as shallow water waves throughout the entire ocean; the depth of the ocean is always shallower than half of their wavelength. As shallow water waves, their speed depends on water depth, but they can still travel at speeds over 750 km/hr (Figure 12.1.4.1)!

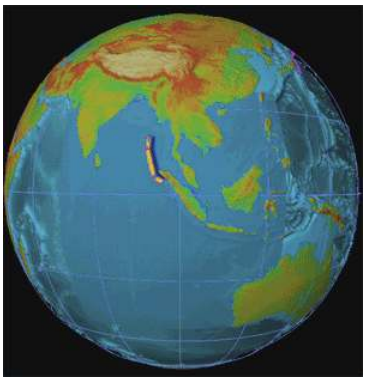


Figure 12.1.4.1 Animation of the spread of tsunamis created during the 2004 Indonesia earthquake (NOAA Center for Tsunami Research (NCTR) [Public domain]).

When tsunamis approach land, they behave just like any other wave; as the depth becomes shallower, the waves slow down and the wave height begins to increase. However, contrary to popular belief, tsunamis do not arrive on shore as giant, cresting waves. Since their wavelength is so long, it is impossible for their height to ever exceed 1/7 of their wavelength, so the waves don’t actually curl or break. Instead, they usually hit the shore as sudden surges of water causing a very rapid increase in sea level, like that of an enormous rise in tide. It may take several minutes for the wave to pass, during which time sea level can rise to 40 m higher than usual.

Large tsunamis occur every 2-3 years, with very large, damaging events happening every 15-20 years. The most devastating tsunami in terms of loss of life resulted from a magnitude 9 earthquake in Indonesia in 2004 (Figure 12.1.4.2), which created waves up to 33 m tall and left about 230,000 people dead in Indonesia, Thailand, and Sri Lanka. In 2011 a 9.0 magnitude earthquake in Japan triggered a tsunami up to 40.5 m high, which resulted in over 18,000 deaths. This earthquake also caused the Fukishima nuclear accident, and moved Japan about 8 inches closer to the U.S.



Figure 12.1.4.2 A village in Sumatra following the Indonesia Tsunami in December 2004 (U.S. Navy photo by Photographer's Mate 2nd Class Philip A. McDaniel [Public domain via Wikimedia Commons]).

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CHAPTER OVERVIEW

13: Tides

13.1: Tides

13.1.1: Tidal Forces

13.1.2: Dynamic Theory of Tides

13.1.3: Tide Classification

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13.1: Tides

Learning Objectives

After reading this chapter you should:

- understand that tides are just very long waves, with crests and troughs
- understand Newton's Law of Universal Gravitation and how it applies to tides
- understand why most places on Earth experience two tides per day, not just the one predicted from gravitational attraction between the Earth and moon (i.e. inertial force)
- understand how the Earth, sun and moon interact to create spring and neap tides
- understand why the gravitational pull of the sun on tides is less than the pull of the moon
- understand why tides do not occur at the same time every day
- understand why amphidromic circulation occurs as a result of tides
- know the difference between diurnal, semi-diurnal, and mixed tides
- know the phases of a tidal current
- know what causes a tidal bore

The previous chapter discussed various types of waves at sea and along the shore. However, at least in terms of wavelength the largest waves in the ocean are the tides, where one wavelength stretches halfway around the Earth. The crests of these long waves represent the high tides, while the troughs create low tides.

You probably learned when you were younger that the basic cause of the tides is the gravitational attraction between the Earth and moon. This is a very old idea, as the Greek scientist Pytheas first made the connection between the tides and the moon back in 300 B.C. However, we now know that the tides involve a lot more than just the Earth and the moon, and it was Isaac Newton's work with gravitational forces in the 1600s that led to our modern understanding of tidal cycles.

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13.1.1: Tidal Forces

Our modern understanding of tide formation stems from Isaac Newton'sLaw of Universal Gravitation, which states that any two objects have a gravitational attraction to each other. The magnitude of the force is proportional to the masses of the objects, and inversely proportional to the square of the distance between the objects, according to the equation in Figure 13.1.1.1.

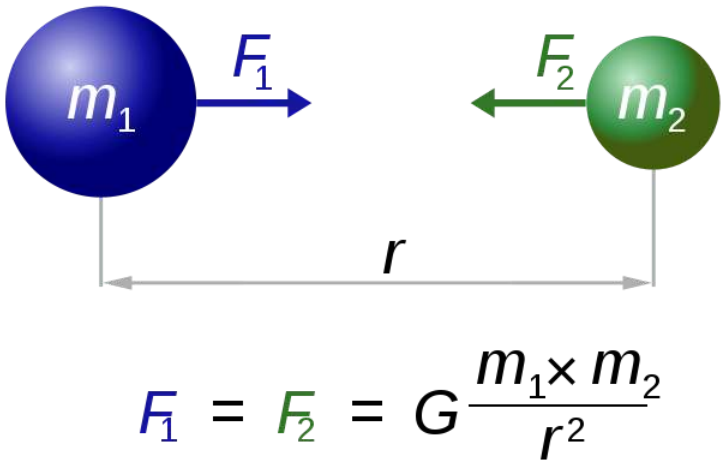


Figure 13.1.1.1 Newton's Law of Universal Gravitation. The gravitational force between two objects (F) is calculated as the product of the two masses (m_1 and m_2) divided by the distance between them (r) squared. G is the universal gravitational constant; 6.67408×10^{-11} m^3 kg^{-1} s^{-2} (By I, Dennis Nilsson [CC BY 3.0], via Wikimedia Commons).

In the case of tides, there are a few other factors that modify this equation so that the distance (r) is cubed rather than squared, giving distance an even greater impact on tidal forces. But for our purposes, the important lesson is that the greater the masses of the objects, the greater the gravitational force, and the farther the objects are from each other, the weaker the force.

Such a gravitational force exists between the Earth and moon, attempting to pull them towards each other. Since the water covering Earth is fluid (unlike the solid land that is more resistant to tidal forces), this gravitational force pulls water towards the moon, creating a "bulge" of water on the side of the Earth facing the moon (Figure 13.1.1.2). This bulge always faces the moon, while the Earth rotates through it; the regions of Earth moving through the bulge experience a high tide, while those parts of the Earth away from the bulge experience a low tide.

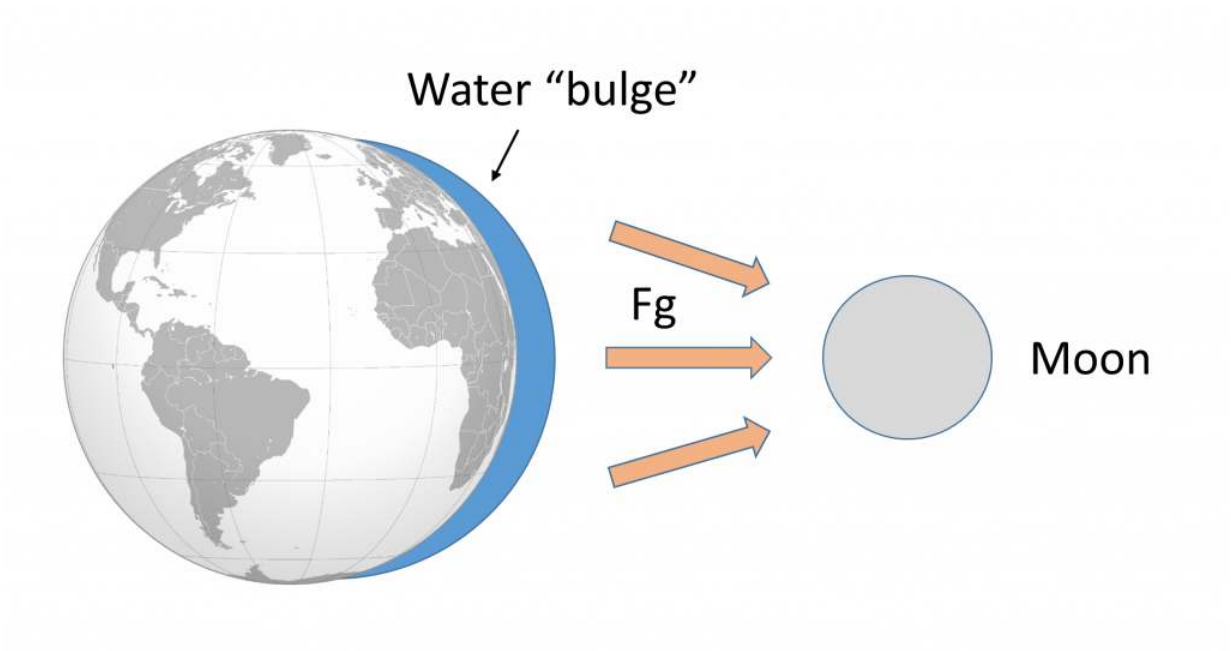


Figure 13.1.1.2 Gravitational forces between the Earth and moon cause a bulge of water to appear on the side of the Earth facing the moon (PW).

If the tides were this simple, everywhere on Earth would see one high tide per day, as there would only be a bulge of water on the side closest to the moon. However, if you have ever looked at tide charts, or lived near the ocean, you probably know that in most places there are two high tides and two low tides per day. Where is this second high tide “bulge” coming from?

The gravitational force between the Earth and moon might be expected to draw the two objects closer together, however, this is not happening. This is because the inward gravitational force is opposed by outward forces that keep the Earth and moon apart. The outward force is an **intertial force** created by the rotation of the Earth and moon. Contrary to popular belief, the moon is not simply rotating around the Earth; in fact, the Earth and moon are both rotating around each other. Imagine the Earth and moon as equal-sized objects revolving around a point at their center of mass. If both objects had the same mass, the center of rotation would be a point equidistant between the two objects. But since the mass of the Earth is 82 times greater than the mass of the moon, the center of revolution must be closer to the Earth. As an analogy, think about two people on a see-saw. If the people are of roughly equal size, they can sit on either end of the see-saw at it will rotate around a point at equal distance between them. But if the two people have very different masses, such as a large adult and a small child, the larger person must move closer to the pivot point for the see-saw to rotate effectively. In the same way, the center of rotation between the Earth and the moon (the **barycenter**) must be located closer to the Earth. In fact, the center of rotation lies *within* the Earth, about 1600 km below the surface. As the Earth and moon rotate around the barycenter, the moon travels much farther than the Earth, giving the impression that the moon is rotating around Earth (Figure 13.1.1.3).

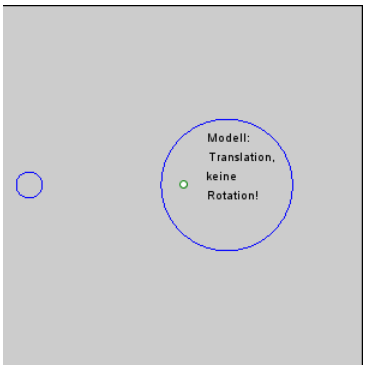


Figure 13.1.1.3 Rotation of the Earth and moon around the barycenter (white dot). Because of its larger mass, the center of rotation lies closer to the Earth than to the moon (By Modalanalytiker (Own work) [CC BY-SA 3.0 or GFDL (http://www.gnu.org/copyleft/fdl.html)], via Wikimedia Commons).

The rotation of the Earth-moon system creates an outward inertial force, which balances the gravitational force to keep the two bodies in their orbits. The inertial force has the same magnitude everywhere on Earth, and is always directed away from the moon. Gravitational force, on the other hand, is always directed towards the moon, and is stronger on the side of the Earth closest to the moon. Figure 13.1.1.4 describes how these forces combine to create the tidal forces. At point O in the center of the Earth, the gravitational force (F_g) and the inertial force (F_r) are equal, and cancel each other out. On the side of Earth closest to the moon, the inward gravitational force (F_g) is greater than the outward inertial force (F_r); the net resulting force (A) is directed towards the moon, and creates a bulge of water on the side facing the moon. On the side of Earth opposite the moon, the outward inertial force is greater than the inward gravitational force; the net resulting force (C) is directed away from the moon, creating a water bulge directed away from the moon.

Now, as the Earth rotates through a 24 hour day, each region passes through two bulges, and experiences two high tides and two low tides per day. This represents Newton's **Equilibrium Theory of Tides,** where there are two high tides and two low tides per day, of similar heights, each six hours apart. But as with everything else in oceanography, reality is much more complex than this idealized situation.

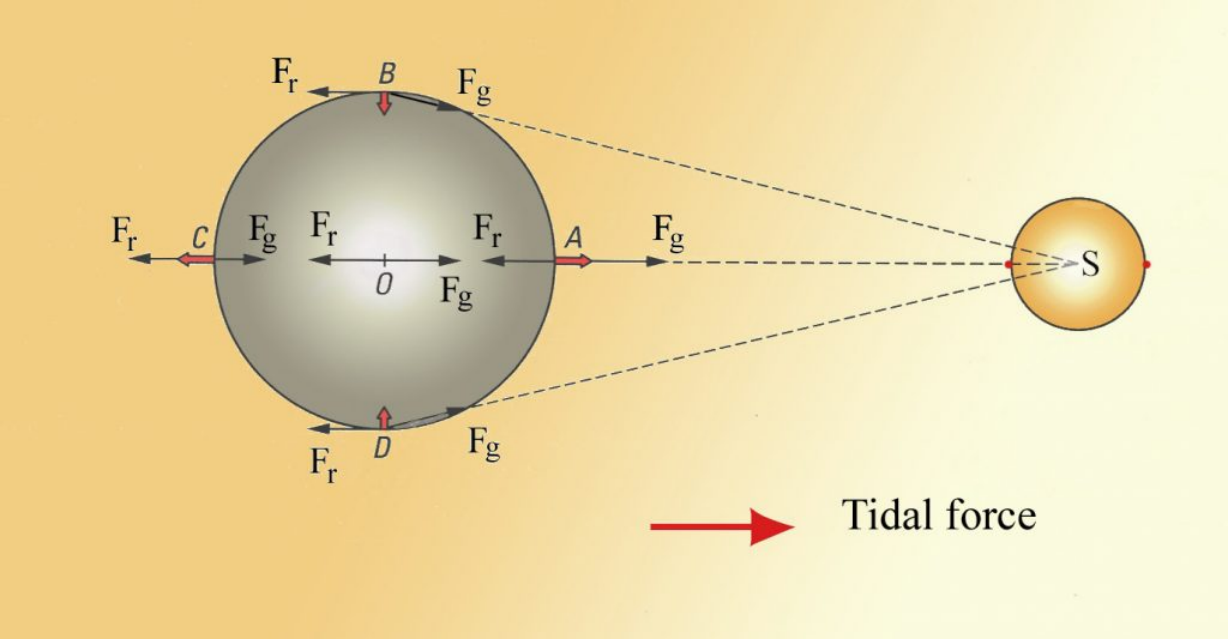


Figure 13.1.1.4 Gravitational force (F_g) are strongest closer to the moon and weaker opposite the moon. Inertial forces (F_r) are equal throughout the Earth and directed away from the moon. The tidal forces A and C are the result of the interaction between F_g and F_r and create water bulges on both sides of the Earth, leading to two high tides per day (Vitold Muratov (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

Some of the additional complexity is because in addition to the moon, the sun also exerts tide-affecting forces on Earth. The solar gravitational and inertial forces arise for the same reasons described above for the moon, but the magnitudes of the forces are different. The sun is 27 million times more massive than the moon, but it is 387 times farther away from the Earth. Despite its larger mass, because the sun is so much farther away than the moon, the sun's gravitational forces are only about half as strong as the moon's (remember that distance is cubed in the gravity equation). The sun thus creates its own, smaller water bulges, independent of the moon's, that contribute to the creation of tides.

When the sun, Earth and moon are aligned, as occurs during new and full moons, the solar and lunar bulges are also aligned, and add to each other (constructive interference; see section 10.2) creating an especially high tidal range; high high tides and low low tides (Figure 13.1.1.5). This period of maximum tidal range is called a **spring tide**, and they occur every two weeks.

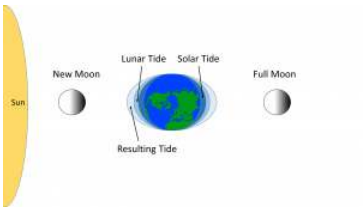


Figure 13.1.1.5 Spring tides with high tidal ranges occur when the solar and lunar tides are added together during full and new moons when the Earth, sun and moon are aligned (PW).

When the sun, Earth and moon are at 90° to each other, the solar and lunar bulges are out of phase, and cancel each other out (destructive interference). Now the tidal range is small, with low high tides and high low tides (Figure 13.1.1.6). These are **neap tides**, and occur every two weeks, when the moon is in its 1/4 and 3/4 phases (Figure 13.1.1.7).

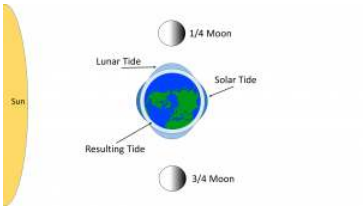


Figure 13.1.1.6 Neap tides are created during 1/4 and 3/4 moons when the Earth, sun and moon are perpendicular to each other. The solar and lunar tides cancel each other out, resulting in a small tidal range (PW).

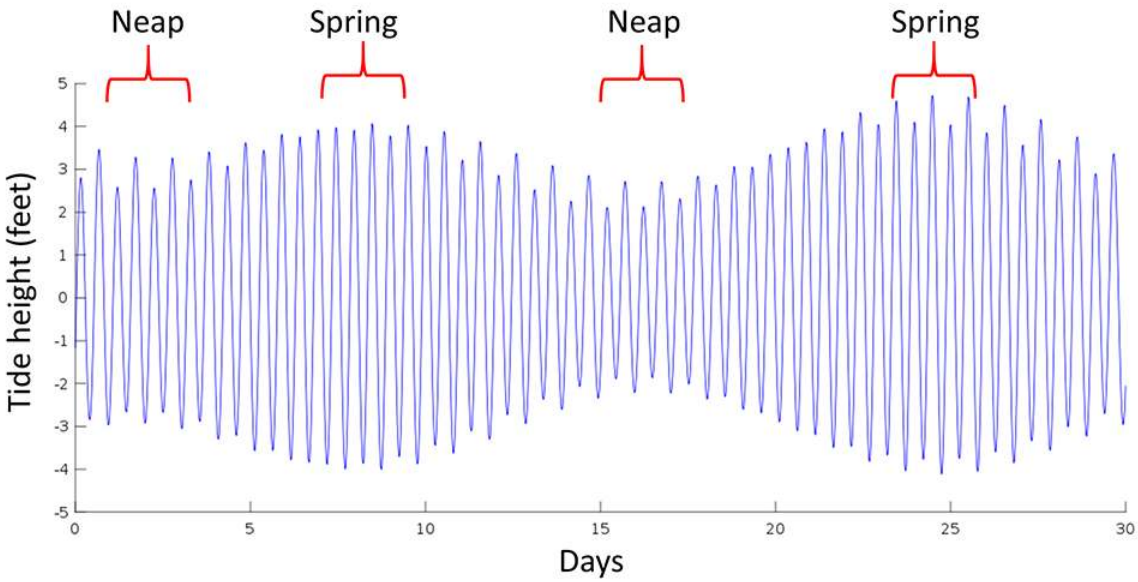


Figure 13.1.1.7 30 days of tidal data from Bridgeport, CT, USA, showing spring and neap tidal ranges (Modified by PW from Cody Logan (clpo13), Public Domain via Wikimedia Commons).

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13.1.2: Dynamic Theory of Tides

The Equilibrium Theory of tides predicts that each day there will be two high and two low tides, each one occurring at the same time day after day, with each pair producing tides of similar heights. While this view provides a basic explanation for the primary forces that generate the tides, it does not take into account such variables as the effects of the continents, the depth of the water, and many other factors. In all, there are almost 400 variables that must be incorporated into predicting the tides! The **Dynamic Theory** of tides takes these other factors into account, and shows that the tides are much more complicated and variable from place to place than the Equilibrium Theory would suggest. For example, some areas receive only one high and one low tide per day (see section 11.3). Furthermore, the tidal range varies greatly across the globe; in the Mediterranean Sea, there can be a difference of only 10 cm between high and low tides, while the Bay of Fundy in Canada experiences a tidal range of up to 17m (56 ft) every day (Figure 13.1.2.1).



Figure 13.1.2.1 Tidal range in the Bay of Fundy, Canada. Both photographs were taken on the same day in July 2003 (By Dylan Kereluk from White Rock, Canada (Flickr) [CC BY 2.0], via Wikimedia Commons).

Examination of any tide chart will show that the tides don't occur at same time each day; in fact, each tidal peak occurs about 50 minutes later than it did in the previous day. This is due to the orbit of the moon around the Earth. Imagine a high tide that occurs at a particular location (X) at 1:00 pm (Figure 13.1.2.2). The high tide occurs as location X moves through the bulge of water facing the moon. It will take the Earth 24 hours to complete one revolution, to bring location X back to site of the water bulge that caused that high tide. However, during those 24 hours, the moon has also moved as it orbits the Earth, so the high tide bulge has moved beyond its original location. The Earth thus has to rotate an additional distance for location X to reach the bulge and experience that same high tide. Because it takes the moon about 28 days to orbit the Earth, the moon gets "ahead" of the Earth's rotation by about 50 minutes per day. Therefore, it takes location X 24 hours and 50 minutes to rotate through the same tidal bulge, and as a result, the tidal peaks occur about 50 minutes later each day. In our example, an afternoon high tide at 1:00 pm on one day would be followed by a high tide at about 1:50 pm the following day. This 24 hour and 50 minute cycle is referred to as a **tidal day**.

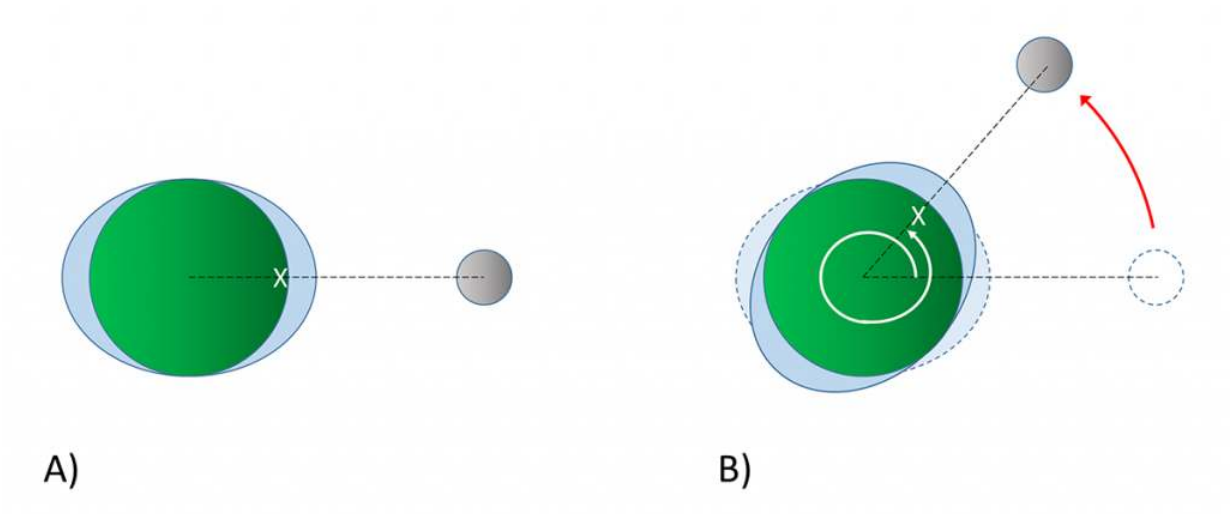


Figure 13.1.2.2 In A) a high tide is occurring at point X on Earth's surface. 24 hours later, X has made a complete rotation and is back in its original position. However, the moon has moved during that time (B), so X must travel an additional distance (white arrow) to once again become aligned with the moon and experience a high tide. For this reason, corresponding tides occur approximately 60 minutes later each day (PW).

The motion of the moon impacts the tidal cycles in other ways. As the moon orbits the Earth, its orbital plane is at an angle relative to the rotational plane of Earth. This angle, or **declination**, means than the moon fluctuates between an angle of 28.5º north of the equator, to 28.5º south of the equator roughly every two weeks (the cycle from maximum to minimum and back takes about 27 days). Figure 13.1.2.3 illustrates a case where the moon is at its maximum declination 28.5º north of the equator, creating its corresponding tidal maxima. A point on the Earth at the latitude indicated by the red line would experience two high tides as it rotated through 24 hours, at points A and B. But the two high tides would not be of equal heights; the high tide at A would be higher than the high tide at B. This helps create a mixed semi-diurnal tide; two high tides of different heights per day (see section 11.3).

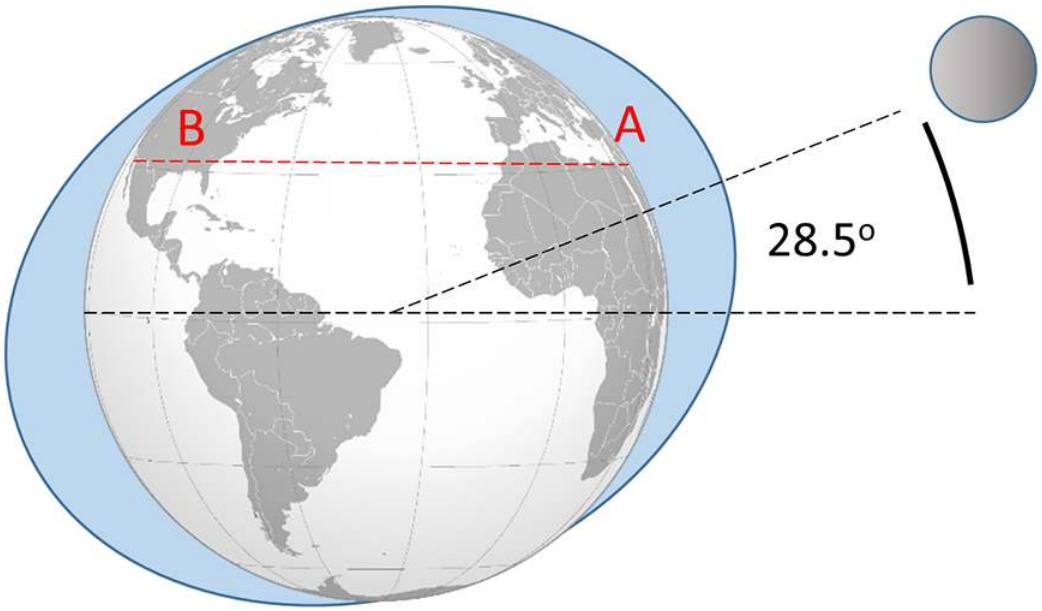


Figure 13.1.2.3 The effect of the moon declination on tide heights. The moon oscillates between 28.5º north and 28.5º south of the equator every two weeks, leading to uneven tidal heights each day at a particular latitude (PW).

Finally, the continents and the bottom topography of the oceans have an impact on the tides that are experienced in an area. Because the tides are essentially waves with extremely long wavelengths extending halfway across the Earth, they behave as

shallow water waves, and they are influenced and refracted by the bottom contours, leading to regional tidal variations. When the tidal crests encounter land, they are are reflected, and the wave moves back out to sea, theoretically until it encounters another continent on the opposite side of the ocean basin. The crest is once again reflected, and the water oscillates back and forth as a standing wave across the ocean basin. However, because of the scale over which these tidal waves move, we must take into account the influence of the Coriolis Effect. As the tidal crest is reflected back across the ocean basin, its path is deflected by the Coriolis force; to the right in the Northern Hemisphere, and to the left in the Southern Hemisphere (see section 9.1). Using the Northern Hemisphere as an example, imagine a tidal crest that has reached land on the western side of an ocean basin. It would have a tendency to be reflected and move across the basin towards the east. But the Coriolis force deflects the movement to the right, causing the crest to instead head south. When the crest hits land in the south, it would now tend to reflect towards the north, but once again the Coriolis deflection to the right kicks in, and the wave instead moves to the east. From the east the reflected wave is deflected to the north, and so on. The result of all of this is that instead of a simple standing wave moving back and forth across the ocean, the tidal crest follows a circular pattern around the ocean basin, counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. This is analogous to shaking a pan full of water in a circular manner, and watching the water follow a similar circular path as it sloshes around inside. This large scale circular rotation pattern of tides is called **amphidromic circulation** (Figure 13.1.2.4). The rotation occurs around a central **amphidromic point** or **node**, that shows little tidal variation, while the largest tidal ranges occur on the edges of the circulation pattern. In Figure 13.1.2.4 the amphidromic points are indicated by the dark blue areas where the white lines converge, like spokes from a bicycle wheel, and the dark red and brown areas show the regions of maximum tidal heights. The tidal maxima will rotate around the amphidromic points, taking about 12 hours for a complete rotation, leading to two high and two low tides per day in many places. If a tidal maximum is occurring along one of the white lines in Figure 13.1.2.4 at a certain time in the Northern Hemisphere, one hour later that high tide will have moved to the white line to the left (counterclockwise), and so on until it completes a rotation. In the Southern Hemisphere, the tide will move to the line to the right for clockwise rotation.

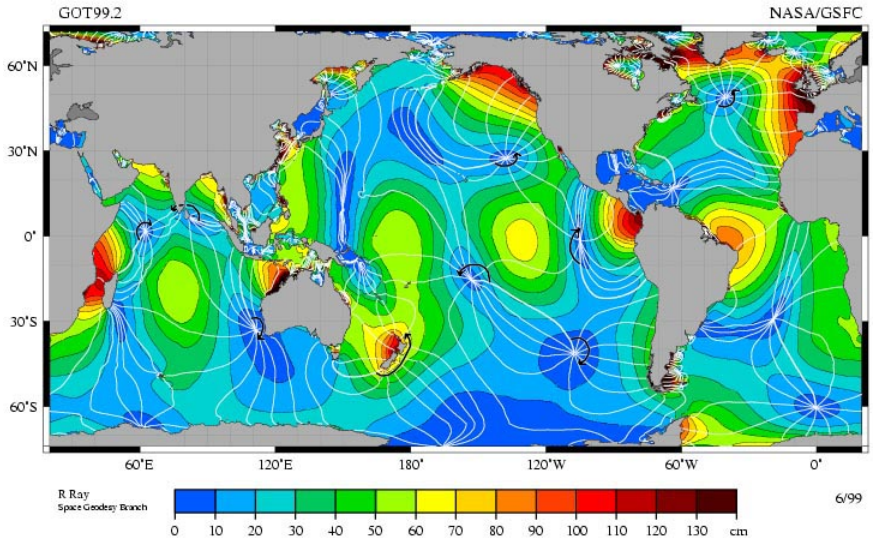


Figure 13.1.2.4 Amphidromic circulation. Amphidromic points are represented where the white lines converge in areas of minimal tidal range. Tidal crests rotate around the amphidromic points, clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere. See text for details (NASA, Public Domain via Wikimedia Commons).

The result of all of these variables is that the tides will not always occur twice each day, at the same time and with equal heights as the Equilibrium Theory of tides may suggest. Instead, each region of the oceans has a unique set of factors that contribute to the types of tides it will experience. The major types of tides are discussed in the next section.

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13.1.3: Tide Classification

With so many variables playing a role in the production of tides, it is understandable that not every place on Earth will experience exactly the same tidal conditions. There are three primary classifications for tides, depending on the number and relative heights of tidal cycles per day.

A **diurnal tide** consists of only one high tide and one low tide per day (Figure 13.1.3.1). “Diurnal” refers to a daily occurrence, so a situation where there is only one complete tidal cycle per day is considered a diurnal tide. Diurnal tides are common in the Gulf of Mexico, along the west coast of Alaska, and in parts of Southeast Asia.

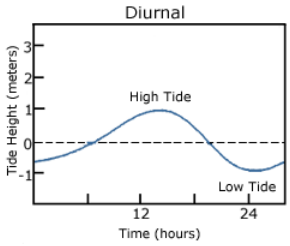


Figure 13.1.3.1 A diurnal tide, with one high and one low tide per day (By NOAA [Public domain], via Wikimedia Commons).

A **semidiurnal tide** exhibits two high and two low tides each day, with both highs and both lows of toughly equal height (Figure 13.1.3.2). “Semidiurnal” means “half of a day”; one tidal cycle takes half of a day, therefore there are two complete cycles per day. Semidiurnal tides are common along the east coasts of North America and Australia, the west coast of Africa, and most of Europe.

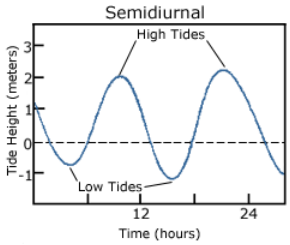


Figure 13.1.3.2 A semi-diurnal tide, with two high and two low tides per day, each of roughly equal heights (By NOAA [Public domain], via Wikimedia Commons).

Mixed semidiurnal tides (or **mixed tides**), have two high tides and two low tides per day, but the heights of each tide differs; the two high tides are of different heights, as are the two low tides (Figure 13.1.3.3). The differences in height may be the result of amphidromic circulation, the angle of the moon, or any of the other variables discussed in section 11.2. Mixed semidiurnal tides are found along the Pacific coast of North America.

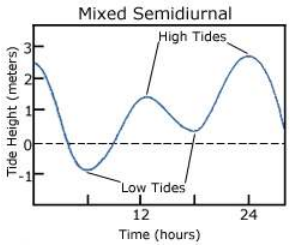


Figure 13.1.3.3 A mixed semi-diurnal tide, with two high and two low tides per day, each with a different height (By NOAA [Public domain], via Wikimedia Commons).

Figure 13.1.3.4 shows the distribution of the various tide types throughout the world.

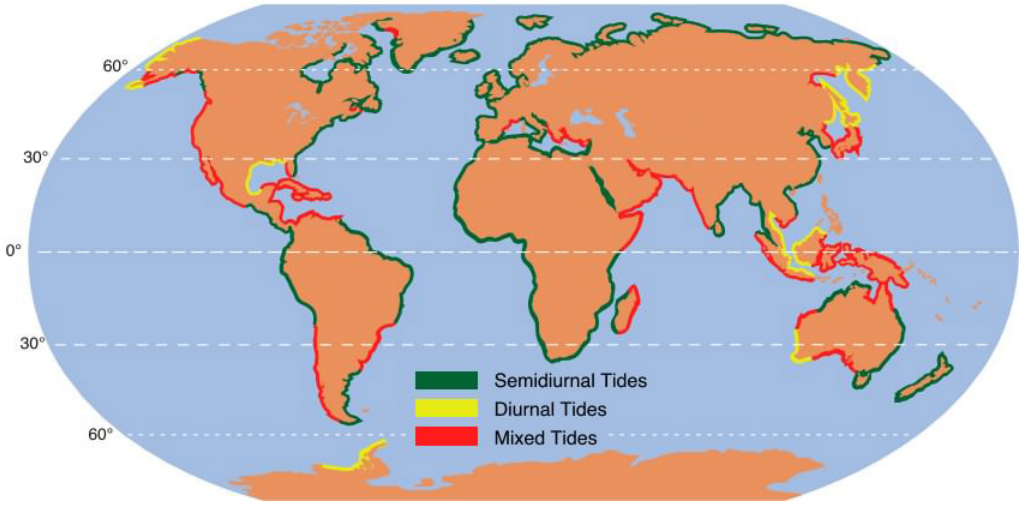


Figure 13.1.3.4 Global distribution of the different types of tides (By KVDP (Own work) [Public domain], via Wikimedia Commons).

Tidal Currents

The movement of water with the rising and falling tide creates tidal currents. As the tide rises, water flows into an area, creating a **flood current**. As the tide falls and water flows out an **ebb current** is created. **Slack water**, or **slack tides** occur during the transition between incoming high and outgoing low tides, when there is no net water movement.

The strength of a tidal current depends on the volume of water that enters and exits with each tidal cycle (the **tidal volume** or **tidal prism**), and the area through which the water flows. A large tidal volume moving through a large area may create only a weak tidal current, as the volume is spread over a wide area. On the other hand, a narrow area may produce a strong tidal current even if the tidal volume is small, as all of the water is forced through a small area. It follows that the strongest tidal currents will result from a large tidal range moving through a narrow area.

Tidal bores occur where rivers meet the ocean. If the incoming tidal current is stronger than the river outflow, the tidal bore appears as a wave, or moving wall of water that moves up the river as the tide comes in (Figure 13.1.3.5).



Figure 13.1.3.5 A tidal bore near Silverdale in the United Kingdom (Arnold Price [CC BY-SA 2.0], via Wikimedia Commons).

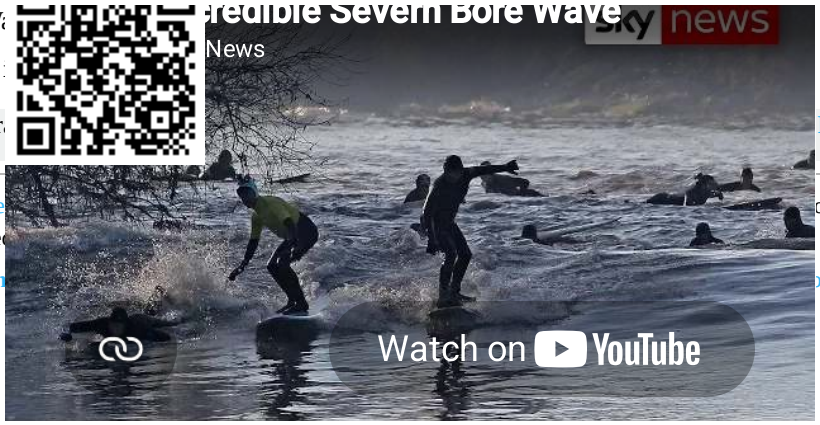
In many cases these tidal bores may move through a river or inlet for many kilometers, and if they are large enough they can form continually breaking waves that surfers can ride much farther and longer than a traditional ocean wave, such as the Severn Bore in England, shown in the video below.



Incredible Severn Bore Wa

Additional links for more

- For an even more dr River



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- **11.3: Tide Classification** oceanography.

CHAPTER OVERVIEW

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14.1.7: Sea Level Change

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14.1: Coastal Oceanography

Learning Objectives

- After reading this chapter you should:

- recognize the various zones of a beach
- understand how the relationship between swash and backwash determines the composition of a beach
- understand the concept of longshore transport
- know the different erosional and depositional structures created by longshore transport
- understand the issues associated with different forms of hard stabilization: groins, jetties etc.
- know what an estuary is
- know the four types of geological estuaries, and how they form
- know the four types of estuaries based on salinity and mixing patterns

For most people, their image of the coast is the place where the land meets the sea, most likely in the form of a beach. But it is more than just the narrow strip along the water line; technically the term “coast” refers to the range of land over which the ocean has an effect on climate, foliage, and other environmental processes. This range may extend for tens of kilometers inland from the water’s edge. Furthermore, what we recognize as the coast today, may not have been a coastal area in the past, as sea level has varied from about 6 m above to 125 m below its current height over the past two million years.

This chapter begins with the features of coastal regions, the processes that shape the coastline, and how humans try to control these processes. Following that, we will examine the different types of estuaries that are found in coastal areas.

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14.1.1: Beaches

For most people, when they think of coastal areas they picture a beach, and the beach that they imagine is probably a typical sandy beach composed of quartz sand grains. But beaches are comprised of whatever types of sediments are dominant in the local area. For example, parts of Hawaii and Iceland are famous for their black sand beaches, made up of eroded basalt and other volcanic materials. The beautiful tropical white sand beaches we see in travel ads are largely composed of the crushed calcium carbonate remains of coral skeletons (much of which has been chewed up and excreted by a fish before we happily run our toes through it!) Other beaches may lack sand altogether and instead be dominated by small shells, or larger rocks or pebbles (Figure 14.1.1.1).



Figure 14.1.1.1 Various beach substrates. Clockwise from top left: Punaluu Black Sand Beach, Hawaii, USA (Diego Delso [CC BY-SA 3.0], via Wikimedia Commons); Shell Beach, Shark Bay, Western Australia (Brian W. Schaller (Own work) [CC BY-NC-SA 3.0], via Wikimedia Commons); white coral sand beach in the Maldives (http://www.elitedivingagency.com/, [CC BY-SA 3.0] via Wikimedia Commons); rocky beach at Killbear Provincial Park in Ontario, Canada (John Vetterli (originally posted to Flickr as Beach) [CC BY-SA 2.0], via Wikimedia Commons).

The shoreline is divided up into multiple zones (Figure 14.1.1.2). The **backshore** is the region of the beach above the high tide line, which is only submerged under unusually high wave conditions, such as during storms. The **foreshore** lies between the high tide and low tide lines; it is submerged during high tide and is exposed during low tide. The **nearshore** extends from the low tide line to the depth where wave action is no longer influenced by the bottom, i.e. to where the depth exceeds the wave base. Finally, the **offshore** zone represents the depths beyond the nearshore region.

Along the beach itself, the area above the high tide line is called the **berm**, which is usually dry and relatively flat. The berm often ends with a berm crest or berm **scarp**, which is a steeper wall carved out by wave action that leads down to the foreshore. The foreshore has a number of other names, including the **beach face**, the **intertidal** or **littoral zone**, and if the area is fairly flat, the **low tide terrace**. Just off shore from the beach there are often **longshore bars** and longshore troughs running parallel to the beach. The longshore bars are accumulations of sand that are deposited by wave action and longshore currents. The decrease in depth above longshore bars is what often causes waves to start to break well before reaching the beach.

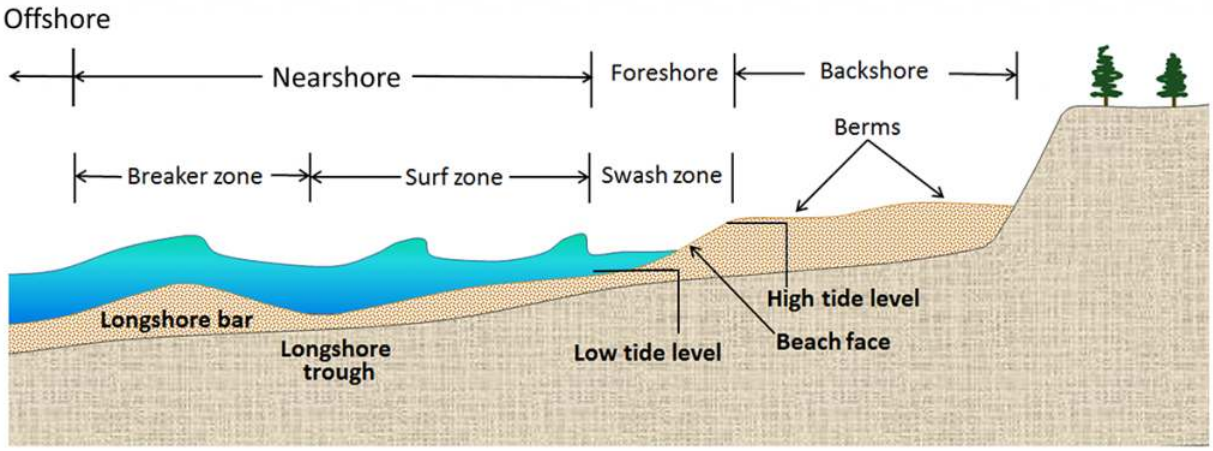


Figure 14.1.1.2 The zones of a typical beach (Modified by PW from Steven Earle, "Physical Geology").

The sand or other particles that make up the beach are distributed by wave action. The water that moves over a beach through incoming waves is called **swash**, and it also contains suspended sand grains that can get deposited on the beach. Some of the swash percolates into the sand while the rest of the water washes back out as **backwash** as the wave recedes. Backwash removes sand from the beach and returns it to the ocean. Sand will therefore be deposited or eroded depending on which process is dominant. If wave action is light, a lot of incoming water gets absorbed by the sand, so swash dominates. Under heavier waves the beach becomes saturated with water, so less can be absorbed, and backwash is dominant. This leads to seasonal cycles in beach structure; waves are heavier during the winter as a result of stormier conditions at sea, so backwash dominates and sand is removed from the beach and deposited offshore in longshore bars. In the summer the waves are gentler, swash dominates, and the sand is transported from the longshore bar and deposited on the shore to create a wider, sandy beach (Figure 14.1.1.3).

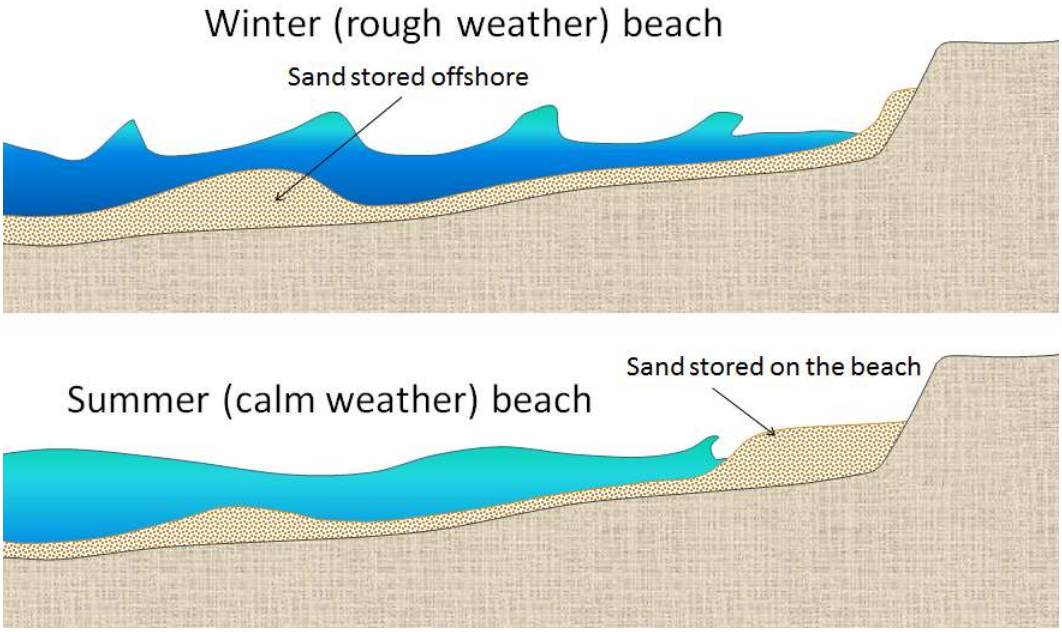


Figure 14.1.1.3 The differences between summer and winter on beaches in areas where the winter conditions are rougher and waves have a shorter wavelength but higher energy. In winter, sand from the beach is stored offshore (Steven Earle, "Physical Geology").

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14.1.2: Longshore Transport

We learned previously that refraction causes waves to approach parallel to shore. However, most waves still reach the shore at a small angle, and as each one arrives, it pushes water along the shore, creating what is known as a **longshore current** within the surf zone (the areas where waves are breaking) (Figure 14.1.2.1). Longshore currents can move up to 4 km/hr, strong enough to carry people with them, as everyone knows who has been swimming in the ocean only to look up and see that they have been carried far down the beach from their towel!

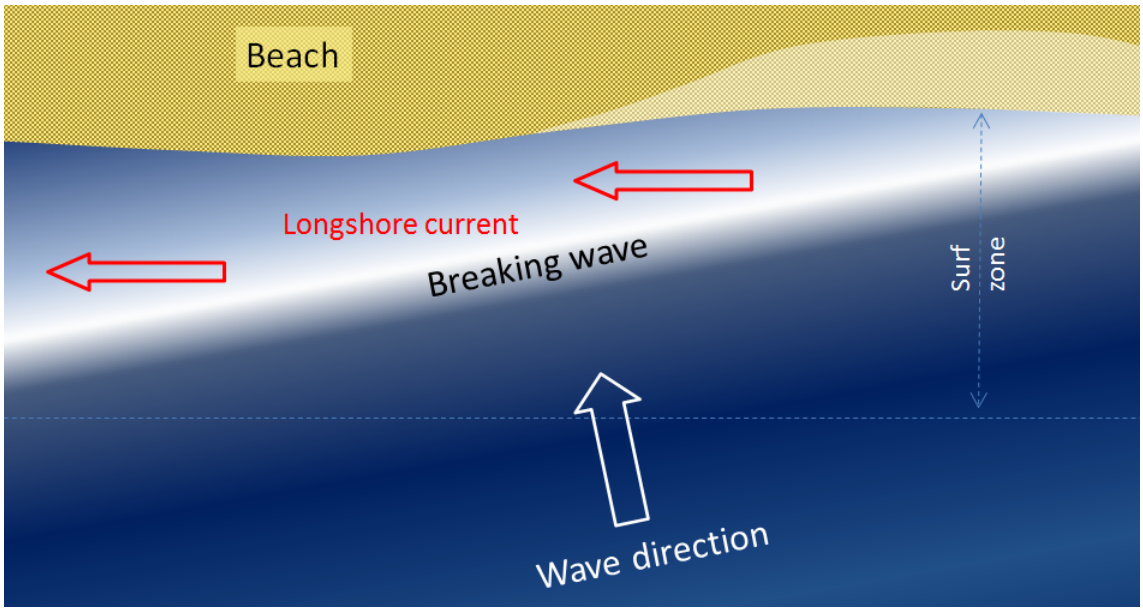


Figure 14.1.2.1 Longshore currents are caused by waves approaching shore at a small angle, moving water parallel to the shore (Steven Earle, "Physical Geology").

Another important effect of waves reaching the shore at an angle is that when they wash up onto the beach, they do so at an angle, but when that same wave water washes back down the beach, it moves straight down the slope of the beach (Figure 14.1.2.2). The upward-moving water, known as the swash, pushes sediment particles along the beach, while the downward-moving water, the backwash, brings them straight back. With every wave that washes up and then down the beach, particles of sediment are moved along the beach in a zigzag pattern.

The combined effects of sediment transport within the surf zone by the longshore current and sediment movement along the beach by swash and backwash is known as **longshore transport**, or **littoral drift**. Longshore transport moves a tremendous amount of sediment along coasts (both oceans and large lakes) around the world, and it is responsible for creating a variety of depositional features. The net movement of sediment due to longshore transport is to the south along both coasts of the continental United States, because the storms and high winds that originally create the swell tend to occur at higher latitudes and move to the south.

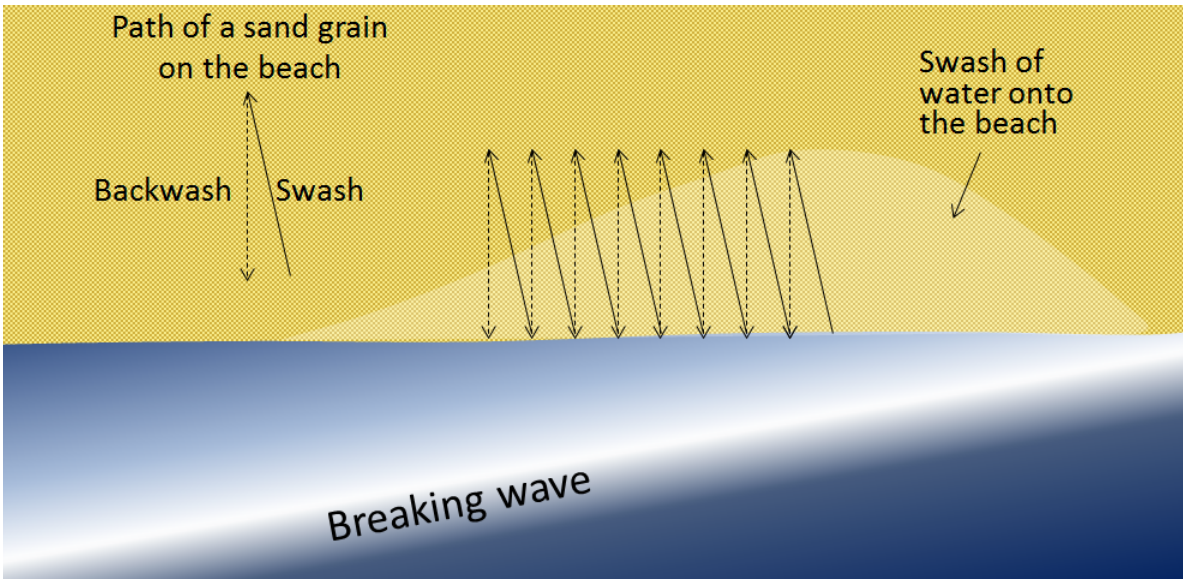


Figure 14.1.2.2 The zigzag pattern of sediment movement along a beach creating longshore transport. In this figure the longshore transport moves particles to the left (Steven Earle, "Physical Geology").

A **rip current** (often incorrectly called a "rip tide"; they are not really related to tides) is another type of current that develops in the nearshore area, and has the effect of returning water that has been pushed up to the shore by incoming waves or accumulated through longshore currents, particularly converging longshore currents. Rip currents often occur where there is a channel between sandbars that makes it easier for the retreating water to escape. As shown in Figure 14.1.2.3, rip currents flow straight out from the shore, and because the water is directed through a narrow space, the current can be very strong. The currents lose strength quickly just outside of the surf zone, but they can be dangerous to swimmers who get caught in them and are pulled away from shore. Swimmers caught in a rip current should not try to swim directly back to shore, as it is difficult to fight the current and the swimmer can quickly tire. Instead, swim parallel to the beach for a short distance until you are outside of the rip current, and then you can easily swim to shore.

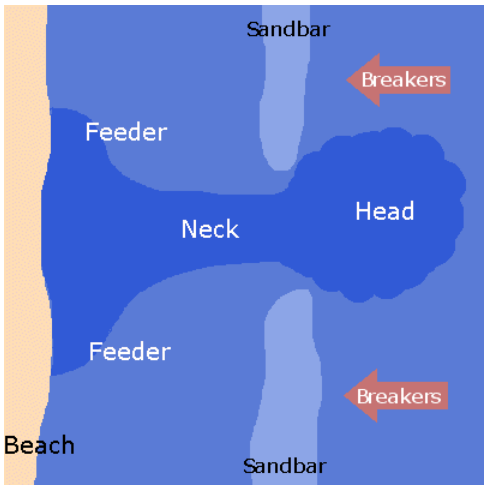


Figure 14.1.2.3 Creation of a rip current from wave action and longshore transport. Water accumulates on the beach, and then rushes out to sea through a narrow channel, creating a strong current (National Weather Service, Wilmington, NC (NOAA) [Public domain], via Wikimedia Commons).

Rip currents are visible in Figure 14.1.2.4, a beach at Tunquen in Chile near Valparaiso. As is evident from the photo, the rips correspond with embayments in the beach profile. Three of them are indicated with arrows, but it appears that there may be several others farther along the beach.

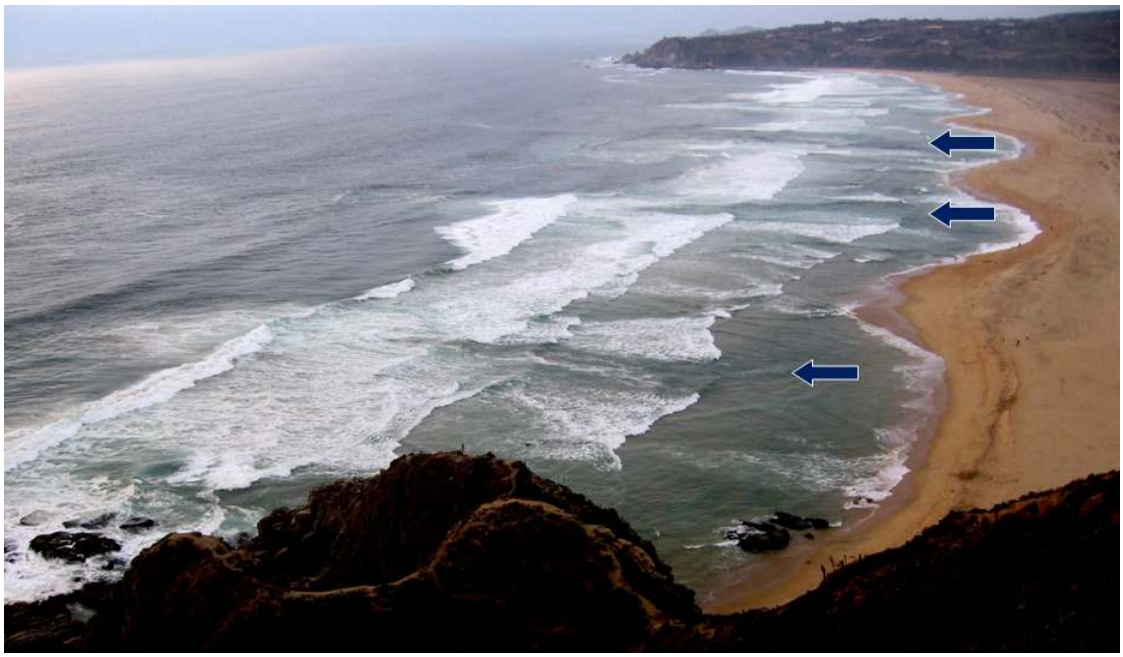


Figure 14.1.2.4 Rip currents along a beach in Chile, indicated by the arrows. Longshore currents converging in a curved beach have nowhere to go but straight back out to sea, creating a rip current (Steven Earle, “Physical Geology”).

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14.1.3: Landforms of Coastal Erosion

Large waves crashing onto a shore bring a tremendous amount of energy that has a significant eroding effect, and several unique erosion features commonly form on rocky shores with strong waves.

When waves approach an irregular shore, they are slowed down to varying degrees, depending on differences in the water depth, and as they slow, they are bent or refracted. In Figure 14.1.3.1, wave energy is represented by the blue arrows. That energy is evenly spaced out in the deep water, but because of refraction, the energy of the waves is being focused on the **headlands**. On irregular coasts, the headlands receive much more wave energy than the intervening bays, and thus they are more strongly eroded. The result of this is **coastal straightening**, where an irregular coast will eventually become straightened, although that process may take millions of years.

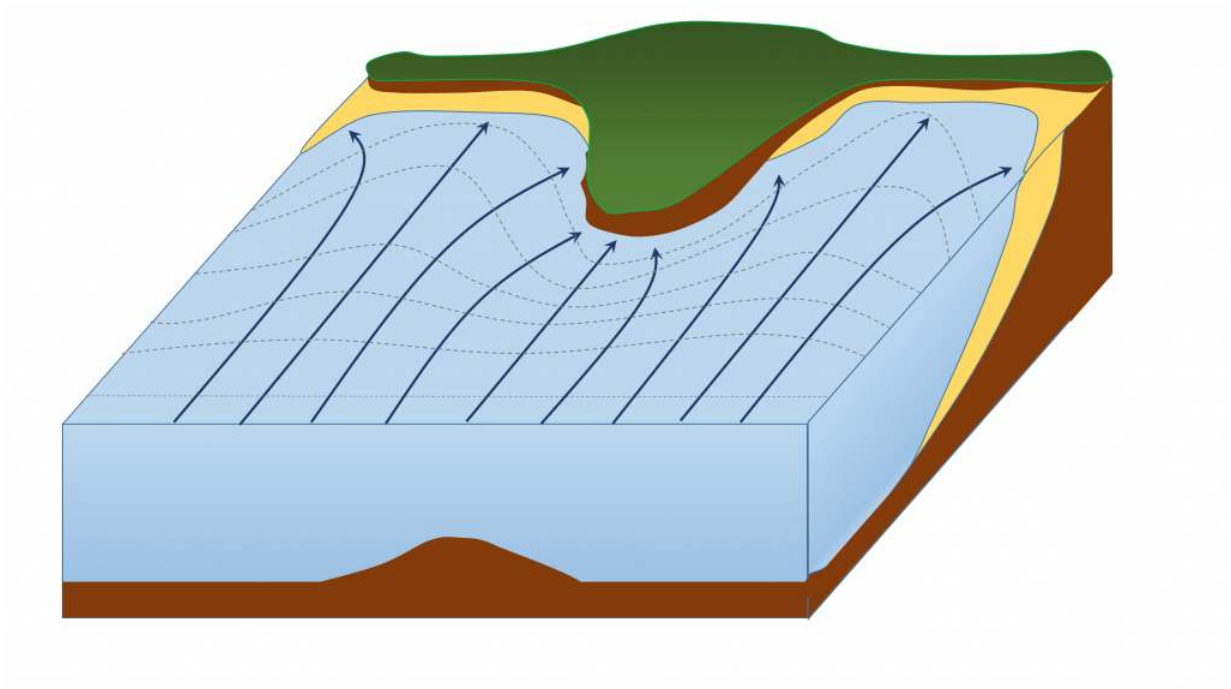


Figure 14.1.3.1 The approach of waves (blue lines) towards a coastal headland. The blue arrows represent wave energy; most of that energy is focused on the headlands, causing greatest erosion in this area (PW).

Wave erosion is greatest in the surf zone, where the wave base is impinging strongly on the seafloor and where the waves are breaking. The result is that the substrate in the surf zone is typically eroded to a flat surface known as a **wave-cut platform** (or **wave-cut terrace**) (Figure 14.1.3.2). A wave-cut platform extends across the intertidal zone.



Figure 14.1.3.2 A wave-cut platform in bedded sedimentary rock on Gabriola Island, B.C. The wave-eroded surface is submerged at high tide (Steven Earle, “Physical Geology”).

Arches and **sea caves** form as a result of the erosion of relatively non-resistant rock. Wave action and strong longshore currents can carve a cave into a headland, and if the erosion extends all the way through, it becomes an arch. If a hole develops in the ceiling of a cave, a **blowhole** can be created, shooting water into the air when waves crash in the cave. An arch in the Barachois River area of western Newfoundland, Canada, is shown in Figure 14.1.3.3. This feature started out as a sea cave, and then, after being eroded from both sides, became an arch. During the winter of 2012-2013, the arch collapsed, leaving a small stack at the end of the point.



Figure 14.1.3.3 Top: An arch in tilted sedimentary rock at the mouth of the Barachois River, Newfoundland, July 2012. Bottom: The same location in June 2013. The arch has collapsed and a small stack remains (Photo: Dr. David Murphy, used with permission in Steven Earle, “Physical Geology”).

The tower of rock left behind from a collapsed arch is called a **sea stack** (Figure 14.1.3.4). But sea stacks can also form during the formation of wave-cut platforms or other features, when relatively resistant rock that does not get completely eroded remains behind to form the stack.



Figure 14.1.3.4 A sea stack, likely created from the collapse of a sea arch (Doug Lee [CC BY-SA 2.0], via Wikimedia Commons).

Figure 14.1.3.5 summarizes the process of transformation of an irregular coast into a straightened coast with **sea cliffs** (wave-eroded escarpments) and the remnants of stacks, arches, and wave-cut platforms. The next stages of this process would be the continued landward erosion of the sea cliffs and the complete erosion of the stacks and wave-cut platforms in favor of a continuous and nearly straight sandy beach.

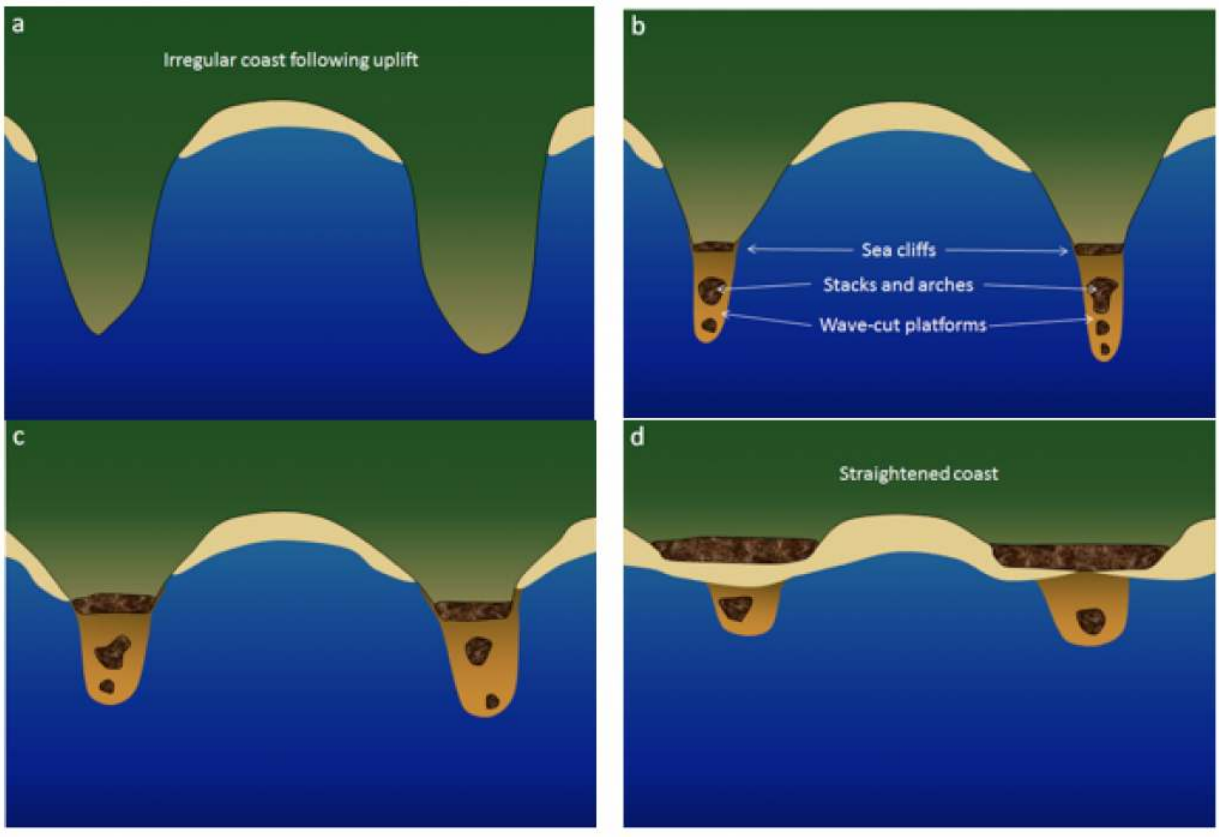


Figure 14.1.3.5 Evolution of a straightened coast through the erosion to stacks and arches, sea cliffs, and wave-cut platforms (Steven Earle, “Physical Geology”).

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14.1.4: Landforms of Coastal Deposition

Some coastal areas are dominated by erosion, an example being the Pacific coast of North America, while others are dominated by deposition, examples being the Atlantic and Caribbean coasts of the United States. But on almost all coasts, both deposition and erosion are happening to varying degrees most of the time, although in different places. On deposition-dominant coasts, the coastal sediments are still being eroded from some areas and deposited in others.

On coasts that are dominated by depositional processes, most of the sediment being deposited typically comes from large rivers. Much of the sediment is immediately deposited at the mouth of the river, creating large fan-shaped **deltas**. An obvious example is where the Mississippi River flows into the Gulf of Mexico at New Orleans; another is the Yellow (Huang He) River in China (Figure 14.1.4.1).



Figure 14.1.4.1 The Yellow River delta in China, created by one of the most sediment-laden rivers on Earth (NASA [Public domain], via Wikimedia Commons).

The evolution of sandy depositional features on sea coasts is primarily influenced by waves and currents, especially longshore currents. As sediment is transported along a shore, either it is deposited on beaches, or it creates other depositional features. A **spit**, for example is an elongated sandy deposit that extends out into open water in the direction of a longshore current (Figure 14.1.4.2).



Figure 14.1.4.2 Farewell Spit, on the northern shore of New Zealand's South Island (By NASA/GSFC/METI/ERSDAC/JAROS, and U.S./Japan ASTER Science Team (NASA's Earth Observatory) [Public domain], via Wikimedia Commons).

A spit that extends across a bay to the extent of closing, or almost closing it off, is known as a **baymouth bar** (Figure 14.1.4.3). Most bays have streams flowing into them, and since this water has to get out, it is rare that a baymouth bar will completely close the entrance to a bay.



Figure 14.1.4.3 Left: Illustration of a baymouth bar and tombolo (Steven Earle, "Physical Geology"). Right: A baymouth bar at the mouth of the Klamath River in northern California (Linda Tanner, https://www.flickr.com/photos/goingslo/5827465324, Creative Commons CC-BY 2.0).

Tombolos are common where islands are abundant, and they typically form where there is a wave shadow behind a nearshore island (Figure 14.1.4.4). This becomes an area with reduced energy, and so the longshore current slows and sediments accumulate. Eventually enough sediments accumulate to connect the island to the mainland with a tombolo (Figure 14.1.4.5).

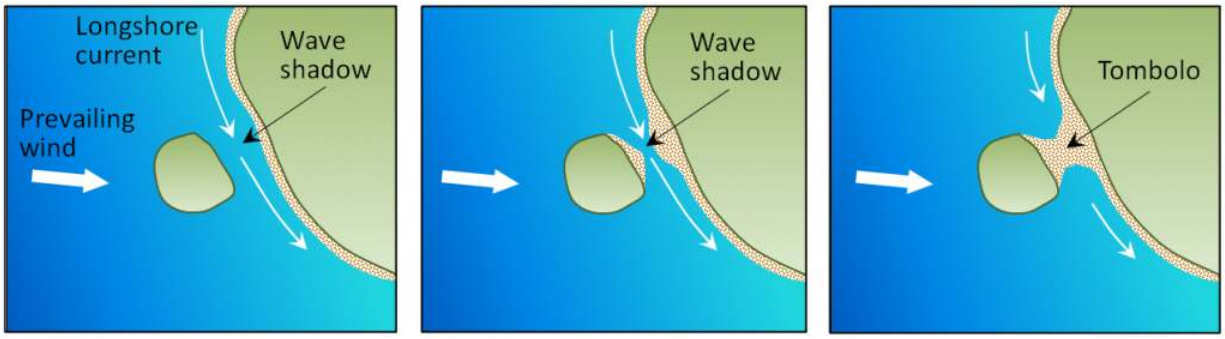


Figure13.4.4 Formation of a tombolo. In the wind shadow of an island, there is little wave action, so the sediment moved by longshore transport gets deposited, eventually linking the island to the mainland (Steven Earle, “Physical Geology”).



Figure 14.1.4.5 A stack (with a wave-cut platform) connected to the mainland by a tombolo, Leboeuf Bay, Gabriola Island, British Columbia (Steven Earle, “Physical Geology”).

In areas where coastal sediments are abundant and coastal relief is low (because there has been little or no recent coastal uplift), it is common for **barrier islands** to form (Figure 14.1.4.6). Barrier islands are elongated islands composed of sand that form offshore from the mainland, potentially reaching several kilometers wide and hundreds of kilometers long. They are common along the U.S. Gulf Coast from Texas to Florida, and along the U.S. Atlantic Coast from Florida to Massachusetts. The islands often form as the result of sediment moving offshore through river discharge, while wave action works to push the sediment back towards the shore. The resulting sediment buildup is then stretched into long barrier islands by longshore transport.



Figure 14.1.4.6 Assateague Island on the Maryland coast, U.S. This barrier island is about 60 km long and only 1 km to 2 km wide. The open Atlantic Ocean is to the right and the lagoon is to the left. This part of Assateague Island has recently been eroded by a tropical storm, which pushed massive amounts of sand into the lagoon. (http://soundwaves.usgs.gov/2014/04/i...ial_ViewCV.jpg).

Mature barrier islands contain a number of ecological zones. Beginning on the ocean side of the island there is a beach, consisting of the zones previously described. Behind the beach lie dunes that are built up by sand transported by wind. The ocean side of the dunes are home to grasses and other plants which help stabilize the sand from erosion, and also help slow down the wind to allow sand to settle and accumulate. Beyond the dunes lies a more heavily vegetated barrier flat, covered by larger shrubs and trees that are tolerant to the high winds and salty conditions. As the land slopes down on the side of the island facing the mainland, the low-lying areas transition into a salt marsh or mud flat habitat, which is protected from wave action, but is influenced by tidal changes. The mud flats are colonized by grasses, which slow down the movement of water and lead to increased sediment deposition, building up the land in the marsh. Different species of grasses eventually dominate the different elevations of the salt marsh, depending on their tolerance for submersion in seawater. These salt marshes are very important habitats for many invertebrates, birds, and juvenile fish. Between the island and the mainland lies a lagoon, which usually contains brackish water from the mixing of fresh water runoff from the land and the seawater within a somewhat enclosed space. Barrier islands, although attractive locations for beach houses, are not permanent structures, and people should be wary of building on them. Over time, the erosion on the seaward side of the island, and the expansion of the marsh on the landward side, causes the island to slowly move towards the mainland, eventually closing off the lagoon. Maintaining dune grasses is one way to slow this movement, and as we will see in the next section, people have developed a number of other strategies to try to curtail the natural erosion of beaches.

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14.1.5: Human Interference with Shorelines

The continued erosion and deposition of coastal sediments is a natural process, with features forming and disappearing as sea level and other conditions change. However, we have also come to enjoy and rely on many of these beaches and other coastal features for commerce, recreation, and living space. So from our perspective, we often see the transient nature of the coast as a threat to our activities, and as a result we have developed a number of ways to try and influence the erosion process, usually through **hard stabilization**, or the building of structures to stop the flow of sand. While some of these efforts have been successful, many others have actually exacerbated the problem, as we will see below.

Groins (or groynes) are barriers built perpendicular to the shore (Figure 14.1.5.1 *left*). Groins are built to interrupt longshore transport and trap sand upstream of the groin, which they do very well. But downstream of the groin the source of replacement sand is cut off, while sand continues to be removed, so erosion can become even more pronounced on that side of the groin. To prevent that erosion, another groin must be built downstream of the first one, which then creates its own erosion problems, leading to another groin, and so on. Eventually a beach may become covered in a series of groins, called a **groin field**, all trying to stabilize the natural flow of sand (Figure 14.1.5.1 *right*).



Figure 14.1.5.1 Left: A groin on Bolinas Beach, California. Longshore transport would be moving from right to left, accumulating sand on the right of the groin (Beatrice Murch, https://www.flickr.com/photos/blmurch/952523971, CC BY-SA 2.0). Right: A groin field. Longshore transport would move from the bottom of the picture towards the top (https://pxhere.com/en/photo/1048378, Public Domain, CC0).

Jetties are like longer groins, often built to protect the mouths of harbors to prevent them from filling with sand. Because they are longer they can trap more sand than groins, and they also can contribute to increased erosion on the downstream side (Figure 14.1.5.2). If too much sand accumulates upstream of the jetty it can spread past the jetty and into the mouth of the harbor, in which case the jetty may need to be extended.



Figure 14.1.5.2 The jetties protecting Santa Cruz harbor, California. Sand has accumulated on the left (north) side of the harbor, and the beach is eroded on the right (south) side (Google Maps, Map Data: Google).

Breakwaters are walls that are usually built parallel to the shore. Their purpose is not necessarily to interfere with sediment transport, but instead to protect the areas behind them from heavy wave action, so they are often deployed at the mouth of a harbor, or to protect the boats in a mooring field. But breakwaters do have an unintended impact on sediment distribution. Longshore transport continues to move sand along the beach, but once it gets behind the breakwater the lack of wave action interrupts the flow, and the sand settles and accumulates. The beach grows behind a breakwater, until eventually they may become connected (Figure 14.1.5.3). With the longshore transport interrupted, increased erosion can occur downstream of the breakwater.



Figure 14.1.5.3 A series of breakwaters along the coast in Skagen, Denmark. In this case, sand has accumulated to the point that they now connect the breakwaters to land (Google Maps; Imagery: DigitalGlobe, Aerodata,International Surveys, TerraMetrics, Data SIO, NOAA, US Navy, NGA, GEBCO, Map Data: Google).

Santa Monica Pier

The beach around the Santa Monica Pier in California provides a good example of the effects of breakwaters on a sandy shore. A breakwater was constructed in the early 1930s to protect the pier and the boats that moored near it. Following the construction, the once-straight beach became much wider behind the breakwater as sand accumulated in the absence of strong wave action (*below left*). Now that the breakwater is no longer in place, the bulge in the shoreline is gone, and the beach is much straighter once again (*below right*).



Left; aerial image of Santa Monica Pier and breakwater from 1936 (Courtesy of Santa Monica Public Library Image Archives/ Spence Air Photos). Right; Santa Monica Pier in 2011 (© JCS, CC BY-SA 3.0, via Wikimedia Commons).

Seawalls are constructed at the top of the surf zone, where the waves crash against the shore. The walls are designed as a barrier between the waves and the shore, to prevent the land from being eroded (Figure 14.1.5.4). They are often utilized in beachfront property to prevent the ground under a home from being undermined by the waves. However, as with the other forms of hard stabilization we have discussed, seawalls are not without their own environmental consequences. The sudden release of wave energy on a seawall can create turbulence, which undermines the sediment at the base of the wall and causes it to erode. Furthermore, on a softer, natural coastline some of the wave energy is absorbed or dissipated, but with a hard seawall most of the wave energy is reflected, leading to stronger longshore currents and faster erosion. In many places where seawalls have been built the beaches are getting steeper, and erosion rates have increased, with the potential for seawalls to collapse along with whatever they are supporting. Because of this, some coastal communities are phasing out seawall construction to try to return to more natural beach fronts.



Figure 14.1.5.4 A sea wall at Horsey Gap, UK (Evelyn Simak [CC BY-SA 2.0], via Wikimedia Commons).

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14.1.6: Estuaries

Estuaries are partially enclosed bodies of water where the salt water is diluted by fresh water input from land, creating **brackish** water with a salinity somewhere between fresh water and normal seawater. Estuaries include many bays, inlets, and sounds, and are often subject to large temperature and salinity variations due to their enclosed nature and smaller size compared to the open ocean.

Estuaries can be classified geologically into four basic categories based on their method of origin. In all cases they are a result of rising sea level over the last 18,000 years, beginning with the end of the last ice age; a period that has seen a rise of about 130 m. The rise in sea level has flooded coastal areas that were previously above water, and prevented the estuaries from being filled in by all of the sediments that have been emptied into them.

The first type is a **coastal plain estuary**, or **drowned river valley**. These estuaries are formed as sea level rises and floods an existing river valley, mixing salt and fresh water to create the brackish conditions where the river meets the sea. These types of estuaries are common along the east coast of the United States, including major bodies such as the Chesapeake Bay, Delaware Bay, and Narragansett Bay (Figure 14.1.6.1). Coastal plain estuaries are usually shallow, and since there is a lot of sediment input from the rivers, there are often a number of depositional features associated with them such as spits and barrier islands.

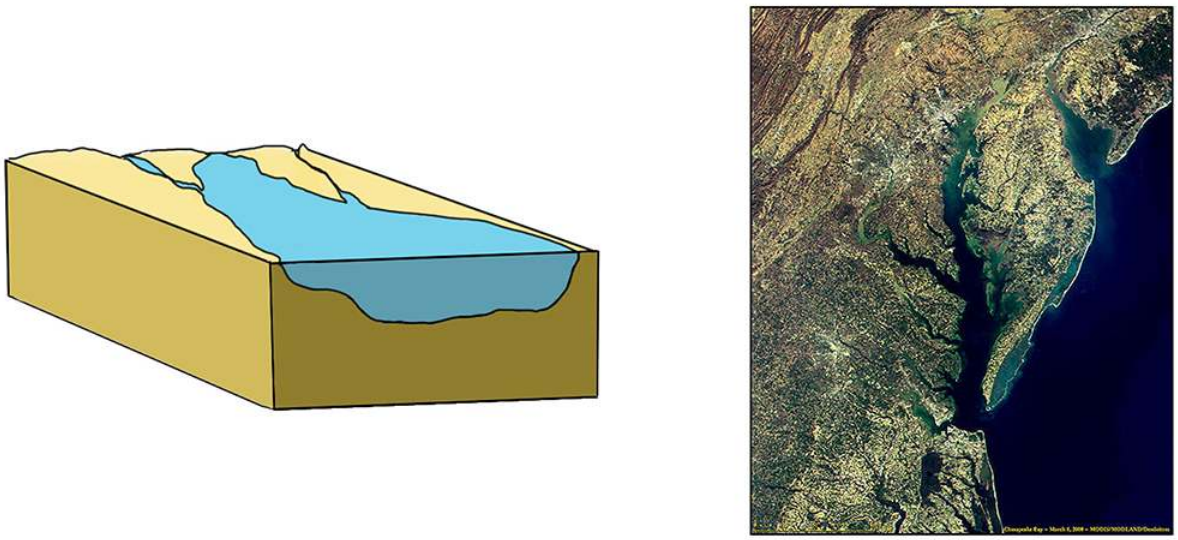


Figure 14.1.6.1 A coastal plain estuary. Sea level has risen and flooded what was once a river valley. The satellite image shows Chesapeake Bay and Delaware Bay, two coastal plain estuaries (left: JR, right: NASA, Public Domain via Wikimedia Commons).

The presence of sand bars, spits, and barrier islands can lead to **bar-built estuaries**, where a barrier is created between the mainland and the ocean. The water that remains inside the sand bar is cut off from complete mixing with the ocean, and receives freshwater input from the mainland, creating estuarine conditions (Figure 14.1.6.2).

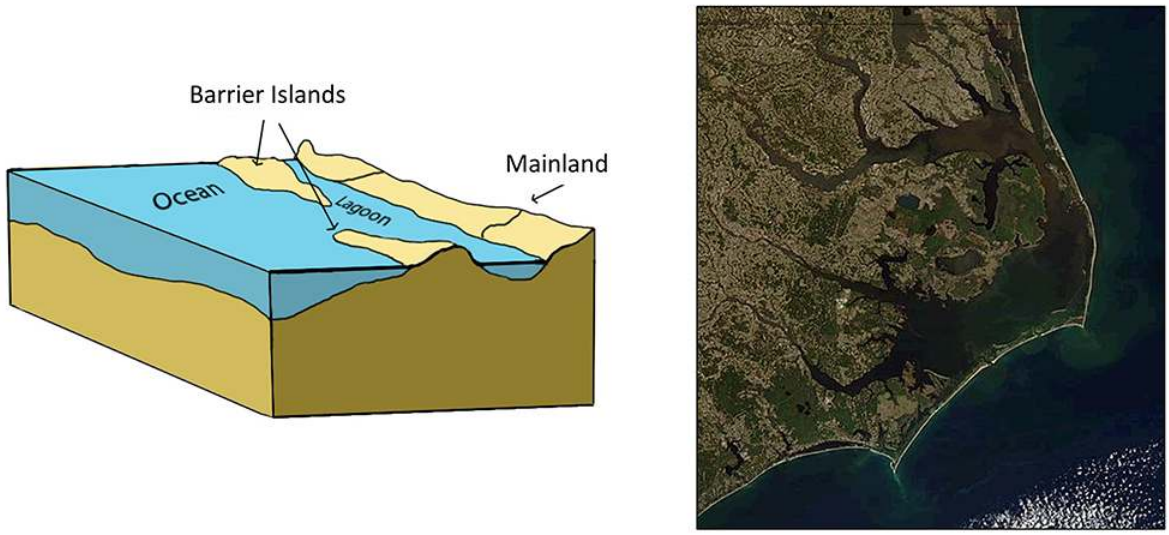


Figure 14.1.6.2 A bar-built estuary. Sand bars and barrier islands have partially isolated a lagoon from the rest of the ocean. Freshwater input into the lagoon from the mainland creates brackish conditions in the estuary. At right is a satellite image of Pamlico Sound, North Carolina, a bar-built estuary surrounded by spits and barrier islands (left: JR, right: NASA, Public Domain via Wikimedia Commons).

Fjords are estuaries formed in deep, U-shaped basins that were carved out by advancing glaciers. When the glaciers melted and retreated, sea level rose and filled these troughs, creating deep, steep-walled fjords (Figure 14.1.6.3). Fjords are common in Norway, Alaska, Canada, and New Zealand, where there are mountainous coastlines once covered by glaciers.

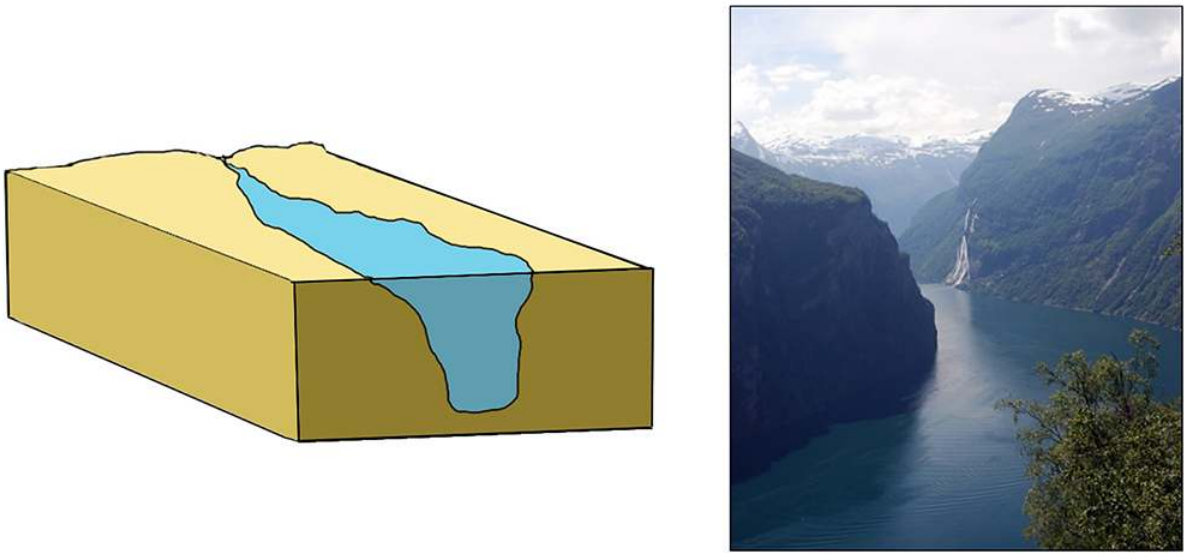


Figure 14.1.6.3 A fjord is a deep estuary that was carved out by glacial movements. At right is Geirangerfjord, Norway (left: JR, right: Fgmedia, [CC-BY-SA-3.0], via Wikimedia Commons).

Tectonic estuaries are the result of tectonic movements, where faulting causes some sections of the crust to subside, and those lower elevation sections then get flooded with seawater. San Francisco Bay is an example of a tectonic estuary (Figure 14.1.6.4).

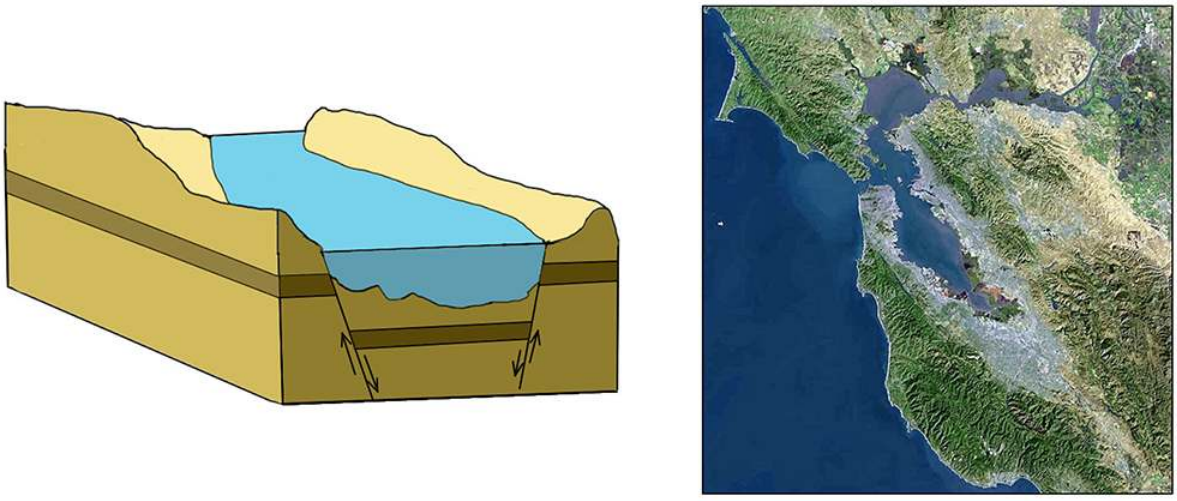


Figure 14.1.6.4 A tectonic estuary, formed from the subsidence of crust along fault lines, and the subsequent filling by seawater. San Francisco Bay is a tectonic estuary, shown at right (left: JR, right: USGS, Public Domain, via Wikimedia Commons).

Estuaries are also classified based on their salinity and mixing patterns. The amount of mixing of fresh and salt water in an estuary depends on the rate at which fresh water enters the head of the estuary from river input, and the amount of seawater that enters the estuary mouth as a result of tidal movements. The input of fresh water is reflected in the **flushing time** of the estuary. This refers to the time it would take for the in-flowing fresh water to completely replace all the fresh water currently in the estuary. Seawater input is measured by the **tidal volume**, or **tidalprism**, which is the average volume of sea water entering and leaving the estuary during each tidal cycle. In other words, it is the volume difference between high and low tides. The interaction between the flushing time, tidal volume, and the shape of the estuary will determine the extent and type of water mixing within the estuary.

In a **vertically mixed**, or **well-mixed** estuary there is complete mixing of fresh and salt water from the surface to the bottom. In a particular location the salinity is constant at all depths, but across the estuary the salinity is lowest at the head where the fresh water enters, and is highest at the mouth, where the seawater comes in. This type of salinity profile usually occurs in shallower estuaries, where the shallow depths allow complete mixing from the surface to the bottom.

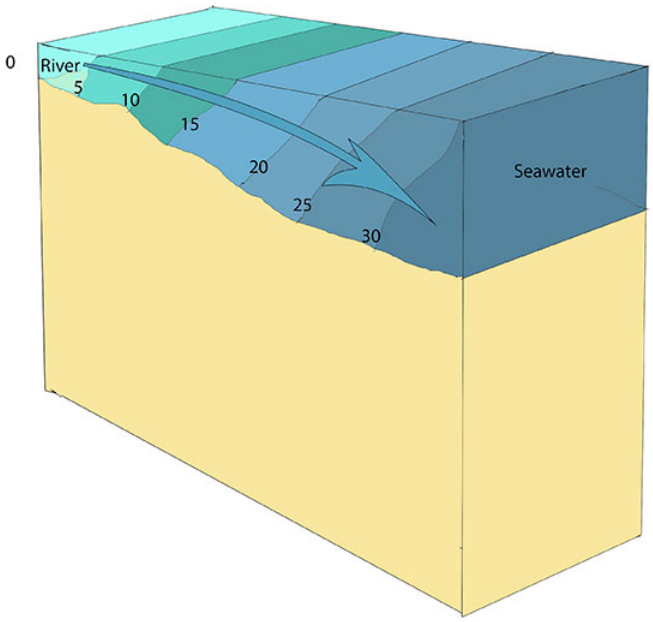


Figure 14.1.6.5 A well-mixed estuary. The shallow basin allows nearly complete mixing or fresh and seawater from top to bottom. Salinities are in ppt (JR).

Slightly stratified or **partially mixed** estuaries have similar salinity profiles to vertically mixed estuaries, where salinity increases from the head to the mouth, but there is also a slight increase in salinity with depth at any point. This usually occurs in deeper estuaries than those that are well-mixed, where waves and currents mix the surface water, but the mixing may not extend all the way to the bottom.

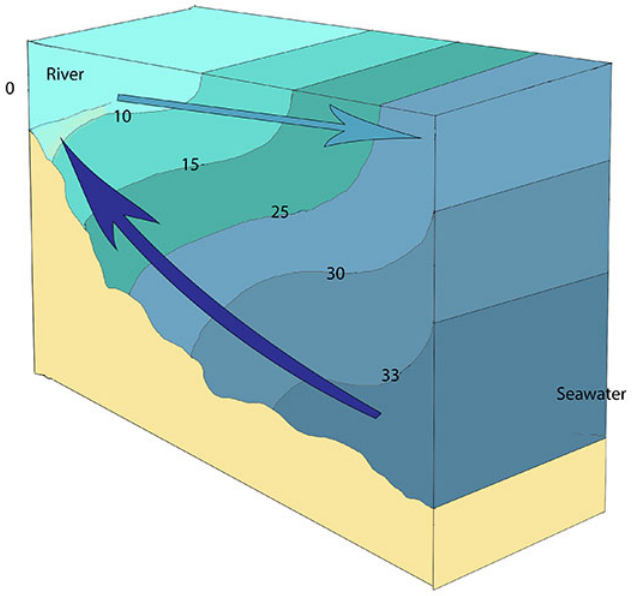


Figure 14.1.6.6 A slightly-stratified estuary. Generally deeper than a well-mixed estuary, the inflow of seawater (dark blue arrow) and fresh water (light blue arrow) create an estuary where salinity increases with depth, and at the surface when moving from the head to the mouth of the estuary. Salinities are in ppt (JR).

A **salt wedge** estuary occurs where the outflow of fresh water is strong enough to prevent the denser ocean water to enter through the surface, and where the estuary is deep enough that surface waves and turbulence have little mixing effect on the deeper water. Fresh water flows out along surface, salt water flows in at depth, creating a wedge shaped lens of seawater moving along the bottom. The surface water may remain mostly fresh throughout the estuary if there is no mixing, or it can become brackish depending on the level of mixing that occurs.

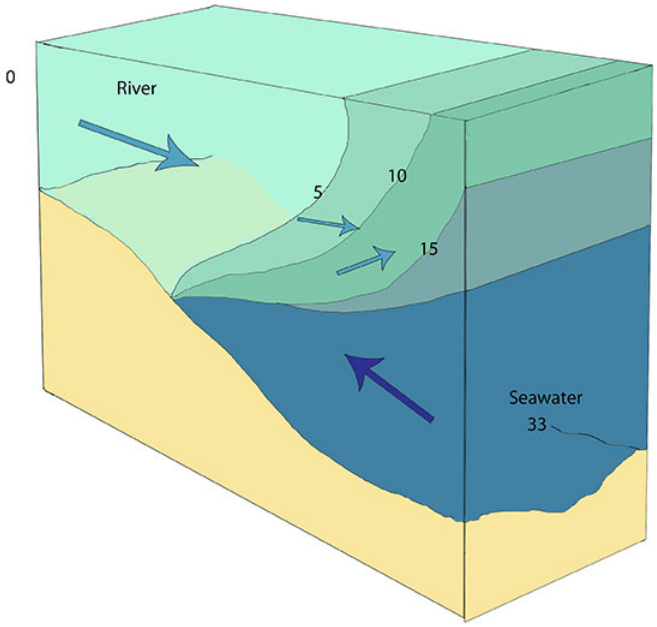


Figure 14.1.6.7 A salt wedge estuary. Strong river outflow (light blue arrows) creates a layer of mostly fresh water that sits on top of a wedge of encroaching seawater along the bottom (dark blue arrow). Salinities are in ppt (JR).

Highly stratified profiles are found in very deep estuaries, such as in fjords. Because of the depth, mixing of fresh and salt water only occurs near the surface, so in the upper layers salinity increases from the head to the mouth, but the deeper water is of standard ocean salinity.

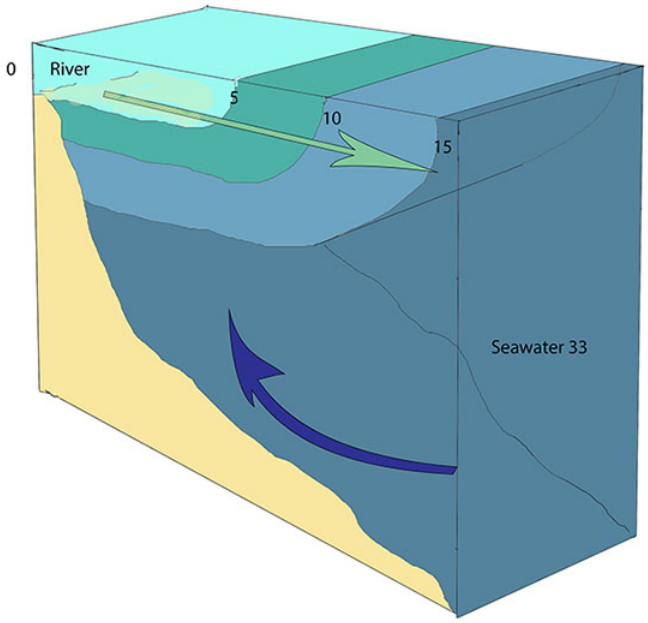


Figure 14.1.6.8 A highly stratified estuary. Strong river outflow and a deep basin prevent mixing between surface and bottom water, creating an estuary that is vertically stratified. Salinities are represented in ppt (JR).

Estuaries are very important commercially, as they are home to the majority of the world's metropolitan areas, they serve as ports for industrial activity, and a large percentage of the world's population lives near estuaries. Estuaries are also very important biologically, especially in their role as the breeding grounds for many species of fish, birds, and invertebrates.

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14.1.7: Sea Level Change

Sea level change has been a feature on Earth for billions of years, and it has important implications for coastal processes, estuaries, and both erosional and depositional features. There are two main mechanisms of sea level change, eustatic and isostatic, as described below.

Eustatic sea level changes are global sea level changes related to changes in the volume of water in the ocean. These can be due to changes in the volume of glacial ice on land, thermal expansion of the water, or to changes in the shape of the seafloor caused by plate tectonic processes. For example, seafloor spreading widens an ocean basin, thus changing its volume and affecting sea level.

Over the past 20,000 years, there has been approximately 125 m of eustatic sea level rise due to glacial melting. Most of that took place between 15,000 and 7,500 years ago during the major melting phase of the North American and Eurasian Ice Sheets (Figure 14.1.7.1). At around 7,500 years ago, the rate of glacial melting and sea level rise decreased dramatically, and since that time, the average rate has been in the order of 0.7 mm/year.

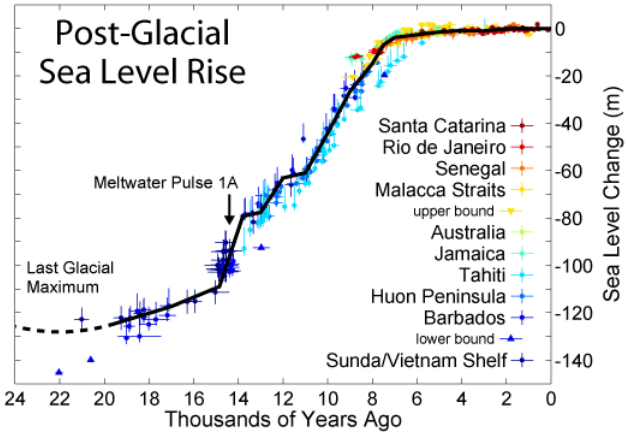


Figure 14.1.7.1 Sea level rise resulting from the melting of glacial ice over the past 24,000 years (Robert A. Rhode, CC BY-SA 3.0, via Wikimedia Commons).

Anthropogenic climate change led to accelerating sea level rise starting around 1870. Since that time, the average rate has been 1.1 mm/year, but it has been gradually increasing. Since 1992, the average rate has been 3.2 mm/year (Figure 14.1.7.2). Much of this is due to increased glacial melting as the global climate gets warmer, but a large part is due to **thermal expansion** of the water. As water warms, the molecules gain more kinetic energy and move faster and farther apart; the result is that the same amount of water now takes up more space. So even without the input of new water from melting ice, warming ocean temperatures will cause sea level to rise.

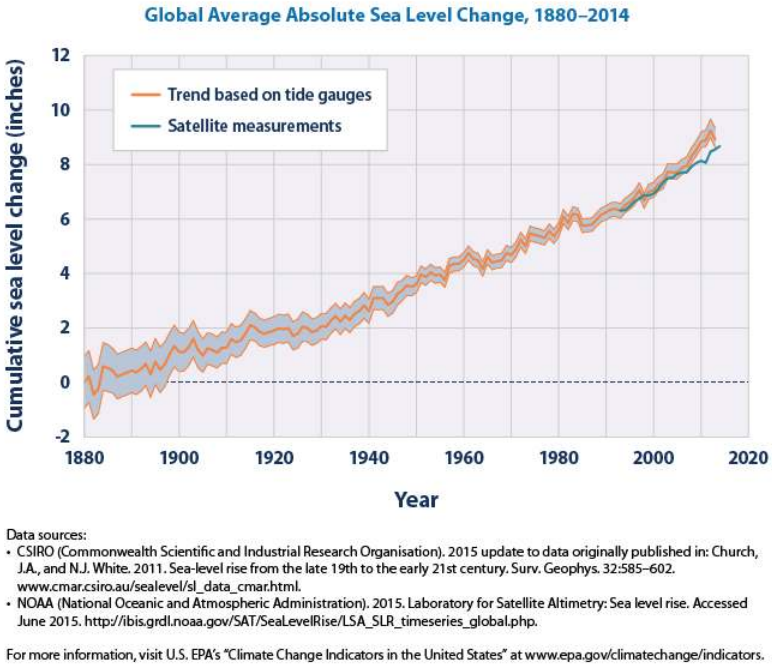


Figure 14.1.7.2 Average absolute sea level change, which refers to the height of the ocean surface, regardless of whether nearby land is rising or falling. Satellite data are based solely on measured sea level, while the long-term tide gauge data include a small correction factor because the size and shape of the oceans are changing slowly over time (US EPA [Public domain], via Wikimedia Commons).

Isostatic sea level changes are local changes caused by subsidence or uplift of the crust related either to changes in the amount of ice on the land, or to growth or erosion of mountains. Almost all of Canada and parts of the northern United States were covered in thick ice sheets at the peak of the last glaciation. Following the melting of this ice, there has been an isostatic rebound of continental crust in many areas. This ranges from several hundred meters of rebound in the central part of the Laurentide Ice Sheet (around Hudson Bay) to 100 m to 200 m in places such as Vancouver Island and the mainland coast of British Columbia. In other words, although global sea level was about 130 m lower during the last glaciation, the glaciated regions were depressed at least that much in most places, and more than that in places where the ice was thickest. Tectonic processes, such as the uplift of crust, can also cause localized changes in sea level.

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CHAPTER OVERVIEW

15: Introduction to Marine Life

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15.2: Explaining the Diversity of Life
15.3: Life in the Ocean
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15.1: How do we categorize living things?

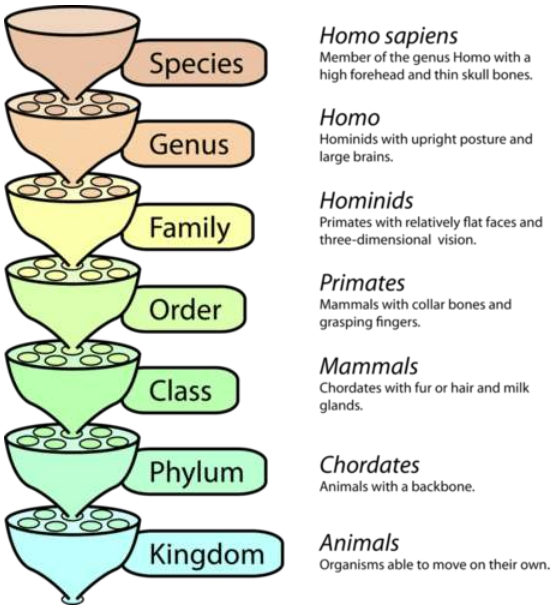
There are millions and millions of species, so classifying organisms into proper categories can be a difficult task. To make it easier for all scientists to do, a classification system had to be developed.

Linnaean Classification

The evolution of life on Earth over the past 4 billion years has resulted in a huge variety of species. For more than 2,000 years, humans have been trying to classify the great diversity of life. The science of classifying organisms is called **taxonomy.** Classification is an important step in understanding the present diversity and past evolutionary history of life on Earth.

All modern classification systems have their roots in the **Linnaean classification** system. It was developed by Swedish botanist Carolus Linnaeus in the 1700s. He tried to classify all living things that were known at his time. He grouped together organisms that shared obvious physical traits, such as number of legs or shape of leaves. For his contribution, Linnaeus is known as the "father of taxonomy." You can learn more about Linnaeus and his system of classification by watching the video at this link: http://teachertube.com/viewVideo.php?video_id=169889.

The Linnaean system of classification consists of a hierarchy of groupings, called **taxa**(singular, taxon). Taxa range from the kingdom to the species (see **Figure** below). The **kingdom** is the largest and most inclusive grouping. It consists of organisms that share just a few basic similarities. Examples are the plant and animal kingdoms. The **species** is the smallest and most exclusive grouping. It consists of organisms that are similar enough to produce fertile offspring together. Closely related species are grouped together in a **genus.**



Linnaean Classification System: Classification of the Human Species. This chart shows the taxa of the Linnaean classification system. Each taxon is a subdivision of the taxon below it in the chart. For example, a species is a subdivision of a genus. The classification of humans is given in the chart as an example.

Binomial Nomenclature

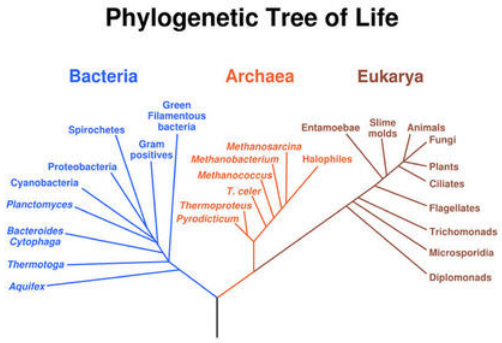
Perhaps the single greatest contribution Linnaeus made to science was his method of naming species. This method, called **binomial nomenclature**, gives each species a unique, two-word Latin name consisting of the genus name and the species name. An example is *Homo sapiens*, the two-word Latin name for humans. It literally means "wise human." This is a reference to our big brains.

Why is having two names so important? It is similar to people having a first and a last name. You may know several people with the first name Michael, but adding Michael's last name usually pins down exactly whom you mean. In the same way, having two names uniquely identifies a species.

Revisions in Linnaean Classification

Linnaeus published his classification system in the 1700s. Since then, many new species have been discovered. The biochemistry of many organisms has also become known. Eventually, scientists realized that Linnaeus's system of classification needed revision.

A major change to the Linnaean system was the addition of a new taxon called the domain. A**domain** is a taxon that is larger and more inclusive than the kingdom. Most biologists agree there are three domains of life on Earth: Bacteria, Archaea, and Eukaryota (see **Figure** below). Both Bacteria and Archaea consist of single-celled prokaryotes. Eukaryota consists of all eukaryotes, from single-celled protists to humans. This domain includes the Animalia (animals), Plantae (plants), Fungi (fungi), and Protista (protists) kingdoms.



This phylogenetic tree is based on comparisons of ribosomal RNA base sequences among living organisms. The tree divides all organisms into three domains: Bacteria, Archaea, and Eukarya. Humans and other animals belong to the Eukarya domain. From this tree, organisms that make up the domain Eukarya appear to have shared a more recent common ancestor with Archaea than Bacteria.

Summary

- Classification is an important step in understanding life on Earth.
- All modern classification systems have their roots in the Linnaean classification system.
- The Linnaean system is based on similarities in obvious physical traits. It consists of a hierarchy of taxa, from the kingdom to the species.
- Each species is given a unique two-word Latin name.
- The recently added domain is a larger and more inclusive taxon than the kingdom.

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15.2: Explaining the Diversity of Life

The theory of evolution by natural selection describes a mechanism for species change over time. That species change had been suggested and debated well before Darwin. The view that species were static and unchanging was grounded in the writings of Plato, yet there were also ancient Greeks that expressed evolutionary ideas.

In the eighteenth century, ideas about the evolution of animals were reintroduced by the naturalist Georges-Louis Leclerc, Comte de Buffon and even by Charles Darwin's grandfather, Erasmus Darwin. During this time, it was also accepted that there were extinct species. At the same time, James Hutton, the Scottish naturalist, proposed that geological change occurred gradually by the accumulation of small changes from processes (over long periods of time) just like those happening today. This contrasted with the predominant view that the geology of the planet was a consequence of catastrophic events occurring during a relatively brief past. Hutton's view was later popularized by the geologist Charles Lyell in the nineteenth century. Lyell became a friend to Darwin and his ideas were very influential on Darwin's thinking. Lyell argued that the greater age of Earth gave more time for gradual change in species, and the process provided an analogy for gradual change in species.

In the early nineteenth century, Jean-Baptiste Lamarck published a book that detailed a mechanism for evolutionary change that is now referred to as inheritance of acquired characteristics. In Lamarck's theory, modifications in an individual caused by its environment, or the use or disuse of a structure during its lifetime, could be inherited by its offspring and, thus, bring about change in a species. While this mechanism for evolutionary change as described by Lamarck was discredited, Lamarck's ideas were an important influence on evolutionary thought. The inscription on the statue of Lamarck that stands at the gates of the Jardin des Plantes in Paris describes him as the "founder of the doctrine of evolution."

Charles Darwin and Natural Selection

The actual mechanism for evolution was independently conceived of and described by two naturalists, Charles Darwin and Alfred Russell Wallace, in the mid-nineteenth century. Importantly, each spent time exploring the natural world on expeditions to the tropics. From 1831 to 1836, Darwin traveled around the world on *H.M.S. Beagle*, visiting South America, Australia, and the southern tip of Africa. Wallace traveled to Brazil to collect insects in the Amazon rainforest from 1848 to 1852 and to the Malay Archipelago from 1854 to 1862. Darwin's journey, like Wallace's later journeys in the Malay Archipelago, included stops at several island chains, the last being the Galápagos Islands (west of Ecuador). On these islands, Darwin observed species of organisms on different islands that were clearly similar, yet had distinct differences. For example, the ground finches inhabiting the Galápagos Islands comprised several species that each had a unique beak shape (Figure 15.2.1). He observed both that these finches closely resembled another finch species on the mainland of South America and that the group of species in the Galápagos formed a graded series of beak sizes and shapes, with very small differences between the most similar. Darwin imagined that the island species might be all species modified from one original mainland species. In 1860, he wrote, "Seeing this gradation and diversity of structure in one small, intimately related group of birds, one might really fancy that from an original paucity of birds in this archipelago, one species had been taken and modified for different ends."[1]

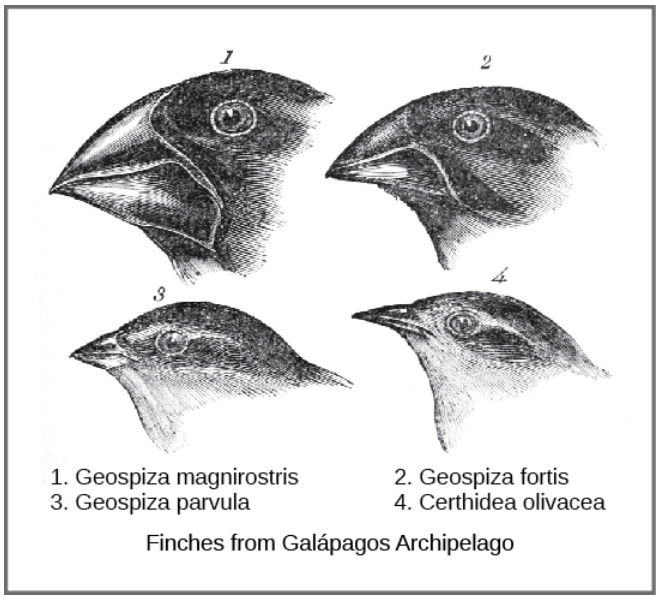


Figure 15.2.1: Darwin observed that beak shape varies among finch species. He postulated that the beak of an ancestral species had adapted over time to equip the finches to acquire different food sources. This illustration shows the beak shapes for four species of ground finch: 1. Geospiza magnirostris (the large ground finch), 2. G. fortis (the medium ground finch), 3. G. parvula (the small tree finch), and 4. Certhidea olivacea (the green-warbler finch).

Wallace and Darwin both observed similar patterns in other organisms and independently conceived a mechanism to explain how and why such changes could take place. Darwin called this mechanism natural selection. Natural selection, Darwin argued, was an inevitable outcome of three principles that operated in nature. First, the characteristics of organisms are inherited, or passed from parent to offspring. Second, more offspring are produced than are able to survive; in other words, resources for survival and reproduction are limited. The capacity for reproduction in all organisms outstrips the availability of resources to support their numbers. Thus, there is a competition for those resources in each generation. Both Darwin and Wallace's understanding of this principle came from reading an essay by the economist Thomas Malthus, who discussed this principle in relation to human populations. Third, offspring vary among each other in regard to their characteristics and those variations are inherited. Out of these three principles, Darwin and Wallace reasoned that offspring with inherited characteristics that allow them to best compete for limited resources will survive and have more offspring than those individuals with variations that are less able to compete. Because characteristics are inherited, these traits will be better represented in the next generation. This will lead to change in populations over generations in a process that Darwin called "descent with modification."

Papers by Darwin and Wallace (Figure 15.2.2) presenting the idea of natural selection were read together in 1858 before the Linnaean Society in London. The following year Darwin's book, *On the Origin of Species*, was published, which outlined in considerable detail his arguments for evolution by natural selection.



(a) (b)

Figure 15.2.2: (a) Charles Darwin and (b) Alfred Wallace wrote scientific papers on natural selection that were presented together before the Linnean Society in 1858.

Demonstrations of evolution by natural selection can be time consuming. One of the best demonstrations has been in the very birds that helped to inspire the theory, the Galápagos finches. Peter and Rosemary Grant and their colleagues have studied Galápagos finch populations every year since 1976 and have provided important demonstrations of the operation of natural selection. The Grants found changes from one generation to the next in the beak shapes of the medium ground finches on the Galápagos island of Daphne Major. The medium ground finch feeds on seeds. The birds have inherited variation in the bill shape with some individuals having wide, deep bills and others having thinner bills. Large-billed birds feed more efficiently on large, hard seeds, whereas smaller billed birds feed more efficiently on small, soft seeds. During 1977, a drought period altered vegetation on the island. After this period, the number of seeds declined dramatically: the decline in small, soft seeds was greater than the decline in large, hard seeds. The large-billed birds were able to survive better than the small-billed birds the following year. The year following the drought when the Grants measured beak sizes in the much-reduced population, they found that the average bill size was larger (Figure 15.2.3). This was clear evidence for natural selection (differences in survival) of bill size caused by the availability of seeds. The Grants had studied the inheritance of bill sizes and knew that the surviving large-billed birds would tend to produce offspring with larger bills, so the selection would lead to evolution of bill size. Subsequent studies by the Grants have demonstrated selection on and evolution of bill size in this species in response to changing conditions on the island. The evolution has occurred both to larger bills, as in this case, and to smaller bills when large seeds became rare.

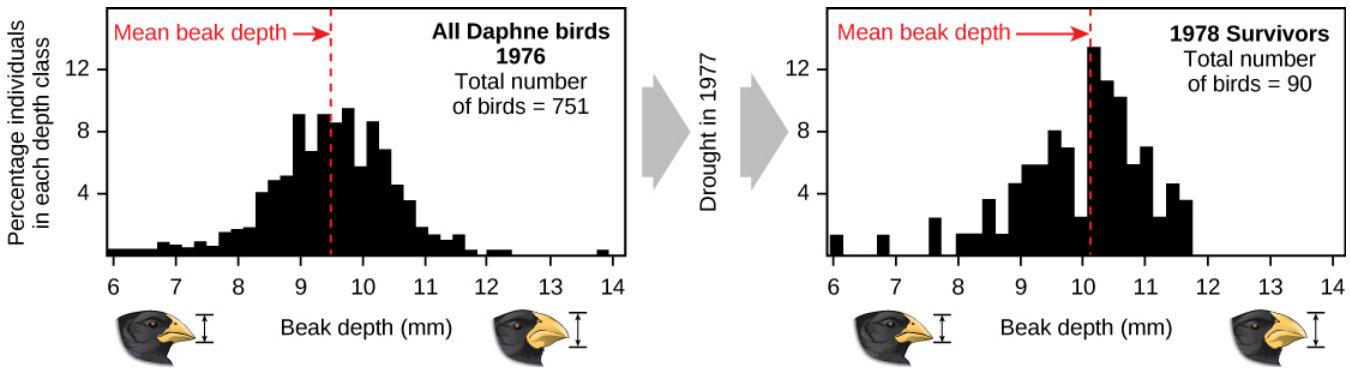


Figure 15.2.3: A drought on the Galápagos island of Daphne Major in 1977 reduced the number of small seeds available to finches, causing many of the small-beaked finches to die. This caused an increase in the finches' average beak size between 1976 and 1978.

Variation and Adaptation

Natural selection can only take place if there is variation, or differences, among individuals in a population. Importantly, these differences must have some genetic basis; otherwise, selection will not lead to change in the next generation. This is critical because variation among individuals can be caused by non-genetic reasons, such as an individual being taller because of better nutrition rather than different genes.

Genetic diversity in a population comes from two main sources: mutation and sexual reproduction. Mutation, a change in DNA, is the ultimate source of new alleles or new genetic variation in any population. An individual that has a mutated gene might have a different trait than other individuals in the population. However, this is not always the case. A mutation can have one of three outcomes on the organisms' appearance (or phenotype):

- A mutation may affect the phenotype of the organism in a way that gives it reduced fitness—lower likelihood of survival, resulting in fewer offspring.
- A mutation may produce a phenotype with a beneficial effect on fitness.
- Many mutations, called neutral mutations, will have no effect on fitness.

Mutations may also have a whole range of effect sizes on the fitness of the organism that expresses them in their phenotype, from a small effect to a great effect. Sexual reproduction and crossing over in meiosis also lead to genetic diversity: when two parents reproduce, unique combinations of alleles assemble to produce unique genotypes and, thus, phenotypes in each of the offspring.

A heritable trait that aids the survival and reproduction of an organism in its present environment is called an adaptation. An adaptation is a "match" of the organism to the environment. Adaptation to an environment comes about when a change in the range of genetic variation occurs over time that increases or maintains the match of the population with its environment. The variations in finch beaks shifted from generation to generation providing adaptation to food availability.

Whether or not a trait is favorable depends on the environment at the time. The same traits do not always have the same relative benefit or disadvantage because environmental conditions can change. For example, finches with large bills were benefited in one climate, while small bills were a disadvantage; in a different climate, the relationship reversed.

Patterns of Evolution

The evolution of species has resulted in enormous variation in form and function. When two species evolve in different directions from a common point, it is called divergent evolution. Such divergent evolution can be seen in the forms of the reproductive organs of flowering plants, which share the same basic anatomies; however, they can look very different as a result of selection in different physical environments, and adaptation to different kinds of pollinators (Figure 15.2.4).



(a) (b)

Figure 15.2.4: Flowering plants evolved from a common ancestor. Notice that the (a) dense blazing star and (b) purple coneflower vary in appearance, yet both share a similar basic morphology. (credit a, b: modification of work by Cory Zanker)

In other cases, similar phenotypes evolve independently in distantly related species. For example, flight has evolved in both bats and insects, and they both have structures we refer to as wings, which are adaptations to flight. The wings of bats and insects, however, evolved from very different original structures. When similar structures arise through evolution independently in different

species it is called convergent evolution. The wings of bats and insects are called analogous structures; they are similar in function and appearance, but do not share an origin in a common ancestor. Instead they evolved independently in the two lineages. The wings of a hummingbird and an ostrich are homologous structures, meaning they share similarities (despite their differences resulting from evolutionary divergence). The wings of hummingbirds and ostriches did not evolve independently in the hummingbird lineage and the ostrich lineage—they descended from a common ancestor with wings.

Footnotes

1. 1 Charles Darwin, *Journal of Researches into the Natural History and Geology of the Countries Visited during the Voyage of H.M.S. Beagle Round the World, under the Command of Capt. Fitz Roy, R.N*, 2nd. ed. (London: John Murray, 1860), http://www.archive.org/details/journalofresea00darw.
2. 2 Sahar S. Hanania, Dhia S. Hassawi, and Nidal M. Irshaid, "Allele Frequency and Molecular Genotypes of ABO Blood Group System in a Jordanian Population," *Journal of Medical Sciences* 7 (2007): 51-58, doi:10.3923/jms.2007.51.58

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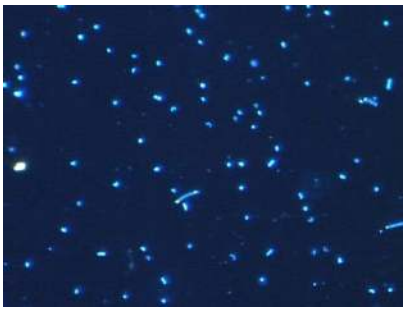
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15.3: Life in the Ocean

The traditional system of categorizing life defines five Kingdoms: animal, plant, bacteria, fungi and protist. And just as all five can be found across the continents, they exist and flourish in the sea as well. Here, life is adapted to various hazards including temperature, pressure, darkness, and salinity. The greatest depths are where the plant kingdom begins to show itself less while the other kingdoms exist in a quiet dark stillness. Their network of interactions changes with each type of ocean environment and results in a colorful diversity in life throughout the oceans.

Bacteria

Bacteria are single-celled organisms lacking membrane-bound nuclei. They are found all throughout the water column, some being aerobic and some anaerobic. They exist in the water, in sediment, and of course, in many organisms throughout the ocean. Marine bacteria play different roles when compared with bacteria in other habitats. Some marine organisms harbor light-emitting bacteria, which forms a symbiotic relationship working to create an effect called bioluminescence, and the light produced could be used to hunt prey, defend against predators, and also find mates[1]. Other bacteria, like cyanobacteria, photosynthesize and are said to be the original source of oxygen on the early planet. They build stromalites, which also provide scientists with a key piece of evidence as to the Earth's past geological history. In addition to that, deep-sea bacteria that thrive on minerals from hydrothermal vents also develop important relationships with the surrounding communities, such as providing essential food sources.

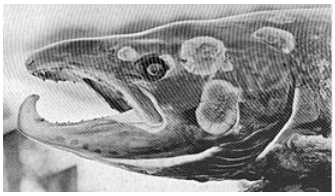


Marine Bacteria: Courtesy of UCSB

Fungi

Marine fungi mainly have continental origins, many being transported out to sea by drifting organic matter from land. They exist largely as parasites on the other kingdoms of life. They can be found living within sand grains, inside coral, and in/outside plants and animals. Many marine fungi have been found to originate from mangrove swamps in which floating pieces of driftwood harbor fungi that may disperse down the water column. Examples include:

- *Rhyzophydium littoreum*
- *Arthopyrenia halodytes*
- *Fusarium*
- *Candida albicans*



Salmon with fungal disease: Courtesy of Wikipedia

Plants

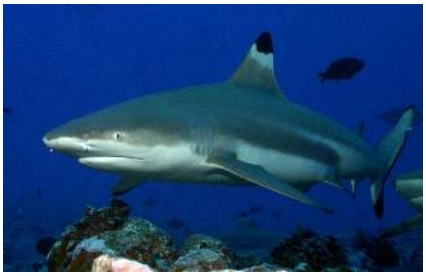
Plants are **autotrophic** organisms. That means that they can build their own organic matter, rather than consuming it. Plants use sunlight to build this organic matter through the process of photosynthesis. Therefore, they can be called photoautotrophs. All plants are multicellular, which means their bodies are composed of multiple cells. Underwater plants are adapted to higher levels of salinity and are found all throughout the upper ocean. Here, plants are adapted in similar ways to their terrestrial counterparts. Both have developed ways to anchor themselves and deal with wind or water currents.



Mangroves: Courtesy of Wikipedia

Animals

In comparison to plants, animals cannot create their own organic matter. Therefore, they must eat. Organisms that must consume organic matter are referred to as **heterotrophs**. The most visible kingdom of life in the oceans is the animal kingdom. Marine animals are adapted to a variety of hardships at all ranges across the ocean. The main food web lies between members of the animal kingdom, where the other kingdoms function mainly as food sources or recyclers. Marine animals are both predators and herbivores, creating a complex diversity in the way ecosystems must interact to sustain themselves. Most animals are invertebrates, or organisms that lack a vertebral column. Vertebrates include fish, seabirds, sea turtles, and marine mammals. Some important Phyla of animals are described in the next sections.



Blacktip Reef Shark, Courtesy of Tracking Sharks

Protista

Protists cannot be classified as plants, animals, or fungi, and many scientists no longer include Protista as a Kingdom. Nevertheless, it is an easy way to categorize some organisms that don't fit into the other Kingdoms. This group includes both unicellular and multicellular organisms, and organisms that are autotrophic and those that are heterotrophic. They are found anywhere containing liquid water and exist in the oceans as plankton and algae. Many marine protists provide a food source for the upper levels of the oceanic food web. Familiar marine protists include:

- macroalgae
- dinoflagellates
- diatoms
- coccolithophores



Protists: Courtesy of UMD

1.education.nationalgeographic....scence/?ar_a=1

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15.4: Phylum Porifera

Porifera

Sponges are unique creatures. They are in the Phylum Porifera and there are about 5,000 different known species. They are one of the simplest forms of multi-cellular animals and come in a variety of different colors, shapes, and sizes. Sponges lack organs and a nervous system. They are sessile organisms, attached to reef surfaces via a holdfast. They get their food through filter feeding when the ocean currents go through their pores. There are 4 different classes of sponges; Calcarea (calcareous- has spicules*), Hexactinellida (horn sponges), Demospongiae (coralline), and Sclerospongiae (glass sponges). Each class of sponge is composed of different organic materials. If you want to learn more about sponges and their structure, this is a really informative article about them!



“Venus Flower Basket Glass Sponges” by NOAA [CC by 2.0]

Spicules are “sharp spikes made of calcium carbonate located in the mesohyl. Spicules form the skeleton of many sponges”. Spicules can be a form of defense for the sponges to deter or hurt predators.

Sponges are some of the oldest organisms on earth having been around up to 500 million years! Each individual cell in a sponge can transform to do any function within the sponge. This is a unique and helpful trait to have. An article from Oceana.org even says that “a sponge destroyed in a blender can reform itself as the cells swim back together and take on the form and job needed for recovery”.

Most sponges are also hermaphrodites meaning that each adult can act as either male or female in the reproductive process. Some sponges can also go through asexual reproduction to produce clones when a piece of sponge breaks off and grows in another location!

Sponge fibers help filter water through the organism. This is important because they filter feed on plankton and bacteria while attached to the ocean floor

There are many different types of rope sponge. Row Pore Rope Sponge, Scattered Pore Rope Sponge, Thin Rope Sponge, and so on. They are a tropical sponge that come in a variety of colors, sizes, and depth ranges. This Marine Species Identification portal has some amazing information on marine species.



“A red rope sponge (Amphimedon compressa) in Hawa’ii” by Laurie Minor-Penland [CC BY 2.0]

Tube sponges are another type of tropical sponge. They have a tube shape and also come in a variety of colors, shapes, and sizes. Below is the Yellow Tube Sponge, also known as *Aplysina fistularis*. The Yellow Tube Sponge has a large opening at the top of it and is attached at the bottom. If they are exposed to the atmosphere, they will turn purple and black and die.



“Aplysina fistularis– Yellow Tube Sponge” by Nick Hobgood [CC BY 3.0]

The Barrel Sponge is another common type of tropical sponge. They are called Barrel Sponges because of the way they look. They are bowl shaped and have very large openings (osculum) at the top of them. They grow to be at least 6 feet wide, making them one of the largest sponge species in their habitats. They are homes for a variety of marine life such as shrimps, crabs, gobies, and cardinalfishes.



“Xestospongia muta, the Barrel Sponge” by NOAA [CC by 2.0]

Fun fact: Giant tube sponges can live up to 2000 years old!!

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15.5: Phylum Cnidaria

Cnidarians are organisms found exclusively in aquatic habitats. Of the over 10,000 aquatic Cnidarian species discovered, most inhabit marine environments. Cnidarians exhibit two major body forms, polyp or medusae. Polyps consist of a body stalk with a tubular shape with a single opening and multiple tentacles that surround this opening which serves as a mouth and anus. Polyps point upward for filter-feeding. The tentacles are covered with cnidocytes (stinging cells). Cnidocytes are the most important characteristic that distinguishes organisms as belonging to the Cnidarian phylum.Polyps are sessile and remain attached to a substrate. The medusa form is most notably observed in the umbrella like form of jellies. Similar to polyps, they consist of a body with a single opening surrounded by tentacles, but the gelatinous layer is much thicker and the mouth is usually oriented towards the substrate when swimming. Many Cnidarians have the ability to switch between these two morphs over the course of their life.

The photo below depicts a living hard coral polyp with its cnidocyte-covered tentacles protruding from the calcareous cup.



"Coral Polyp under dissecting Microscope, 2019" (Cnidocytes are visible as bumps on the tentacles) by Jason Charbonneau [CC by 4.0]

The image below show the medusa form of Cnidarians, as seen in jellies. Check out the Monterey Bay Aquarium's fantastic Live Jelly Cam!



Sea Nettle (*Chrysaora fuscescens*) Jellyfish in captivity in the Monterey Bay Aquarium [CC BY SA 2.5]

Reference

[1] Brusca, R.C., & Brusca, G.J.(2003). Invertebrates. Sunderland, MA: Sinauer Associates, Inc.

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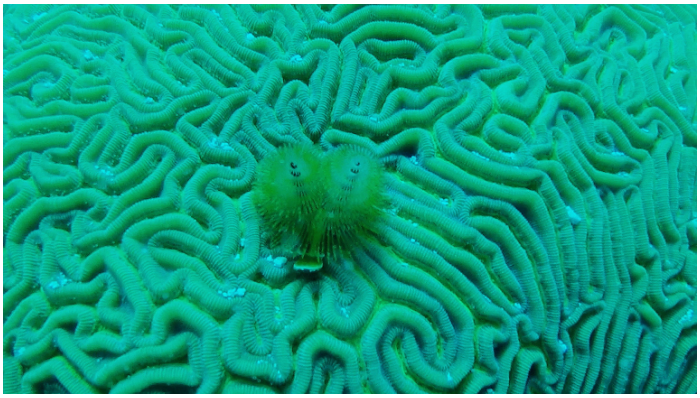
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15.6: Phylum Annelida

Annelids

The Phylum Annelida is defined by its morphology as segmented worms. They express closed circulatory systems in the same fundamental manner that *Homo sapiens* do. Annelids are categorized taxonomically into many classes. A few of these classes are: Class Polychaeta, Class Clitellate and Class Echiuran. Common organisms in the Class Clitellate include Oligochaeta (earthworms) and Branchiobdellida ad Hirudinea (leeches). Ecologically, Annelids are subdivided into two classifications. Errent polychaetes express a multitude of parapodia appendages, defined heads complete with proboscis and are able to extend their bodies in order to have a greater physical surface area for gas exchange. The latter group being Sessile Tube Dwelling Annelids. These organisms contain the segmented part of their bodies within an external tube. The portion of their being that is extended is a feeding appendage that filters through the water. A great example of a Sessile Tube Dwelling Annelid is the Christmas-tree worm. Each Christmas-tree worm has two feeding appendages to filter through the water. They also have the ability to retract quickly when the worm feels threatened. Some are even photosensitive, which means when natural light is obscured by a potential predator, they automatically retreat. Sessile annelids are also inclusive of medusa worms and deep-sea thermal vents.



“Christmas Tree Worm on Brain Coral” by Alana Olendorf [CC by 4.0]

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15.7: Phylum Arthropoda

Arthropoda

All members of the Phylum Arthropoda have a distinct, rigid exoskeleton of chitin. Arthropoda are also known as the most taxonomically diverse phylum on the planet. They occupy nearly every known habitat on earth. Estimates state that there are anywhere from 30-100 million different species in this phylum. The segmentation of Arthropod bodies is different from that of Annelids in that they consist of subsections composed of fused segments referred to as tagmatization. Tagmata are the specific sections (i.e. Head, Thorax, Abdomen), which may vary from species to species. Some important features of arthropods are their open circulatory system, molting, coelomate, protostome and the fact that they live in almost all habitats on earth. Arthropods will molt their exoskeletons, this process of molting is also called ecdysis. Their exoskeletons have three main functions: protection from predators, prevention from desiccation, and locomotion (attachment sites for mussels). Sclerotization is the hardening of the procuticle after molting (also called tanning). Arthropods are also commonly noted for their jointed appendages. Examples of marine Arthropods include lobster, banded coral shrimp, cleaner shrimp. barnacles, horseshoe crabs, copepods, and other microcrustaceans that form the zooplankton.


"Coral Banded Shrimp" by hjk_888 [CC BY-NC-ND 2.0]

The most important difference that is noticed between certain Arthropods such as crayfish and crabs is the morphology of and arrangements of body parts. Crabs and crayfish both have chelipeds, but crabs have a distinguishable single structure prosoma. As with all Arthropods, both types of creatures are contained by chitinous exoskeletons that they are able to shed and regrow throughout their life-cycles.


"Rode amerikaanse rivierkreeft, Red Swamp Crayfish, Procambarus clarkii 04" by Luc hoogenstein [CC by 4.0]

Barnacles are interesting, even among the variant Phylum of life that is Arthropods. As with all other Arthropods, they are distinguished by a chitinous exoskeleton, however unlike many of their relatives, a barnacle also has an outer cup made of calcium carbonate that resembles a tiny volcano in structure. Barnacles also have internal calcareous plates that can close. This is not only for protection from predators, but also to keep the barnacles from drying out if they are not submerged in the water. There are over 1,400 known species of barnacles, most of which have the notable characteristics of secreting a natural 'glue' that allows them to adhere to the various surfaces that they inhabit. The incredible part is, through scientific observation and testing, the tensile strength of these natural epoxies have been measured at over 5,000 lbs/in^2. The manner in which barnacles feed is through cirri. This is a 'feather-like' appendage that combs through the water for zooplankton and other microorganisms that they can capture and ingest.

 “Barnacle Remains under dissection microscope, 2019” by Jason Charbonneau [CC by 4.0]

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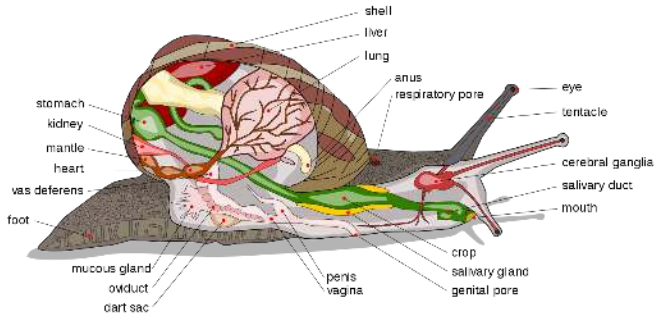
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15.8: Phylum Mollusca

Molluscs

The phylum Mollusca is defined by several special characteristics. These defining characteristics include a mantle with a mantle cavity, a shell (except where lost), visceral mass, foot, and radula. The odontophore is in the mouth of most mollusks and it supports the radula (a ribbon of teeth). In many molluscs, it moves forward while the radula contacts the food, allowing the mollusc to feed. Mollusca can be found in freshwater, marine and terrestrial habitats. More features of molluscs include bilateral symmetry, soft or unsegmented bodies, respiration via ctenidium, ganglia/nerve comprised nervous system, haemocoel body cavity, etc. [1].



"Anatomy of a Common Land Snail", by Jeff Dahl [CC BY-SA 4.0]

Although when most people think of a mollusc they imagine a typical clam or snail with the shell on the exterior of the organism, there are actually variations of these characteristics. One example of this is the Flamingo Tongue. The Flamingo Tongue is a small marine snail. It is a fascinating organism,in that their mantle, when they are at rest, covers the outer shell on the exterior of the organism. When threatened the Flamingo Tongue will retract all of the exposed tissue back into the shell in a very interesting manner.



"Flamingo Tongue, 2018" by Alana Olendorf [CC by 4.0]

Polyplacophorans

Chitons are the common name for species in the Class Polyplacophora. Chitons are considered more primitive in relation to other groups within the phylum such as bivalves or cephalopods. There are 8 dorsal plates lining the dorsal surface of these organisms. There are many primitive species of the class Polyplacophora that are still extant. However, only a few species of Monoplacophorans still exist in the world today.

 “Dorsal side of chiton specimen under dissection microscope 2019” by Jason Charbonneau [CC by 4.0]

Gastropods

Gastropoda is a very diverse class of molluscs. The subclass Prosobranchia, are often identified by their coiled, cone shaped or tubular shells [1]. The mantel cavity is typically located on the anterior of the organism. Important distinctions of these organisms are the variations or the absence of radula. These organisms utilize Nephridia for excretion of nitrogenous waste. One characteristic of gastropods is the presence of a calcareous operculum. This structure acts as a shielded plate that protects the organism by covering the opening when it is retracted into its shell.

In the photo below, the dark oval in the center is the operculum. They are calciferous and rough to the touch, protecting the opening to the shell like a man-hole cover.

 “Gastropod under dissection microscope” by Jason Charbonneau [CC by 4.0]

Nudibranchs

Nudibranchs are typically categorized as sea snails which lack shells. Often they are richly colored and captivating to the eye. Their magnificent beauty is the result of aposematism which is the bright coloration exhibited by these organisms that warns predators that they are toxic, distasteful or dangerous. The toxicity of these organisms differentiates from species to species, with some being exponentially more dangerous than others. Their toxicity is usually contingent upon their evolutionary specialization and their genus’s specific niche.


“Flabellina iodinea” by Jerry Kirkhart [CC by 2.0]

Bivalves

Class Bivalvia consists of molluscs that have two connected shells such as Oysters, Clams, Mussels, Scallops and many more. Most bivalves are enjoyed as delicacies, despite the fact that they are benthic level filter-feeders. As others in the phylum Mollusca, bivalves have a shell that is made up of deposits of Calcium Carbonate. These deposits are derived from substances in the water and harden over-time.


“Clam Shells” by Jason Charbonneau [CC by 4.0]

Visible among the many aspects of an open clam are the posterior and anterior adductor muscles. The function of these muscles is to hold the shell closed as a defense from predators. The strength of these muscles given their relative size is immense. The mantle is a layer of tissue that overlays the visceral mass of these organisms and is directly connected to the shell. The foot of the bivalve is directly responsible for its movement. The muscular foot will emerge when the shell is opened and pushes the organism along or into the benthos or substrate.

Cephalopods

Cephalopods exhibit several similarities but also distinguishable differences from other molluscs. They sometimes exhibit a calcium carbonate shell. Squids and Octopus lack this feature, but the more primitive Nautilus does have this feature. Species in the Class Cephalopoda contain a large closed circulatory system and prehensile arms/tentacles that encompass a mouth, complete with a beak and radula. One of the most important evolutionary aspects of these organisms are their large complex eyes. These eyes are specialized for improved sight at depths where little ambient light reaches. The mantle of these organisms forms a sizable ventral cavity containing ctenidia. A portion of the mantle also forms a muscular funnel. Water is taken up and forced through these chambers under pressure creating a unique form of jet propulsion. There are over 900 living species of cephalopods that inhabit the world today.


“Nautilus shell” by Jason Charbonneau [CC by 4.0]


“Squid, December 9, 2013” [CC 0 Public Domain]

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1. Brusca, Gary J., Brusca, Richard C. 2003, Invertebrates 2nd ed. ISBN 0-87893-097-3

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15.9: Phylum Echinodermata

Echinodermata

The Phylum Echinodermata is distinguished by characteristics such as spiny-skin, pentaradial symmetry, and an endoskeleton composed of calcareous ossicles. Given that all species in this phylum are exclusively marine dwellers, they also evolved a specialized water-vascular system. This includes several canals, that comprise part of a hydraulic system for functions such as the extension of limbs, movement, nutrient distribution, gas exchange and feeding [4]. Another fascinating aspect of Echinoderms is that the water vascular system also includes the tube feet with powerful suction capability. Due to the structural makeup of these organisms, largely in part to their calcareous plates, they dry and preserve well as both specimens and fossils.


“Live sea urchin specimen in salt water tank, 2019” by Jason Charbonneau [CC by 4.0]

Asteroidea

The photo below features a dried sea star. Sea star is a broad ecological term that refers to the entire Class Asteroidea. Sea stars typically have five arms, which they rely upon for movement; these animals also utilize their limbs and the various forms of tube feet on them for the capture and drawing in of prey. Their diet includes but is not limited to: gastropods, bivalves and many annelid worms. Interestingly enough, most Asteroidea in nature protrude their stomach outside of the body through the mouth to digest prey externally. This allows them to digest organisms they have captured and rest on top of. The aboral surface of the specimen shows the spiky-skin described above, jutting out from calcareous plated dermis. The small white structure, slightly off center to the left is the sieve plate or madreporite. This structure is the exterior opening to the water-vascular system of the sea star.


“Aboral view, sea star, 2019” by Jason Charbonneau [CC by 4.0]

The oral surface shows many interesting distinctions. Among them, ambulacral grooves lining the arms. Lining the ambulacral groove are several dozen podia, which are also called tube feet. These podia are the primary mechanism for movement and predation. In the aboral view shown below, the tube feet as the tiny black structures that appear sunken into the organism (this is due to the fact that this specimen is long dead).


“Oral view, sea star, 2019” by Jason Charbonneau [CC by 4.0]

Ophiuroidea

The class Ophiuroidea (brittle stars and basket stars) is related to the asteroids but they are more morphologically slender, the arms do not contain tube feet, and they are distally stretched from the central disc. In the case of the basket stars, the structure of their limbs is highly branched, resembling capillary systems, or an entire conglomeration of roots. The complexity of these limbs may be attributed to evolutionary advantages from an increase in the surface area to volume ratio for gas exchange and feeding. The ability to reach out and capture large zooplankton prey without excess expenditure of energy is favorable to these organisms. The brittle star on the other hand exhibits an almost-serpentine like movement of the arms, giving them the nickname serpent stars. The diet of these organisms is very generalist, as they will consume essentially anything that they can come in contact with.


“Brittle Star, 2019” [CC 0 Public Domain]

A special note in the story of the brittle star comes from recent reports that some brittle star species may actually share a symbiotic relationship with corals. An observational study in 2010 showed deep sea corals in the Gulf of Mexico covered in a substance called floc, which is speculated to be an odd combination of trace amounts of oil, dispersants, and excess mucus from stressed or sickly corals. Brittle stars were observed brushing off this substance which would have likely otherwise lead to the death of some very old coral reef colonies [Smithsonian-Ocean]. The extent and mechanism of this relationship is still not fully understood.

Echinoidea

Echinoidea is the class of Echinoderms that includes sea urchins, sand dollars, sea biscuits and others. The spines observed on these organisms are actually mobile, which serves to enhance protection, feeding, and aid in movement. Echinoidea are encased in an endoskeleton commonly called a test. Much like other endoskeleton exhibiting marine dwellers, the test is comprised of calcium carbonate [3]. Other morphological characteristics to note are that sea urchins exhibit longer spines than other members of the class such as heart urchins and sand dollars where the latter are very short relative to the organism's size [4].


"Oral surface of dried sea urchin, 2019" by Jason Charbonneau [CC by 4.0]


"Aboral view of a heart urchin test, 2019" by Jason Charbonneau [CC by 4.0]

Holothuroidea

Sea cucumbers are a unique, divergent class from the rest of the Echinodermata. Unlike their counterparts, their 'skeletal' structure is greatly reduced, resulting in a soft body that is malleable and has many captivating traits. Without any arms, class Holothuroidea relies on small tentacles that surround the mouth. They prey on minuscule food items that are afloat in the oceans or that rest on the benthic level of the ocean floor [4].


"Sea cucumber suspended in preservitive, 2019" by Jason Charbonneau [CC by 4.0]

Some unique aspects of sea cucumbers lie in their rather interesting defense mechanisms. If touched or squeezed, they will rapidly shrink in size and project water from several pores all over their body which hardens as they shrink. They can also expel all of their internal organs allowing predators to feed on them, but later they can regenerate these parts whichis nothing short of astonishing. They also have mutable connective tissue and can quickly change the shape and texture of their body.

Crinoidea

The Class Crinoidea includes the feather stars and sea lilies. The defining characteristic of this class is that they anchor themselves to a substrate through the use of cirri. These cirri are attached to a long stalk which keeps them in place, as most of the species comprising this phylum are sessile. There is an exception to this general rule in the case of *Analcidometra armata*, commonly called the swimming crinoid. This is a rare group of Crinoids that may be observed across the Caribbean; they use their ten arms to gently manipulate water as a medium in order to become mobile [Marine Species Identification Portal]. These Crinoids are threatened species and if seen they should not be touched!


"Feather star, 2006" by Nick Hobgood [CC by 3.0]

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15.10: Phylum Chordata

Four diagnostic features characterize species in the phylum Chordata: 1) The notochordis a malleable rod running the length of the organism's body, to which the rest of the skeletal structure relies upon for foundational support; 2) the presence of a tail extending past the anus; 3) a hollow, dorsal nerve cord (becomes the spinal cord in humans!); and 4) pharyngeal gill slits, with the ability to be modified for specialized functions in mature vertebrates. Though this might sound outlandish, but since we are chordates, even all human fetuses early in development have gills!



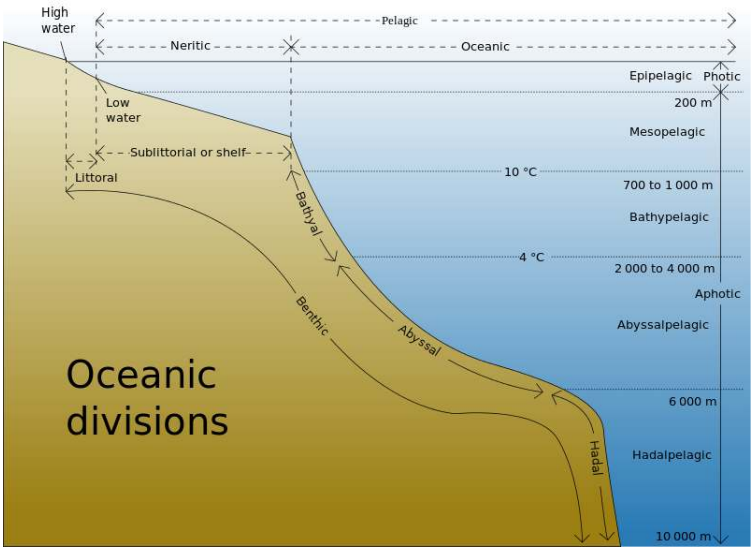
"9-Week Human Embryo, 2000" by Ed Uthman [CC by 2.0]

There are many classes that comprise the phylum Chordata. These classifications and their relationships are constantly reassessed with taxonomic research and our continued development of our understanding of life on Earth. There are many marine chordates. They include invertebrates, such as sea squirts, and vertebrates, such as fish, seabirds, sea turtles, and marine mammals.

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15.11: Zones of Marine Environments



Profile view of ocean zones (WikiMedia)

Pelagic

This zone, also known as the open ocean, is comprised of oceanic water that is not in direct contact with the shore or sea floor. The pelagic zone is subdivided into vertical zones, based on factors such as sunlight amount. While these waters are known as the "deserts of the sea", a wide variety of organisms still call the open ocean their home.

Epipelagic

This zone, known as the "sunlit" zone, begins on the surface of the water and extends to the dark, sunlight-less area where photosynthesis ceases in production. This layer is heated by the sun and is mixed vertically downward by winds on the surface. This mixing allows for a distribution of heat in this layer around the world, from super warm in the Persian Gulf (97 degrees Fahrenheit) to super cold up at the North Pole (28 degrees Fahrenheit) (National Weather Service)!

At the bottom of this layer lies the thermocline. The thermocline is a region of rapidly decreasing temperatures with increasing depth. The depth of the thermocline varies in space and time and does not remain uniform across the globe as it is dependent on the season and year. The variability of this line can be altered by unusual weather patterns, like a monsoon, or local impacts (Pusparini et al., 2017).

Mesopelagic

The mesopelagic zone reaches 200-1000m in depth. Technical deep-sea divers reach the 300m mark (National Weather Service). The mesopelagic zone is home to some of the most diverse fishes and behaviors. It is estimated that around 90% of all global fish biomass occurs in this zone (Broad, 2015). The populations of these organisms are so large that WWII hydroacoustic researchers were not able to penetrate the zone because of the high biomass of organisms with swim bladders. Common fish of the Mesopelagic include hatchetfish, lanternfish, barreleye fish, and bristlemouth fish (Shields, 2017). Many of the fish in this region are on the smaller side, but the megamouth shark (*Megachasma pelagios*) stands out because of its unusually massive size (Oceana). Reaching up to 13-18 feet long, megamouth shark are large filter feeders that swim with their mouth open to collect food (Oceana).

Bathypelagic

The bathypelagic zone is known as the "midnight" zone for its characteristically lightless waters. At 1,000-4,000 meters below, no light is able to penetrate the ocean water this deep, preventing any primary production. The only source of light in this zone comes

from organisms using bioluminescence (National Weather Service). Bioluminescence is the ability for animals to glow in the dark, caused by the release of energy from chemical reactions in the form of light (Roberts).

Temperatures this deep are fairly constant and hover around 39 degrees Fahrenheit. This is only 7 degrees warmer than the temperature that freshwater freezes at! As a portion of the deep sea, there are no currents that affect this zone, rendering it as a relatively stable environment. The pressure this far deep would amount to 5,800 lbs per square inch, and it would be hard to believe life can withstand this much force (National Weather Service). However, there have been multiple soft-bodied organisms found this far below, as well as some bony fish (OceanScape)! Scientists have found that marine life in the bathypelagic zone can have a mass 10 times greater than predicted (Broad, 2015). The organisms that live in the Midnight zone survive off of fallen organic matter from the zones above (OceanScape).



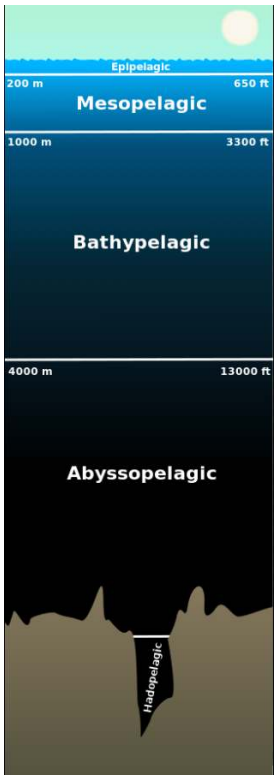
An organism that interacts with the Bathypelagic zone is the Sloane viperfish, who migrates to surface waters during the night from its location in the Bathypelagic

zone. (Wikimedia)

Abyssopelagic

The name for the abyssopelagic zone comes from the Greek word *abyss* which means “no bottom”. Early oceanographers believed that this zone did not reach the ocean’s floor. With the development of ocean floor mapping technologies, researchers know now that the abyssopelagic zone contains three-quarters of the ocean floor’s total area. Extending from roughly 4,000-6,00 meters, the abyssopelagic zone is pitch-black and home to many organisms that are specialized to live in high pressure, cold, and lightless conditions (National Weather Service).

In contrast to the epipelagic zone where photosynthesis provides the main food resources for many ocean organisms, the abyssopelagic marine life rely on detritus (dead marine organisms) to sink towards this deep layer for resources. Caused by increasing water pressure, the abyssopelagic zone is much less explored than the shallower ocean zones, and it is frequent that scientists discover new species when exploring this area. The most common characteristics of species that live in this zone include slow metabolic rates, slow consumption, flexible stomachs, large mouths, and bioluminescence. Interestingly, although these animals have unique adaptations to their environment, many belong to the same groups of continental shelf species (Brennan, 2018).



Epipelagic, Mesopelagic, Bathypelagic, and Abyssopelagic zones are depicted. (Wikimedia)

Benthic

Unlike its sister zones, the benthic (from Greek: “Depth of the sea”) zone is characterized not by its depth but by the body of water itself. The benthic zone is the lowest level of the ocean, encompassing the sediment surface and the water level right above it, starting from the shoreline and expanding to the deepest part of the ocean floor. Due to its large surface area, it is difficult to generalize the physical characteristics of the benthic zone; however, the most important and notable part of the zone lies with its ecological properties.

The benthic zone is home to a massive amount of oceanic life referred to as “benthos”, or bottom-dwellers. Out of all marine species, 98% of them can be found on the ocean floor, making the benthic zone the lifeblood of diversity in the ocean, comprised of mostly scavengers or detritivores, organisms that feed off of dead organic material. Benthic organisms found in the ocean floor itself, such as clams or macroinvertebrate, are called “infauna”. Stationary organisms or those that move along the ocean floor, such as polyps or sea stars, are called “epifauna”. Organisms like shrimp and flounders that move about freely in the water above the benthic zone but live at the bottom of the zone are called nektobenthos. Many organisms found in the benthic zone are highly attuned to life under deep-water pressure and little to no light availability, as these are the conditions found in the abyssal zone of the ocean. Due to their extreme adaptations, benthos do not travel far up the water column and use dead or decaying organic matter that drifts down from higher levels as a main energy source.

Helpful Tools:

Videos:

1. https://www.youtube.com/watch?v=UwVNkfCov1k: Just how deep does the ocean go?

More on the Benthic Zone: Helpful websites and reading sources

1. https://www.cabi.org/isc/datasheet/107788
2. https://watersheds.ca/wp-content/uploads/2016/02/Benthic-zone-Final.pdf

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12. en.Wikipedia.org/wiki/Deep_s...dus_sloani.jpg

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CHAPTER OVERVIEW

16: Plankton, Productivity and Energy Transfer

16.1: Primary Production

16.2: The Producers

16.3: Factors Influencing Production

16.4: Patterns of Primary Production

16.5: Food Chains and Food Webs

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16.1: Primary Production

Primary production is the creation of new organic matter from inorganic substrates, and it is this organic matter that serves as the base of the food web for most marine consumers. Primary production generally refers to the process of **photosynthesis,** or the utilization of light energy to produce chemical fuels that is undertaken by plants and algae according to the reaction:

$$6\,CO_2 + 6\,H_2O \longrightarrow C_6H_{12}O_6 + 6\,O_2 \tag{16.1.1}$$

Here, powered by light energy, carbon dioxide and water combine to produce glucose and oxygen. However, primary production is also carried out by bacteria in the absence of light through **chemosynthesis**. Instead of light providing the energy for the reaction, the energy comes from the oxidation of inorganic materials, such as hydrogen sulfide (see section 4.11 on hydrothermal vents). Here we will concentrate on photosynthesis because it plays a much larger role in total oceanic productivity than chemosynthesis.

The organisms responsible for oceanic primary production include a wide diversity of marine plants and algae. While many people may be more familiar with the larger seagrasses and macroalgae (seaweeds), by far the greatest amount of photosynthesis in the ocean comes from microscopic algae, the **phytoplankton**. The term “plankton” refers to organisms that drift with the currents, and the phytoplankton are the free-floating algae that undergo photosynthesis (contrast this with the **zooplankton**, who are the drifting animals). In section 7.2 we will take a closer look at the organisms responsible for oceanic primary production.

The total amount of organic material created by the producers is called the **gross primary productivity**, or **total production**. However, the primary producers consume a portion of this organic matter themselves through respiration, so the total amount that is left to support the consumers in the ocean is called **net production** (gross productivity – respiration). Gross production can be divided into two components, new production and regenerated production. **New production** is supported by nutrients brought in from outside of the local ecosystem through processes such as upwelling or ocean currents. **Regenerated production** results from the recycling of nutrients within an ecosystem.

Overall, marine productivity is similar to terrestrial production. Marine net production is about 35-50 billion metric tons per year, while terrestrial production reaches 50-70 billion tons per year. However, the biomass responsible for that production in the ocean is about 1-2 billion metric tons, compared to 600-1000 billion metric tons of biomass in terrestrial systems. So the oceans are producing almost as much organic material as terrestrial producers, but are doing it from only a fraction of the amount of producer biomass. One reason for this discrepancy is that the phytoplankton are constantly being consumed, while much of the terrestrial biomass is much longer-lived than the plankton.

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16.2: The Producers

Although the phytoplankton are microscopic in size compared to marine plants and macroalgae like seaweeds and seagrasses, they account for by far the greatest amount of photosynthesis in the oceans; about 95% of all marine primary productivity. Most of the production by phytoplankton comes from three groups, the diatoms, dinoflagellates, and coccolithophores.

Diatoms are single-celled algae consisting of cellular material inside a shell, or test, made of silica, a component of glass. Diatoms are relatively large, reaching up to about 1 mm in diameter, and come in a range of shapes, from circular disks to elongated or triangular forms (Figure 16.2.1). In some species of diatoms individual cells link together into multicellular chains. Diatoms are very efficient producers, with up to 55% of the absorbed sunlight energy incorporated into carbohydrate formation; this is one of the highest photosynthetic efficiencies known. Diatoms are most abundant in coastal and cold, nutrient-rich waters. Where diatoms are bountiful, the underlying sediments are rich in their silica shells, creating siliceous sediment and diatomaceous earth (see section 12.3).

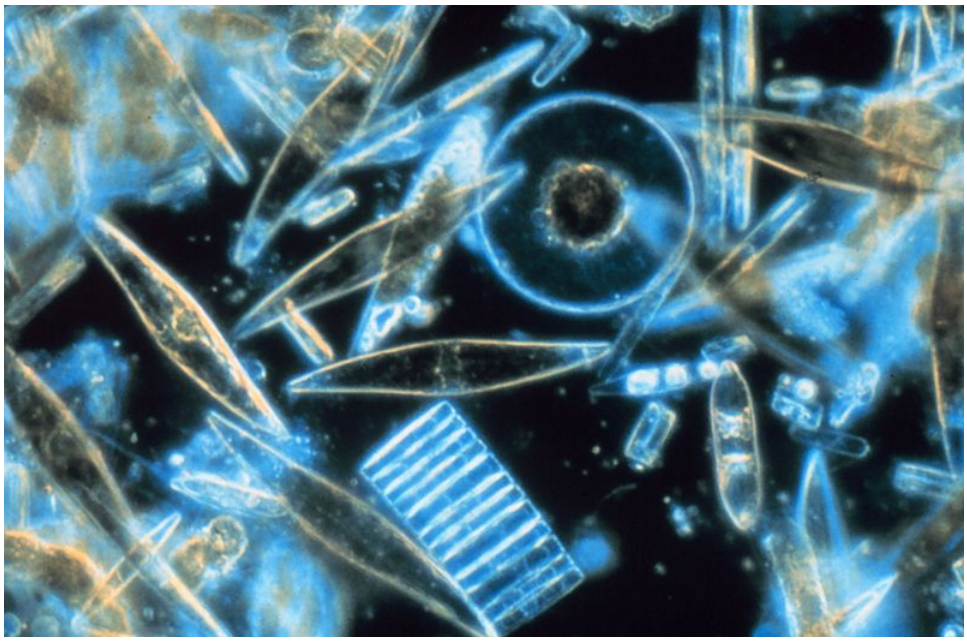


Figure 16.2.1 A variety of diatom species from the Southern Ocean, seen under a light microscope (Prof. Gordon T. Taylor, Stony Brook University, [Public domain], via Wikimedia Commons).

Dinoflagellates are another form of single-shelled photosynthetic algae that are generally smaller than diatoms, most with sizes in the 0.015 – 0.04 mm range. Most dinoflagellates have a characteristic pair of flagella (hence the name); small whip-like “tails” that they use for locomotion. Usually there is one flagellum that trails from the body to provide forward movement, and another that encircles the cell to make it spin as it moves. Unlike diatoms, dinoflagellates do not have a mineralized shell. Instead, many are covered by cellulose, which easily decomposes in seawater, so their shells do not really contribute to sediment formation (Figure 16.2.2). While most dinoflagellates undergo photosynthesis, some species will also ingest prey.

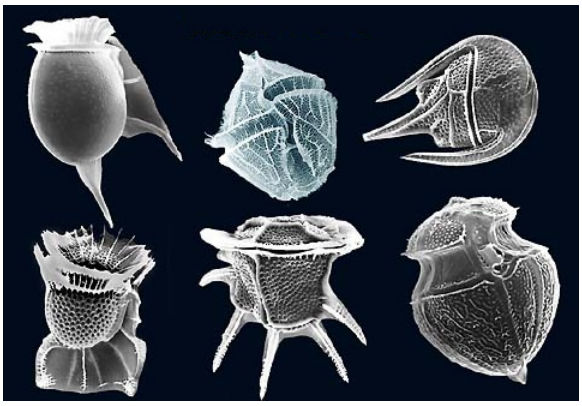


Figure 16.2.2 Electron microscope image of several dinoflagellate species. In some of the species, a groove can be seen running around the circumference of the organism; this groove is usually home to one of the flagella (fickleandfreckled, https://www.flickr.com/photos/fickle...led/6939384773, CC-BY 2.0).

A third, much smaller type of phytoplankton includes the **coccolithophores**, which range from about 5-100 micrometers wide. As with diatoms and dinoflagellates, these are single-celled photosynthetic algae that contribute significantly to oceanic primary production, but their cellular material is encased in a very different kind of shell. The test (shell) is made up of a number of interlocking circular plates composed of calcium carbonate that link together to form a sphere (Figure 16.2.3). Coccolithophores are most abundant in warm, open ocean waters, and their sinking tests can lead to calcium carbonate sediments in some parts of the ocean (see section 12.3).

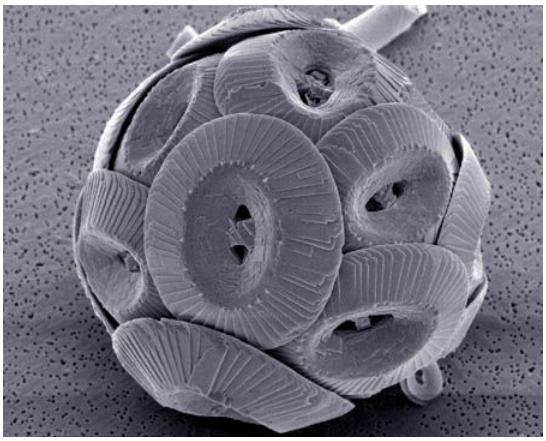


Figure 16.2.3 Electron microscope image of the North Atlantic coccolithophore *Coccolithus pelagicus* (Richard Lampitt, Jeremy Young, The Natural History Museum, London (http://planktonnet.awi.de/) [CC BY 2.5], via Wikimedia Commons).

Recent evidence suggests that another group of organisms, the **bacteria**, or **picoplankton**, may be very important primary producers. Although they are very small, on the order of 0.2-2 micrometers long, they can be found in very high concentrations, and may be responsible for up to 70% of all productivity in some parts of the ocean.

Harmful algal blooms

Primary production provides plentiful food resources for ocean consumers, so a high abundance of phytoplankton is a good thing, right? As in many other cases, too much of a good thing can sometimes be dangerous, and an overabundance of dinoflagellates or diatoms can often create serious concerns. These events are referred to as **harmful algal blooms**, or HABs. HABs can occur for a number of reasons, although a common one is an overabundance of nutrients, which is often due to excessive terrestrial runoff of fertilizers or other nitrogen- and phosphate-containing materials. These conditions lead to an explosion in algal populations that can change the color of the water if the cells are in high enough concentrations. Figure 16.2.4 shows a massive bloom that contained so many dinoflagellate cells that it turned the water reddish-brown, a so-called "**red tide**." (It has been suggested that Biblical references to seas being "turned to blood" may have actually been describing red tide events).



Figure 16.2.4 A red tide caused by dinoflagellates near the Scripps Institution of Oceanography Pier, La Jolla California (Alejandro Díaz, Public Domain via Wikimedia Commons).

These massive algal blooms can have some serious consequences. For one, when all of this algae eventually dies off and sinks, their decomposition uses up the dissolved oxygen in the water, leaving anoxic or hypoxic conditions that can lead to the mass die-off of fish and invertebrates. Dinoflagellates and diatoms are also capable of producing toxins themselves. These phytoplankton get eaten by fish, shellfish and other organisms, and in high abundances the toxins become concentrated in the tissues of the consumers. When humans or other higher-level consumers then eat these organisms, the toxins are concentrated enough to cause sickness or even death. For example, some dinoflagellates produce a toxin that causes **paralytic shellfish poisoning**, which can occur in humans as soon as 30 minutes after eating infected shellfish. This toxin attacks the nervous system, producing symptoms of dizziness, nausea, slurred speech, loss of feeling, and uncoordinated movements, and can ultimately be fatal. Diatoms produce a toxin called domoic acid that causes **amnesic shellfish poisoning**, leading to memory loss, seizures, and potentially death. Domoic acid poisoning also affects marine animals; it is thought to have been responsible for an event in Capitola, California in 1961 where flocks of seabirds acted crazily, even attacking humans. This event inspired the Alfred Hitchcock movie “The Birds.”

Additional links for more information:

- For more on harmful algal blooms, visit NOAA’s HAB site: http://www.noaa.gov/what-is-harmful-algal-bloom

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16.3: Factors Influencing Production

For terrestrial plants, many factors affect productivity, including light, temperature, nutrients, soil, and water. For phytoplankton, soil is obviously not needed, and water availability is not an issue. Temperature is generally more stable in the ocean than on land, so for phytoplankton, productivity comes down to the availability of light and nutrients.

Light

Since light is vital for photosynthesis, phytoplankton and other primary producers are limited to the uppermost layers of the ocean where light is abundant enough to sustain the reaction. As depth increases, light intensity decreases until there reaches a depth where photosynthesis can no longer occur (Figure 16.3.1). The region through which sufficient light for photosynthesis can penetrate is called the photic or euphotic zone, which extends down to about 200 m (section 6.5).

In addition to undergoing photosynthesis, phytoplankton also respire, consuming some of the organic compounds they produce. Rates of respiration are not light dependent, and respiration occurs at all depths and light levels. Therefore, as depth increases the rate of photosynthesis declines as light is diminished, until a point is reached where the rate of photosynthesis equals the respiration rate (Figure 16.3.1). This depth is the **compensation depth**, and it marks the lower level of the photic zone, and represents the depth where net primary production ends. Below this depth, there is net respiration.

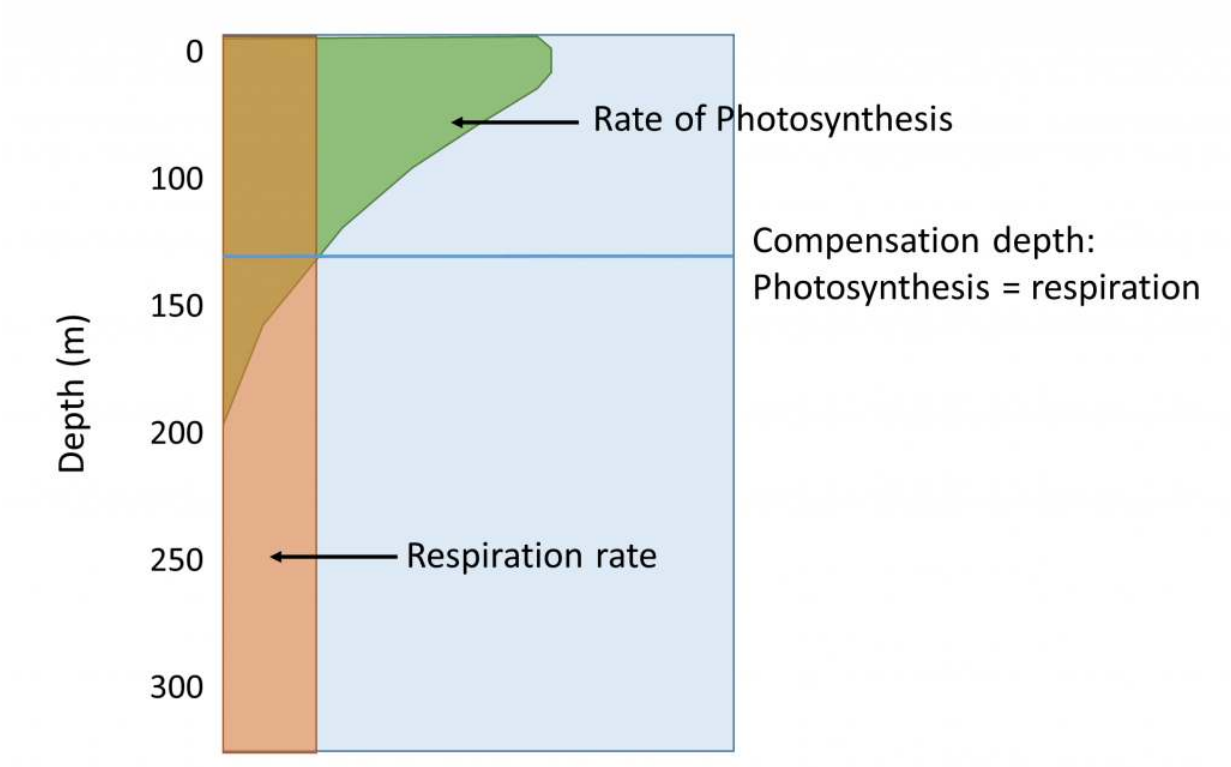


Figure 16.3.1 As depth increases the rate of photosynthesis declines as light becomes limited. The rate of respiration remains consistent at all depths. The depth where photosynthesis equals respiration is the compensation depth (PW).

Nutrients

Nutrients are required by all of the marine primary producers. The major nutrients required by phytoplankton are nitrogen and phosphorus, in the forms of nitrate NO_3^-, nitrite NO_2^{2-}, ammonium NH_4^+, and phosphate PO_4^{3-}. Many phytoplankton, particularly the diatoms, also require silica, SiO_2, for shell formation. All of these nutrients occur in very small amounts in seawater, so they are often the limiting factors for phytoplankton growth in most situations, particularly the nitrogen compounds. For example, agricultural soil contains 0.5% nitrogen in the upper meter of soil, while surface ocean water contains about 0.00005% nitrogen, 1/10,000 the amount in soil.

As we saw in section 5.6, nutrients are not distributed evenly throughout the water column (Figure 16.3.2). Near the surface nutrients are quickly utilized by phytoplankton as they become available, so surface waters are nutrient-poor. But as the phytoplankton are consumed or die they are recycled into particles of organic matter, such as fecal pellets or carcasses, that sink into deeper water. Once in deep water, decomposition of these materials releases the nutrients back into the water column. Because there are no producers to utilize them at depth, nutrients are more abundant in deeper water.

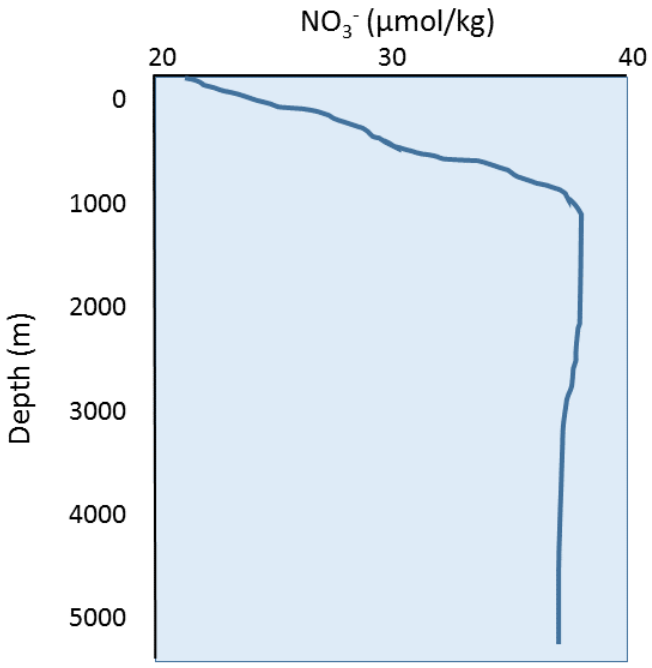


Figure 16.3.2 Representative nutrient profile in the open ocean. While this profile shows nitrate abundance, the profile is similar for other nutrients such as phosphate and silica (PW).

These deep water nutrients are out of reach of the phytoplankton at the surface. The thermocline and density stratification of the water column generally prevents the nutrient-rich deep water from mixing with the surface water. However, under certain conditions this nutrient-rich deep water may be brought to the surface through the process of upwelling (see section 9.5). Where upwelling occurs there is usually high productivity as the phytoplankton can take advantage of the input of nutrients.

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16.4: Patterns of Primary Production

Primary productivity varies both geographically and seasonally. Geographically, phytoplankton abundance generally decreases as you move from coastal to oceanic waters (Figure 16.4.1). Coastal waters are more productive than the central ocean for two main reasons. First, runoff from land often contains a high abundance of nutrients which get deposited in coastal waters and stimulate production. Second, the shallower bottom along the continental shelf can trap nutrients and prevent them from sinking to greater depths. It is easier for these nutrients to be brought back to the surface when they remain trapped in the shallows. Conversely, the central ocean generally has very low primary production, as these areas are far removed from any terrestrial sources of nutrients, and the great depth prevents the deep nutrients from returning to the surface.

Global averages for ocean surface primary production are about 75-150 g C/m^2/yr. Some highly productive areas include the California coast (200-300 g C/m^2/year), the Southern Ocean (200-400 g C/m^2/year), and the coast of Peru (200-400 g C/m^2/year), all regions with significant upwelling. The central ocean, by contrast, produces less than 50 g C/m^2/year.

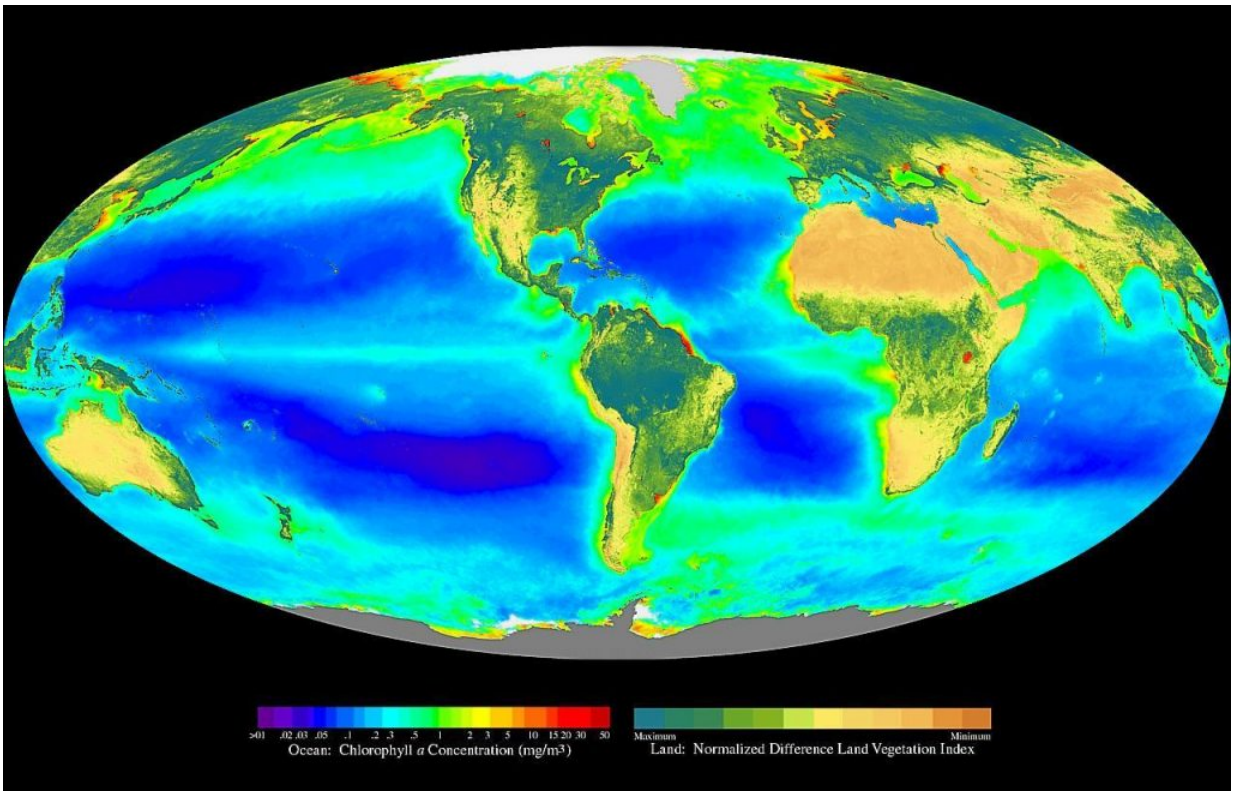


Figure 16.4.1 Global surface ocean primary productivity, as measured by chlorophyll concentration (Provided by the SeaWiFS Project, Goddard Space Flight Center and ORBIMAGE [Public domain], via Wikimedia Commons).

Regional and seasonal changes in primary production are due to a combination of the availability of light and the amount of nutrients provided by water mixing above the thermocline. In tropical regions sunlight is plentiful throughout the year, so light is not a limiting factor. The surface water is always warm and there is always a pronounced thermocline, leading to highly stratified water that prevents the nutrient-rich bottom water from reaching the surface (section 6.2). Thus productivity in tropical water is always nutrient-limited, and productivity is low throughout the year (Figure 16.4.2). Because tropical water is nutrient-poor with little phytoplankton production, the water is very clear, as is the case with water in the central ocean.

At the poles, the water is uniformly cold at all depths, so there is not much of a thermocline and little stratification, allowing mixing to occur year-round (section 6.2). This mixing distributes nutrients throughout the water column, so that for much of the year productivity will not be nutrient-limited. However, the polar regions may experience several months with little or no light during the winter, and the fluctuation in light levels leads to variation is seasonal productivity. In the winter months, mixing is occurring and nutrients are abundant, but there is no light, so there is no productivity. By late spring the sunlight returns, and combined with the abundance of nutrients, a spring/summer bloom of phytoplankton occurs (Figure 16.4.2). By late summer, the

nutrients have been depleted and zooplankton have been grazing on the phytoplankton, so the bloom begins to decline. In the autumn, light levels decline and prevent further production throughout the winter. But during the winter the mixing is distributing nutrients throughout the water, ready for the sun to return and stimulate a bloom in the following summer.

In temperate regions there is much more seasonal variation in the depth and intensity of the thermocline (see section 6.2). The thermocline is shallower and stronger in the summer, and is deeper and weaker in the winter, so mixing of deep and surface water is more pronounced in the winter months. As with the poles, this winter mixing creates nutrient-rich water during the winter, but the lack of light limits winter productivity. When light levels increase in the spring, the combination of abundant light and nutrients creates a spring bloom of productivity. By late summer there is still plenty of light, but the nutrients have been depleted by the phytoplankton bloom, and the summer thermocline has prevented further mixing, so productivity declines. In the autumn, cooler temperatures weaken the thermocline, and increasing storms cause a deeper mixed layer to form, bringing nutrients back to the surface. At the same time, there is still sufficient light available that a smaller autumn bloom occurs (Figure 16.4.2). But this bloom is short-lived, as light declines throughout the autumn and into the winter. Again, there is little production during the winter due to light limitations, but the winter storms and deep thermocline recharge the water with nutrients for the next spring bloom.

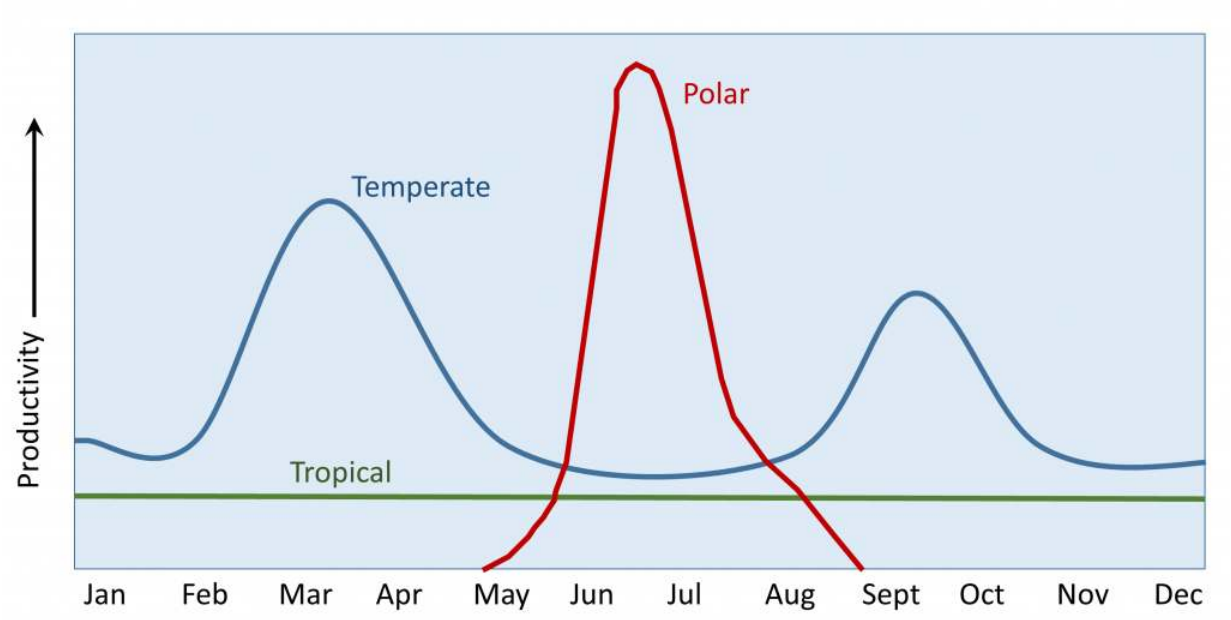


Figure 16.4.2 Seasonal patterns of productivity in the Northern Hemisphere. Tropical regions are always nutrient-limited and show low productivity. Polar regions are light limited in the winter and only display production during the late spring and summer months when light is available. Northern temperate regions have a spring bloom, and a smaller autumn bloom (PW).

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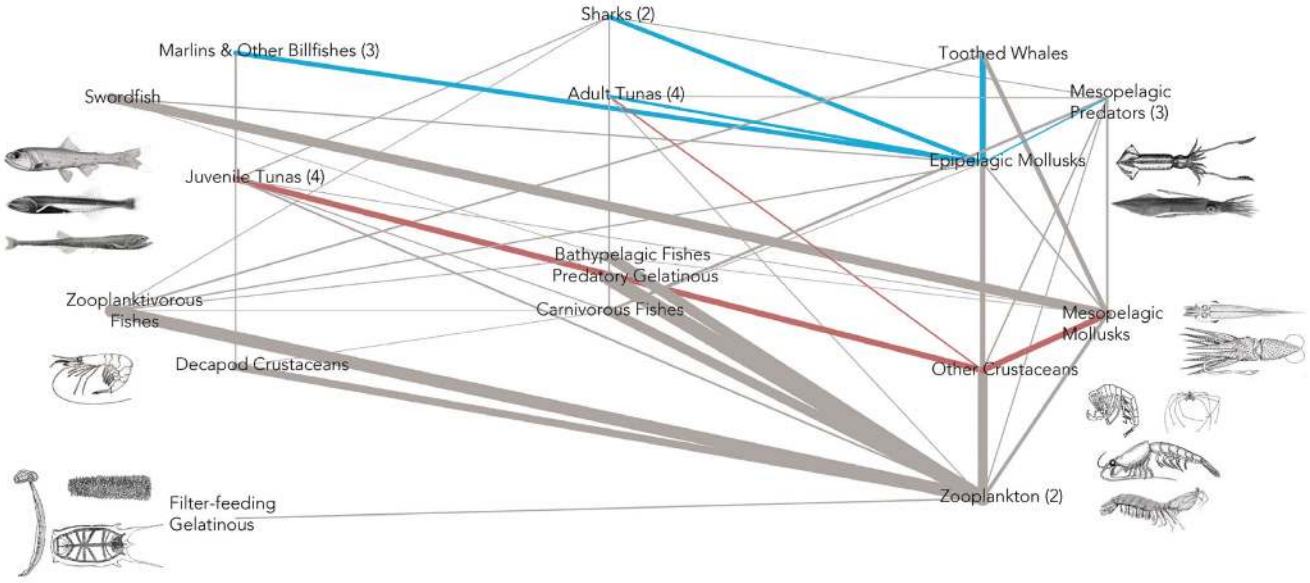
16.5: Food Chains and Food Webs

Trophic Levels

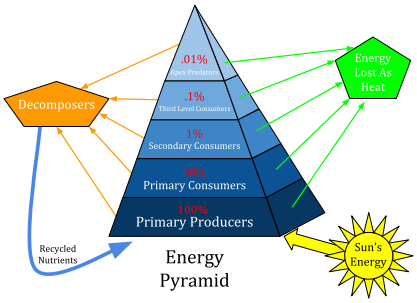
A food chain follows one path of energy and materials between species, and can be used to show feeding relationships. Here is an example of a marine food chain:

Phytoplankton --> Zooplankton --> Small Fish --> Porpoises --> Orca whales

Organisms are placed into different trophic levels, according to what they consume. Trophic levels include different categories of organisms such as producers, consumers, and decomposers. Producers are the basic trophic level while top predators are the peak level. Producers are autotrophs, meaning they produce their own food through photosynthesis or chemosynthesis. Consumers are animals that eat producers and are split into many different categories: primary consumers, secondary consumers, tertiary consumers, and more. Primary consumers are plant-eating herbivores; secondary consumers eat the herbivores; tertiary consumers can consume both the primary and secondary consumers. There may be more levels of consumers until eventually the top predator is reached. In the example above, phytoplankton are the primary producers, zooplankton are primary consumers, small fish are secondary consumers, porpoises are tertiary consumers and orca whales are quaternary consumers. It is important to note that consumers can be carnivores, animals that eat other animals, and also omnivores, animals that consume many types of food.



An example of a food web is depicted above. As you can see, a food web is more complex and is a whole system of connected food chains. Decomposers are also part of the food web and include organisms that feed on all varieties of dead plants and animals which are responsible for returning nutrients to the environment. Food webs are more complex than food chains, yet equally as useful in understanding the processes of ecological communities. Some food webs may be more complex than others but the concepts are always constant. A food web shows the flow of nutrients between different types of organisms which can help us understand the energy transfer as described before. Food webs begin with autotrophs and continue with heterotrophs, but due to their codependence, changes in abundance of one kind of organism may affect the others. For example, if the amount of phytoplankton were to suddenly decline dramatically, so would the number of heterotrophs that depend on the phytoplankton as a food source (known as 'bottom up' control of food webs).

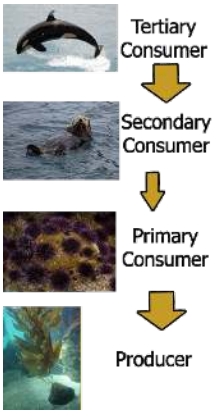


Energy Transfer

The amount of energy that flows through the different trophic levels of a food web is usually displayed as a pyramid (see above). This pyramid shows much more information than just a diagram for the hierarchy between predator and prey. The amount of area in each trophic level displays the amount of energy present in the biomass. Producers clearly occupy the largest area on the pyramid, and therefore the largest amount of energy. Autotrophs convert solar and chemical energy into the biologically usable form, glucose. In the form of glucose, energy is able to be introduced into the food web system and which is then transferred through consumption. Since autotrophs are the source of energy into the food web it makes sense that they contain the highest quantity of energy within their biomass. Although the individual primary producer is relatively tiny in itself, the vast number of producers results in their combined biomass being the largest trophic group in the ocean. The combination of their vast biomass and retention of much of the energy they produce is what leads producers to occupy such a wide base of the energy pyramid. As you move up the trophic levels the amount of energy gained from consumption decreases by a factor of approximately 10 per level. This means that primary consumers only receive 10% of the energy from primary producers when consumed. The most shocking is that the apex predator in the example above represents the quaternary predator and therefore only receives .01% of the energy that was produced by the original primary producer. Energy, then, gets trapped in unusable forms (eg. fiber and bone) and is used in metabolizing material into useable forms, leading to an overall loss.

Trophic Cascades

When you examine a food web, you can observe how all the food chains interact in one community. When observing a single food chain, you can see the path in which energy and nutrients get passed along through a specific community. Since a food chain is much more simplistic than a food web, it can be used to predict the response of an ecosystem due to changes in population of a single species. Trophic cascades are one way in which a food chain can be used to predict changes in an ecosystem. A trophic cascade occurs when one species has a change in population size, resulting in changes in populations of other species within the food chain. A classic example of a trophic cascade is the example to the right describing the relationship between orcas, sea otters, sea urchins, and kelp forests along the coasts of Alaska. It was found that an increase in killer whale predation was drastically reducing sea otter populations. Without a strong sea otter presence, sea urchin predation was low resulting in increased herbivory on kelp forests. With urchins free from significant predation, entire kelp forests were consumed resulting in urchin barrens which ultimately shifted the entire dynamic of these ecosystems. The use of food webs to predict changes in ecosystems through trophic cascades is essential to understand the full effects of humans on the natural world. They can help us better understand the best response to these cascades such as the one described above.



Foundation Species and Keystone Species

There are also organisms in a food web that are known as *keystone species* and *foundation species*. Foundation species tend to be primary producers and play a large role in the community due to their ability to build or provide a structure that other organisms

inhabit (eg. mussels and kelp forests). These foundation species include organisms that impact a community by creating physical changes in the environment. Keystone species can be located anywhere in the food web, and play a vital role in the maintenance of the community. These organisms exert major influence over a community's structure by ecological roles or niches. Keystone species are defined as those that have a disproportionate impact on an ecosystem relative to their abundance. Sea otters do not have a very high biomass in a kelp forest but through predation of other species (such as sea urchins), they have a strong impact on the health and biomass of the kelp.

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6. http://lh4.ggpht.com/-HafbEZnW9Xk/USIFDYBkA5I/AAAAAAAABO8/QEvw0xZbrwA/food-web-and-food-chain-compared_thumb%25255B4%25255D.jpg?imgmax=800
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CHAPTER OVERVIEW

17: Seaweeds and Marine Plants

17.1: Kelp Forest Ecosystems

17.2: Mangrove Ecosystems

17.3: Seagrass Beds

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17.1: Kelp Forest Ecosystems



Figure 1. depicts a kelp forest ecosystem commons.wikimedia.org/wiki/F...-_DSC06945.JPG

Distribution

Kelp forests are mostly found in cool, shallow, nutrient-rich water near coasts. Kelp is a brown alga (*Phaeophyceae*), which requires access to light in order to photosynthesize - this is the reason for their abundance in shallow coastal waters (Sanctuaries.noaa.gov 2015). Kelp is one of the fastest growing organisms in the world, growing anywhere from 2 centimeter to 1 meter each day, depending on species and environmental conditions (Montereybayaquarium.org 2015).

Kelp can be found on the coast of every continent besides Antarctica (*Figure 1)*. Kelp forest ranges include areas of cold water, generally coinciding with coastal upwelling (Sanctuaries.noaa.gov 2015). The eastern Pacific Ocean along the west coast of North America, is an ideal area for kelp forests due to the presence of upwelling bringing cold water and nutrients to the surface from deep ocean water. Towards the poles, kelp forest development is limited by the availability of light at higher latitudes, while kelp towards the equator tends to be limited by the presence of warmer waters (Steneck et al. 2002).

Characteristics

Kelp forests often draw comparisons to terrestrial forests, however the two ecosystems are distinct in that kelp is an algae - not a member of the Plant Kingdom. The structure of all kelp bears three universal morphological characteristics: holdfasts, stipes, and fronds (*figure 2*).

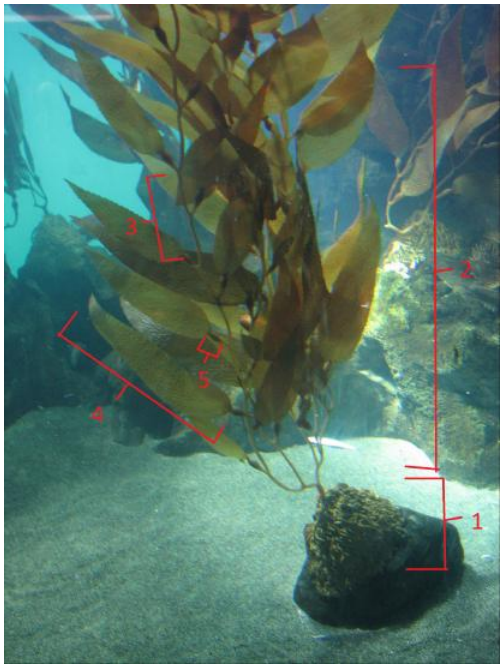


Figure 2. Diagram of kelp anatomy. Labels characteristics found in many species of kelp; 1. Holdfast 2. Frond 3. Stipe 4. Blade 5. Gas bladder. Note that gas bladders are not present in all species of kelp. upload.wikimedia.org/wikiped...rocystis_4.JPG

The holdfasts, which serve as an anchor for kelp, must be attached to hard substrates such as rocks or sand. Although they resemble roots on terrestrial plants, holdfasts do not transport water or nutrients through their stipe. The stipe, analogous to the stem, gives structural support to the algae. Fronds are blades extending from the stipe. The blades serve as the photosynthetic and nutrient uptake center for the organism (Connor and Baxter 1989).



Figure 3. Pictured above is bull kelp. The single pneumatocyst is clearly shown at the end of the stipe, with fronds extending from the end.
commons.wikimedia.org/wiki/F...eocystis_2.JPG

In addition to the universal characteristics of all kelp, many species have also developed pneumatocyst - also known as a "gas bladder" **.** Pneumatocysts are gas-filled bladders that provide the kelp with buoyancy that lifts the blades towards the surface, allowing for more photosynthetic productivity. The most common species of kelp found in Pacific forests are giant kelp (*Macrocystis pyrifera*) and bull kelp (*Nereocystis luetkeana*), both of which contain pneumatocysts.

Giant kelp is common in southern areas of the coast and bull kelp in the northern. The reason for this schism is due to giant kelp being perennial, meaning that the same organism will grow for multiple years – in this case, up to seven years – while bull kelp is an annual, which means it will complete its entire life cycle in one year. Winter storms farther north rip up kelp, washing it onto the shore, which makes bull kelp, and its one-year life cycle, much better suited to northern areas.

Ecology

Habitat

Kelp forests are an important foundation species , which provide large areas of habitat within the ocean (Oceanservice.noaa.gov 2015). Kelp forests have a wide variety of inhabitants within their boundaries: invertebrates frequently graze on the blades, fish find shelter in the fronds, and invertebrates - such as brittle stars, sea stars, anemones, sponges and tunicates - live in the holdfast (Connor and Baxter 1989). Kelp is also the single largest source of fixed carbon within their ecosystems (Graham 2004). Pieces of decomposing kelp sink deeper into the ocean, providing nutrients to deep sea organisms. Thus, kelp derived particulate organic matter (POM) is an important food source for fauna both within and outside the kelp forest (Norderhaug et al., 2003).

Since kelp is capable of attenuating waves and dampening their energy, many animals, such as mammals and birds, are able to seek refuge in the forests (Mork 1994). Various species of kelp tolerate ocean storms differently (Montereybay.noaa.gov 2015). In fact,

along the California coast giant kelp has been known to dominate in years with less turbulent sea conditions, while bull kelp is more dominant in open waters and years of turbulent conditions (Harrold et al., 1988).

Sea Otters and Urchin Barrens

Sea otter (*Enhydra lutris*) are one of the keystone species within kelp forests - meaning that they have a substantial impact within an ecosystem despite their relatively small population. Otters' diets consist mainly of invertebrates such as purple sea urchins (*Strongylocentrotus purpuratus*) which inhabit kelp forests along North America's west coast. Urchins feed on kelp and can potentially damage kelp forests by chewing through kelp holdfasts. Otters are the primary predator of urchins, keeping their populations at sustainable levels and maintaining the ecological balance of the kelp forest. Without otters a trophic cascade can occur; leading to destructive levels of grazing by urchin, causing a phenomenon known as an "urchin barren" (Sala et al., 1998, Siversten 2006). Due to kelp's role as a foundation species, significant populations of organisms that depend on their habitat are negatively impacted (Dayton 1985).

Fore more information:

Giant Kelp:

www.montereybayaquarium.org/a...ae/giant-kelp/

Bull Kelp:

www.montereybayaquarium.org/a...gae/bull-kelp/

Urchin Barren:

http://www.hindawi.com/journals/jmb/2012/492308/

Kelp Forests:

http://oceanservice.noaa.gov/facts/kelplives.html

sanctuaries.noaa.gov/about/ec.../kelpdesc.html

Kelp Forest Ecosystems:

http://montereybay.noaa.gov/sitechar/kelp1.html

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17.2: Mangrove Ecosystems



Mangrove Ecosystems

Mangrove swamps are a complex and diverse compilation of resilient species that tolerate extremely challenging conditions of living in shallow seawater. There are approximately 70 species of mangrove trees that act as the foundation of the ecosystem, providing a habitat for numerous species in their branches above water and their tangled roots below water. Mangrove trees have adapted to living in high salinity conditions by developing a filtration system in their roots, and to stabilizing the soft sediments below by creating a complex root matrix to support their often massive weight above water. Many species of birds nest, roost, and feed in the canopy of the Mangroves. Various insects, snakes, lizards, and frogs make their homes in the branches, as well. Much like a coral reef or rocky intertidal shore, mangroves also provide habitat for many invertebrate and vertebrate marine species. Crabs, anemones, mussels, barnacles, oysters, and even juvenile reef fishes can be found clinging to or swimming through the complex system of roots. The mangrove provides shelter from predators as well as protection from physical forces such as high velocities of water flow for these small critters. It is important to protect this unique type of ecosystem because it provides shelter to an incredible diversity of marine and terrestrial species.

Mangroves provide billions of dollars worth of free services to humans. Through photosynthesis, they absorb a large chunk of carbon dioxide from the atmosphere, helping to reduce green house gas effects. They filter waste water and pollution along developed coastlines and provide clean water for numerous marine species. Additionally, their sturdy root system stabilizes shores from erosion or strong storms and can even create new land by trapping sediments. Mangroves also provide direct benefits to humans in the forms of medicines, seafood, fruits, fiber, and wood. However, in the past decade alone, over 35% of the world's mangroves have been destroyed largely due to human activity. Coastal development, climate change, and aquaculture have removed mangroves at an alarming rate that is more rapid than deforestation of tropical rainforests, yet have received far less attention from activists and media. Scientists estimate that all mangroves could be lost within the next 100 years if conservation efforts are not made. The loss of mangroves would cause a collapse of balance in many other ecosystems, especially coral reefs. Juvenile coral reef fish are raised within the mangrove root system, and without the mangroves, coral reef diversity will be adversely affected. Mangrove conservation should be one of the priorities of marine science in the near future.

online resources

- http://ocean.si.edu/mangroves
- www.mnh.si.edu/exhibits/ocean...e_removal.html

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17.3: Seagrass Beds

Another important primary producer in the ocean is seagrass. Please go to this website to learn more about seagrass ecosystems: https://ocean.si.edu/ocean-life/plants-algae/seagrass-and-seagrass-beds

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CHAPTER OVERVIEW

18: Coral Reefs

18.1: Coral Reefs- An Introduction

18.2: Different Types of Corals

18.3: Reef Types and How Coral Reefs are Formed

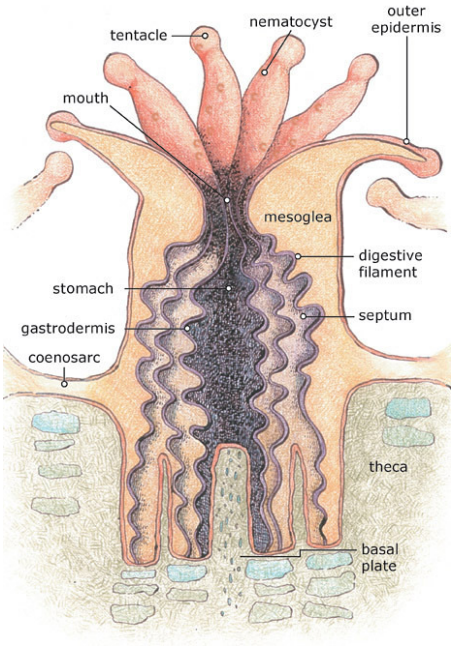
18.4: Symbiotic Relationships in Coral Reef Ecosystem

18.5: Importance of Coral Reefs

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18.1: Coral Reefs- An Introduction

What are corals you ask? " A little bit algae, a little bit rock, and a lot animal." Corals are close cousins to sea anemones and jellyfish, meaning they all belong to the same phylum of Cnidarians.Coral is a marine invertebrate that is composed of simple structures known as polyps. The polyp has one opening which is the coral's mouth that is surrounded by a ring of tentacles. These tentacles have stinging cells called cnidocytes, used for food and protection. These cells fire tiny toxic barbs called nematocysts when they are touched to ward off predators or capture food. Inside the polyp are reproductive and digestive tissues. Coral can live as an individual polyp, but most corals exist as large colonies of interconnected polyps.



"Coral Polyp" by Wikimedia Commons under CC by 2.0

How do Corals Survive?

Sunlight: Corals grow in shallow water where the sunlight can reach them. The algae that live inside of them, zooxanthellae, need sunlight to survive, since the coral animal depends on the zooxanthellae, corals need sunlight to survive.

Clear Water: Clear water is needed for corals to survive because it lets the sunlight in. When the water is opaque the corals will not thrive.

Warm Water Temperature: Most corals, though not all, require warm waters to grow and survive. Corals general live in water temperature of 68-90°F

Clean Water: Corals are very sensitive to pollution and sediments that can be in the sea. Sediments can create cloudy water conditions which won't allow for sunlight to get through which will harm the polyps. Corals do not typically do well in environments where there is little sunlight, except for the deep-sea corals. Another issue that may arise is the addition of too many nutrients. An overflow of nutrients in the water can cause things such as algae to grow, in turn taking over the reef. Sediments and pollutants can also cause direct harm to coral tissue.

Salt Water: Corals require a certain balance between the concentration of salts and water (salinity).

Nutrients: Though zooxanthellae do provide some nutrients, the coral animals still need to receive other sources of nutrition. Corals can also grab passing nutrients (like zooplankton) by using their tentacles to snag passing organisms flowing by.

Coral Reef Functions:

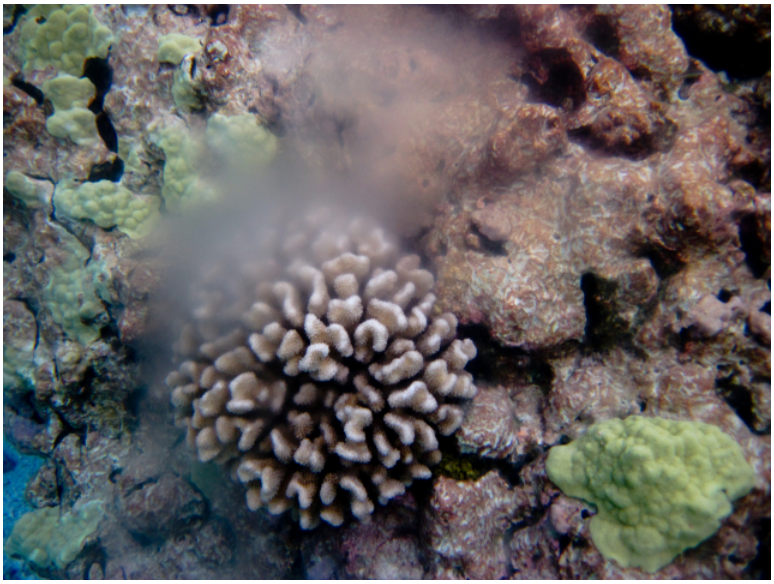
Protection: Functions as a form of protection for the shores from harsh waves and storms

Habitat: Provides shelter and safety for many organisms

Nutrients: Many forms of heterotrophic marine life are supported from the primary production of the coral zooxanthellae. Carbon and nitrogen-fixing is an important part of the marine food chain. Nutrient cycling within the coral polyps is very efficient and contributes to the high production of reefs.

How do Corals Reproduce?

Some species have distinct male or female polyps while other coral can alternate gender. Many coral reef species only reproduce once or twice a year. For some coral that only have single-gender polyps, certain events (i.e. full moon) trigger the polyps to release massive amounts of sperm or eggs that flood into the seas, where the gametes eventually come together. Most coral species spawn by releasing eggs and sperm into the water; but the period of spawning depends on the type of coral. When an egg and sperm meet, the fertilized egg eventually develops into a larva called planula. Planula can form by two different ways, either fertilized within the body of a polyp or fertilized outside the polyp, externally in the water. Fertilization of an egg within the body of the coral polyp happens when sperm that is released through the mouth of another polyp, is taken in by a polyp with an egg. When the larva is matured enough the mother will spit it out via the mouth. If fertilization is external, corals will eject large amounts of egg and sperm into the water which eventually find each other. The release of larvae or gametes is known as coral spawning. Most species also reproduce asexually, budding new polyps or fragments, that grow either on an older polyp or drift around before landing on a new surface.



"Cauliflower coral spawning, French Frigate Shoals" by USFWS – Pacific Region under CC by 2.0

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18.2: Different Types of Corals

What Types of Coral Exist?

When people think about coral they often think about the stony calcified coral they get at a gift shop during a vacation to a tropical island. However, these are only the skeletons of stony corals, and are only one type of coral that exists. There are also deep water and soft corals. Each of these types of corals is unique structurally and in how they interact with their surroundings. There are over two thousand different types of coral which form colonies that play a key role in marine environments.

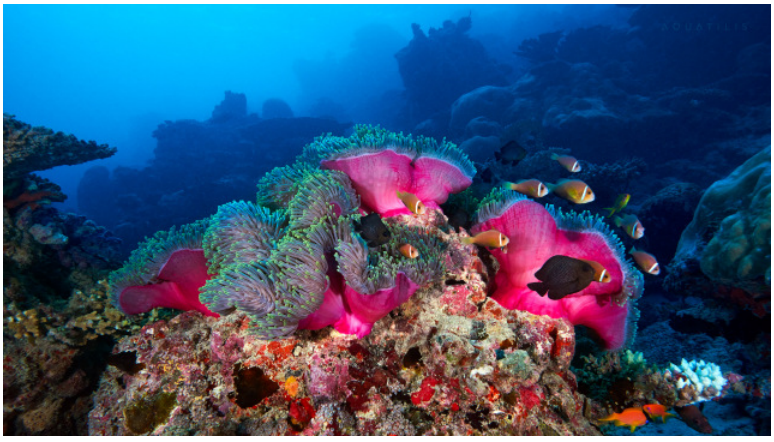
1. Hard Corals



"Hard Corals" by Amal FM under CC BY-SA 2.0.

Hard coral is a fundamental part of building the coral reef. The polyps of hard corals secrete skeletons of calcium carbonate (Limestone), that will eventually become rock. Hard coral includes species such as brain coral and elkhorn coral. Hard coral are considered to be hermatypes, which are reef-building corals. These corals require a certain type of algae called zooxanthellae that live within their tissues, this important mutualism is necessary for their survival. These organisms are what give corals their colors, which is dependent upon where they are living. The shape and appearance of each coral are dependent upon its species type, location, depth, water movement, and many other factors. Hard coral polyps are very similar to sea anemones, as they also have stinging cells called cnidocytes. Hard coral species can be found in all of the world's oceans, yet their populations are expected to decrease due to global changes and ocean acidification.

As mentioned earlier, hard corals build coral reefs. A coral reef begins when the coral polyp attaches itself to a rock on the seafloor. This one coral polyp then begins to divide or bud into thousands and thousands of clones of itself. Polyps and their calcareous skeletons connect to one another which creates a colony that will act as one. As this colony grows over hundreds and thousands of years it will join with other colonies and become a reef. Read more in our chapter on coral reef formation.



"Maldives reef" by Alexander Semenov under CC by 2.0

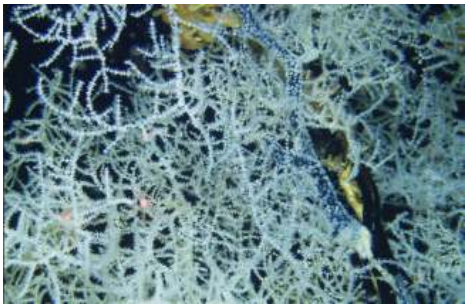
2. Soft Corals



“SOFT CORALS” BY NAJIM ALMISBAH UNDER CC BY-SA 2.0

Soft corals often resemble plants and trees. They are soft and bendable so they do not have stony skeletons like hard coral. For protection and support, they grow woodlike cores for stability. These cores are made of structural proteins such as gorgonin and other protein similar to those of nails and horns of other animals. These types of coral are referred to as Ahermatypes(non-reef building corals). Examples of soft coral in the Bahamas and Caribbean include sea fingers or sea whips. This type of coral also does not always have a symbiotic relationship with zooxanthellae. Though many utilize their presence, soft corals will typically eat any type of passerby out of the water column. These types of corals do not produce calcium carbonate, rather they contain spiny skeletal sclerites. These give the softer corals some protection and support. Soft species also prefer to live in nutrient-rich waters with less intense light. In other parts of the world, fleshy true soft corals have no rigid internal skeleton at all.

3. Deep-Sea Corals



DEEP SEA CORAL, WHITE “BLACK CORAL” *LEIOPATHES GLABERRIMA* BY NOAA PHOTO LIBRARY UNDER CC BY 2.0

Deep-sea coral can be found in the dark depths of up to 6,000 m (20,000 ft) below the surface of the ocean. These corals live in icy cold water with little to no sunlight. These corals, like shallow species, can exist as single polyps or multiple, living in complex colonies made up of different species. Since these species do not require sunlight or warm water, they are able to grow in a vast array of waters around the world. They have even been found in waters as cold as -1-degree celsius. Since these corals live without sunlight they contain no zooxanthellae. This means these corals must obtain their energy and nutrients elsewhere. They do this by trapping tiny organisms in passing currents. The main reason many scientists had no idea of the existence of these deep-sea corals, is because, for many years, the oceans deep depths were inaccessible. In addition to being surprisingly diverse, scientists have also discovered that deep-sea corals are amongst the oldest marine organisms on record. Since corals are constantly growing, regenerating new polyps, some coral reefs have been actively growing for almost 40,000 years. To your and my surprise, scientists have identified nearly as many deep-sea corals as shallow-water species.

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18.3: Reef Types and How Coral Reefs are Formed

Have you ever wondered how coral reefs are formed? It has been established by the scientists in the Tropical Marine Biology field that coral reefs can be divided into three types: Fringing reefs, Barrier reefs, and Atolls which all require very specific conditions to develop. Coral reefs begin to form when free-swimming coral larvae attach to a rock or other submerged object on the edges of islands. As the corals grow, over very long periods of time, they form into a reef. The three types of reef represent stages in development of a coral reef over time.

1. Fringing Reefs:

Fringing reefs grow near the coastline around islands and continents. These types of reefs are the most common type we see and are considered to be the youngest of the 3 types of coral reefs. They are separated from the shore by shallow lagoons. The first stage of the formation is when the coral larvae attach itself to rocks or soil near the coasts. In certain parts of the world, they tend to form where volcanoes have also formed due to the shallow sloped walls being ideal to make shore reefs. The larvae become polyps and excrete calcium carbonate, which forms their exoskeleton. The secreted calcium carbonate sediments on the rocks and provides a substrate for more polyps coming to attach themselves. As more and more polyps attach, and layer over time, they create a coral reef. Calcareous Algae also add their sediments to the structure. Other organisms with calcareous skeletons, also add their own remains to the reef as they die and sink. Since fringing reefs are the youngest and grow onshore, they tend to have less diversity of species within that ecosystem. Some locations of these reefs include Kenya, Australia and other parts of Africa shores.



Fringing Reef in Eilat, Israel
Photo by: Mark A. Wilson [CC by 2.0]

2. Barrier Reefs:

As the name states, these reef types border along coastlines with a very wide and deep lagoon separating the two structures. The term “barrier” comes from the more shallower coral that sticks out of the water and creates a barrier or wall-like structure. One of the most famous and largest barrier reefs is “The Great Barrier Reef” in Australia. This reef is around 1200 miles and consists of many complex reefs making it up. This type of reef is formed when the fringing reefs slowly combine with each other and form a borderline along the coast. The calcium carbonate structures attract more polyps and the spaces are filled up. It forms a line along the coast and a ring around an island. Large barrier reefs are the rarest type of coral reef and only seen in a select few places on earth such as Belize and parts of the southern Pacific Ocean. Smaller barrier reefs are seen in places where islands are in an earlier stage of becoming submerged.



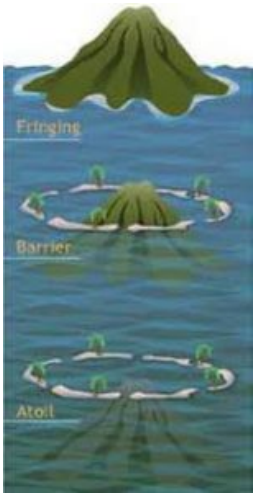
Great Barrier Reef by Studio Sarah Lou by Flickr

Most of the reefs surrounding the Turks and Caicos Islands are barrier reefs that formed from the precipitation of $CaCO_3$ on the remnants of continental debris resulting from the separation of North America from South Africa 200 million years ago (continental drift). Dissolved minerals in the seawater form small particles called ooids that become cemented together to form oolite rock which makes up most of the Turks and Caicos islands and bank.

3. Atoll Reefs:

An Atoll Reef is a ring-like shaped coral reef or small islands of reefs in a circle with a lagoon in the middle and are usually located in the middle of the sea. These types of reefs are formed when an island has sunk completely (or nearly) in the middle of the ocean from rising sea levels around a pre-existing structure (these islands are often the tops of underwater volcanoes). Most atoll reefs can be found in the Pacific ocean where the waters are warm and salty. They can grow between the sizes of 12.6 km^2 (Great Chagos Bank) and as small as 1.5 km^2 (Ontong Java reef). More animals are found in an atoll than in earlier stage reefs. Over a million species can live in one square kilometer of coral reef. The three types of reef from fringing to barrier to atoll, form in this sequential order and represent stages of development of a reef.

Watch a great animation clip of how an atoll forms.



"Formation of Coral" by NOAA Ocean Service Education via Wikimedia Commons

Below is a link to a video created by three undergraduates at Keene State College who discuss more about the different types of coral reefs:



A YouTube element has been excluded from this version of the text. You can view it online here: http://pb.libretexts.org/marine/?p=35

4. Patch reefs

Patch reefs are small, isolated reefs that grow from the open bottom of the island platform or continental shelf. They usually occur between fringing reefs and barrier reefs. They vary in size and rarely reach the surface of the water.


Patch reefs in Hoopers Bay area by James St. John Via Flickr

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18.4: Symbiotic Relationships in Coral Reef Ecosystem

Coral reefs are home for many organisms such as sponges, fish including large nurse sharks and reef sharks to groupers, clownfish, eels, snappers, and parrotfish, jellyfish, anemones, crustaceans, other invertebrates and algae. So, how do coral reefs support such a huge weight on their shoulders? It is the symbiotic relationship that is formed when two different species interact with each other. These interactions create a balance within the ecosystem because at least one of the species is gaining from it. The other species may also gain from the relationship, be unaffected or even get harmed from the relationship. Symbiotic relationships are very common in the ocean, especially near coral reefs. There are three main types of symbiotic relationships. They are mutualism, parasitism, and commensalism. Mimicry is also frequently seen amongst coral reef organisms.

Types of Mutualism:

There are two primary types of mutualism: obligate mutualism and facultative mutualism.

Mutualism, or a mutualistic relationship, by definition, is when two organisms of different species work together so that each is benefiting from the relationship.

- Obligate mutualism is when neither organism can live without one another.

An example of obligate mutualism is the relationship between ants and Acacia plants. While the plant provides shelter and food for the ants, the ants actually defend the plant from organisms such as other herbivores that may eat the plant, as well as remove any other species of plants that may limit the plant's growth.



"Acacia Ants" photo via Wikimedia Commons under 2.0

A more specific example of obligate mutualism that is more related to this topic would be the relationship between hard coral and algae (zooxanthellae). The relationship between coral and zooxanthellae (algae), is one of the most important mutualistic relationships within the coral reef ecosystem. Zooxanthellae are microscopic, photosynthetic algae that reside inside the coral. The hard coral provides protection, as well as compounds needed for photosynthesis to occur. In return for their protection for herbivores and other organisms, zooxanthellae photosynthesize organic compounds from the sun, and then pass the nutrients, glucose, glycerol, and amino acids, which are the products of photosynthesis, to their coral hosts, essentially giving the coral reefs their beautiful colors. The corals then use those nutrients to produce proteins, fats, carbohydrates, and calcium carbonate. This is so important, in fact, approximately 90% of the nutrients produced during the photosynthesis in zooxanthellae is transferred to the coral for their use. Zooxanthellae also aid in the excretion, or removal of waste such as carbon dioxide and nitrogen. Ultimately, without algae, coral would starve to death (coral bleaching), and if algae didn't have protection, they would be more vulnerable to several herbivores and other organisms. This relationship is so important, that if this mutualistic relationship did not exist, it would be very likely coral reefs would not even exist. Therefore, making this relationship obligate mutualism, as mentioned before.

Image result for coral and zooxanthellae symbiotic relationship " CORAL POLYP " BY EMAZE

The mutualistic relationship between anemones and clownfish is also another commonly known relationship. Clownfish are found in warmer waters of the Indian and Pacific oceans. Of the over 1,000 anemone species that live in the ocean, only 10 species coexists with the 26 species of tropical clownfish. Within these species, only select pairs of anemone and clownfish are compatible. Sea anemones are actually predators, with stinging polyps, that attach themselves to rocks, the ocean floor, or even coral. They

patiently wait for fish to swim by close enough to get entangled in their poisonous tentacles. The toxins paralyze their prey, and the tentacles guide the prey into the anemone's mouth. However, clownfish are the exception and actually call the anemone home. Clownfish are coated with a mucus layer that essentially makes them immune to the deadly sting of the anemone. Therefore, clownfish are able to live within the anemone's tentacles, while also gaining protection from predators, and the clownfish helps feed the anemone by either letting them eat their leftovers, or by also luring fish over to the anemone, so that the anemone can catch them with their poisonous tentacles, and eat them for dinner (or maybe lunch).



"Clownfish and Sea Anemone" by Samuel Chow under Flickr

- Facultative mutualism the other type of mutualism, is when species benefit from one another, but do not necessarily fully depend on one another.

An example of facultative mutualism is the relationship between certain types of our gut bacteria, or the bacteria that live in our digestive tracts, and us humans. When we eat food, bacteria use some of the nutrients from that food we are actually digesting, and in return, they help us digest our food.

Again, a more specific example of facultative mutualism that is more so related to the coral reef ecosystem, is the relationship between shrimp or smaller fish and large marine organisms. The shrimp or cleaning fish remove materials, such as parasites, off of the larger marine organisms, in which they get a meal from, and the larger marine organisms have potentially harmful parasites removed!



Goatfish, *Mulloidichthys flavolineatus* at Kona, Hawaii is being cleaned by two cleaner Wrasses, *Laborites phthirophagus photo via* Wikimedia Commons under CC 3.0

As mentioned before, earlier on in the post, smaller fish or cleaner shrimp, such as the Bluehead Wrasse or Spanish Hogfish remove parasites and other materials off larger marine organisms such as fish, sharks, and rays. In most cases, these smaller fish would typically be the larger marine organism's prey, however, in this case, these larger organisms gain the benefit of having these parasites removed, that could potentially cause harm, while the smaller fish or shrimp get a meal.

Cleaner fish and larger fish share a mutualistic relationship. This is because the cleaner fish eats harmful parasites and other small sources of food off of the large fish. This gives the cleaner fish a meal, the larger fish is helped because it no longer has these parasites on them. Often times larger fish wait in "**cleaning stations"** for the cleaner fish to come and get these things off of them. Some small shrimp can also be cleaners. The picture below shows a cleaner shrimp cleaning a large fish at a cleaning station that would normally eat the shrimp if it wasn't for this mutualism.



"DANGEROUS DINING" BY CHRIS LEWIS UNDER VIMEO

Another facultative mutualistic relationship is between the root-fouling sponge called *Tedania inis*, and red mangrove called *Rhizophora mangle*. In this relationship, the red mangrove provides the sponge with carbon that was produced by the mangrove, and the nitrogen the sponge releases gets eaten up by the mangrove to enhance growth.

Mutualism also occurs between spider crabs and algae. This relationship benefits both of these species because the greenish-brown algae live on the spiders back, which helps the spider crab blend into the shallow areas of the ocean floor where they live. In return, the algae benefit from a good place to live.



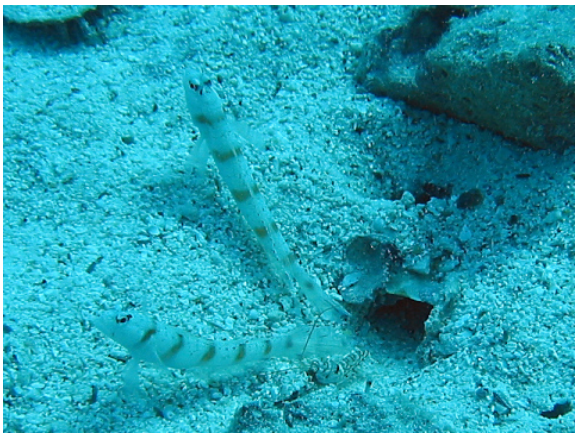
"JAPANESE SPIDER CRAB" BY (OVO) UNDER FLICKR

Another example is the relationship between the Boxer Crab and anemones. In this relationship, the Boxer Crab carries around two anemones that sting and it uses them for protection. The anemones are benefited because since the crab carries them around, it allows them to be mobile which increases their options for finding food.



"BOXING (POM POM) CRAB" BY LIQUIDGURU UNDER VIMEO

One last mutualistic relationship is the relationship between a goby (*Nes Longus and Ctenogobius saepepallens*) and a snapping shrimp (Alpheus floridanus). The shrimp dig a decent sized burrow in the floor of the ocean, and the goby will then live in the entrance of that burrow. When the shrimp exits the burrow, it will stay in contact with the goby through its antennae, and depending on the species of the goby, it will either signal to the shrimp of approaching predators by darting headfirst back into the burrow or by flicking its caudal tail. Ultimately, the goby gets a free place to live and hide from potential predators, while in return the shrimp gets a look-out individual while it hunts for food!



"Goby fish with shrimp" photo via Wikimedia Commons under public domain

Parasitism:

Parasitism is not a mutualistic relationship because only one of the species is benefited. The parasite gains from the relationship while the other species involved is harmed. Ectoparasites live on the outside of the host body, whereas endoparasites live inside the host.

One example of a ectoparasitic relationship is between fish lice and small fish hosts. The fish can be killed if there are too many fish lice attached to it. The lice benefit from the fish by feeding off of their bodily fluids.



"SEA LICE ON SALMON" BY 7BARRYM0RE UNDER **PUBLIC DOMAIN**

Isopods can also cause be involved in a parasitic relationship. Some isopods will eat the fishes tongue and then live in the fishes mouth so they can eat whatever the fish is attempting to eat.



"BETTY IN MOUTH" BY UNIVERSITY OF SALFORD PRESS OFFICE UNDER FLICKR

Commensalism:

Commensalism is a relationship where one species benefits from another species. The other species is neither harmed nor helped in this relationship. There are many examples of commensalism in the ocean.

One example of commensalism among marine life is jellyfish and small fish. The small fish will typically hide inside of the jellyfish's stinging tentacles if the stinging does not affect them. The tentacles provide protection for the fish from larger predators. This relationship has no effect on the jellyfish.



"BABY FISH TAKE SHELTER IN JELLYFISH" BY EARTH TOUCH NEWS NETWORK

Another relationship is between shrimp and a featherstar. The shrimp will blend in with the featherstar and use it for protection. As you can see in the picture below, it is very difficult to find the shrimp hiding in there.



"Shrimp in Featherstar" by prilfish under Flickr

Mimicry:

One type of Mimicry is when one organism that is harmless evolves to look similar to another organism that is poisonous. This stops predators from eating them because they think they are the poisonous species. They can also use mimicry to appear larger than they really are.

The four-eye butterflyfish uses a large eyespot in order to appear larger to predators.



"CHAETODON CAPISTRATUS1" BY CHRIS HUSS UNDER ***PUBLIC DOMAIN***

Another example of mimicry is between the Sabre-tooth Blenny and Cleaner Wrasses. The Cleaner Wrasse have a mutualistic relationship with larger fish so they don't get eaten, and the Sabre-tooth Blenny takes advantage of this relationship by evolving to look very similar to the Cleaner Wrasse. Instead of cleaning the larger fish, the Sabre-tooth Blenny will take a bite out the the large fish's flesh and swim away. This is an example of aggressive mimicry.



"BLUESTREAK CLEANER WRASSE" BY NEMO'S GREAT UNCLE UNDER FLICKR "SABRE-TOOTH BLENNY" BY FISH INDEX

On the top, there is a Bluestreak cleaner wrasse and on the bottom is a sabre-tooth blenny. You can see how similar they look and how fish could mistake them.

What's the Big Deal?



"General characteristics of a large marine ecosystem (Gulf of Alaska)" photo via Wikimedia Commons under Public Domain

Mutualistic relationships, whether obligate or facultative mutualism, are an integral part of sustaining a coral reef ecosystem, and without them, the coral reefs would simply not exist. These mutualistic relationships define a largely intricate number of connections and relationships which deeply rely on one other, and where one could start to deteriorate, another could as well. It is more important now than ever that we sustain healthy coral reefs to support these intricate relationships, communities, and ecosystems.

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18.5: Importance of Coral Reefs

Why Are Reefs Important?

Coral reefs are very complex ecosystems that provide valuable habitat for fish and other animals with their beautiful and unique structures. These structures provide shelter for many organisms such as fish, marine worms, clams and many other animals and plants that all play a vital role in the coral reef ecosystem. Coral reefs are important for a variety of reasons which we will discuss below.



Coral Outcrop Flynn Reef by Toby Hudson via Wikimedia [CC by 2.0]

Biodiversity:

Coral reefs are often thought of as a busy city; the buildings being made of coral and the thousands of organisms inhabiting this city acting like the humans interacting with each other and performing daily jobs. Coral reefs provide protection and shelter to nearly one-quarter of all known marine species and have evolved into one of the largest and most complex ecosystems known to humans. Coral reefs are home to over 4,000 species of fish, 700 species of coral and thousands of other plants and animals. This diversity of species provides a large gene pool giving communities more resilience during extreme environmental conditions and climate change. This is important to the overall health of an ecological community. With greater species diversity, the impact of losing any one species to extinction will be less. The enormous diversity of coral reef organisms also provides potential for new medicines or other products that may be developed from biochemicals that these organisms produce. Most coral reef organisms have not been studied for their potential benefits to medicine and industry.

Coastal Protection:

Coral reefs act as a natural barrier protecting coastal beaches, cities, and communities from the waves of the ocean. Nearly 200 million people depend on coral reefs to protect them from storm surges and waves. Without coral reefs, many buildings would become vulnerable to storm damage.



Reefs Protect Costallines by NOAA via Flicker [CC by 2.0]

Food:

Coral reefs are an important food source for the people who live near the reefs and are crucial for worlds fisheries providing them with a significant source of protein. In developing countries, the reef is said to contribute to one-quarter of the total fish catch providing food resources for tens of millions of people.



"Commercial Fisherman" by NOAA [CC by 2.0]

Medicine:

Many of the compounds now being used in human medicines are found on the coral reef with the potential of more to be discovered. A number of organisms found on reefs produce chemical compounds that have been isolated for human applications. Scientists have developed treatments for a variety of illnesses such as cardiovascular diseases, skin cancer, ulcers, and leukemia. Other compounds can help with reducing inflammation, kill viruses and relax muscles. Not only do the organisms inhabiting the coral reef provide medical treatments but the coral's unique skeletal structure has been used for bone-grafting material.

Tourism:

Coral reef ecosystems are among the most biologically diverse and economically valuable ecosystems on Earth, as they not only support local but global economies. Through tourism (i.e. snorkeling, scuba diving, swimming) and fisheries, coral reefs generate billions of dollars, as well as jobs in more than 100 countries around the world. The annual value of the ecosystem services provided by coral reefs to millions of people is estimated to be over $375 billion. The coral reefs can indirectly bring economic value to these countries by letting visitors enjoy beaches, eat local seafood, paddleboard and sail. All of this is possible due to coral reefs acting as a buffer against waves, storms, and floods. The downside is when tourism harms the coral reefs it not only affects the organisms and the coral reefs, but local people and their economy who rely on the income from tourism.



Snorkel by Angelique800326 via Wikimedia [CC by 2.0]

A good example can be found in Bonaire, a small Caribbean island. Bonaire earns about $23 million (USD) annually from coral reef activities, yet managing its marine park costs less than $1 million per year. A study conducted in 2002 estimated the value of coral reefs at $10 billion, with direct economic benefits of $360 million per year. For residents of coral reef areas who depend on income from tourism, reef destruction creates a significant loss of employment in the tourism, marine recreation, and sport fishing industries. This large amount of money of revenue generated is being threatened by the degradation of coral reefs. As you can see

there is a positive feedback loop occurring because of this situation. Many components of tourism, including recreational activities, are the cause of damage to the reefs but ironically it has been shown that ecotourism is as well.

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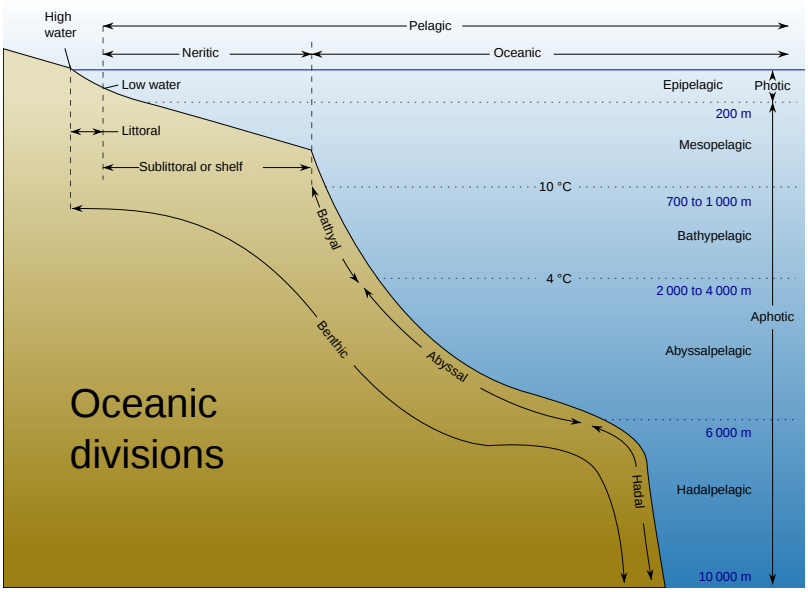
CHAPTER OVERVIEW

19: Marine Communities

19.1: Marine Community Zones
19.2: Intertidal Zone- Rocky Shore
19.3: Intertidal Zone- Sandy / Mudflat shores
19.4: Subtidal Zone
19.5: Oceanic Zone
19.6: Deep Sea

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19.1: Marine Community Zones



Our discussion of the various communities that exist in the ocean will be separated according to 4 ocean zones. Of course, there is much community diversity than defined in this way, but this is a good starting point.

1) The **intertidal zone**, which is also called the littoral zone, is the section of the ocean on the shore that transitions between being submerged and emerged according to the tide. It begins at the high tide line and ends at the low tide line. This zone is the most extreme and also the most influenced from processes on land.

2) The **subtidal zone**, also referred to as the sublittoral zone, begins at the low tide line and continues over the continental shelf, usually to a depth of about 200 meters. Although this zone is nearly always submerged, the uppermost portion may occasionally be exposed during extreme low tide events. Because it is close to shore, this zone experiences changes attributed to processes on land.

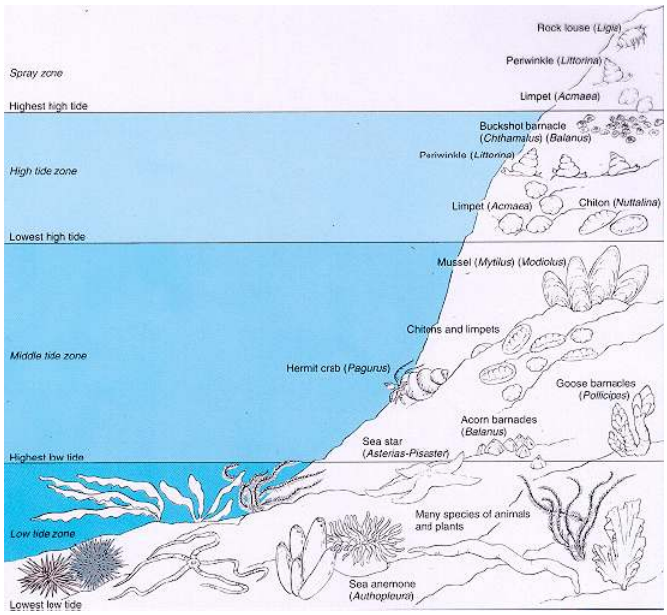
3) The **oceanic zone** exists outside of the continental shelf, where the water depth exceeds 200 meters. Benthic communities in this area do not include photosynthetic organisms. The only photosynthetic organisms in this zone are plankton. Many nekton in this zone migrate between deeper and shallower waters.

4) The **deep sea** is the zone beginning at about 1000 meters depth. Here, there is no light and photosynthetic organisms are not present. Primary producers in this community rely on chemosynthesis, or the production of organic matter using inorganic compounds as the source of energy.

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19.2: Intertidal Zone- Rocky Shore

The incredibly diverse Rocky Intertidal Ecosystems are located on shorelines with hard substrates, usually large rocks with pockets of sand. You can find many **epifuana**, or organisms that live on the surface of hard substrates. This is a very extreme environment that has selected for organisms that are uniquely adapted to tolerate both exposure to air and high wave action, among other physical environmental factors. Frequent disturbance makes this a stressful habitat to live in, but intensity of stress varies among the high-, mid-, and low-intertidal zones. The high-intertidal zone, ending at the landward region of the intertidal zone, experiences the most extreme conditions and can be exposed to air and sunlight for long periods of time between high and low tide. Organisms that live in this region, including barnacles, limpets, chitons, and other encrusting species, are tough critters that have to withstand desiccation (loss of water) and sun exposure (over-heating) or extreme cold depending on the location and season. The mid-intertidal zone is submerged and exposed for equal time. In geographic areas that have semi- or mixed-diurnal tides, such as the western coast of the US, intertidal zones are exposed to two high and two low tides per day. This subjects organisms living in the mid-intertidal zone, including mussels and fucoid algae, to four periods of intense wave action per day. These organisms must develop strong attachment techniques to avoid dislodgment from the rocky shore, which would ultimately lead to death. The low-intertidal zone, the least stressful zone that is submerged in water for the majority of the day, is home to eel grass, sponges, and tunicates. Without the imminent threat of forceful waves, many delicate invertebrates thrive in the low intertidal.



Online resources

- http://montereybay.noaa.gov/sitechar/rock3.html
- http://dusk.geo.orst.edu/oceans/intertidal.html

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19.3: Intertidal Zone- Sandy / Mudflat shores

There are also intertidal zones that are characterizes by soft substrates, such as sand or mud. Because this substrate is soft, we find many burrowing organisms, referred to as **infauna**. Examples of infauna include clams and some types of snails, worms and small crustaceans. By digging into the sand or mud, these organisms are protected from predators and can remain moist when the tide is low.

Mudflats (tidal flats) are areas in which coastline shores are separated from the destructive forces of the ocean's waves. They are created through the deposit of sediments originating from the oceans and rivers. During the event of a high tide, sea water level rises to engulf the entire area bringing in nutrients from the oceans. As the sea water extracts during the low tide, a vast amount of nutrients are left behind making the mudflat a rich feeding ground for a plethora of organisms. The ecosystem that exists within the area heavily revolves around this nutrient delivery system. The abundance of nutrients make mudflats one of the world's most productive ecosystems, and as such, produces one of the most diversified range of living organisms.



The notorious sand bubbler crab is a species that depends on the mechanism that brings in nutrients from the sea. They only emerge to feed during periods of low tide, right after the sea water has retreated back into the ocean. They sift rapidly through the fine sediments (sand) for food and throw aside the rest in the form of a small ball. The entire surface of the area in which these crabs inhabit can be covered in these balls just as the high tide is about to come in, to which now the crabs are the ones retreating back underneath the sands.

You can watch this crab in action by viewing this video: https://www.youtube.com/watch?v=6XJtq2d_lFs

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19.4: Subtidal Zone

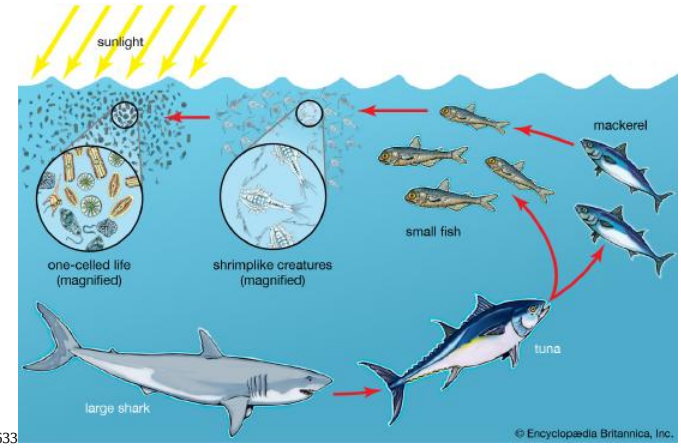
The subtidal zone it the zone close to shore, but constantly submerged. Coastal waters, i.e. the intertidal and subtidal zones, experience the most extreme changes due to land processes. For example, the salinity can fluctuate between high salinity in times of hot, dry winds, to low salinity during times of heavy rain and runoff. Temperature extremes can also exist, as the temperature of the air changes seasonally. Of course, subtidal communities are less affected than intertidal communities, but they still experience these changes.

Previously, this textbook has discussed important subtidal ecoystems- kelp forest and algae beds, seagrasses and coral reefs. These ecosystems are found in shallower areas of the subtidal zone, where sunlight penetrates to the seafloor. Beyond the shallower subtidal, communities resemble that of oceanic communities, in which the primary producers are phytoplankton.

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19.5: Oceanic Zone

As previously stated, the oceanic zone encompasses the ocean that lies beyond the continental shelf and exists where the water depth exceeds 200 meters. At the surface of this zone is the euphotic zone, or the zone that receives enough sunlight to support photosynthesis. Here, the primary producers are phytoplankton.



121633_5s1y121633

An example of an oceanic food chain is shown above. Phytoplankton are eaten by zooplankton, which are eaten by small fish, which are eaten by mackerel, which are eaten by tuna, which are eaten by large sharks.

An interesting event occurs everyday in the oceanic zone. At night zooplankton migrate toward the surface to feed. They are followed by larger predators. When the sun rises, they all sink back to deeper, darker waters, to avoid predation. This migration is referred to as Diel Vertical Migration, as it happens on a daily scale. This migration occurs all of the world, including in some lakes. As these organisms return to deeper waters, they defecate, producing food sources for deep sea organisms.

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19.6: Deep Sea

Introduction

The deep sea is an environment that is unfriendly to humankind. As a result, it represents one of the least explored areas on Earth. It has even been said that the deep sea is less explored than the surface of the moon. The deep sea region begins about 1000 meters beneath the ocean's surface, located between the thermocline and the seafloor (Gage 1991). Here, light is limited, and after a depth of 100-500 meters, it disappears completely. Temperatures in the deep sea can significantly drop -- nearing 0°C at the very bottom. However, its temperature typically ranges between 2-3°C. The salinity level is relatively stable, with concentrations ranging between 34-35 parts per thousand (ppt). On the other hand, its pressure steadily rises with depth, increasing approximately 10 atmospheric pressure per meter (Gage 1991; George 2013).

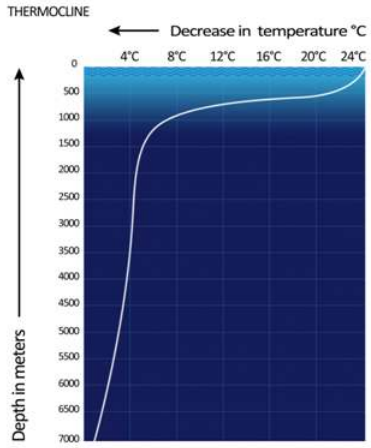


Figure 1. https://upload.wikimedia.org/wikiped...HERMOCLINE.png
This figure depicts the thermocline. Temperature of the surface water remains high until it reaches the thermocline where the temperature drops rapidly as the lack of heat from the sun dissipates. Temperature continues to decline, though much less rapidly, at lower depths.

Life at the Depths

While many scientists believed the deep sea was devoid of life due to its harsh conditions, multiple expeditions have demonstrated the opposite. They found that the deep sea possessed intricate and lively ecosystems, despite being thousands of meters below the surface. In the 1870s, the Challenger expedition revealed the existence of a variety of deep water species. More recently, in 1977, scientists aboard DSV Alvin discovered diverse deep sea communities around hydrothermal vents.

Caused by the absence of light, life in the deep sea faces multiple challenges, such as having difficulty producing energy, attracting prey, and finding mates. At these depths, primary producers can no longer rely on the sun for energy, so they use alternative sources, such as falling detritus from above. These falling particles are known as marine snow, consisting of dead or decaying plant, animal material, fecal matter, and other organic and inorganic particulate matter (Silver 2015). The continuous rain of marine snow provides food for many deep-sea creatures, giving them plenty of carbon and nitrogen to sustain the scavengers of the deep and fuel the rest of the ecosystem.

Physical Adaptations

The animals of the deep sea have adapted to survive in their habitats' extremely harsh conditions: no light, low amounts of food, low amounts of oxygen, high pressure, and cold temperatures. Advantageous physiological changes have developed to enable them to survive in this kind of environment.

Many organisms in the deep sea have adapted a useful physiological trait to aid in the environment's low light conditions: bioluminescence. Bioluminescence is the production and emission of light by a living organism. The ability to produce their own light allows organisms to lure prey, attract mates, and find paths in the water. Additionally, it can be used for defense, mimicry, and counter-illumination (George 2013). Certain squid species can flash warning lights or secret bioluminescent ink in order to defend themselves and ward off predators. Mimicry can be used to match the light pattern of an otherwise fearsome or undesirable organism in order to fend off predators as well. Counter-illumination employs the use of lights on the underside of an organism to match the light coming from above, allowing an organism to appear dark from above (to match the dark water below it) and light from below (to match the light coming through the water above). A common example of attraction via bioluminescence can be demonstrated by angler fish. Angler fish have an appendage that extends from their head with a small light on the end. As they wave this appendage back and forth, they attract and lure prey towards their mouth (Gage 1991).

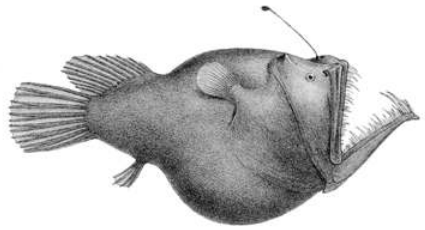


Figure 2. upload.wikimedia.org/wikiped...nglerfish).jpg

This image depicts the anglerfish with a light lure. This is a demonstration of the evolutionary adaptation bioluminescence.

Many deep sea fishes are either completely blind or have developed enlarged eyes, which enable them to detect the faintest of lights. To compensate for the loss of vision, they are well-equipped with a lateral line system, tactile organs, and other sensory structures (e.g. fin rays) (Pandev, 2007).

Body color also has become an adaptation for camouflage. Animals' bodies are often transparent, black, or red. The absence of red light at these depths keeps red-colored animals hidden from predators and prey.

Piezophiles are organisms that thrive in areas with high pressure. In order for them to inhabit such extreme environments, they evolved various mechanisms to counteract the effects of the elevated pressures. They are able to handle such pressure by remaining at a small size and eliminating excess cavities in their bodies that would otherwise collapse.

Many marine organisms have lost appendages and organs to conserve energy output, compensate for the extreme conditions, or because they simply do not use them anymore. The most common organ missing is the swim bladder; this is to escape the high stress of pressure present with extreme depth. In other cases, body parts have become enlarged and exaggerated (sensory organs, eyes, or mouths) in order to adapt to low light or scarce food (Pandey, 2007). An example of exaggerated and reduced body parts is the gulper eel, (*Eurypharynx pelecanoides*) which has tiny fins but also an enormous mouth, much larger than its body. Its mouth is loosely hinged and can open wide enough to swallow animals bigger than itself. Since food can be sparse in the deep sea, being able to eat whatever comes along, despite its size, is a huge advantage. The gulper eel has also adapted to having highly reduced pectoral fins because of their lack of use.



Figure 3. upload.wikimedia.org/wikiped...urypharynx.jpg
This image depicts the gulper eel with a large mouth and reduced fins.

To Learn More:

What's Hiding at the Most Solitary Place on Earth? The Deep Sea - Kurzgesagt Video

https://youtu.be/PaErPyEnDvk

Marine Snow

https://www.youtube.com/watch?v=EF4IAGAXZsM

Angler Fish and Gulper Eel

https://www.youtube.com/watch?v=FswYwyke7cc

Hatchetfish

https://www.youtube.com/watch?v=XmjflLIB9J4

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CHAPTER OVERVIEW

20: Marine Vertebrates

20.1: Ichthyology

20.2: Seabirds

20.3: Sea Turtles

20.4: Marine Mammals

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20.1: Ichthyology

Ichthyology is the study of fish. The description below separates fish into 3 groups- Agnatha, bony fish, and cartilaginous fish.

Agnatha

Moving into the vertebrate category, it is important to start off with the superclass Agnatha, more commonly known as the jawless fish. Two of the classes comprising Agnatha are Cyclostomata (lampreys) and Myxini (hagfishes). It is commonly accepted that the evolution of vertebrates began with the segmentation of a vertebral column. This gave rise to a 'backbone' [1]. This vertebral column typically incorporates and/or replaces the primitive notochord. Further vertebrate adaptations include the development of sensory organs, a complex neural system, and a brain encased in a skull. Vertebrate Chordates also exhibit bilateral symmetry, having a closed circulatory system with a chambered heart. The degree to which the heart is chambered (i.e. one, two, three and four chambered hearts), varies with taxonomic class.

Lampreys (Cyclostomata) are jawless fish that are parasitic on other fish. As juveniles they derive their nutrition from filter feeding on plankton and particles floating in fresh water. Lamprey juveniles wait until maturity to migrate into salt water environments. In the ocean, lampreys may often be observed attached to larger reef fish and megafauna. They press their mouth to their host and using a tongue, they draw blood and tissue out of their victims.



"Sea Lamprey, 2015" by Joanna Gilkeson [CC by 2.0]

The Myxini Class are commonly known as hagfish and reside solely in marine environments. They are similar in structure to eels and are jawless. Myxini can be either consumers or detritivores, feeding on the flesh of weakened or already dead fish. They have also been known to prey upon small invertebrates. Just like the lamprey, hagfish tongues resemble a rasp, and are similar to serrations on a knife. One of the defense mechanisms exhibited by the Class Myxini resembles that of many amphibians in that when threatened, they will release mass amounts of high-viscosity fluids. These fluids help to distract, escape from, confuse or deter potential predators [1].



"Eptatretus soutii, 2005" by Stan Shebs [CC by 3.0]

Osteichthyes and Chondrichthyes

Fish can be divided into the two main groups, Osteichthyes and Chondrichthyes. Osteichthyes, traditionally considered as an taxonomic class, is now known to be a paraphyletic group. As their name suggests, they are boney fish that dwell in both fresh water and salt water around the globe. Boney fish are comprised of a hard calcareous skeleton and are coated in slippery, sometimes sharp scales. An important feature of boney fish is their lateral line. This is a zone that runs horizontally along the body of the fish and is predominantly used in the detection of vibrations. The lateral line has been attributed to the coordinated and navigational success of schools of fish, in which mass quantities of individuals conglomerate for various reasons. There are currently over 34,000 known species of fish on the planet, and that number is both growing and shrinking [fishbase]. Many fish populations are threatened by egregious over-fishing practices which has caused species diversity and abundance in various ecosystems to move into a downward spiral.



"1. Oncorhynchus mykiss 2. Pontinus nematophthalmus 3. Aracana aurita 4. Peristedion gracile"
by Public Domain [CC 0]

The skeletons of Chondrichthyes are comprised of cartilage rather than bone. Cartilaginous fish include the Rays, Sharks and Chimaeras. The vast majority of sharks are predators, much of their power and deadliness comes from their evolutionary adaptations in their physiology. Their streamlined and highly muscular body lends to their high proficiency as consumers. The flattened bodies of rays contributes to their free-flowing nature through water. Often times rays consume invertebrates that are found in the benthos of the ocean, however they are highly diverse in size, morphology and behavior. Giant Manta Rays can get up to 7m in width and filter feed on masses of zooplankton, whereas the southern stingray may only be a few dozen centimeters in width and feed on the bottom in sandy flats near coral reefs [NOAA].



"Hypanus americanus, 1992" by Barry Peters [CC by 2.0]

Chimaera

The Chimaera are an interesting group of organisms that inhabit the deep sea. Their physiology is atypical and they possess a cross of several characteristics as their name suggests. Rather than a jawline filled with teeth Chimaeras have a flat dental plate. Origins of Chimaeroid marine species can be traced back upwards to 280 million years, predating the earliest dinosaurs of the Triassic period.



“Dunkleosteus Skull, 2014” by Zachi Evenor, courtesy of Vienna History Museum, Austria [CC by 3.0]

A January 2017 discovery of a fossilized *Dwykaselachus oosthuizeni* skull showed that there are few structural differences in ancient Chimaeras compared to modern Chimaeras. CT scans of the fossil showed significant cranial nerves, inner-ear structure, and nostrils, which are all exhibited by modern Chimaeras [UChicago Medicine]. An incredible aspect of this would pertain to Earth’s projected geological history. The presence of modern-day Chimaeras with little difference from ancient species means that these organisms survived two mass extinction events, showing the perseverance of the beings that dwell in the ocean depths.



“Modern Deep Sea Chimaera” by NOAA, Public Domain [CC 0]

In Text Reference: 1. Peachey, Donna & Gordon, The Biocam Museum of Life Series. Kelowna, B.C. Canada VIY 7N8 Box 417 PBC, 2000

The information in this chapter in thanks to content contributions from Jason Charbonneau

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20.2: Seabirds



In addition to fish, the ocean is also home to a variety of birds. These birds are referred to as seabirds because they spend a considerable amount of time around the sea. Many seabirds spend their entire lives in the ocean, only coming to land to reproduce. Some seabirds dive and swim below the surface, such as penguins, whose wings are modified into flippers, or cormorants, whose webbed feet act as paddles. Others stay at the surface, only hunting in the uppermost meter of the water.

Please go to this website to read more about seabirds: https://ca.audubon.org/what-s-seabird

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20.3: Sea Turtles



The term sea turtle describes turtles that spend their lives in the ocean. There are 7 species of sea turtle, which are well adapted for swimming. Most only come on land to breed, and some migrate vast distances. Please go to this website to read more about sea turtles: https://www.seeturtles.org/sea-turtle-facts

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20.4: Marine Mammals



Marine mammals are mammals that spend either all or most of their lives in the ocean. Examples include polar bears, sea otters, sea lions, manatees and whales. Please go to this website to learn about marine mammals: https://www.noaa.gov/education/resource-collections/marine-life/marine-mammals

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CHAPTER OVERVIEW

21: Climate Change

21.1: Greenhouse Effect

21.2: Climate Change

21.3: Carbon Dioxide, pH, and Ocean Acidification

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21.1: Greenhouse Effect

The balance of incoming and outgoing heat on Earth is referred to as its **heat budget**. As with any budget, to maintain constant conditions the budget must be balanced so that the incoming heat equals the outgoing heat. The heat budget of Earth appears below (Figure 21.1.1).

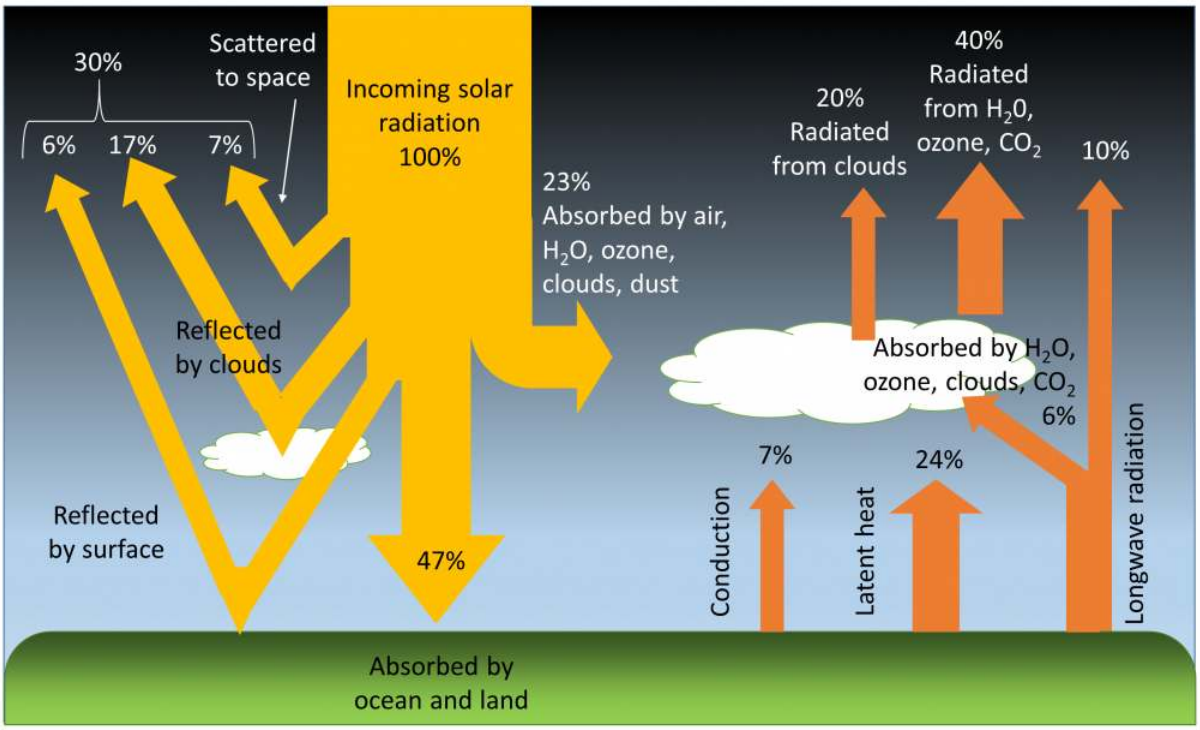


Figure 21.1.1 Earth's heat budget. Of all of the solar radiation reaching Earth, 30% is reflected back to space and 70% is absorbed by the Earth (47%) and atmosphere (23%). The heat absorbed by the land and oceans is exchanged with the atmosphere through conduction, radiation, and latent heat (phase change). The heat absorbed by the atmosphere is eventually radiated back into space (PW).

Of all of the solar energy reaching the Earth, about 30% is reflected back into space from the atmosphere, clouds, and surface of the Earth. Another 23% of the energy is absorbed by the water vapor, clouds, and dust in the atmosphere, where it is converted into heat. Just under half (47%) of the incoming solar radiation is absorbed by the land and ocean, and this energy heats up the Earth's surface. The energy absorbed by the Earth returns to the atmosphere through three processes; conduction, radiation, and latent heat (phase change)(Figure 21.1.1).

Conduction is the transfer of heat through direct contact between the surface and the atmosphere. Air is a relatively poor thermal conductor (which means it is a good insulator), so conduction represents only a small part of the energy transfer between the Earth and the atmosphere; equal to about 7% of the incoming solar energy.

All bodies with a temperature above absolute zero (-273° C) radiate heat in the form of longwave, infrared **radiation** (see the electromagnetic spectrum in section 6.5). The warmed Earth is no exception, and about 16% of the original solar energy is radiated from the Earth to the atmosphere (Figure 21.1.1). Some of this radiated energy will dissipate into space, but a significant amount of heat will be absorbed by the atmosphere. This is the basis for the **greenhouse effect** (Figure 8.12). In the greenhouse effect, shortwave solar radiation passes through the atmosphere and reaches the Earth's surface where it gets absorbed. When the radiation is re-emitted by the Earth, it is now in the form of long wavelength, infrared radiation, which does not easily pass through the atmosphere. Instead, this infrared radiation is absorbed by the atmosphere, particularly by the greenhouse gases such as CO_2, methane, and water vapor. As a result, the atmosphere heats up. Without the greenhouse effect, the average temperature on Earth would be about -18° C, which is too cold for liquid water, and therefore life as we know it could not exist!

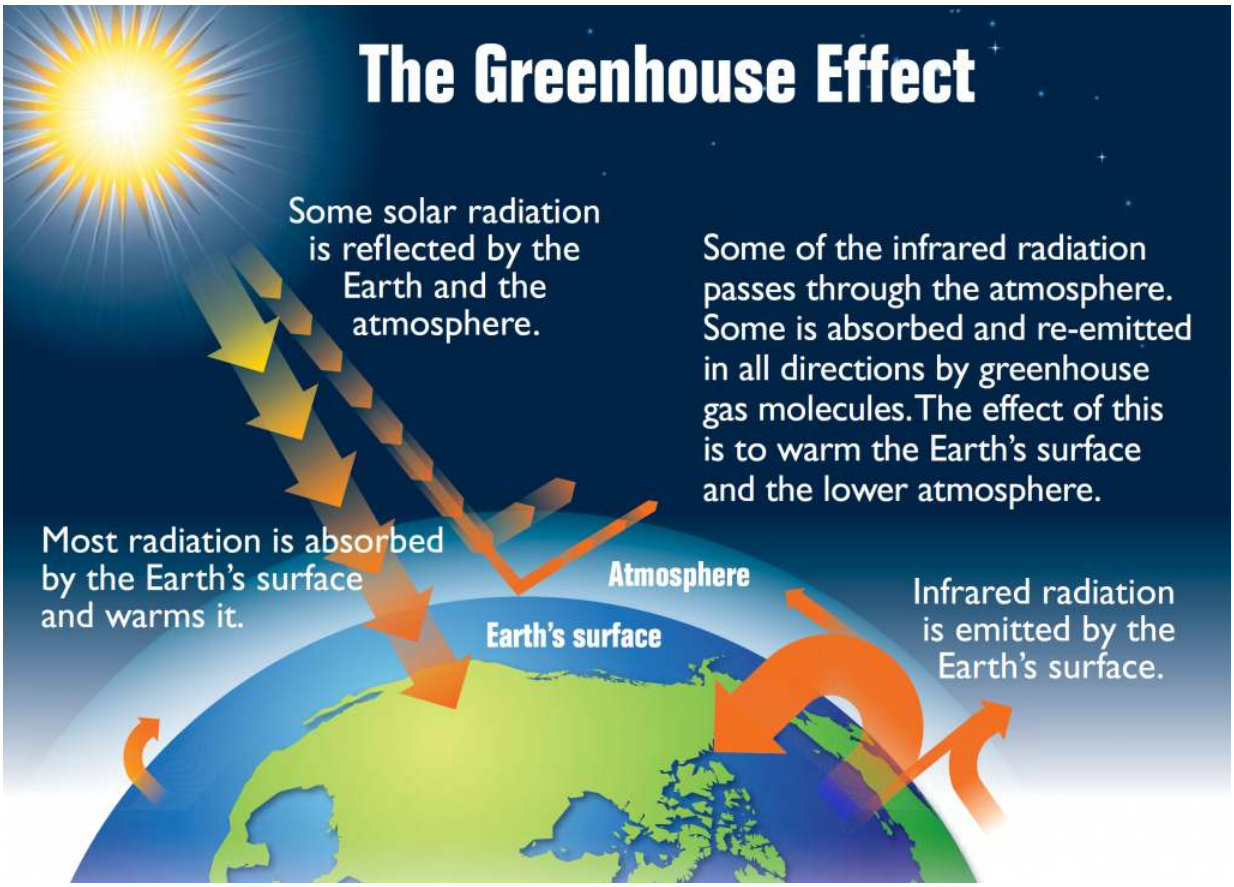


Figure 21.1.2 An explanation of the greenhouse effect (By US EPA [Public domain], via Wikimedia Commons).

There is a great deal of concern about the greenhouse effect across the globe; not because of the presence of the effect itself, but because the effect is intensifying, causing climate change or global warming. Since the Industrial Revolution the atmospheric concentrations of the major greenhouse gases, particularly CO_2 and methane, have increased dramatically due to industrialization, the burning of fossil fuels, and deforestation. At the same time, there has been rapid warming of the global climate; CO_2 concentrations have increased more than 25% and global temperature has risen by 0.5^{o} C over the past century. Unless production of these greenhouse gases is curbed, this rapid warming trend may continue, with potentially dire consequences.

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21.2: Climate Change

If one thing has been constant about Earth's climate over geological time, it is its constant change. In the geological record, we can see this in the evidence of glaciations in the distant past, and we can also detect periods of extreme warmth by looking at the isotope composition of seafloor sediments. Not only has the climate changed frequently, the temperature fluctuations have been very significant. Today's mean global temperature is about 15°C. However, during its coldest periods, the global mean was as cold as -50°C, while at various times during the Paleozoic and Mesozoic and during the Paleocene–Eocene thermal maximum, it was close to 30°C.

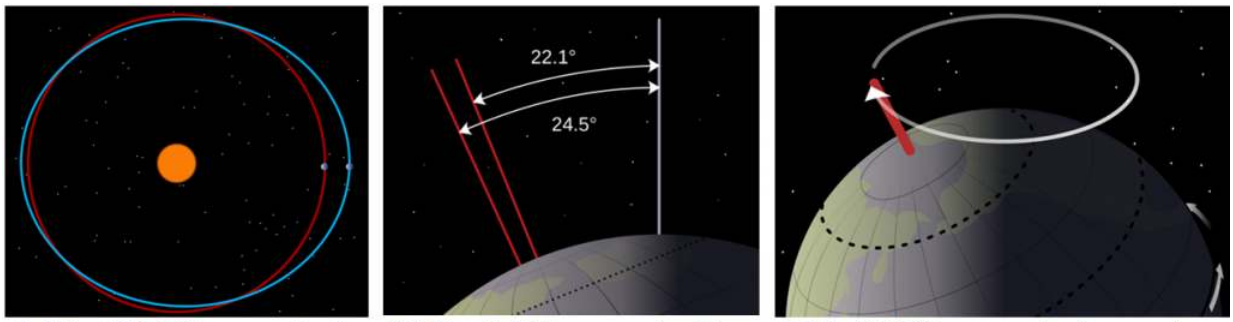
There are two parts to climate change, the first one is known as **climate forcing**, which is when conditions change to give the climate a little nudge in one direction or the other. The second part of climate change, and the one that typically does most of the work, is what we call a **feedback**. When a climate forcing changes the climate a little, a whole series of environmental changes take place, many of which either exaggerate the initial change (**positive feedback**), or suppress the change (**negative feedback**).

An example of a climate forcing mechanism is the increase in the amount of carbon dioxide (CO_2) in the atmosphere that results from our use of fossil fuels. CO_2 traps heat in the atmosphere and leads to climate warming. Warming changes vegetation patterns; contributes to the melting of snow, ice, and permafrost; causes sea level to rise; reduces the solubility of CO_2 in sea water; and has a number of other minor effects. Most of these changes contribute to more warming. Melting of permafrost, for example, is a strong positive feedback because frozen soil contains trapped organic matter that is converted to CO_2 and methane (CH_4) when the soil thaws. Both these gases accumulate in the atmosphere and add to the warming effect. On the other hand, if warming causes more vegetation growth, that vegetation should absorb CO_2, thus reducing the warming effect, which would be a negative feedback. Under our current conditions — a planet that still has lots of glacial ice and permafrost — most of the feedbacks that result from a warming climate are positive feedbacks and so the climate changes that we cause get naturally amplified by natural processes.

Natural Climate Forcing

Natural climate forcing has been going on throughout geological time. A wide range of processes has been operating at widely different time scales, from a few years to billions of years. The longest-term natural forcing variation is related to the evolution of the Sun. Like most other stars of a similar mass, our Sun is evolving. For the past 4.6 billion years, its rate of nuclear fusion has been increasing, and it is now emitting about 40% more energy (as light) than it did at the beginning of geological time. A difference of 40% is big, so it's a little surprising that the temperature on Earth has remained at a reasonable and habitable temperature for all of this time. The mechanism for that relative climate stability has been the evolution of our atmosphere from one that was dominated by CO_2, and also had significant levels of CH_4 — both greenhouse gasses — to one with only a few hundred parts per million of CO_2 and just under 1 part per million of CH_4. Those changes to our atmosphere have been no accident; over geological time, life and its metabolic processes have evolved (such as the evolution of photosynthetic bacteria that consume CO_2) and changed the atmosphere to conditions that remained cool enough to be habitable.

The position of the Earth relative to the Sun is another important component of natural climate forcing. Earth's orbit around the Sun is nearly circular, but like all physical systems, it has natural oscillations. First, the shape of the orbit changes on a regular time scale (close to 100,000 years) from being close to circular to being very slightly elliptical. But the circularity of the orbit is not what matters; it is the fact that as the orbit becomes more elliptical, the position of the Sun within that ellipse becomes less central or more **eccentric** (Figure 21.2.1a). **Eccentricity** is important because when it is high, the Earth-Sun distance varies more from season to season than it does when eccentricity is low.



(a) The 100,000 year cycle of eccentricity of the Sun within the Earth's orbit. The blue orbit is more elliptical than the red one, and the Sun is offset from the centre of the ellipse

(b) The 41,000 year cycle of the angle of tilt (obliquity) of the Earth's axis (red lines) compared to a line perpendicular to the plane of the Earth's orbit (light blue line).

(c) the 20,000 year cycle of precession of the Earth's rotational axis

Figure 21.2.1 Components of the Milankovitch cycles, which influence global climate over thousands of years (Steven Earle, "Physical Geology").

Second, Earth rotates around an axis through the North and South Poles, and that axis is at an angle to the plane of Earth's orbit around the Sun (Figure 21.2.1b). The angle of tilt (also known as **obliquity**) varies on a time scale of 41,000 years. When the angle is at its maximum (24.5°), Earth's seasonal differences are accentuated. When the angle is at its minimum (22.1°), seasonal differences are minimized. The current hypothesis is that glaciation is favored at low seasonal differences as summers would be cooler and snow would be less likely to melt and more likely to accumulate from year to year. Third, the direction in which Earth's rotational axis points also varies, on a time scale of about 20,000 years (Figure 21.2.1c). This variation, known as **precession**, means that although the North Pole is presently pointing to the star Polaris (the pole star), in 10,000 years it will point to the star Vega. The importance of eccentricity, tilt, and precession to Earth's climate cycles (now known as **Milankovitch Cycles**) was first pointed out by Yugoslavian engineer and mathematician Milutin Milankovitch in the early 1900s. Milankovitch recognized that although the variations in the orbital cycles did not affect the total amount of insolation (light energy from the Sun) that Earth received, it did affect where on Earth that energy was strongest.

Volcanic eruptions don't just involve lava flows and exploding rock fragments; various particulates and gases are also released, the important ones being sulphur dioxide and CO_2. Sulphur dioxide is an aerosol that reflects incoming solar radiation and has a net cooling effect that is short lived (a few years in most cases, as the particulates settle out of the atmosphere within a couple of years), and doesn't typically contribute to longer-term climate change. Volcanic CO_2 emissions can contribute to climate warming but only if a greater-than-average level of volcanism is sustained over a long time (at least tens of thousands of years). It is widely believed that the catastrophic end-Permian extinction (at 250 Ma) resulted from warming initiated by the eruption of the massive Siberian Traps over a period of at least a million years.

Ocean currents are important to climate, and currents also have a tendency to oscillate. Glacial ice cores show clear evidence of changes in the Gulf Stream that affected global climate on a time scale of about 1,500 years during the last glaciation. The east-west changes in sea-surface temperature and surface pressure in the equatorial Pacific Ocean, known as the El Niño Southern Oscillation or ENSO varies on a much shorter time scale of between two and seven years. These variations tend to garner the attention of the public because they have significant climate implications in many parts of the world. The strongest El Niños in recent decades were in 1983, 1998, and 2015 and those were very warm years from a global perspective. During a strong El Niño, the equatorial Pacific sea-surface temperatures are warmer than normal and heat the atmosphere above the ocean, which leads to warmer-than-average global temperatures.

Climate Feedbacks

As already stated, climate feedbacks are critically important in amplifying weak climate forcings into full-blown climate changes. Since Earth still has a very large volume of ice, mostly in the continental ice sheets of Antarctica and Greenland, but also in alpine glaciers and permafrost, melting is one of the key feedback mechanisms. Melting of ice and snow leads to several different types of feedbacks, an important one being a change in albedo, or the reflectivity of a surface. Earth's various surfaces have widely differing

albedos, expressed as the percentage of light that reflects off a given material. This is important because most solar energy that hits a very reflective surface is not absorbed and therefore does little to warm Earth. Water in the oceans or on a lake is one of the darkest surfaces, reflecting less than 10% of the incident light, while clouds and snow or ice are among the brightest surfaces, reflecting 70% to 90% of the incident light. When sea ice melts, as it has done in the Arctic Ocean at a disturbing rate over the past decade, the albedo of the area affected changes dramatically, from around 80% down to less than 10%. Much more solar energy is absorbed by the water than by the pre-existing ice, and the temperature increase is amplified. The same applies to ice and snow on land, but the difference in albedo is not as great. When ice and snow on land melt, sea level rises. (Sea level is also rising because the oceans are warming and that increases their volume). A higher sea level means a larger proportion of the planet is covered with water, and since water has a lower albedo than land, more heat is absorbed and the temperature goes up a little more. Since the last glaciation, sea-level rise has been about 125 m; a huge area that used to be land is now flooded by heat-absorbent seawater. During the current period of anthropogenic climate change, sea level has risen only about 20 cm, and although that doesn't make a big change to albedo, sea-level rise is accelerating.

Most of northern Canada, Alaska, Russia, and Scandinavia has a layer of permafrost that ranges from a few centimeters to hundreds of meters in thickness. Permafrost is a mixture of soil and ice and it also contains a significant amount of trapped organic carbon that is released as CO_2 and CH_4 when the permafrost breaks down. Because the amount of carbon stored in permafrost is in the same order of magnitude as the amount released by burning fossil fuels, this is a feedback mechanism that has the potential to equal or surpass the forcing that has unleashed it. In some polar regions, including northern Canada, permafrost includes methane hydrate, a highly concentrated form of CH_4 trapped in solid form. Breakdown of permafrost releases this CH_4. Even larger reserves of methane hydrate exist on the seafloor, and while it would take significant warming of ocean water down to a depth of hundreds of meters, this too is likely to happen in the future if we don't limit our impact on the climate. There is strong isotopic evidence that the Paleocene–Eocene thermal maximum was caused, at least in part, by a massive release of sea-floor methane hydrate.

There is about 45 times as much carbon in the ocean (as dissolved bicarbonate ions, HCO_3^-) as there is in the atmosphere (as CO_2), and there is a steady exchange of carbon between the two reservoirs (see section 5.5). But the solubility of CO_2 in water decreases as the temperature goes up. In other words, the warmer it gets, the more oceanic bicarbonate that gets transferred to the atmosphere as CO_2. That makes CO_2 solubility another positive feedback mechanism. Vegetation growth responds positively to both increased temperatures and elevated CO_2 levels, and so in general, it represents a negative feedback to climate change because the more the vegetation grows, the more CO_2 is taken from the atmosphere. But it's not quite that simple, because when trees grow bigger and more vigorously, forests become darker (they have lower albedo) so they absorb more heat. Furthermore, climate warming isn't necessarily good for vegetation growth; some areas have become too hot, too dry, or even too wet to support the plant community that was growing there, and it might take centuries for something to replace it successfully. All of these positive (and negative) feedbacks work both ways. For example, during climate cooling, growth of glaciers leads to higher albedos, and formation of permafrost results in storage of carbon that would otherwise have returned quickly to the atmosphere.

Anthropogenic Climate Change

When we talk about anthropogenic climate change, we are generally thinking of the industrial era, which really got going when we started using fossil fuels (coal to begin with, and later oil and natural gas) to drive machinery and trains, and to generate electricity. That was around the middle of the 18th century. The issue with fossil fuels is that they involve burning carbon that was naturally stored in the crust over hundreds of millions of years as part of Earth's process of counteracting the warming Sun.

A rapidly rising population, the escalating level of industrialization and mechanization of our lives, and an increasing dependence on fossil fuels have driven the anthropogenic climate change of the past century. The trend of mean global temperatures since 1880 is shown in Figure 21.2.2. For approximately the past 55 years, the temperature has increased at a relatively steady and disturbingly rapid rate, especially compared to past changes. The average temperature now is approximately 0.8°C higher than before industrialization, and two-thirds of this warming has occurred since 1975.

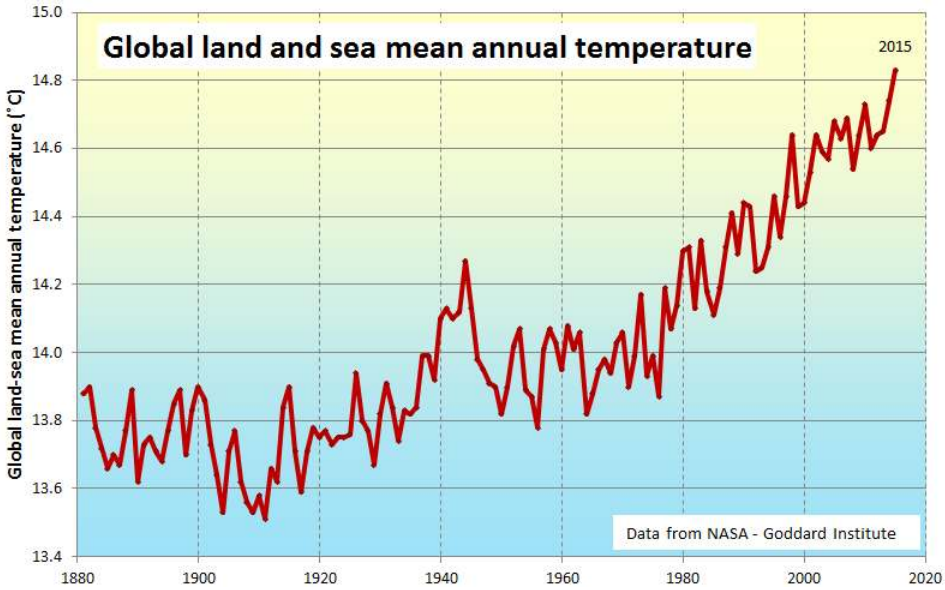


Figure 21.2.2 Global mean annual temperatures for the period from 1880 to 2015 (Steven Earle, "Physical Geology", by SE from data at NASA at: http://data.giss.nasa.gov/gistemp/ta...LB.Ts+dSST.txt).

The Intergovernmental Panel on Climate Change (IPCC), established by the United Nations in 1988, is responsible for reviewing the scientific literature on climate change and issuing periodic reports on several topics, including the scientific basis for understanding climate change, our vulnerability to observed and predicted climate changes, and what we can do to limit climate change and minimize its impacts. Figure 21.2.3, from the fifth report of the IPCC, issued in 2014, shows the relative contributions of various greenhouse gases and other factors to current climate forcing, based on the changes from levels that existed in 1750.

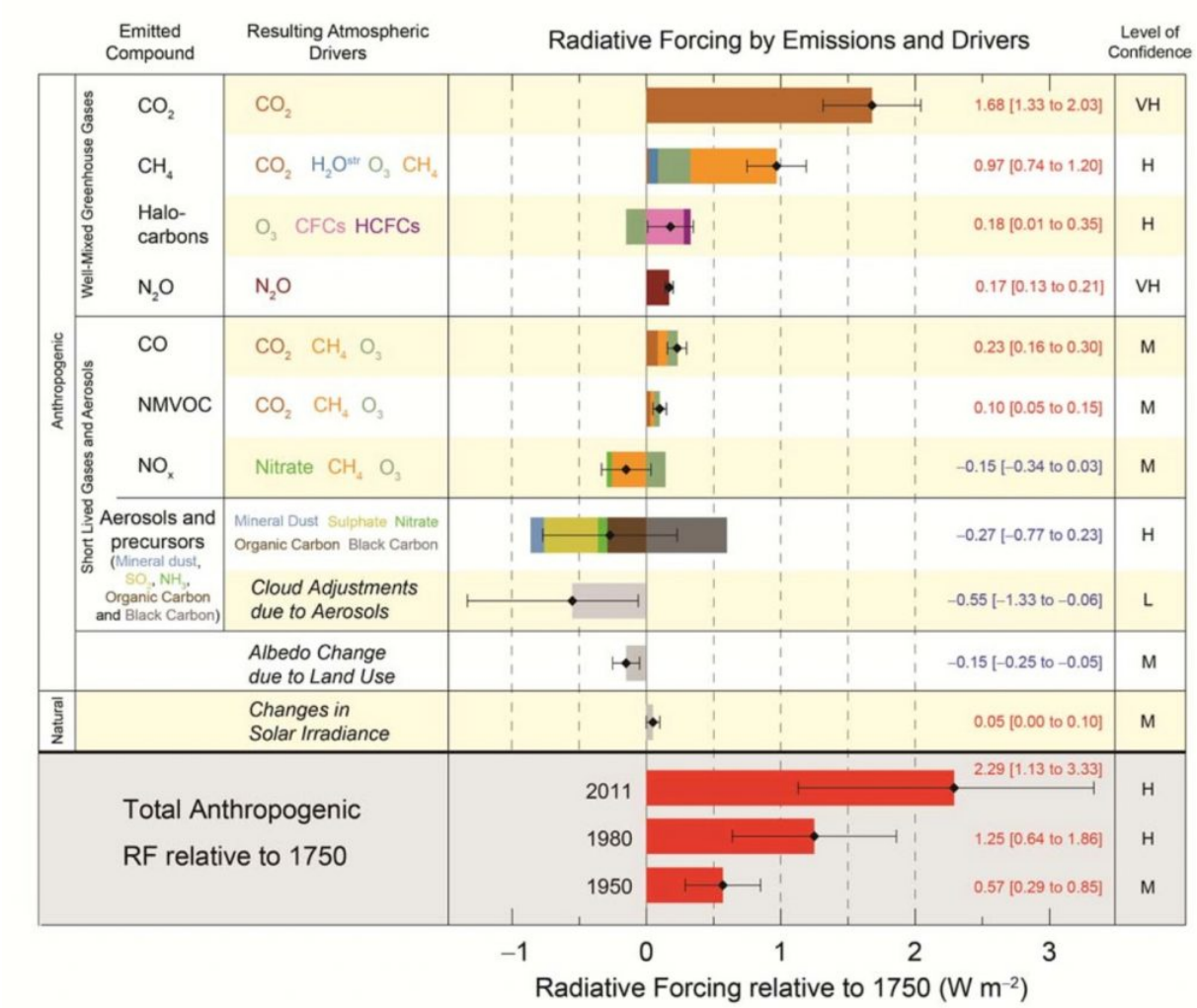


Figure 21.2.3 The relative importance of factors that are contributing to anthropogenic warming (from http://www.ipcc.ch/report/graphics/i...0-%20WG1&f=SPM).

The biggest anthropogenic contributor to warming is the emission of CO_2, which accounts for 50% of positive forcing. CH_4 and its atmospheric derivatives (CO_2, H_2O, and O_3) account for 29%, and the halocarbon gases (mostly leaked from air-conditioning appliances) and nitrous oxide (N_2O) (from burning fossils fuels) account for 5% each. Carbon monoxide (CO) (also produced by burning fossil fuels) accounts for 7%, and the volatile organic compounds other than methane (NMVOC) account for 3%. CO_2 emissions come mostly from coal- and gas-fired power stations, motorized vehicles (cars, trucks, and aircraft), and industrial operations (e.g., smelting), and indirectly from forestry. CH_4 emissions come from production of fossil fuels (escape from coal mining and from gas and oil production), livestock farming (mostly beef), landfills, and wetland rice farming. N_2O and CO come mostly from the combustion of fossil fuels. In summary, close to 70% of our current greenhouse gas emissions come from fossil fuel production and use, while most of the rest comes from agriculture and landfills. Figure 21.2.4 shows the IPCC's projections for temperature increases over the next 100 years as a result of these increasing greenhouse gases.

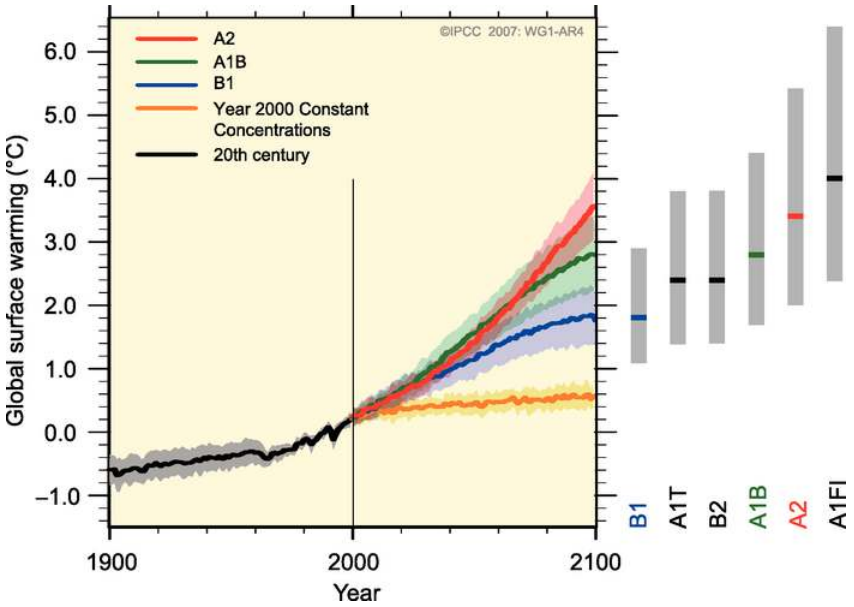


Figure 21.2.4 Projected global temperature increases for the 21st century based on a range of different IPCC scenarios of future political and technological variables (from https://www.ipcc.ch/publications_and...re-spm-5-l.png).

Impacts of Climate Change

We've all experienced the effects of climate change over the past decade. However, it's not straightforward for climatologists to make the connection between a warming climate and specific weather events, and most are justifiably reluctant to ascribe any specific event to climate change. In this respect, the best measures of climate change are those that we can detect over several decades, such as the temperature changes shown in Figure 21.2.2, or the sea level rise shown in Figure 21.2.5. As already stated, sea level has risen approximately 20 cm since 1750, and that rise is attributed to both warming (and therefore expanding) seawater and melting glaciers and other land-based snow and ice (melting of sea ice does not contribute directly to sea level rise as it is already floating in the ocean).

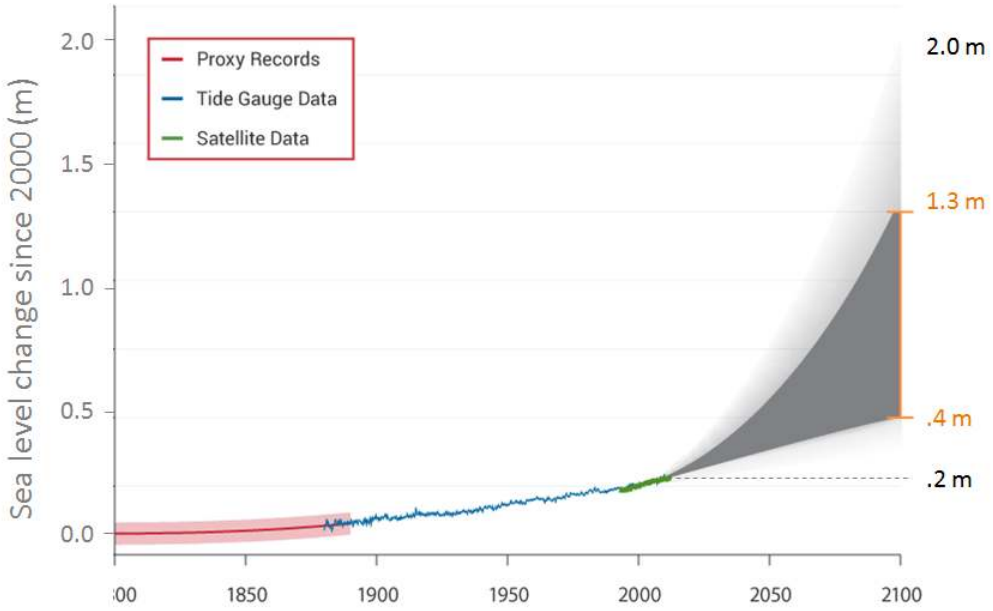


Figure 21.2.5 Projected sea-level increases to 2100, showing likely range (grey) and possible maximum (Stephen Earle, "Physical Geology", adapted by SE from: http://nca2014.globalchange.gov/repo...ntro-section-2 based on data in Parris et al., 2012, NOAA).

Projections for sea level rise to the end of this century vary widely. This is in large part because we do not know which of the above climate change scenarios (Figure 21.2.4) we will most closely follow, but many are in the range from 0.5 m to 2.0 m. One of the problems in predicting sea level rise is that we do not have a strong understanding of how large ice sheets, such as Greenland and

Antarctica, will respond to future warming. Another issue is that the oceans don't respond immediately to warming. For example, with the current amount of warming, we are already committed to a future sea level rise of between 1.3 m and 1.9 m, even if we could stop climate change today. This is because it takes decades to centuries for the existing warming of the atmosphere to be transmitted to depth within the oceans and to exert its full impact on large glaciers. Most of that committed rise would take place over the next century, but some would be delayed longer. And for every decade that the current rates of climate change continue, that number increases by another 0.3 m. In other words, if we don't make changes quickly, by the end of this century we'll be locked into 3 m of future sea level rise. In a 2008 report, the Organization for Economic Co-operation and Development (OECD) estimated that by 2070 approximately 150 million people living in coastal areas could be at risk of flooding due to the combined effects of sea level rise, increased storm intensity, and land subsidence. The assets at risk (buildings, roads, bridges, ports, etc.) are in the order of $35 trillion ($35,000,000,000,000). Countries with the greatest exposure of population to flooding are China, India, Bangladesh, Vietnam, U.S.A., Japan, and Thailand. Some of the major cities at risk include Shanghai, Guangzhou, Mumbai, Kolkata, Dhaka, Ho Chi Minh City, Tokyo, Miami, and New York.

One of the other risks for coastal populations, besides sea level rise, is that climate warming is also associated with an increase in the intensity of tropical storms (e.g., hurricanes or typhoons), which almost always bring serious flooding from intense rain and storm surges. Some recent examples are New Orleans in 2005 with Hurricane Katrina, and New Jersey and New York in 2012 with Hurricane Sandy. Tropical storms get their energy from the evaporation of warm seawater in tropical regions. In the Atlantic Ocean, this takes place between 8° and 20° N in the summer. Figure 21.2.6 shows the variations in the sea-surface temperature (SST) of the tropical Atlantic Ocean (in blue) versus the amount of power represented by Atlantic hurricanes between 1950 and 2008 (in red). Not only has the overall intensity of Atlantic hurricanes increased with the warming since 1975, but the correlation between hurricanes and sea-surface temperatures is very strong over that time period.

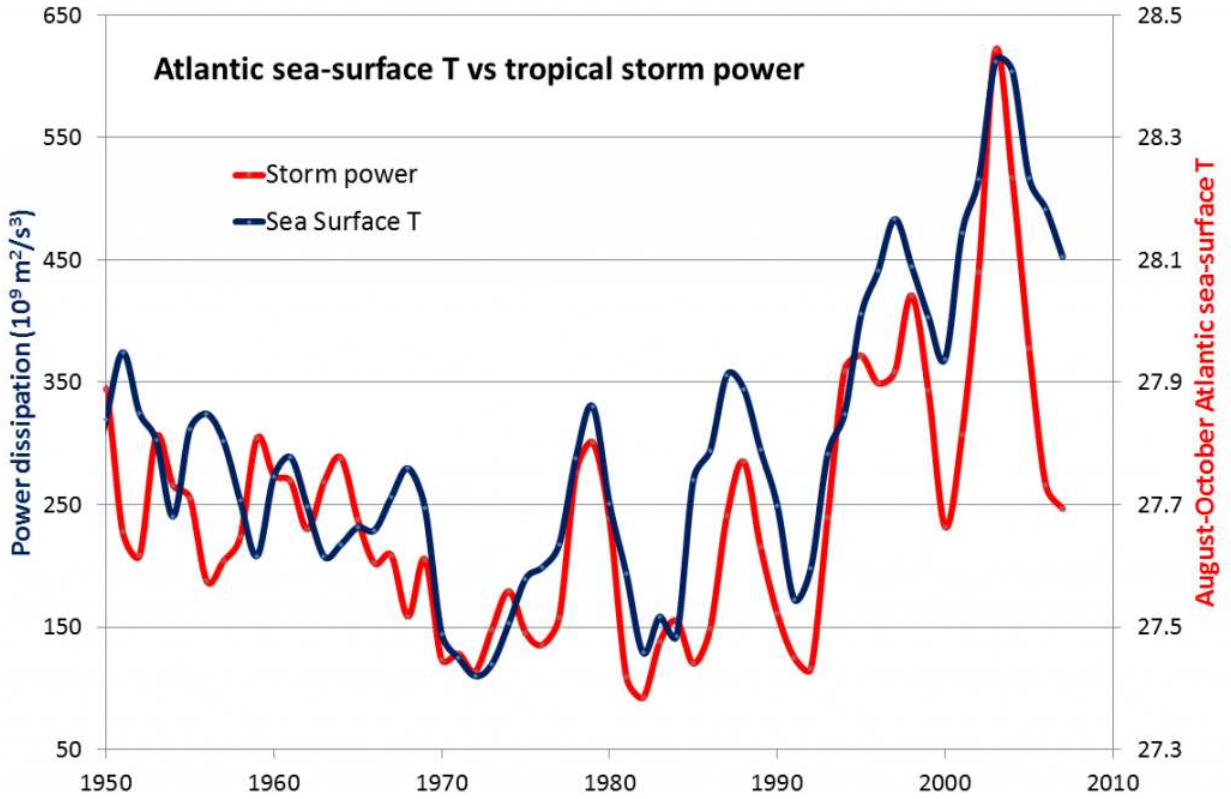


Figure 21.2.6 Relationship between Atlantic tropical storm cumulative annual intensity and Atlantic sea-surface temperatures (Steven Earle, "Physical Geology", by SE from data at: http://wind.mit.edu/~emanuel/Papers_data_graphics.htm).

The geographical ranges of diseases and pests, especially those caused or transmitted by insects, have been shown to extend toward temperate regions because of climate change. West Nile virus and Lyme disease are two examples that already directly affect North Americans, while dengue fever could be an issue in the future (dengue became a "nationally notifiable condition" in the United States in 2010). For several weeks in July and August of 2010, a massive heat wave affected western Russia, especially the area southeast of Moscow, and scientists have stated that climate change was a contributing factor. Temperatures soared to over 40°C, as

much as 12°C above normal over a wide area, and wildfires raged in many parts of the country. Over 55,000 deaths are attributed to the heat and to respiratory problems associated with the fires. A summary of the impacts of climate change on natural disasters is given in Figure 21.2.7. The major types of disasters related to climate are floods and storms, but the health implications of extreme temperatures are also becoming a great concern. In the decade 1971 to 1980, extreme temperatures were the fifth most common natural disasters; by 2001 to 2010, they were the third most common.

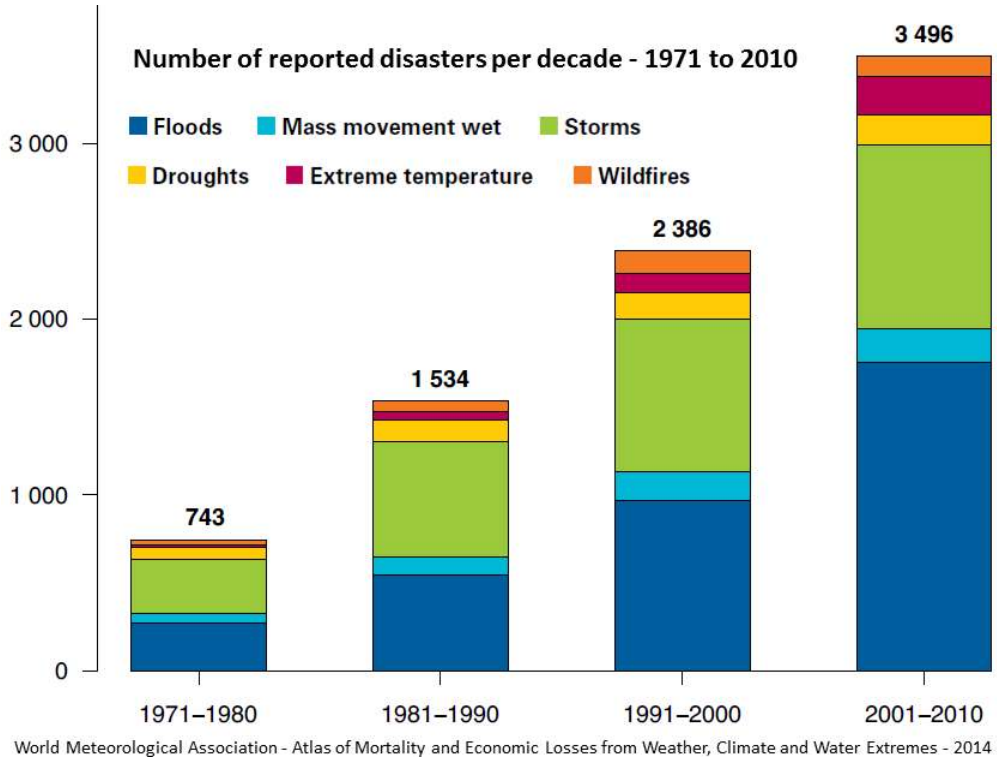


Figure 21.2.7 Numbers of various types of disasters between 1971 and 2010 (From WMO atlas of mortality and economic Losses from weather, climate and water extremes, 2014).

*"Physical Geology" by Steven Earle used under a CC-BY 4.0 international license. Download this book for free at http://open.bccampus.ca

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21.3: Carbon Dioxide, pH, and Ocean Acidification

Figure 21.3.1 Representative carbon dioxide profiles for the Pacific and Atlantic oceans (PW).

The anthropogenic addition of carbon dioxide into the atmosphere has an additional effect on ocean chemistry. About 30% of the carbon dioxide that is added to the atmosphere is absorbed by the oceans. This additional carbon dioxide affects the pH of the ocean. We previously learned the following:

When CO_2 gas dissolves in the ocean, it interacts with the water to produce a number of different compounds according to the reaction below:

$$CO_2 + H_2O \leftrightarrow H_2CO_3 \leftrightarrow H^+ + HCO_3^- \leftrightarrow 2H^+ + CO_3^{2-}$$

CO_2 reacts with water to produce carbonic acid (H_2CO_3), which then dissociates into bicarbonate (HCO_3^-) and hydrogen ions (H^+). The bicarbonate ions can further dissociate into carbonate (CO_3^{2-}) and additional hydrogen ions (Figure 21.3.2).

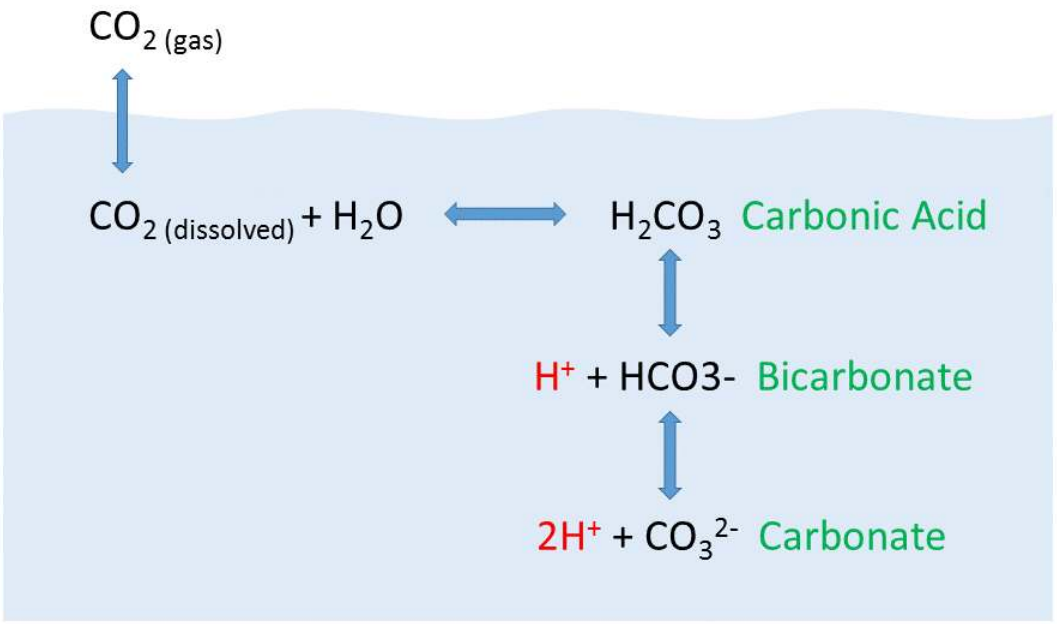


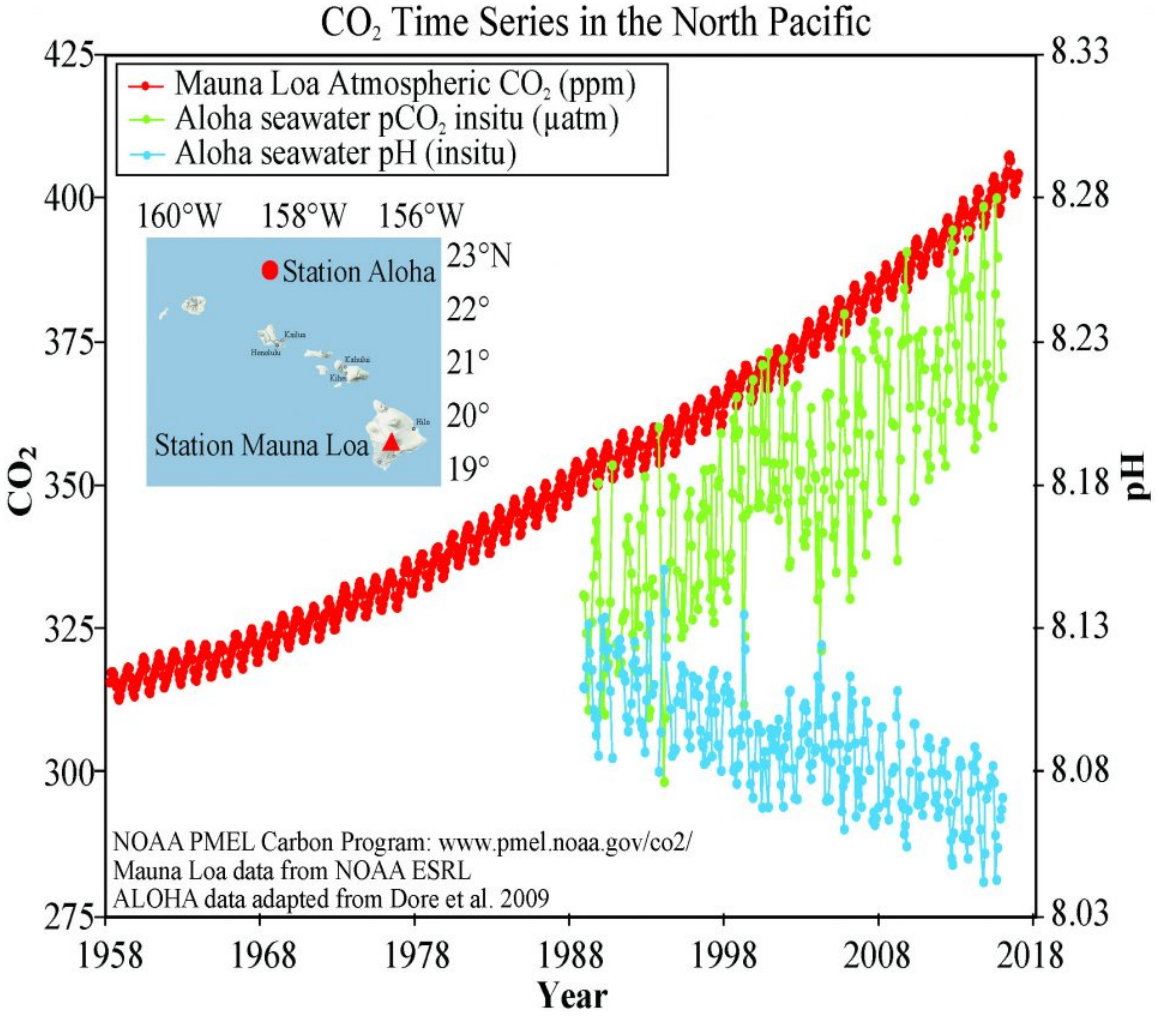
Figure 21.3.2 The fate of dissolved carbon dioxide in the oceans. Most of the carbon ends up in the form of bicarbonate (PW).

Carbon dioxide and the other carbon compounds listed above play an important role in buffering the pH of the ocean. Currently, the average pH for the global ocean is about 8.1, meaning seawater is slightly basic. Because most of the inorganic carbon dissolved in the ocean exists in the form of bicarbonate, bicarbonate can respond to disturbances in pH by releasing or incorporating hydrogen ions into the various carbon compounds. If pH rises (low $[H^+]$), bicarbonate may dissociate into carbonate, and release more H^+ ions, thus lowering pH. Conversely, if pH gets too low (high $[H^+]$), bicarbonate and carbonate may incorporate some of those H^+ ions and produce bicarbonate, carbonic acid, or CO_2 to remove H^+ ions and raise the pH. By shuttling H^+ ions back and forth between the various compounds in this equation, the pH of the ocean is regulated and conditions remain favorable for life.

CO_2 and Ocean Acidification

In recent years there has been rising concern about the phenomenon of **ocean acidification**. As described in the processes above, the addition of CO_2 to seawater lowers the pH of the water. As anthropogenic sources of atmospheric CO_2 have increased since the Industrial Revolution, the oceans have been absorbing an increasing amount of CO_2, and researchers have documented a decline in ocean pH from about 8.2 to 8.1 in the last century. This may not appear to be much of a change, but remember that since pH is on a logarithmic scale, this decline represents a 30% increase in acidity. It should be noted that even at a pH of 8.1 the ocean is not actually acidic; the term “acidification” refers to the fact that the pH is becoming lower, i.e. the water is moving towards more acidic conditions.

Figure 21.3.3 presents data from observation stations in and around the Hawaiian Islands. As atmospheric levels of CO_2 have increased, the CO_2 content of the ocean water has also increased, leading to a reduction in seawater pH. Some models suggest that at the current rate of CO_2 addition to the atmosphere, by 2100 ocean pH may be further reduced to around 7.8, which would represent more than a 120% increase in ocean acidity since the Industrial Revolution.



Data: Mauna Loa (ftp://aftp.cmdl.noaa.gov/products/trends/co2/co2_mm_mlo.txt) ALOHA (http://hahana.soest.hawaii.edu/hot/products/HOT_surface_CO2.txt)
Ref: J.E. Dore et al, 2009. Physical and biogeochemical modulation of ocean acidification in the central North Pacific. *Proc Natl Acad Sci USA* **106**:12235-12240.

Figure 21.3.3 Changes in atmospheric CO_2 (red), seawater CO_2 (green) and pH (blue) in the Hawaiian Islands (NOAA PMEL).

Why is this important? Declining pH can impact many biological systems. Of particular concern are organisms that secrete calcium carbonate shells or skeletons, such as corals, shellfish, and may planktonic organisms. At lower pH levels, calcium carbonate dissolves, eroding the shells and skeletons of these organisms (Figure 21.3.4).



Figure 21.3.5 The results of an experiment placing the calcium carbonate shells of pterapods in seawater with a pH of 7.8, the projected ocean pH for the year 2100 under current rates of acidification. The top row shows the shells before the experiment, and the bottom row shows the dissolution of the shells after 45 days of exposure (NOAA).

Not only does a declining pH lead to increased rates of dissolution of calcium carbonate, it also diminishes the amount of free carbonate ions in the water. The relative proportions of the different carbon compounds in seawater is dependent on pH (Figure 21.3.6). As pH declines, the amount of carbonate declines, so there is less available for organisms to incorporate into their shells and skeletons. So ocean acidification both dissolves existing shells and makes it harder for shell formation to occur.

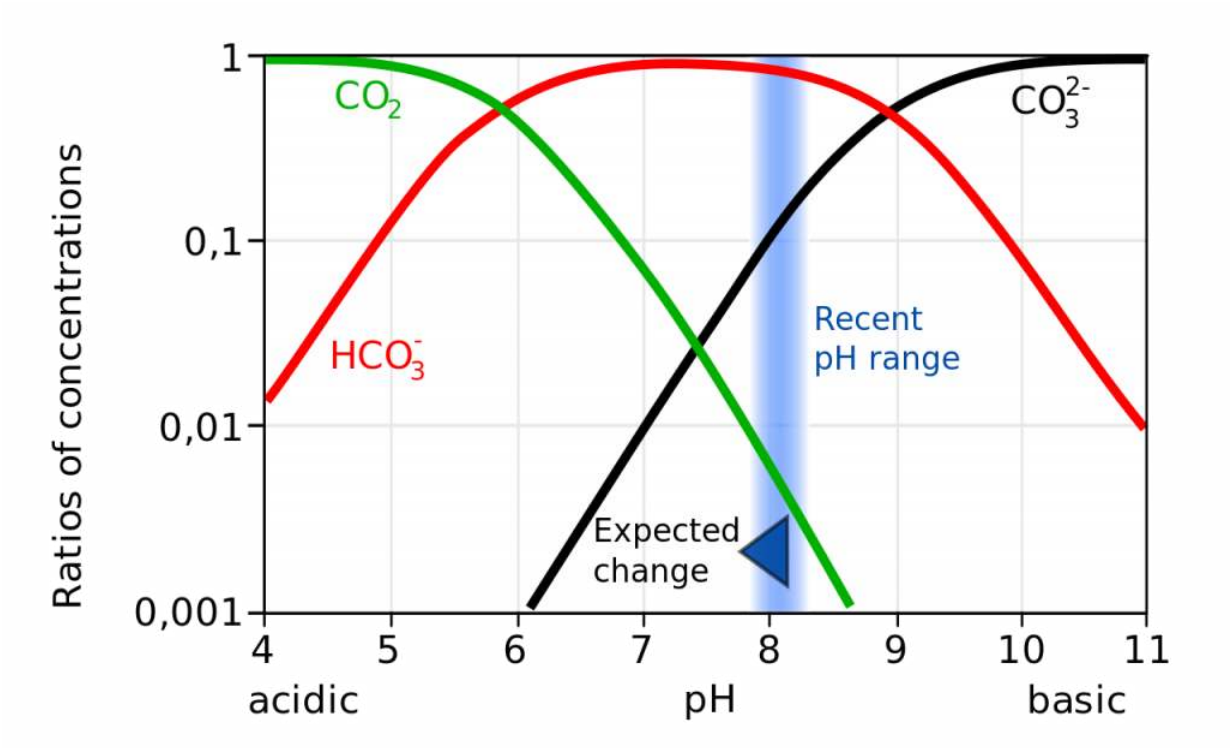


Figure 21.3.6 Proportions of carbon compounds in the ocean at various pH levels. As the ocean pH declines, the proportion of carbonate ions also declines, reducing rates of shell formation (NOAA).

Additional links for more information:

- NOAA Ocean Acidification Program website http://oceanacidification.noaa.gov/

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