INTRODUCTION TO GEOLOGY



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Introduction to Geology

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Licensing

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CHAPTER OVERVIEW

1: Understanding Science

Learning Objectives

- Contrast objective versus subjective observations, and quantitative versus qualitative observations.
- Identify a pseudoscience based on its lack of falsifiability.
- Contrast the methods used by Aristotle and Galileo to describe the natural environment.
- Explain the scientific method and apply it to a problem or question.
- Explain why studying geology is important.
- Describe the steps involved in a reputable scientific study.
- Explain rhetorical arguments used by science deniers.

Think of your favorite science fiction movie. What is it about? Maybe it's about spaceships going to distant planets, or people being cloned in laboratories, or undersea civilizations, or robots that walk among us. These entertaining imaginings are make-believe fantasies, that's why they're called science "fiction." They are not real. But why are they called "science" fiction?

The answer is that science uses a disciplined process to answer questions. In science, "disciplined" does not mean well-behaved. It means following orderly steps in order to come up with the best answers. Science involves observing, wondering, categorizing, communicating, calculating, analyzing, and much more. In order to convert creativity into reality, we need science. In order to travel beyond where anyone has gone before, we need science. In order to understand the world, make sense of it, and conserve it, we need science. In order to confirm our best guesses about the universe and the things in it, we need science. Science fiction stories extend and expand on all the ideas of science and technology in creative ways.

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1.1: What is Science?
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1.1: What is Science?

Scientists seek to understand the fundamental principles that explain natural patterns and processes. Science is more than just a body of knowledge; science provides a means to evaluate and create new knowledge without bias [1]. Scientists use objective evidence over subjective evidence, to reach sound and logical conclusions.

An **objective observation** is without personal bias and the same by all individuals. Humans are biased by nature, so they cannot be completely objective; the goal is to be as unbiased as possible. A **subjective observation** is based on a person's feelings and beliefs and is unique to that individual.



Figure 1.1.1: The Grand Canyon of Yellowstone in Yellowstone National Park. An objective statement about this would be: "The picture is of a waterfall." A subjective statement would be: "The picture is beautiful," or "The waterfall is there because of erosion."

Another way scientists avoid bias is by using quantitative over qualitative measurements whenever possible. A **quantitative** measurement is expressed with a specific numerical value. **Qualitative** observations are general or relative descriptions. For example, describing a rock as red or heavy is a qualitative observation. Determining a rock's color by measuring wavelengths of reflected light or determining its density by measuring the proportions of minerals it contains is quantitative. Numerical values are more precise than general descriptions, and they can be analyzed using statistical calculations. This is why quantitative measurements are much more useful to scientists than qualitative observations.



Figure 1.1.2: Canyons like this, carved in the deposit left by the May 18, 1980 eruption of Mt. St. Helens, are sometimes used by purveyors of pseudoscience as evidence for the Earth being very young. In reality, the unconsolidated and unlithified volcanic ash deposit is carved much more easily than other canyons like the Grand Canyon.

Establishing truth in science is difficult because all scientific claims are **falsifiable**, which means any initial hypothesis may be tested and proven false. Only after exhaustively eliminating false results, competing ideas, and possible variations does a hypothesis become regarded as a reliable scientific theory. This meticulous scrutiny reveals weaknesses or flaws in a hypothesis and is the strength that supports all scientific ideas and procedures. In fact, proving current ideas are wrong has been the driving force behind many scientific careers.

Falsifiability separates science from **pseudoscience**. Scientists are wary of explanations of natural phenomena that discourage or avoid falsifiability. An explanation that cannot be tested or does not meet scientific standards is not considered science, but pseudoscience. Pseudoscience is a collection of ideas that may appear scientific but does not use the scientific method. Astrology is an example of pseudoscience. It is a belief system that attributes the movement of celestial bodies to influencing human behavior. Astrologers rely on celestial observations, but their conclusions are not based on experimental evidence and their statements are not falsifiable. This is not to be confused with *astronomy* which is the scientific study of celestial bodies and the cosmos [2,3].



Figure 1.1.3: Geologists share information by publishing, attending conferences, and even going on field trips, such as this trip to the Lake Owyhee Volcanic Field in Oregon by the Bureau of Land Management in 2019.

Science is also a social process. Scientists share their ideas with peers at conferences, seeking guidance and feedback. Research papers and data submitted for publication are rigorously reviewed by qualified peers, scientists who are experts in the same field. The scientific review process aims to weed out misinformation, invalid research results, and wild speculation. Thus, it is slow, cautious, and conservative. Scientists tend to wait until a hypothesis is supported by an overwhelming amount of evidence from many independent researchers before accepting it as a scientific theory [46].

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1.2: The Scientific Method

Modern science is based on the **scientific method**, a procedure that follows these steps:

- Formulate a question or observe a problem
- Apply objective experimentation and observation
- Analyze collected data and interpret results
- Devise an evidence-based theory
- Submit findings to peer review and/or publication

This has a long history in human thought but was first fully formed by Ibn al-Haytham over 1000 years ago. At the forefront of the scientific method are **conclusions** based on objective **evidence**, not opinion or hearsay [4].

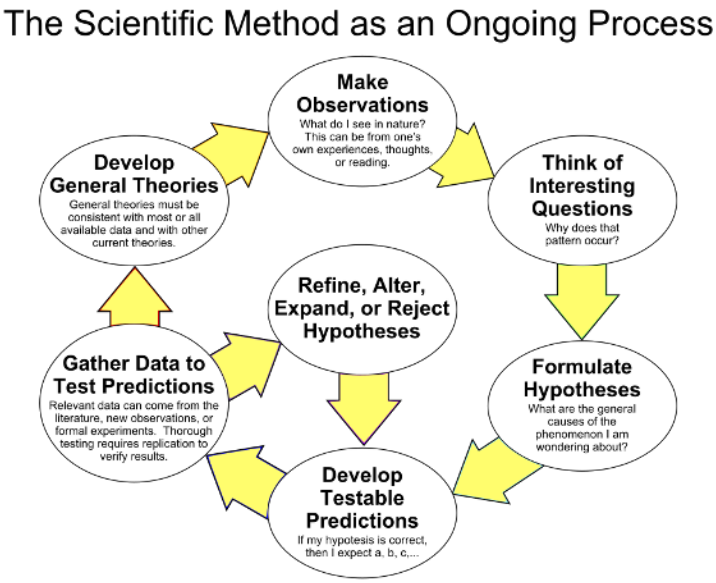


Figure 1.2.1: Diagram of the cyclical nature of the scientific method.

Step 1: Observation, Problem, or Research Question

The procedure begins with identifying a problem or research question, such as a geological phenomenon that is not well explained in the scientific community's collective knowledge. This step usually involves reviewing the scientific literature to understand previous studies that may be related to the question.

Step 2: Hypothesis

Once the problem or question is well defined, the scientist proposes a possible answer, a **hypothesis**, before conducting an experiment or fieldwork. This hypothesis must be specific, falsifiable, and should be based on other scientific work. Geologists often develop multiple working hypotheses because they usually cannot impose strict experimental controls or have limited opportunities to visit a field location [5; 6; 7].

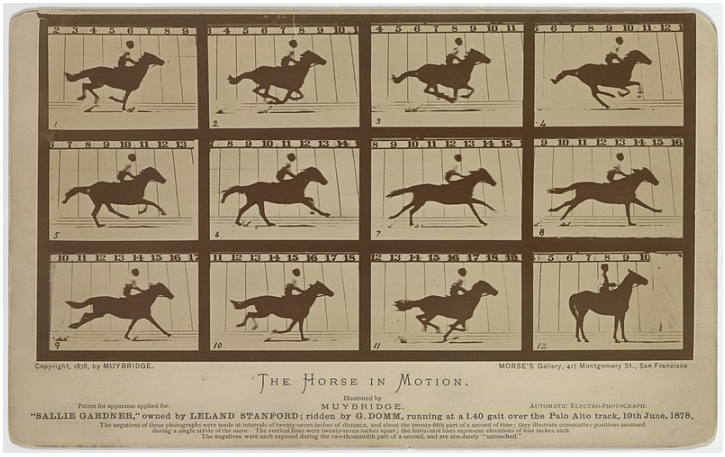


Figure 1.2.2: A famous hypothesis: Leland Stanford wanted to know if a horse lifted all 4 legs off the ground during a gallop since the legs are too fast for the human eye to perceive it. This series of photographs by Eadweard Muybridge proved the horse, in fact, does have all four legs off the ground during the gallop.

Step 3: Experiment and Hypothesis Revision

The next step is developing an **experiment** that either supports or refutes the hypothesis. Many people mistakenly think experiments are only done in a lab; however, an experiment can consist of observing natural processes in the field. Regardless of what form an experiment takes, it always includes the systematic gathering of objective **data**. This data is interpreted to determine whether it contradicts or supports the hypothesis, which may be revised and tested again. When a hypothesis holds up under experimentation, it is ready to be shared with other experts in the field.



Figure 1.2.3: An experiment at the University of Queensland has been going on since 1927. A petroleum product called pitch, which is highly viscous, drips out of a funnel about once per decade. A 9-volt battery is shown for scale.

Step 4: Peer Review, Publication, and Replication

Scientists share the results of their research by publishing articles in scientific journals, such as *Science* and *Nature*. Reputable journals and publishing houses will not publish an experimental study until they have determined its methods are scientifically rigorous and the conclusions are supported by evidence. Before an article is published, it undergoes a rigorous **peer review** by scientific experts who scrutinize the methods, results, and discussion. Once an article is published, other scientists may attempt to replicate the results. This **replication** is necessary to confirm the reliability of the study's reported results. A hypothesis that seemed compelling in one study might be proven false in studies conducted by other scientists. New technology can be applied to published studies, which can aid in confirming or rejecting once-accepted ideas and/or hypotheses.

Step 5: Theory Development

In casual conversation, the word *theory* implies guesswork or speculation. In the language of science, an explanation or conclusion made in a **theory** carries much more weight because it is supported by experimental verification and widely accepted by the scientific community. After a hypothesis has been repeatedly tested for falsifiability through documented and independent studies, it eventually becomes accepted as a scientific theory.

While a hypothesis provides a tentative explanation *before* an experiment, a theory is the best explanation *after* being confirmed by multiple independent experiments. Confirmation of a theory may take years, or even longer. For example, the continental drift hypothesis first proposed by Alfred Wegener in 1912 was initially dismissed. After decades of additional evidence collection by other scientists using more advanced technology, Wegener's hypothesis was accepted and revised as the theory of plate tectonics.

The theory of evolution by natural selection is another example. Originating from the work of Charles Darwin in the mid-19th century, the theory of evolution has withstood generations of scientific testing for falsifiability. While it has been updated and revised to accommodate knowledge gained by using modern technologies, the theory of evolution continues to be supported by the latest evidence.



Figure 1.2.4: Wegener later in his life, ca. 1924-1930.

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1.3: Early Scientific Thought

Western scientific thought began in the ancient city of Athens, Greece [8]. Athens was governed as a democracy, which encouraged individuals to think independently, at a time when most civilizations were ruled by monarchies or military conquerors. Foremost among the early philosopher/scientists to use **empirical thinking** was Aristotle, born in 384 BCE. Empiricism emphasizes the value of evidence gained from experimentation and observation. Aristotle studied under Plato and tutored Alexander the Great. Alexander would later conquer the Persian Empire, and in the process spread Greek culture as far east as India.



Figure 1.3.1: Fresco by Raphael of Plato (left) and Aristotle (right).

Aristotle applied an empirical method of analysis called **deductive reasoning**, which applies known principles of thought to establish new ideas or predict new outcomes. Deductive reasoning starts with generalized principles and logically extends them to new ideas or specific conclusions. If the initial principle is valid, then it is highly likely the conclusion is also valid. An example of deductive reasoning is if A=B, and B=C, then A=C. Another example is if all birds have feathers, and a sparrow is a bird, then a sparrow must also have feathers. The problem with deductive reasoning is if the initial principle is flawed, the conclusion will inherit that flaw. Here is an example of a flawed initial principle leading to the wrong conclusion: if all animals that fly are birds, and bats also fly, then bats must also be birds.

This type of empirical thinking contrasts with **inductive reasoning**, which begins from new observations and attempts to discern underlying generalized principles. A conclusion made through inductive reasoning comes from analyzing measurable evidence, rather than making a logical connection. For example, to determine whether bats are birds, a scientist might list various characteristics observed in birds--the presence of feathers, a toothless beak, hollow bones, lack of forelegs, and externally laid eggs. Next, the scientist would check whether bats share the same characteristics, and if they do not, draw the conclusion that bats are not birds.

Both types of reasoning are important in science because they emphasize the two most important aspects of science: **observation** and **inference**. Scientists test existing principles to see if they accurately infer or predict their observations. They also analyze new observations to determine if the inferred underlying principles still support them [9; 10].



Figure 1.3.2: Drawing of Avicenna (Ibn Sina). He is among the first to link mountains to earthquakes and erosion.

Greek culture was spread by Alexander and then absorbed by the Romans, who help further extend Greek knowledge into Europe through their vast infrastructure of roads, bridges, and aqueducts [11]. After the fall of the Roman Empire in 476 CE, scientific progress in Europe stalled [8]. Scientific thinkers of medieval time had such high regard for Aristotle's wisdom and knowledge they faithfully followed his logical approach to understanding nature for centuries. By contrast, science in the Middle East flourished and grew between 800 and 1450 CE, along with culture and the arts.

Near the end of the medieval period, empirical experimentation became more common in Europe. During the Renaissance, which lasted from the 14th through 17th centuries, artistic and scientific thought experienced a great awakening [12; 13; 14]. European scholars began to criticize the traditional Aristotelian approach and by the end of the Renaissance period, empiricism was poised to become a key component of the scientific revolution that would arise in the 17th century [15].

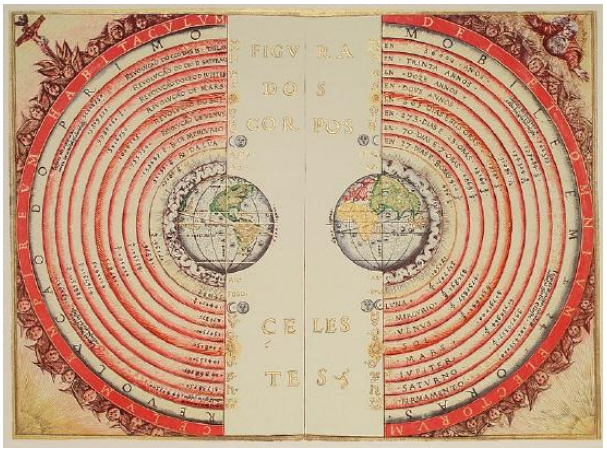


Figure 1.3.3: Geocentric drawing by Bartolomeu Velho in 1568.

An early example of how Renaissance scientists began to apply a modern empirical approach is their study of the solar system. In the second century, the Greek astronomer Claudius Ptolemy observed the Sun, Moon, and stars moving across the sky. Applying Aristotelian logic to his astronomical calculations, he deductively reasoned all celestial bodies orbited around the Earth, which was located at the center of the universe. Ptolemy was a highly regarded mathematician, and his mathematical calculations were widely accepted by the scientific community. The view of the cosmos with Earth at its center is called the **geocentric model**. This geocentric model persisted until the Renaissance period, when some revolutionary thinkers challenged the centuries-old hypothesis.

By contrast, early Renaissance scholars such as astronomer Nicolaus Copernicus (1473-1543) proposed an alternative explanation for the perceived movement of the Sun, Moon, and stars. Sometime between 1507 and 1515, he provided a credible mathematical proof for a radically new model of the cosmos, one in which the Earth and other planets orbited around a centrally located Sun. After the invention of the telescope in 1608, scientists used their enhanced astronomical observations to support this **heliocentric**, Sun-centered, model [16; 17].

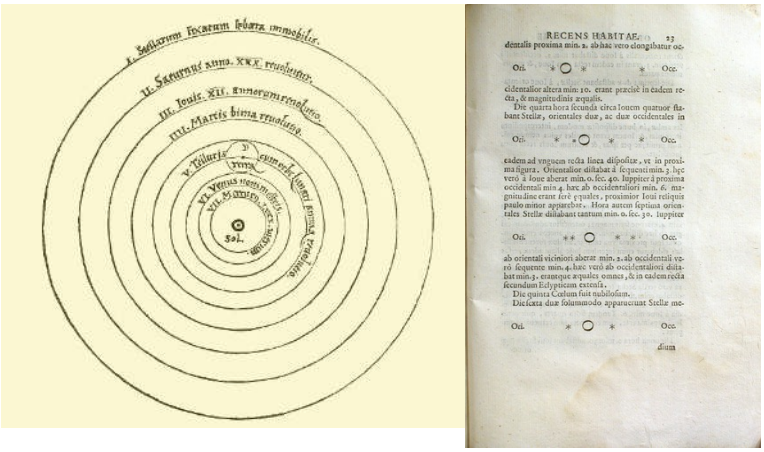


Figure 1.3.4: Copernicus' heliocentric model (left). Galileo's first mention of the moons of Jupiter (right).

Two scientists, Johannes Kepler and Galileo Galilei, are credited with jump-starting the scientific revolution [15]. They accomplished this by building on Copernicus' work and challenging long-established ideas about nature and science.

Johannes Kepler (1571-1630) was a German mathematician and astronomer who expanded on the heliocentric model—improving Copernicus' original calculations and describing planetary motion as elliptical paths. Galileo Galilei (1564 – 1642) was an Italian astronomer who used the newly developed telescope to observe the four largest moons of Jupiter [18]. This was the first piece of direct evidence to contradict the geocentric model since moons orbiting Jupiter could not also be orbiting Earth.

Galileo strongly supported the heliocentric model and attacked the geocentric model, arguing for a more scientific approach to determine the credibility of an idea [19]. Because of this, he found himself at odds with prevailing scientific views and the Catholic Church. In 1633 he was found guilty of heresy and placed under house arrest, where he would remain until his death in 1642 [18; 19].

Galileo is regarded as the first modern scientist because he conducted experiments that would prove or disprove falsifiable ideas and based his conclusions on mathematical analysis of quantifiable evidence—a radical departure from the deductive thinking of Greek philosophers such as Aristotle [15; 18]. His methods marked the beginning of a major shift in how scientists studied the natural world, with an increasing number of them relying on evidence and experimentation to form their hypotheses. It was during this revolutionary time that geologists such as James Hutton and Nicolas Steno also made great advances in their scientific fields of study [15].

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1.4: The Study of Geology

Geologists apply the scientific method to learn about Earth's materials and processes. Geology plays an important role in society; its principles are essential to locating, extracting, and managing natural resources; evaluating environmental impacts of using or extracting these resources; as well as understanding and mitigating the effects of natural hazards.



Figure 1.4.1: A class looks at rocks in Zion National Park.

Geology often applies information from physics and chemistry to the natural world, like understanding the physical forces in a landslide or the chemical interaction between water and rocks. The term comes from the Greek word *geo*, meaning Earth, and *logos*, meaning to think or reckon with.

Why Study Geology?



Figure 1.4.2: Hoover Dam provides hydroelectric energy and stores water for southern Nevada.

Geology plays a key role in how we use **natural resources**—any naturally occurring material that can be extracted from the Earth for economic gain. Our developed modern society, like all societies before it, is dependent on geologic resources. Geologists are involved in extracting fossil fuels, such as coal and petroleum; metals such as copper, aluminum, and iron; and water resources in streams and underground reservoirs inside soil and rocks. They can help conserve our planet's finite supply of **nonrenewable resources**, like petroleum, which are fixed in quantity and depleted by consumption. Geologists can also help manage **renewable resources** that can be replaced or regenerated, such as solar or wind energy, and timber.



Figure 1.4.3: Coal power plant in Helper, Utah.

Resource extraction and usage impact our environment, which can negatively affect human health. For example, burning fossil fuels releases chemicals into the air that are unhealthy for humans, especially children. Mining activities can release toxic heavy metals, such as lead and mercury, into the soil and waterways. Our choices will have an effect on Earth's environment for the foreseeable future. Understanding the remaining quantity, extractability, and renewability of geologic resources will help us better sustainably manage those resources.



Figure 1.4.4: Buildings toppled from liquefaction during a 7.5 magnitude earthquake in Japan.

Geologists also study natural hazards created by geologic processes. **Natural hazards** are phenomena that are potentially dangerous to human life or property. No place on Earth is completely free of natural hazards, so one of the best ways people can protect themselves is by understanding the geology. Geology can teach people about the natural hazards in an area and how to prepare for them. Geologic hazards include landslides, earthquakes, tsunamis, floods, volcanic eruptions, and sea-level rise.



Figure 1.4.5: Oregon's Crater Lake was formed about 7700 years ago after the eruption of Mount Mazama.

Finally, geology is where other scientific disciplines intersect in the concept known as **Earth System Science**. In science, a **system** is a group of interactive objects and processes. Earth System Science views the entire planet as a combination of systems that interact with each other via complex relationships. This geology textbook provides an introduction to science in general and will often reference other scientific disciplines.

Earth System Science includes five basic systems (or spheres), the **geosphere** (the solid body of the Earth), the **atmosphere** (the gas envelope surrounding the Earth), the **hydrosphere** (water in all its forms at and near the surface of the Earth), the **cryosphere** (frozen water part of Earth), and the **biosphere** (life on Earth in all its forms and interactions, including humankind).

Rather than viewing geology as an isolated system, Earth system scientists study how geologic processes shape not only the world but all the spheres it contains. They study how these multidisciplinary spheres relate, interact, and change in response to natural cycles and human-driven forces. They use elements from physics, chemistry, biology, meteorology, environmental science, zoology, hydrology, and many other sciences.

The Geologist's Tools



Figure 1.4.6: Iconic Archaeopteryx lithographica fossil from Germany.

In its simplest form, a geologist's tool may be a rock hammer used for sampling a fresh surface of a rock. A basic toolset for fieldwork might also include:

- Magnifying lens for looking at mineralogical details
- Compass for measuring the orientation of geologic features
- Map for documenting the local distribution of rocks and minerals
- Magnet for identifying magnetic minerals like magnetite
- Dilute solution of hydrochloric acid to identify carbonate-containing minerals like calcite or limestone.

In the laboratory, geologists use optical microscopes to closely examine rocks and soil for mineral composition and grain size. Laser and mass spectrometers precisely measure the chemical composition and geological age of minerals. Seismographs record and locate earthquake activity, or when used in conjunction with ground-penetrating radar, locate objects buried beneath the surface of the earth. Scientists apply computer simulations to turn their collected data into testable, theoretical models. Hydrogeologists drill wells to sample and analyze underground water quality and availability. Geochemists use scanning electron microscopes to analyze minerals at the atomic level, via x-rays. Other geologists use gas chromatography to analyze liquids and gases trapped in glacial ice or rocks.

Technology provides new tools for scientific observation, which leads to new evidence that helps scientists revise and even refute old ideas. Because the ultimate technology will never be discovered, the ultimate observation will never be made. And this is the beauty of science—it is ever-advancing and always discovering something new.

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1.5: Science Denial and Evaluating Sources

Introductory science courses usually deal with accepted scientific theory and do not include opposing ideas, even though these alternate ideas may be credible. This makes it easier for students to understand complex material. Advanced students will encounter more controversies as they continue to study their discipline.

Some groups of people argue that some established scientific theories are wrong, not based on their scientific merit but rather on the ideology of the group. This section focuses on how to identify evidence-based information and differentiate it from pseudoscience.



Figure 1.5.1: Anti-Evolution League at the infamous Tennessee v. Scopes trial.

Science Denial

Science denial happens when people argue that established scientific theories are wrong, not based on scientific merit but rather on subjective ideology—such as for social, political, or economic reasons. Organizations and people use science denial as a rhetorical argument against issues or ideas they oppose. Three examples of science denial versus science are:

1. Teaching evolution in public schools
2. Linking tobacco smoke to cancer
3. Linking human activity to climate change

Among these, denial of climate change is strongly connected with geology and Earth science. A climate denier specifically denies or doubts the objective conclusions of Earth scientists, geologists and climate scientists.



Figure 1.5.2: 2017 March for Science in Washington, DC. This and other similar marches were in response to funding cuts and anti-science rhetoric.

Science denial generally uses three false arguments. The first argument tries to undermine the credibility of the scientific conclusion by claiming the research methods are flawed or the theory is not universally accepted—the science is unsettled. The

notion that scientific ideas are not absolute creates doubt for non-scientists; however, a lack of universal truths should not be confused with scientific uncertainty. Because science is based on falsifiability, scientists avoid claiming universal truths and use language that conveys uncertainty. This allows scientific ideas to change and evolve as more evidence is uncovered.

The second argument claims the researchers are not objective and are motivated by ideology or economic agenda. This is an ***ad hominem* argument** in which a person's character is attacked instead of the merit of their argument. They claim results have been manipulated so researchers can justify asking for more funding. They claim that because the researchers are funded by a federal grant, they are using their results to lobby for expanded government regulation.

The third argument is to demand a balanced view, equal time in media coverage and educational curricula, to engender the false illusion of two equally valid arguments. Science deniers frequently demand equal coverage of their proposals, even when there is little scientific evidence supporting their ideology. For example, science deniers might demand religious explanations to be taught as an alternative to the well-established theory of evolution [39; 40]. Or that all possible causes of climate change be discussed as equally probable, regardless of the body of evidence. Conclusions derived using the scientific method should not be confused with those based on ideologies.

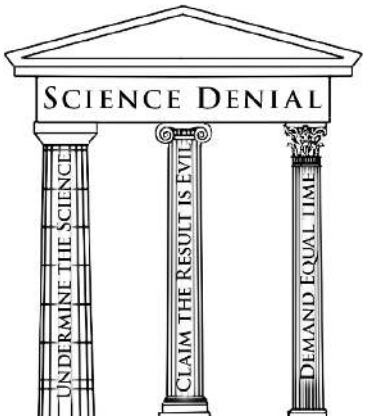


Figure 1.5.3: Three false rhetorical arguments of science denial (Source: National Center for Science Education)

Furthermore, conclusions about nature derived from ideologies have no place in science research and education. For example, it would be inappropriate to teach the flat Earth model in a modern Earth science or geology course because this idea has been disproved by the scientific method. The formation of new conclusions based on the scientific method is the only way to change scientific conclusions. The fact that scientists avoid universal truths and change their ideas as more evidence is uncovered shouldn't be seen as meaning that the science is unsettled. Unfortunately, widespread scientific illiteracy allows these arguments to be used to suppress scientific knowledge and spread misinformation.

In a classic case of science denial, beginning in the 1960s and for the next three decades, the tobacco industry and their scientists used rhetorical arguments to deny a connection between tobacco usage and cancer. Once it became clear scientific studies overwhelmingly found that using tobacco dramatically increased a person's likelihood of getting cancer, their next strategy was to create a sense of doubt about the science. The tobacco industry suggested the results were not yet fully understood and more study was needed. They used this doubt to lobby for delaying legislative action that would warn consumers of the potential health hazards [39, 41]. This same tactic is currently being employed by those who deny the significance of human involvement in climate change.

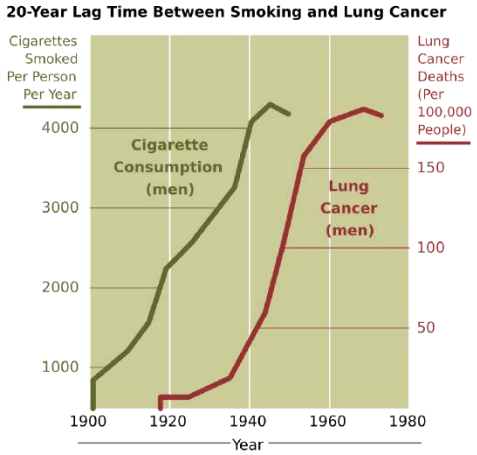


Figure 1.5.4: The lag time between cancer after smoking, plus the ethics of running human trials, delayed the government in taking action against tobacco.

Evaluating Sources of Information

In the age of the internet, information is plentiful. Geologists, scientists, or anyone exploring scientific inquiry must discern valid sources of information from pseudoscience and **misinformation**. This evaluation is especially important in scientific research because scientific knowledge is respected for its reliability [42]. Textbooks such as this one can aid this complex and crucial task. At its roots, quality information comes from the scientific method [43], beginning with the empirical thinking of Aristotle. The application of the scientific method helps produce **unbiased** results. A valid inference or interpretation is based on objective evidence or data. Credible data and inferences are clearly labeled, separated, and differentiated. Anyone looking over the data can understand how the author's conclusion was derived or come to an alternative conclusion. Scientific procedures are clearly defined so the investigation can be replicated to confirm the original results or expanded further to produce new results. These measures make a scientific inquiry valid and its use as a source reputable. Of course, substandard work occasionally slips through and retractions are published from time to time. An infamous article linking the MMR vaccine to autism appeared in the highly reputable journal *Lancet* in 1998. Journalists discovered the author had multiple conflicts of interest and fabricated data, and the article was retracted in 2010.

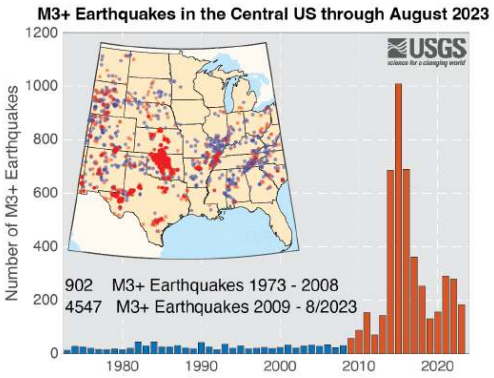


Figure 1.5.5: This graph shows earthquake data. To call these earthquakes induced, due to fracking/drilling, is an interpretation. Since the USGS (a reputable institution) has interpreted these earthquakes as caused by humans, it is a more reliable interpretation.

In addition to methodology, data, and results, the author(s) of a study should be investigated [44]. An author's credibility is based on multiple factors, such as having a degree in a relevant topic or being funded from an unbiased source.

The same rigor should be applied to evaluating the publisher, ensuring the results reported come from an unbiased process [45]. The publisher should be easy to discover. Good publishers will show the latest papers in the journal and make their contact information and identification clear. Reputable journals show their peer review style. Some journals are predatory, where they use unexplained and unnecessary fees to submit and access journals. Reputable journals have recognizable editorial boards. Often, a reliable journal will associate with a trade, association, or recognized open-source initiative.



Figure 1.5.6: Logo for The Geological Society of America (GSA), one of the leading geoscience organizations. They also publish GSA Bulletin, a reputable geology journal.

One of the hallmarks of scientific research is peer review. Research should be transparent to peer review. This allows the scientific community to reproduce experimental results, correct and retract errors, and validate theories. This allows the reproduction of experimental results, corrections of errors, and proper justification of the research to experts.

Citation is not only imperative to avoid plagiarism, but also allows readers to investigate an author's line of thought and conclusions. When reading scientific works, it is important to confirm the citations are from reputable scientific research. Most often, scientific citations are used to reference paraphrasing rather than quotes. The number of times a work is cited is said to measure the influence an investigation has within the scientific community, although this technique is inherently biased [46].

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1.S: Summary

Science is a process, with no beginning and no end. Science is never finished because a full truth can never be known. However, science and the scientific method are the best way to understand the universe we live in. Scientists draw conclusions based on objective evidence; they consolidate these conclusions into unifying models. Geologists likewise understand studying the Earth is an ongoing process, beginning with James Hutton who declared the Earth has "…no vestige of a beginning, no prospect of an end." Geologists explore the 4.5 billion-year history of Earth, its resources, and its many hazards. From a larger viewpoint, earth science can teach people how to develop credible conclusions, as well as identify and stop misinformation.

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CHAPTER OVERVIEW

2: Plate Tectonics

Learning Objectives

By the end of this chapter, students should be able to:

- Describe how the ideas behind plate tectonics started with Alfred Wegener's hypothesis of continental drift.
- Describe the physical and chemical layers of the Earth and how they affect plate movement.
- Explain how movement at the three types of plate boundaries causes earthquakes, volcanoes, and mountain building.
- Identify convergent boundaries, including subduction and collisions, as places where plates come together.
- Identify divergent boundaries, including rifts and mid-ocean ridges, as places where plates separate.
- Explain transform boundaries as places where adjacent plates shear past each other.
- Describe the Wilson Cycle, beginning with continental rifting, ocean basin creation, plate subduction, and ending with ocean basin closure.
- Explain how the tracks of hotspots, places that have continually rising magma, is used to calculate plate motion.

Revolution is a word usually reserved for significant political or social changes. Several of these idea revolutions forced scientists to re-examine their entire field, triggering a paradigm shift that shook up their conventionally held knowledge. Charles Darwin's book on evolution, *On the Origin of Species*, published in 1859; Gregor Mendel's discovery of the genetic principles of inheritance in 1866; and James Watson, Francis Crick, and Rosalind Franklin's model for the structure of DNA in 1953 did that for biology. Albert Einstein's relativity and quantum mechanics concepts in the early twentieth century did the same for Newtonian physics.

The concept of plate tectonics was just as revolutionary for geology. The theory of plate tectonics attributes the movement of massive sections of the Earth's outer layers with creating earthquakes, mountains, and volcanoes. Many earth processes make more sense when viewed through the lens of plate tectonics. Because it is so important in understanding how the world works, plate tectonics is the first topic of discussion in this textbook.

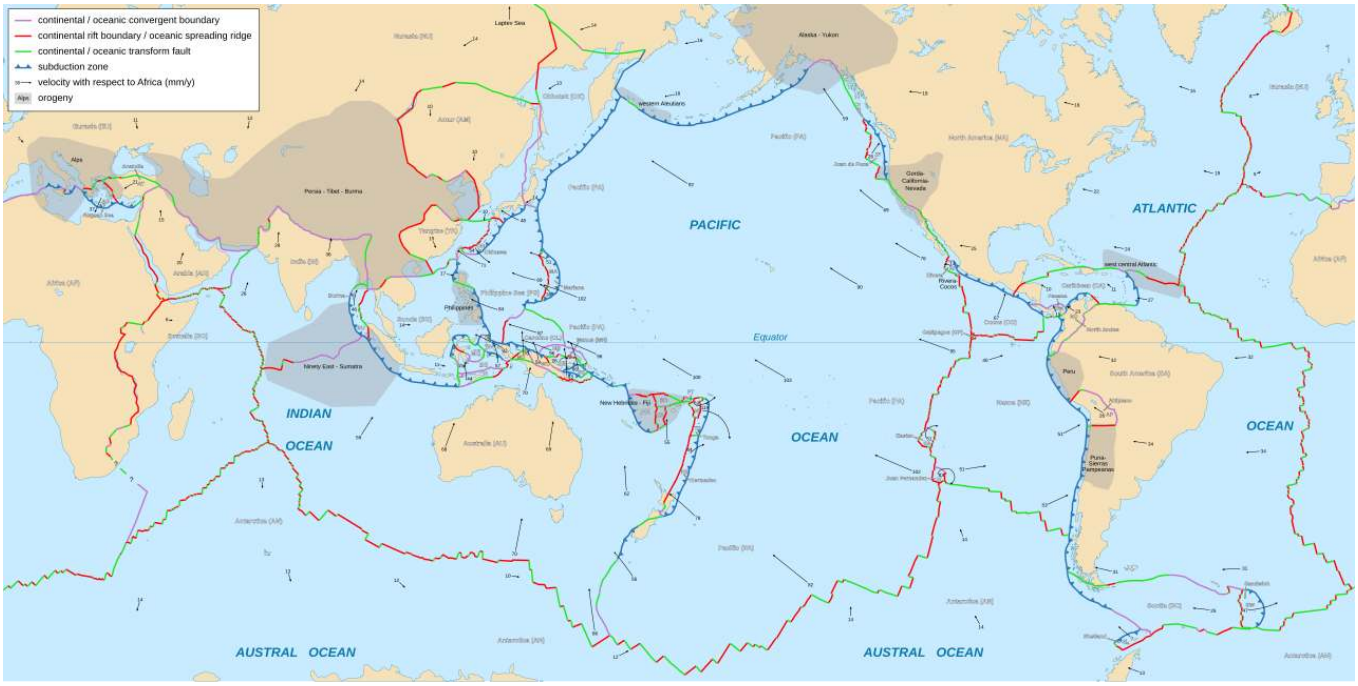


Figure 2.1: Detailed map of all known plates, their boundaries, and movements. (By Eric Gaba; CC BY-SA 2.5 via Wikimedia Commons.)

2.1: Alfred Wegener’s Continental Drift Hypothesis
2.2: Layers of the Earth
2.3: Convergent Boundaries
2.4: Divergent Boundaries
2.5: Transform Boundaries
2.6: The Wilson Cycle
2.7: Hotspots
2.S: Summary

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2.1: Alfred Wegener's Continental Drift Hypothesis

Alfred Wegener (1880-1930) was a German scientist who specialized in meteorology and climatology. His knack for questioning accepted ideas started in 1910 when he disagreed with the explanation that the Bering Land Bridge was formed by isostasy and that similar land bridges once connected the continents [1]. After reviewing the scientific literature, he published a hypothesis stating the continents were originally connected and then drifted apart. While he did not have the precise mechanism worked out, his hypothesis was backed up by a long list of evidence.



Figure 2.1.1: Wegener later in his life, ca. 1924-1930. (By unknown; public domain via Wikimedia Commons.)

Early Evidence for the Continental Drift Hypothesis

Wegener's first piece of evidence was that the coastlines of some continents fit together like pieces of a jigsaw puzzle. People noticed the similarities in the coastlines of South America and Africa on the first world maps, and some suggested the continents had been ripped apart [3]. Antonio Snider-Pellegrini did preliminary work on continental separation and matching fossils in 1858.

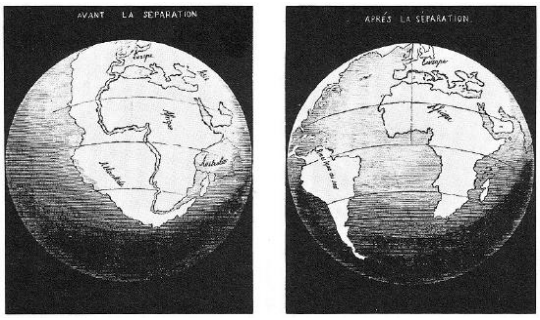


Figure 2.1.2: Snider-Pellegrini's map showing the continental fit and separation, 1858. (By Antonio Snider-Pellegrini; public domain via Wikimedia Commons.)

What Wegener did differently was synthesize a large amount of data in one place. He used the true edges of the continents, based on the shapes of the **continental shelves** [4]. This resulted in a better fit than previous efforts that traced the existing coastlines [5].

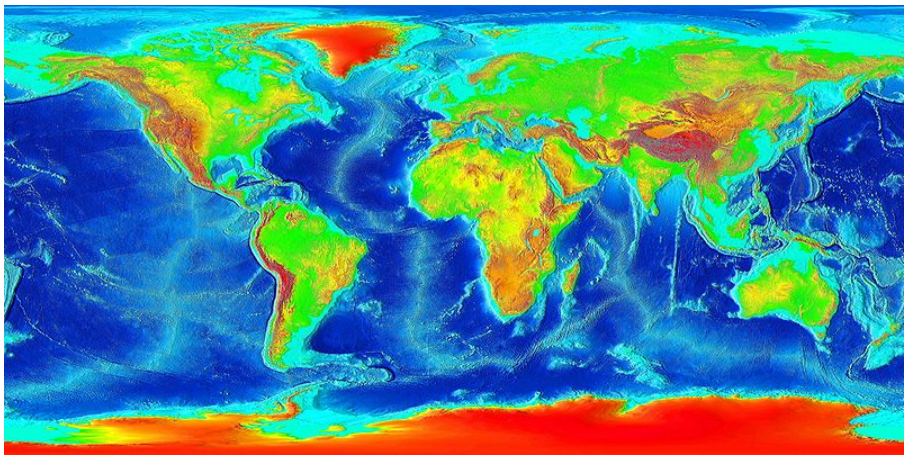


Figure 2.1.3: Map of world elevations. Note the light blue, which are continental shelves flooded by shallow ocean water. These show the true shapes of the continents. (By Tahaisik; CC BY-SA 3.0 via Wikimedia Commons.)

Wegener also compiled evidence by comparing similar rocks, mountains, fossils, and glacial formations across oceans. For example, the fossils of the primitive aquatic reptile *Mesosaurus* were found on the separate coastlines of Africa and South America. Fossils of another reptile, *Lystrosaurus*, were found in Africa, India, and Antarctica. He pointed out these land-dwelling creatures could not have swum across an entire ocean.

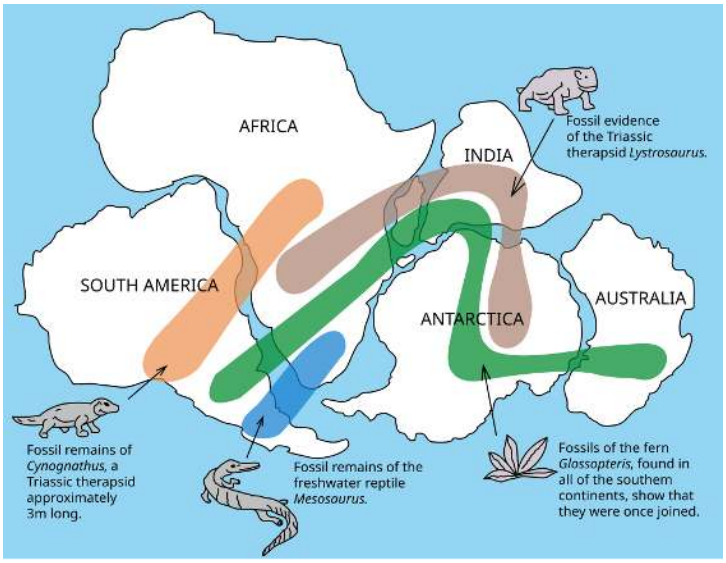


Figure 2.1.4: Image showing fossils that connect the continents of Gondwana (the southern continents of Pangea). (By Osvaldocangaspadilla; public domain via Wikimedia Commons.)

Opponents of **continental drift** insisted trans-oceanic land bridges allowed animals and plants to move between continents [6]. The land bridges eventually eroded away, leaving the continents permanently separated. The problem with this hypothesis is the improbability of a land bridge being tall and long enough to stretch across a broad, deep ocean.

More support for continental drift came from the puzzling evidence that glaciers once existed in normally very warm areas in southern Africa, India, Australia, and Arabia. These climate anomalies could not be explained by land bridges. Wegener found similar evidence when he discovered tropical plant fossils in the frozen region of the Arctic Circle. As Wegener collected more data, he realized the explanation that best fit all the climate, rock, and fossil observations involved moving continents.

Proposed Mechanism for Continental Drift

Wegener's work was considered a fringe science theory for his entire life. One of the biggest flaws in his hypothesis was the inability to provide a mechanism for how the continents moved. Obviously, the continents did not appear to move, and changing the conservative minds of the scientific community would require exceptional evidence that supported a credible mechanism. Other pro-continental drift followers used expansion, contraction, or even the Moon's origin to explain how the continents moved.

Wegener used centrifugal forces and precession, but this model was proven wrong [7]. He also speculated about seafloor spreading, with hints of **convection**, but could not substantiate these proposals [8]. As it turns out, current scientific knowledge revealed convection as the major force in driving plate movements.

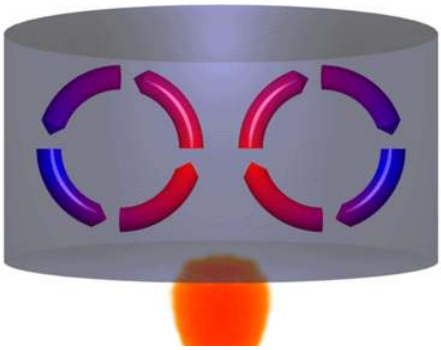


Figure 2.1.5: Animation of the basic idea of convection: an uneven heat source in a fluid causes rising material next to the heat and sinking material far from the heat. (By Oni Lukos; CC BY-SA 3.0 via Wikimedia Commons.)

Development of Plate Tectonic Theory

Wegener died in 1930 on an expedition in Greenland. Poorly respected in his lifetime, Wegener and his ideas about moving continents seemed destined to be lost in history as fringe science. However, in the 1950s, evidence started to trickle in that made continental drift a more viable idea. By the 1960s, scientists had amassed enough evidence to support the missing mechanism—namely, seafloor spreading—for Wegener's hypothesis of continental drift to be accepted as the **theory of plate tectonics**. Ongoing GPS and earthquake data analyses continue to support this theory. The next section provides the pieces of evidence that helped transform one man's wild notion into a scientific theory.

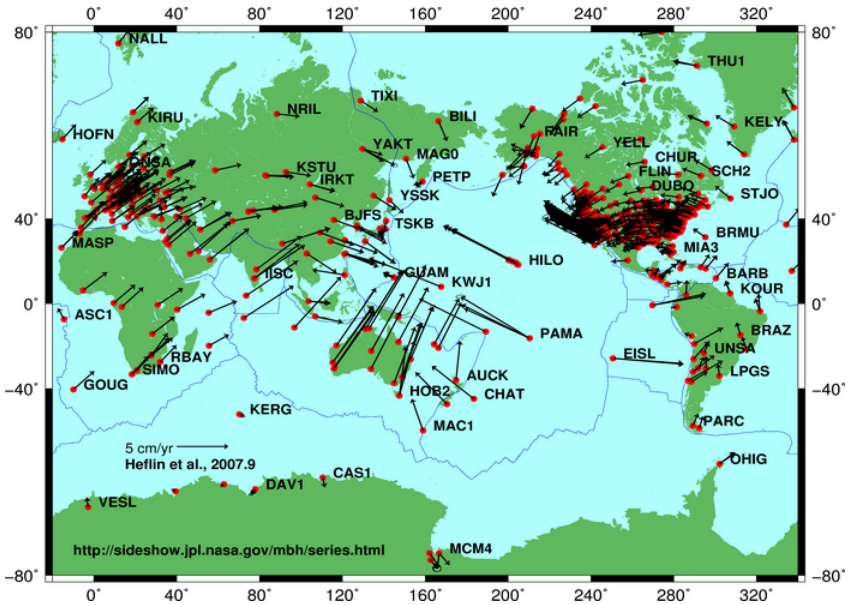


Figure 2.1.6: GPS measurements of plate motions. Black arrows point in the direction of motion and the length represents the amount of movement per year. (By NASA; public domain via Wikimedia Commons.)

Mapping of the Ocean Floors

In 1947 researchers started using an adaptation of SONAR to map a region in the middle of the Atlantic Ocean with poorly-understood topographic and thermal properties [9]. Using this information, Bruce Heezen and Marie Tharp created the first detailed map of the ocean floor to reveal the **Mid-Atlantic Ridge** [10], a basaltic mountain range that spanned the length of the Atlantic Ocean, with rock chemistry and dimensions unlike the mountains found on the continents. Initially, scientists thought the ridge was part of a mechanism that explained the expanding Earth or ocean-basin growth hypotheses [11; 12]. In 1959, Harry Hess proposed the hypothesis of **seafloor spreading** – that the **mid-ocean ridges** represented tectonic plate factories, where a new **oceanic plate**

was being created from these long volcanic ridges. Scientists later included transform faults perpendicular to the ridges to better account for varying rates of movement between the newly formed plates [13]. When earthquake epicenters were discovered along the ridges, the idea that earthquakes were linked to plate movement took hold [14].

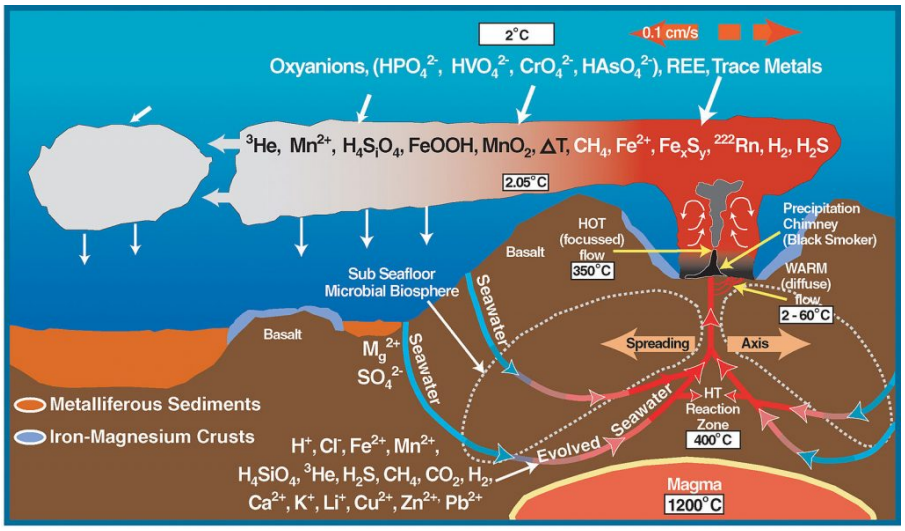


Figure 2.1.7: The complex chemistry around mid-ocean ridges. (By NOAA; public domain via Wikimedia Commons.)

Seafloor sediment, measured by dredging and drilling, provided another clue. Scientists once believed sediment accumulated on the ocean floors over a very long time in a static environment. When some studies showed less sediment than expected, these results were initially used to argue against the continental movement [15; 16]. With more time, researchers discovered these thinner sediment layers were located close to mid-ocean ridges, indicating the ridges were younger than the surrounding ocean floor. This finding supported the idea that the seafloor was not fixed in one place [17].

Paleomagnetism

Paleomagnetism is the study of magnetic fields frozen within rocks, basically a fossilized compass. In fact, the first hard evidence to support plate motion came from paleomagnetism.

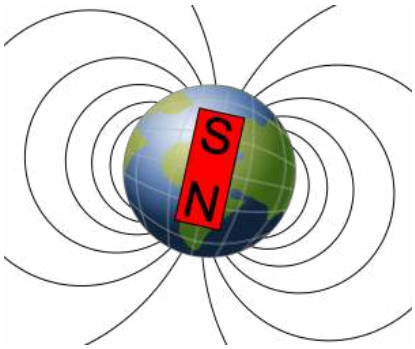


Figure 2.1.8: The magnetic field of Earth, simplified as a bar magnet. Note that Earth's North Pole is actually the south pole of Earth's magnetic field. (By Zureks; public domain via Wikimedia Commons.)

Igneous rocks containing magnetic minerals like magnetite typically provide the most useful data. In their liquid state as magma or lava, the magnetic poles of the minerals align themselves with the Earth's magnetic field. When the rock cools and solidifies, this alignment is frozen into place, creating a permanent paleomagnetic record that includes magnetic inclination, which is related to global latitude, and declination, which is related to magnetic north.

The seafloor was mapped magnetically. Scientists had long known of strange **magnetic anomalies** that formed a striped pattern of symmetrical rows on both sides of mid-oceanic ridges. What made these features unusual was that the north and south magnetic poles within each stripe were reversed in alternating rows [18]. By 1963, Harry Hess and other scientists used these magnetic reversal patterns to support their model for seafloor spreading [19] (see also Lawrence W. Morley [20]).

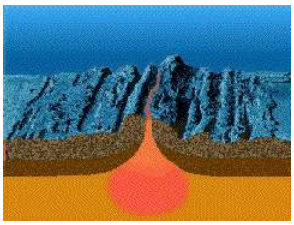


Figure 2.1.9: The iron in the solidifying rock preserves the current magnetic polarity as new oceanic crust forms at mid-ocean ridges. The red and green stripes represent periods of normal and reverse polarity that are mirrored on opposite sides of the ridge. (By USGS; public domain via Wikimedia Commons.)

Scientists had noticed for some time the alignment of magnetic north in many rocks was nowhere close to the Earth's current magnetic north. Some explained this as part of the normal movement of Earth's magnetic north pole. Eventually, scientists realized adding the idea of continental movement explained the data better than the pole movement alone [21].

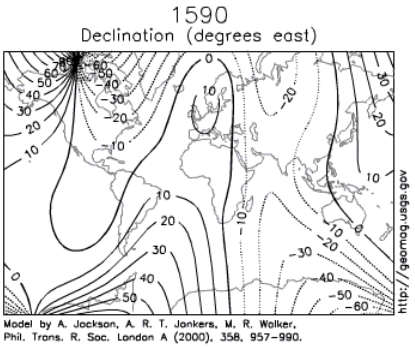


Figure 2.1.10: This animation shows how the magnetic poles have moved over 400 years. As the years pass, the lines show the change in the magnetic declination (the angle between magnetic north and true north) by location. (By USGS; public domain via Wikimedia Commons.)

Wadati-Benioff Zones

Around the same time mid-ocean ridges were being investigated, other scientists linked the creation of ocean trenches and island arcs to seismic activity and tectonic plate movement [22]. Several independent research groups recognized earthquake hypocenters traced the shapes of oceanic plates sinking into the mantle. These deep earthquake zones congregated in planes that started near the surface around ocean trenches and angled beneath the continents and island arcs [23]. Today these earthquake zones called **Wadati-Benioff zones**.

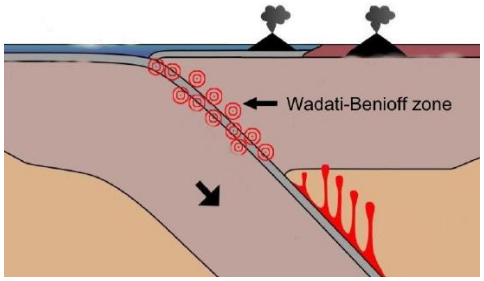


Figure 2.1.11: The Wadati-Benioff zone, showing earthquakes following the subducting slab down. (By Gord-c-lupis; CC BY-SA 3.0 via Wikimedia Commons.)

Based on the mounting evidence, the theory plate tectonics continued to take shape. J. Tuzo Wilson was the first scientist to put the entire picture together by proposing the opening and closing of the ocean basins [24]. Before long, scientists proposed other models showing plates moving with respect to each other, with clear boundaries between them [25]. Others started piecing together complicated histories of tectonic plate movement [26]. The plate tectonic revolution had taken hold.

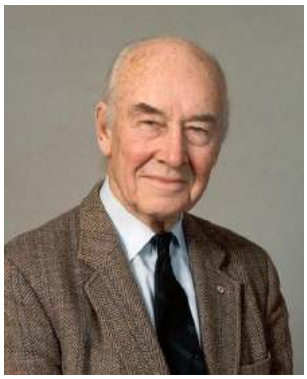


Figure 2.1.12: J. Tuzo Wilson. (By Smorris123 at English Wikipedia; CC BY-SA 3.0 via Wikimedia Commons.)

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2.2: Layers of the Earth

In order to understand the details of plate tectonics, it is essential to first understand the layers of the Earth. Firsthand information about what is below the surface is very limited; most of what we know is pieced together from hypothetical models, and analyzing seismic wave data and meteorite materials. In general, the Earth can be divided into layers based on chemical composition and physical characteristics.

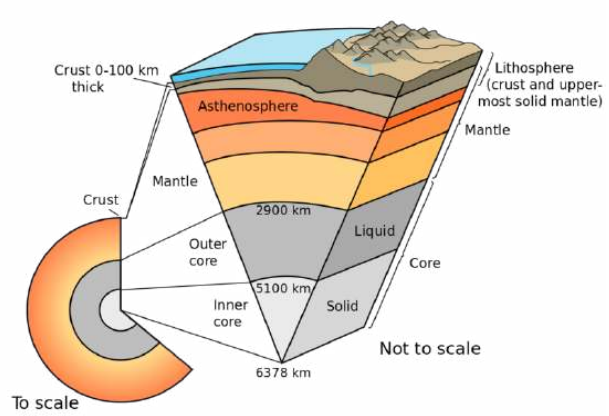


Figure 2.2.1: The layers of the Earth. Physical layers include the lithosphere and asthenosphere; chemical layers are crust, mantle, and core. (By Drlauraguertin; CC BY-SA 3.0 via SEG Wiki.)

Chemical Layers

Certainly, the Earth is composed of countless combinations of **elements**. Regardless of what elements are involved, two major factors—temperature and pressure—are responsible for creating three distinct chemical layers.

Crust

The outermost chemical layer and the one we currently reside on is the **crust**. There are two types of crust. **Continental crust** has a relatively low density and composition similar to granite. **Oceanic crust** has a relatively high density, especially when cold and old, and composition similar to basalt. The surface levels of crust are relatively **brittle**. The deeper parts of the crust are subjected to higher temperatures and pressure, which makes them more **ductile**. Ductile materials are like soft plastics or putty, and they move under force. Brittle materials are like solid glass or pottery, and they break under force, especially when it is applied quickly. Earthquakes, generally occur in the upper crust and are caused by the rapid movement of relatively brittle materials.

The base of the crust is characterized by a large increase in **seismic velocity**, which is a measure of how fast earthquake waves travel through solid matter. Called the Mohorovičić Discontinuity, or **Moho** for short, this zone was discovered by Andrija Mohorovičić (pronounced mo-ho-ro-vee-cheech; audio pronunciation) in 1909 after studying earthquake wave paths in his native Croatia [27]. The change in wave direction and speed is caused by dramatic chemical differences between the crust and mantle. Underneath the oceans, the Moho is found roughly 5 km below the ocean floor. Under the continents, it is located about 30-40 km below the surface. Near certain large mountain-building events known as orogenies, the continental Moho depth is doubled [28].

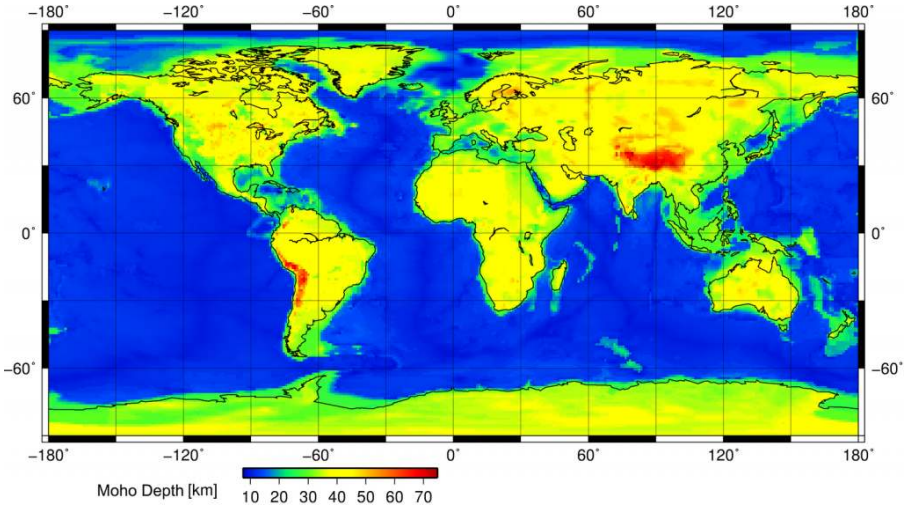


Figure 2.2.2: The global map of the depth of the Moho. (By AllenMcC; CC BY-SA 3.0 via Wikimedia Commons.)

Mantle

The **mantle** sits below the crust and above the core. It is the largest chemical layer by volume, extending from the base of the crust to a depth of about 2900 km [29]. Most of what we know about the mantle comes from seismic wave analysis, though the information is gathered by studying ophiolites and xenoliths. **Ophiolites** are pieces of the mantle that have risen through the crust until they are exposed as part of the ocean floor. **Xenoliths** are carried within magma and brought to the Earth's surface by volcanic eruptions. Most xenoliths are made of **peridotite**, an ultramafic class of igneous rock. Because of this, scientists hypothesize most of the mantle is made of peridotite [30].



Figure 2.2.3: This mantle xenolith containing olivine (green) is chemically weathering by hydrolysis and oxidation into the brown pseudo-mineral iddingsite, which is a complex of water, clay, and iron oxides. The more altered side of the rock has been exposed to the environment longer. (By Matt Affolter; CC BY-SA 3.0 via Wikimedia Commons.)

Core

The **core** of the Earth has both liquid and solid layers, and consists mostly of iron, nickel, and possibly some oxygen [31]. Scientists looking at seismic data first discovered this innermost chemical layer in 1906 [32]. Through a union of hypothetical modeling, astronomical insight, and hard seismic data, they concluded the core is mostly metallic iron [33]. Scientists studying meteorites, which typically contain more iron than surface rocks, have proposed the Earth was formed from meteoric material. They believe the liquid component of the core was created as the iron and nickel sank into the center of the planet, where it was liquefied by intense pressure [34].



Figure 2.2.4: A polished fragment of the iron-rich Toluca Meteorite, with octahedral Widmanstätten Pattern. (By H. Raab; CC BY-SA 3.0 via Wikimedia Commons.)

Physical Layers

The Earth can also be broken down into five distinct physical layers based on how each layer responds to stress. While there is some overlap in the chemical and physical designations of layers, specifically the **core-mantle** boundary, there are significant differences between the two systems.

Lithosphere

Lithos is Greek for stone, and the **lithosphere** is the outermost physical layer of the Earth. It is grouped into two types: oceanic and continental. **Oceanic lithosphere** is thin and relatively rigid. It ranges in thickness from nearly zero in new plates found around mid-ocean ridges, to an average of 140 km in most other locations. **Continental lithosphere** is generally thicker and considerably more plastic, especially at the deeper levels. Its thickness ranges from 40 to 280 km [35]. The lithosphere is not continuous. It is broken into segments called **plates**. A **plate boundary** is where two plates meet and move relative to each other. Plate boundaries are where we see plate tectonics in action—building mountains, triggering earthquakes, and generating volcanic activity.

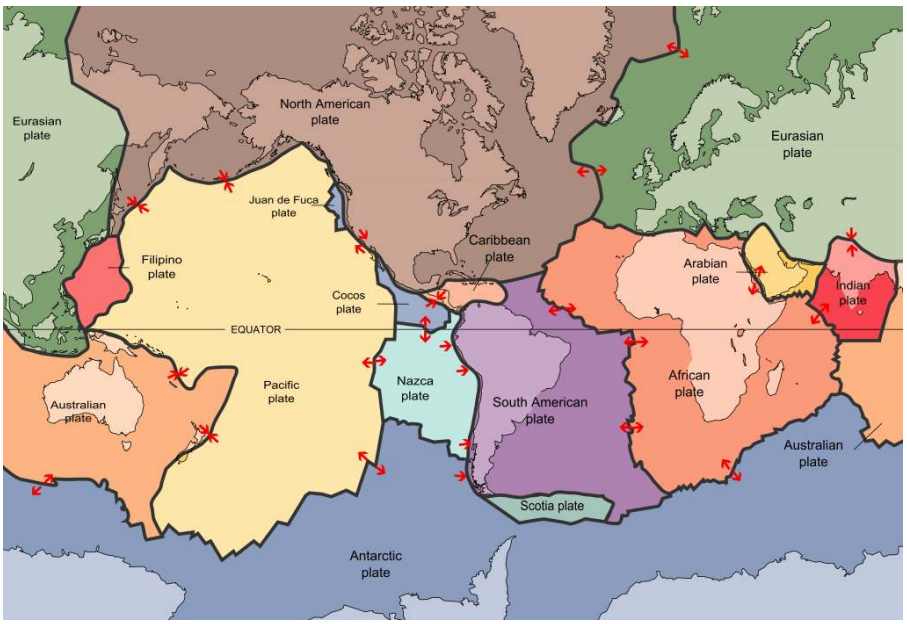


Figure 2.2.5: Map of the major plates with the direction of motion along boundaries indicated with red arrows. (By USGS; public domain via Wikimedia Commons.)

Asthenosphere

The **asthenosphere** is the layer below the lithosphere. *Astheno-* means lacking strength, and the most distinctive property of the asthenosphere is movement. Because it is mechanically weak, this layer moves and flows due to convection currents created by heat coming from the Earth's core [33]. Unlike the lithosphere that consists of multiple plates, the asthenosphere is relatively unbroken. Scientists have determined this by analyzing seismic waves that pass through the layer. The depth at which the

asthenosphere is found is temperature-dependent [36]. It tends to lie closer to the Earth's surface around mid-ocean ridges and much deeper underneath mountains and the centers of lithospheric plates.

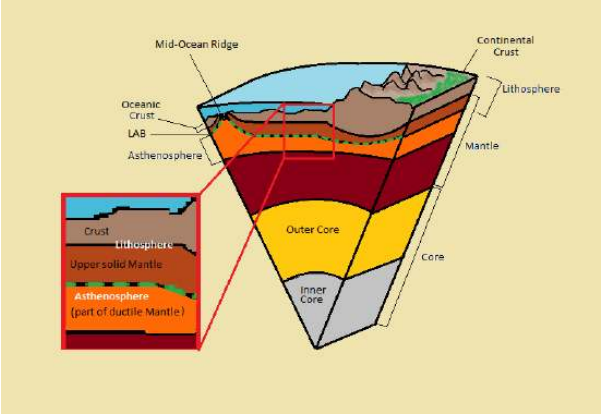


Figure 2.2.5: The lithosphere-asthenosphere boundary changes with certain tectonic situations. (By NealeyS at English Wikipedia; CC BY-SA 3.0 via Wikimedia Commons.)

Mesosphere

The **mesosphere**, sometimes known as the lower mantle, is more rigid and immobile than the asthenosphere. Located at a depth of approximately 410 and 660 km below the Earth's surface, the mesosphere is subjected to very high pressures and temperatures. These extreme conditions create a transition zone in the upper mesosphere where minerals continuously change into various forms or pseudomorphs [37]. Scientists identify this zone by changes in seismic velocity and sometimes physical barriers to movement [38]. Below this transitional zone, the mesosphere is relatively uniform until it reaches the core.

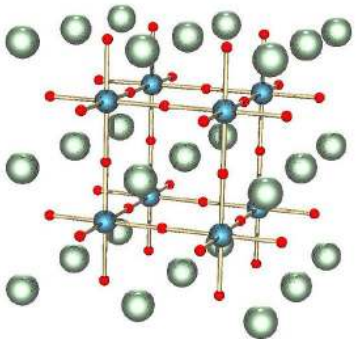


Figure 2.2.6: General perovskite structure. Perovskite silicates (e.g. Bridgmenite, (Mg, Fe)SiO_3) are thought to be the main component of the lower mantle, making it the most common mineral in or on Earth. (By Cadmium at English Wikipedia; public domain via Wikimedia Commons.)

Inner and Outer Core

The **outer core** is the only entirely liquid layer within the Earth. It starts at a depth of 2890 km and extends to 5150 km, making it about 2300 km thick. In 1936, the Danish geophysicist Inge Lehmann analyzed seismic data and was the first to prove a solid **inner core** existed within a liquid outer core [39]. The solid inner core is about 1220 km thick, and the outer core is about 2300 km thick [40].



Figure 2.2.7: Lehmann in 1932. (By Even Neuhaus; public domain via Wikimedia Commons.)

It seems like a contradiction that the hottest part of the Earth is solid, as the minerals making up the core should be liquified or vaporized at this temperature. Immense pressure keeps the minerals of the inner core in a solid phase [41]. The inner core grows slowly from the lower outer core solidifying as heat escapes the interior of the Earth and is dispersed to the outer layers [42].

The Earth's liquid outer core is critically important in maintaining a breathable atmosphere and other environmental conditions favorable for life. Scientists believe the Earth's magnetic field is generated by the circulation of molten iron and nickel within the outer core [43]. If the outer core were to stop circulating or become solid, the loss of the magnetic field would result in Earth getting stripped of life-supporting gases and water. This is what happened, and continues to happen, on Mars [44].

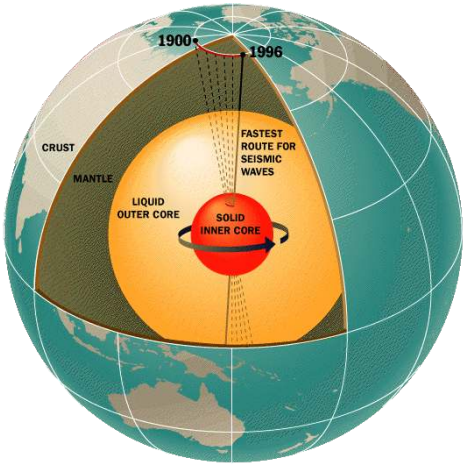


Figure 2.2.8: The outer core's spin most likely generates our protective magnetic field. (By NASA; public domain via Wikimedia Commons.)

Plate Tectonic Boundaries

At **passive margins**, the plates don't move—the continental lithosphere transitions into the oceanic lithosphere and forms plates made of both types. A tectonic plate may be made of both oceanic and continental lithosphere connected by a passive margin. North and South America's eastern coastlines are examples of passive margins. **Active margins** are places where the oceanic and continental lithospheric tectonic plates meet and move relative to each other, such as the western coasts of North and South America. This movement is caused by frictional drag created between the plates and differences in plate densities. The majority of mountain-building events, earthquake activity and active volcanism on the Earth's surface can be attributed to tectonic plate movement at active margins.

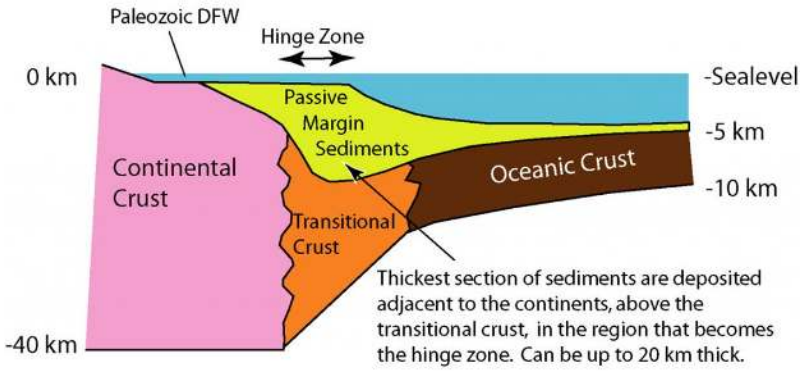


Figure 2.2.9: Passive Margin. (By Joshua Doubek; CC BY-SA 3.0 via Wikimedia Commons.)

In a simplified model, there are three categories of tectonic plate boundaries. Convergent boundaries are places where plates move toward each other. At divergent boundaries, the plates move apart. At transform boundaries, the plates slide past each other.

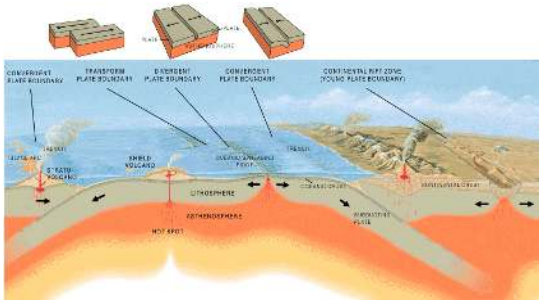


Figure 2.2.10: Schematic of plate boundary types. (By Jose F. Vigil, USGS; public domain via Wikimedia Commons.)

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2.3: Convergent Boundaries

Convergent boundaries, also called **destructive boundaries**, are places where two or more plates move toward each other. Convergent boundary movement is divided into two types, subduction and collision, depending on the density of the involved plates. Continental lithosphere is of lower density and thus more buoyant than the underlying asthenosphere. Oceanic lithosphere is denser than continental lithosphere, and, when old and cold, may even be denser than asthenosphere.

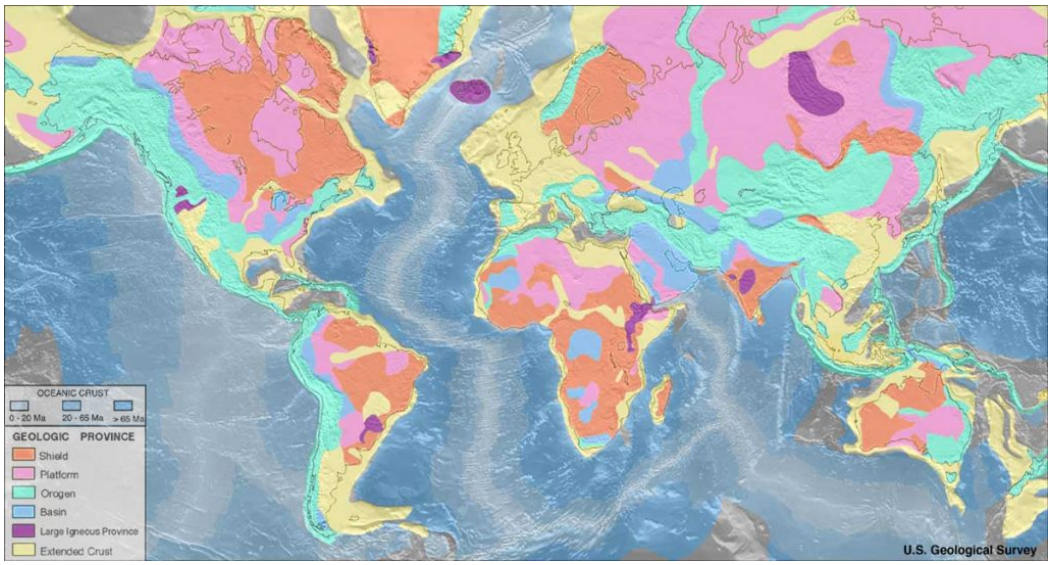


Figure 2.3.1: Geologic provinces of Earth. Shields (orange) and platform (pink) comprise cratons, the stable interior of continents. Orogenies are labeled light blue. (By USGS; public domain via Wikimedia Commons.)

When plates of different densities converge, the higher density plate is pushed beneath the more buoyant plate in a process called subduction. When continental plates converge without subduction occurring, this process is called collision.

Subduction

Subduction occurs when a dense oceanic plate meets a more buoyant plate, like a continental plate or warmer/younger oceanic plate, and descends into the mantle [45]. The worldwide average rate of oceanic plate subduction is 25 miles per million years, which is about a half-inch per year [46]. As an oceanic plate descends, it pulls the ocean floor down into a **trench**. These trenches can be more than twice as deep as the average depth of the adjacent ocean basin, which is usually 3 to 4 km. The Mariana Trench, for example, approaches a staggering 11 km [47].

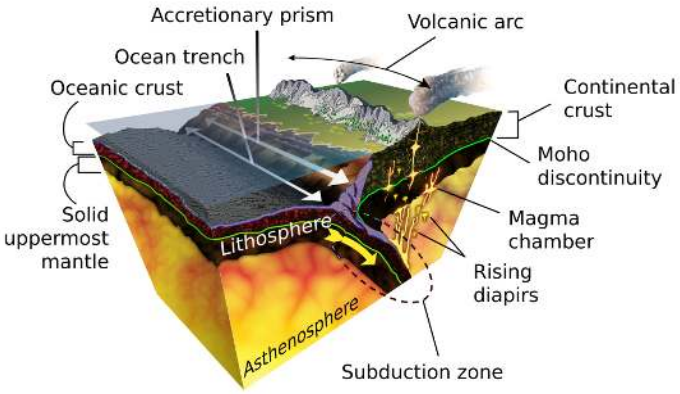


Figure 2.3.2: Diagram of ocean-continent subduction. (By KDS4444; CC BY-SA 4.0 via Wikimedia Commons.)

Within the trench, ocean floor sediments are scraped together and compressed between the subducting and overriding plates. This feature is called the **accretionary wedge**, **mélange**, or **accretionary prism**. Fragments of continental material, including

microcontinents, riding atop the subducting plate may become sutured to the accretionary wedge and accumulate into a large area of land called a **terrane** [48]. Vast portions of California are comprised of accreted terranes [49].

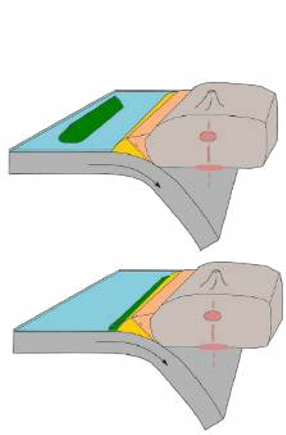


Figure 2.3.3: Microcontinents can become part of the accretionary prism of a subduction zone. (By Paul Inkenbrandt.)

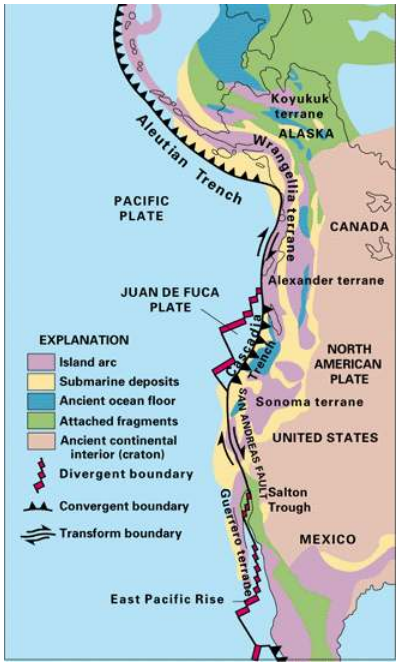


Figure 2.3.4: Accreted terranes of western North America. Everything that is not the “Ancient continental interior (craton)” has been smeared onto the side of the continent by accretion from subduction. (Modified from illustration provided by Oceanus Magazine; original figure by Jack Cook, Woods Hole Oceanographic Institution; adapted by USGS. Used under fair use.)

When the subducting oceanic plate, or **slab**, sinks into the mantle, the immense heat and pressure push volatile materials like water and carbon dioxide into an area below the continental plate and above the descending plate called the **mantle wedge**. The volatiles are released mostly by hydrated minerals that revert to non-hydrated minerals in these higher temperatures and pressure conditions. When mixed with asthenospheric material above the plate, the volatiles lower the melting point of the mantle wedge, and through a process called **flux melting** it becomes liquid magma. The molten magma is more buoyant than the lithospheric plate above it and migrates to the Earth’s surface where it emerges as volcanism. The resulting volcanoes frequently appear as curved mountain chains, volcanic arcs, due to the curvature of the Earth. Both oceanic and continental plates can contain volcanic arcs.

How subduction is initiated is still a matter of scientific debate [50]. It is generally accepted that subduction zones start as passive margins, where oceanic and continental plates come together, and then gravity initiates subduction and converts the passive margin into an active one [51]. One hypothesis is gravity pulls the denser oceanic plate down [52] or the plate can start to flow ductility at a low angle [53]. Scientists seeking to answer this question have collected evidence that suggests a new subduction zone is forming off the coast of Portugal [54]. Some scientists have proposed large earthquakes like the 1755 Lisbon earthquake may even have

something to do with this process of creating a subduction zone [55], although the evidence is not definitive. Another hypothesis proposes subduction happens at transform boundaries involving plates of different densities [56].

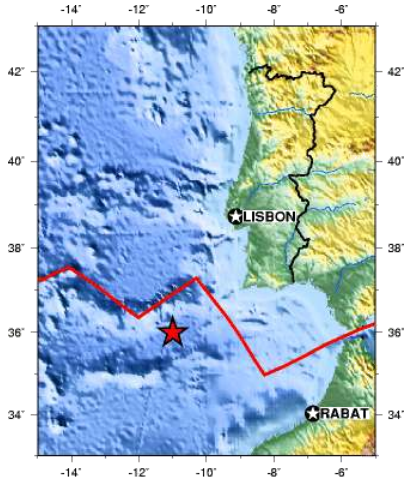


Figure 2.3.5: Location of the large (Mw 8.5-9.0) 1755 Lisbon Earthquake. (By USGS; public domain via Wikimedia Commons.)

Some plate boundaries look like they should be active, but show no evidence of subduction. The oceanic lithospheric plates on either side of the Atlantic Ocean for example, are denser than the underlying asthenosphere and are not subducting beneath the continental plates. One hypothesis is the bond holding the oceanic and continental plates together is stronger than the downward force created by the difference in plate densities.

Subduction zones are known for having the largest earthquakes and tsunamis; they are the only places with fault surfaces large enough to create magnitude-9 earthquakes. These subduction-zone earthquakes not only are very large but also are very deep. When a subducting slab becomes stuck and cannot descend, a massive amount of energy builds up between the stuck plates. If this energy is not gradually dispersed, it may force the plates to suddenly move and release energy along several hundred kilometers of the subduction zone [57]. Because subduction-zone faults are located on the ocean floor, this massive amount of movement can generate giant tsunamis such as those that followed the 2004 Indian Ocean Earthquake and 2011 Tōhoku Earthquake in Japan.

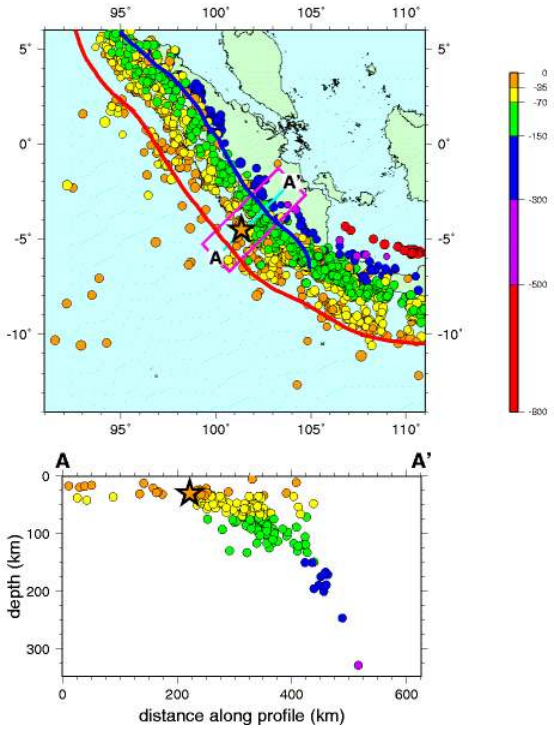


Figure 2.3.6: Earthquakes along the Sunda megathrust subduction zone, along the island of Sumatra, showing the 2006 Mw 9.1-9.3 Indian Ocean earthquake as a star. (By USGS; public domain via Wikimedia Commons.)

All subduction zones have a **forearc basin**, a feature of the overriding plate found between the **volcanic arc** and oceanic trench. The forearc basin experiences a lot of faulting and deformation activity, particularly within the accretionary wedge [58].

In some subduction zones, **tensional forces** working on the continental plate create a **back-arc basin** on the interior side of the volcanic arc. Some scientists have proposed a subduction mechanism called **oceanic slab rollback** creates **extension faults** in the overriding plates [59]. In this model, the descending oceanic slab does not slide directly under the overriding plate but instead rolls back, pulling the overlying plate seaward. The continental plate behind the volcanic arc gets stretched like pizza dough until the surface cracks and collapses to form a back-arc basin. If the extension activity is extensive and deep enough, a back-arc basin can develop into a **continental rifting zone**. These continental divergent boundaries may be less symmetrical than their mid-ocean ridge counterparts [60].

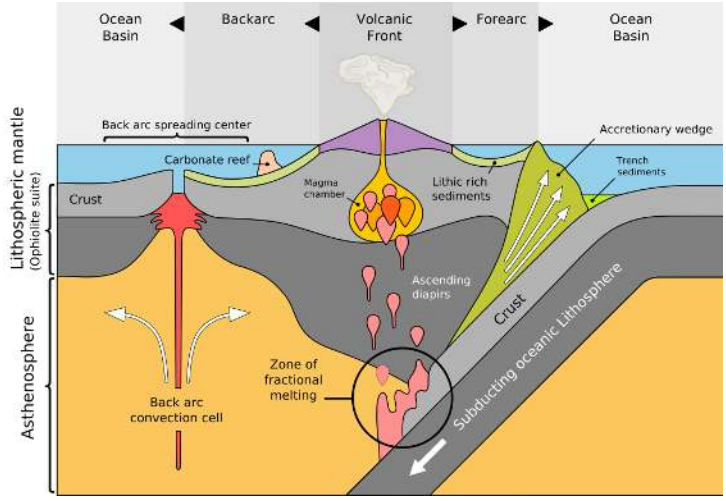


Figure 2.3.7: Various parts of a subduction zone. This subduction zone is ocean-ocean subduction, though the same features can apply to continent-ocean subduction. (By MagentaGreen; CC BY-SA 3.0 via Wikimedia Commons.)

In places where numerous young buoyant oceanic plates are converging and subducting at a relatively high velocity, they may force the overlying continental plate to buckle and crack [61]. This is called **back-arc faulting**. Extensional back-arc faults pull rocks

and chunks of plates apart. Compressional back-arc faults, also known as **thrust faults**, push them together.

The dual spines of the Andes Mountain range include a example of compressional thrust faulting. The western spine is part of a volcanic arc. Thrust faults have deformed the non-volcanic eastern spine, pushing rocks and pieces of a continental plate on top of each other.

There are two styles of thrust fault deformation: **thin-skinned faults** that occur in superficial rocks lying on top of the continental plate and **thick-skinned faults** that reach deeper into the crust. The Sevier Orogeny in the western U.S. is a notable thin-skinned type of deformation created during the Cretaceous Period. The Laramide Orogeny, a thick-skinned type of deformation, occurred near the end of and slightly after the Sevier Orogeny in the same region.

Flat-slab, or shallow, subduction caused the Laramide Orogeny. When the descending slab subducts at a low angle, there is more contact between the slab and the overlying continental plate than in a typical subduction zone. The shallowly-subducting slab pushes against the overriding plate and creates an area of deformation on the overriding plate many kilometers away from the subduction zone [62].

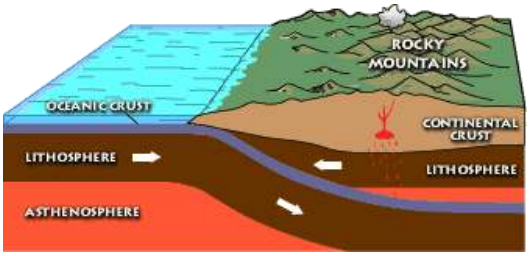


Figure 2.3.8: Shallow subduction during the Laramide Orogeny. (By USGS; public domain via Wikimedia Commons.)

Oceanic-Continental Subduction

Oceanic-continental subduction occurs when an oceanic plate dives below a continental plate. This convergent boundary has a trench and mantle wedge and frequently, a volcanic arc. Well-known examples of continental volcanic arcs are the Cascade Mountains in the Pacific Northwest [63] and the Western Andes Mountains in South America [64].

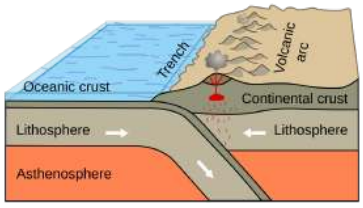


Figure 2.3.9: Subduction of an oceanic plate beneath a continental plate, forming a trench and volcanic arc. (By Booyabazooka from USGS; public domain via Wikimedia Commons.)

Oceanic-Oceanic Subduction

The boundaries of **oceanic-oceanic subduction** zones show very different activity from those involving oceanic-continental plates. Since both plates are made of oceanic lithosphere, it is usually the older plate that subducts because it is colder and denser. The volcanism on the overlying oceanic plate may remain hidden underwater.. If the volcanoes rise high enough the reach the ocean surface, the chain of volcanism forms an island arc. Examples of these island arcs include the Aleutian Islands in the northern Pacific Ocean, Lesser Antilles in the Caribbean Sea, and numerous island chains scattered throughout the western Pacific Ocean [65].

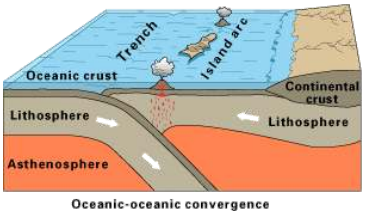


Figure 2.3.10: Subduction of an oceanic plate beneath another oceanic plate, forming a trench and an island arc. (By USGS; public domain via Wikimedia Commons.)

Collisions

When continental plates converge, during the closing of an ocean basin, for example, subduction is not possible between the equally buoyant plates. Instead of one plate descending beneath another, the two masses of continental lithosphere slam together in a process known as **collision** [66]. Without subduction, there is no magma formation and no volcanism. Collision zones are characterized by tall, non-volcanic mountains; a broad zone of frequent, large earthquakes; and very little volcanism.

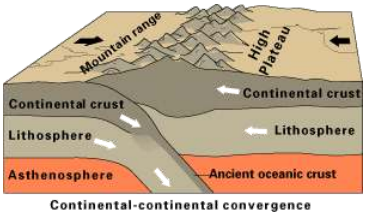


Figure 2.3.11: Two continental plates colliding. (By USGS; public domain via Wikimedia Commons.)

When oceanic crust connected by a passive margin to continental crust completely subducts beneath a continent, an ocean basin closes, and continental collision begins. Eventually, as ocean basins close, continents join together to form a massive accumulation of continents called a **supercontinent**, a process that has taken place in ~500 million-year-old cycles over Earth's history.

The process of collision created **Pangea**, the supercontinent envisioned by Wegener as the key component of his continental drift hypothesis. Geologists now have evidence that continental plates have been continuously converging into supercontinents and splitting into smaller basin-separated continents throughout Earth's existence, calling this process the **supercontinent cycle**, a process that takes place in approximately 500 million years. For example, they estimate Pangea began separating 200 million years ago. Pangea was preceded by earlier supercontinents, one of which being **Rodinia**, which existed 1.1 billion years ago and started breaking apart 800 million to 600 million years ago.

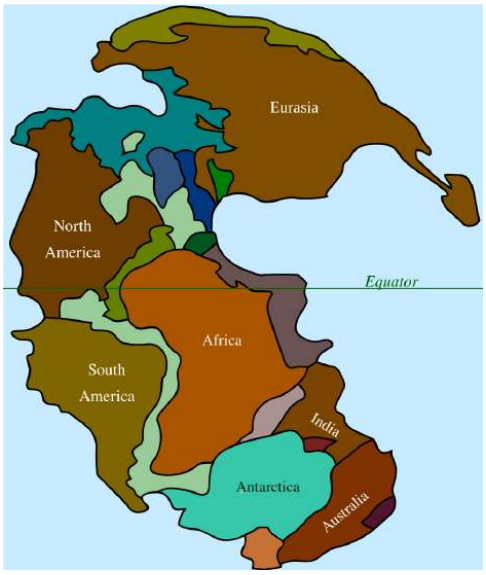


Figure 2.3.12: A reconstruction of the supercontinent Pangea, showing approximate positions of modern continents. (By Astroskiandhike; CC BY-SA 4.0 via Wikimedia Commons.)

A **foreland basin** is a feature that develops near mountain belts, as the combined mass of the mountains forms a depression in the lithospheric plate. While foreland basins may occur at subduction zones, they are most commonly found at collision boundaries. The Persian Gulf is possibly the best modern example, created entirely by the weight of the nearby Zagros Mountains.

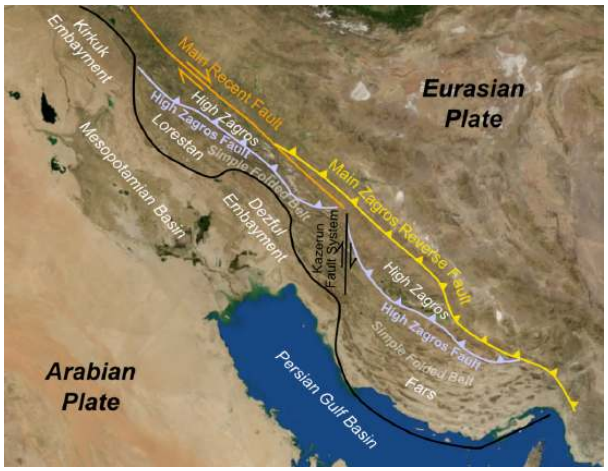


Figure 2.3.13: The tectonics of the Zagros Mountains. Note the Persian Gulf foreland basin. (By Mikenorton; CC BY-SA 3.0 via Wikimedia Commons.)

If continental and oceanic lithosphere are fused on the same plate, it can partially subduct but its buoyancy prevents it from fully descending. In very rare cases, part of a continental plate may become trapped beneath a descending oceanic plate in a process called **obduction** [67]. When a portion of the continental crust is driven down into the subduction zone, due to its buoyancy it returns to the surface relatively quickly.

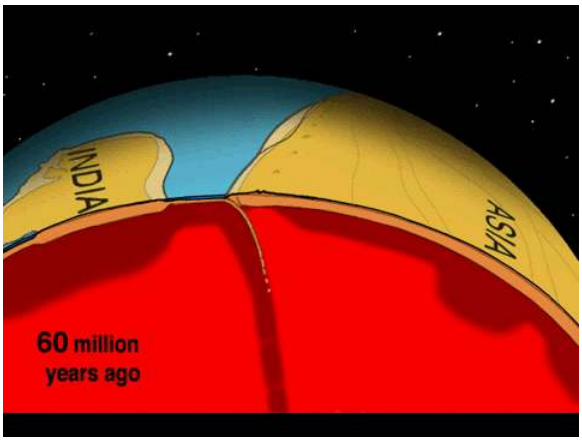
As pieces of the continental lithosphere break loose and migrate upward through the obduction zone, they bring along bits of the mantle and ocean floor and append them on top of the continental plate. Rocks composed of this mantle and ocean-floor material are called ophiolites and they provide valuable information about the composition of the mantle.



Figure 2.3.14: Pillow lavas, which only form underwater, from an ophiolite in the Apennine Mountains of central Italy. (By Matt Affolter; CC BY-SA 3.0 via Wikimedia Commons.)

The area of collision-zone deformation and seismic activity usually covers a broader area because the continental lithosphere is plastic and malleable. Unlike subduction-zone earthquakes, which tend to be located along a narrow swath near the convergent boundary, collision-zone earthquakes may occur hundreds of kilometers from the boundary between the plates.

The Eurasian continent has many examples of collision-zone deformations covering vast areas. The Pyrenees mountains begin in the Iberian Peninsula and cross into France. Also, there are the Alps stretching from Italy to central Europe; the Zagros mountains from Arabia to Iran; and Himalaya mountains from the Indian subcontinent to central Asia.



Animation of India colliding with Asia. (By Raynaldi rji from Tanya Atwater and Peter Molnar; CC BY-SA 4.0 via Wikimedia Commons.)

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2.4: Divergent Boundaries

At **divergent boundaries**, sometimes called **constructive boundaries**, lithospheric plates move away from each other. There are two types of divergent boundaries, categorized by where they occur: **continental rift zones** and **mid-ocean ridges**. Continental rift zones occur in weak spots in the continental lithospheric plate. A mid-ocean ridge usually originates in a continental plate as a rift zone that expands to the point of splitting the plate apart, with seawater filling in the gap. The separate pieces continue to drift apart and become individual continents. This process is known as **rift-to-drift**.

Continental Rifting

In places where the continental plates are very thick, they reflect so much heat back into the mantle it develops strong convection currents that push super-heated mantle material up against the overlying plate, softening it. Tensional forces created by this convective upwelling begin to pull the weakened plate apart. As it stretches, it becomes thinner and develops deep cracks called extension or **normal faults**. Eventually, plate sections located between large faults drop into deep depressions known as **rift valleys**, which often contain keystone-shaped blocks of down-dropped crust known as **grabens**. The shoulders of these grabens are called **horsts**. If only one side of a section drops, it is called a **half-graben**. Depending on the conditions, rifts can grow into very large lakes and even oceans.

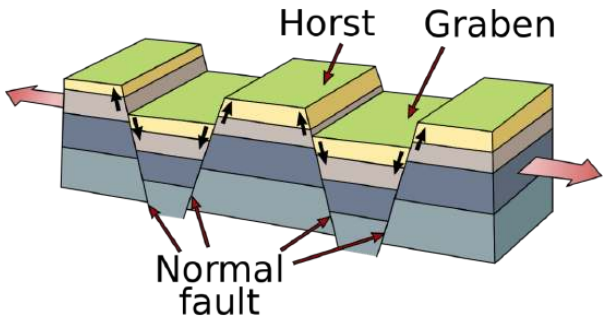


Figure 2.4.1: Normal faults that occur at divergent boundaries can result in horsts and grabens. (By Gregors from USGS; public domain via Wikimedia Commons.)

While seemingly occurring at random, rifting is dictated by two factors. Rifting does not occur in continents with older and more stable interiors, known as **cratons**. When continental rifting does occur, the break-up pattern resembles the seams of a soccer ball, also called a truncated icosahedron. This is the most common surface-fracture pattern to develop on an evenly expanding sphere because it uses the least amount of energy [68].

Using the soccer ball model, rifting tends to lengthen and expand along a particular seam while fizzling out in the other directions. These seams with little or no tectonic activity are called **failed rift arms**. A failed rift arm is still a weak spot in the continental plate; even without the presence of active extension faults, it may develop into a depression called an **aulacogen**. One example of a failed rift arm is the Mississippi Valley Embayment, a depression through which the upper end of the Mississippi River flows. Occasionally connected rift arms do develop concurrently, creating multiple boundaries of active rifting. In places where the rift arms do not fail, for example the Afar Triangle, three divergent boundaries can develop near each other forming a triple junction.



Figure 2.4.2: The Afar Triangle (center) has the Red Sea ridge (center to upper left), Gulf of Aden ridge (center to right), and East African Rift (center to lower left) form a triple junction that are about 120° apart. (By Koba-chan; CC BY-SA 3.0 via Wikimedia Commons.)

Rifts come in two types: narrow and broad. **Narrow rifts** are characterized by a high density of highly active divergent boundaries. The East African Rift Zone, where the horn of Africa is pulling away from the mainland, is an excellent example of an active narrow rift. Lake Baikal in Russia is another. **Broad rifts** also have numerous fault zones, but they are distributed over wide areas of deformation. The Basin and Range region located in the western United States is a type of broad rift. The Wasatch Fault, which also created the Wasatch Mountain Range in the state of Utah, forms the eastern divergent boundary of this broad rift (Animation 1 and Animation 2).



Figure 2.4.3: NASA image of the Basin and Range horsts and grabens across central Nevada. (By NASA; public domain via Wikimedia Commons.)

Rifts have earthquakes, although not of the magnitude and frequency of other boundaries. They may also exhibit volcanism. Unlike the flux-melted magma found in subduction zones, rift-zone magma is created by **decompression melting**. As the continental plates are pulled apart, they create a region of low pressure that melts the lithosphere and draws it upwards. When this molten magma reaches the weakened and fault-riddled rift zone, it migrates to the surface by breaking through the plate or escaping via an open fault. Examples of young rift volcanoes dot the Basin and Range region in the United States. Rift-zone activity is responsible for generating some unique volcanism, such as the Ol Doinyo Lengai in Tanzania. This volcano erupts lava consisting largely of carbonatite, a relatively cold, liquid carbonate mineral [69].

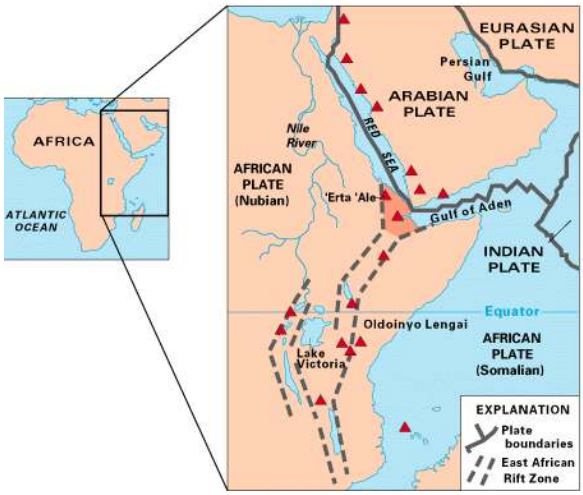


Figure 2.4.4: The narrow East African Rift is dotted with volcanoes, shown as red triangles. (By USGS; public domain via Wikimedia Commons.)

Mid-Ocean Ridges

As rifting and volcanic activity progress, the continental lithosphere becomes more mafic (see Chapter 4) and thinner, with the eventual result transforming the plate under the rifting area into the oceanic lithosphere. This is the process that gives birth to a new ocean, much like the narrow Red Sea emerged with the movement of Arabia away from Africa. As the oceanic lithosphere continues to diverge, a mid-ocean ridge is formed.

Mid-ocean ridges, also known as **spreading centers**, have several distinctive features. They are the only places on Earth that create new oceanic lithosphere. Decompression melting in the rift zone changes asthenosphere material into new lithosphere (magma), which oozes up through cracks in the oceanic plate. The amount of new lithosphere being created at mid-ocean ridges is highly significant. These undersea rift volcanoes produce more lava than all other types of volcanism combined. Despite this, most mid-oceanic ridge volcanism remains unmapped because the volcanoes are located deep on the ocean floor.

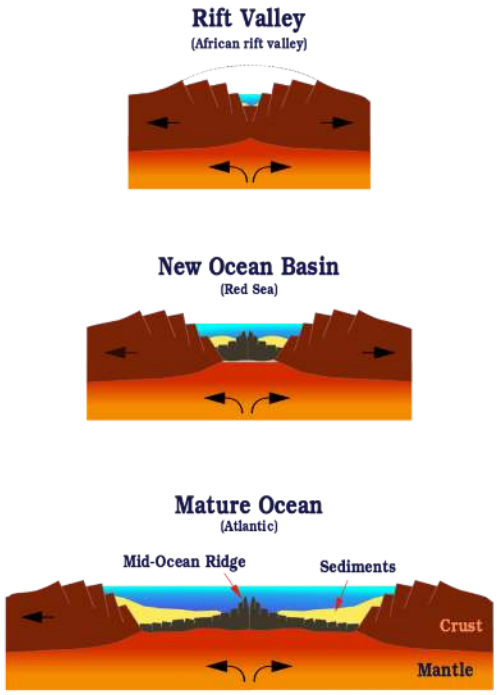


Figure 2.4.5: Progression from rift to mid-ocean ridge. (By Lichtspiel from Hannes Grobe, Alfred Wegener Institute for Polar and Marine Research; CC BY-SA 2.5 via Wikimedia Commons.)

In rare cases, such as a few locations in Iceland, rift zones display the type of volcanism, spreading, and ridge formation found on the ocean floor.

The ridge feature is created by the accumulation of hot lithosphere material, which is lighter than the dense underlying asthenosphere. This chunk of isostatically buoyant lithosphere sits partially submerged and partially exposed to the asthenosphere, like an ice cube floating in a glass of water.

As the ridge continues to spread, the lithosphere material is pulled away from the area of volcanism and becomes colder and denser. As it continues to spread and cool, the lithosphere settles into wide swathes of relatively featureless topography called abyssal plains with lower topography [70].

This model of ridge formation suggests the sections of the lithosphere furthest away from the mid-ocean ridges will be the oldest. Scientists have tested this idea by comparing the age of rocks located in various locations on the ocean floor. Rocks found near ridges are younger than those found far away from any ridges. Sediment accumulation patterns also confirm the idea of **seafloor spreading**. Sediment layers tend to be thinner near mid-ocean ridges, indicating it has had less time to build up.

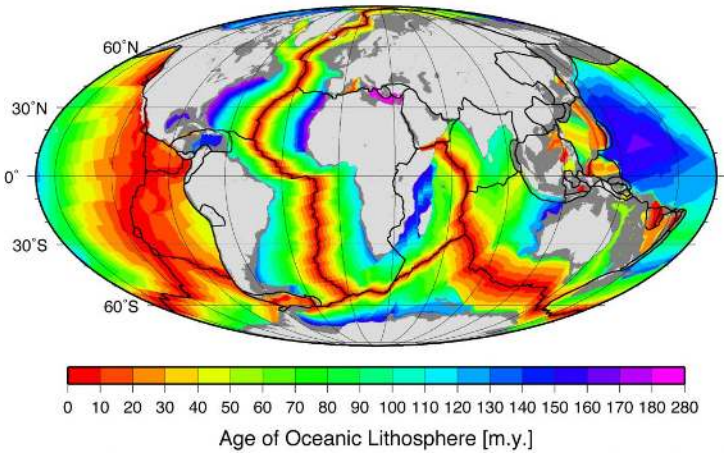


Figure 2.4.6: Age of oceanic lithosphere, in millions of years [m.y.]. Notice the age differences in the Atlantic Ocean along the coasts of the continents. (By Muller, R.D., M. Sdrolias, C. Gaina, W.R. Roest; CC BY-SA 3.0 via Wikimedia Commons.)

As mentioned in the section on paleomagnetism and the development of plate tectonic theory, scientists noticed mid-ocean ridges contained unique magnetic anomalies that show up as symmetrical striping on both sides of the ridge. The Vine-Matthews-Morley hypothesis [20] proposes these alternating reversals are created by the Earth's magnetic field being imprinted into magma after it emerges from the ridge [71]. Very hot magma has no magnetic field. As the oceanic plates get pulled apart, the magma cools below the **Curie point**, the temperature below which a magnetic field gets locked into magnetic minerals. The alternating magnetic reversals in the rocks reflect the periodic swapping of Earth's magnetic north and south poles. This paleomagnetic pattern provides a great historical record of ocean-floor movement, and is used to reconstruct past tectonic activity and determine rates of ridge spreading [72].

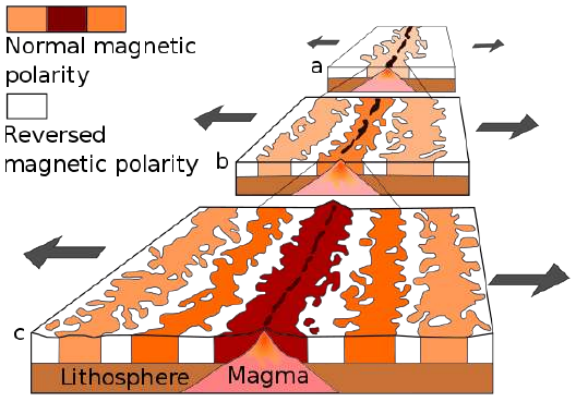


Figure 2.4.7: A time progression (with "a" being oldest and "c" being youngest) showing a spreading center getting wider while recording changes in the magnetic field of the Earth. (By Chmee2 from USGS; public domain via Wikimedia Commons.)

Thanks to their distinctive geology, mid-ocean ridges are home to some of the most unique ecosystems ever discovered. The ridges are often studded with **hydrothermal vents**, deep fissures that allow seawater to circulate through the upper portions of the oceanic plate and interact with hot rock. The super-heated seawater rises back up to the surface of the plate, carrying dissolved gases and minerals, and small particulates. The resulting emitted hydrothermal water looks like black underwater smoke.

Scientists had known about these geothermal areas on the ocean floor for some time. However, it was not until 1977, when scientists piloting a deep-submergence vehicle, the Alvin, discovered a thriving community of organisms clustered around these hydrothermal vents [73]. These unique organisms, which include 10-foot-long tube worms taller than people, live in the complete darkness of the ocean floor deprived of oxygen and sunlight. They use the geothermal energy provided by the vents and a process called bacterial chemosynthesis to feed on sulfur compounds. Before this discovery, scientists believed life on Earth could not exist without photosynthesis, a process that requires sunlight. Some scientists suggest this type of environment could have been the origin of life on Earth [74], and perhaps even extraterrestrial life elsewhere in the galaxy, such as on Jupiter's moon Europa [75].

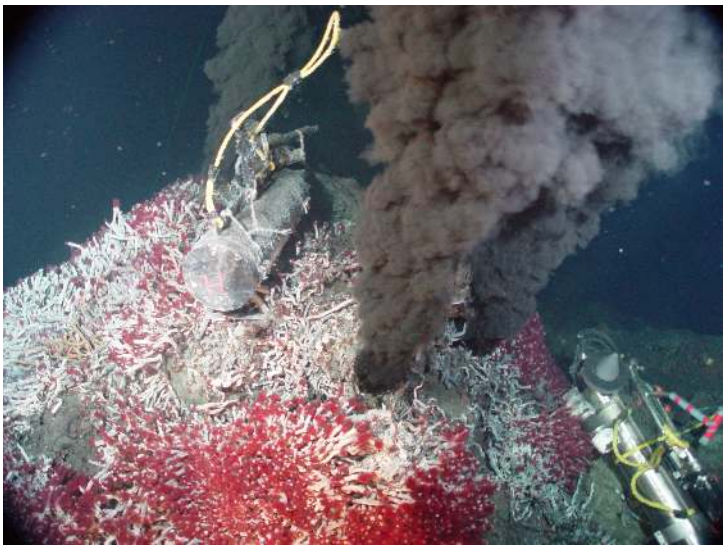


Figure 2.4.8: Black smoker hydrothermal vent with a colony of giant tube worms. (By NOAA; public domain.)

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2.5: Transform Boundaries

A **transform boundary**, sometimes called a **strike-slip** or **conservative boundary**, is where the lithospheric plates slide past each other in the horizontal plane. This movement is described based on the perspective of an observer standing on one of the plates, looking across the boundary at the opposing plate. **Dextral** movement, also known as right-handed or **right-lateral**, describes the opposing plate moving to the right. **Sinistral** movement, also known as left-handed or **left-lateral**, describes the opposing plate moving to the left.

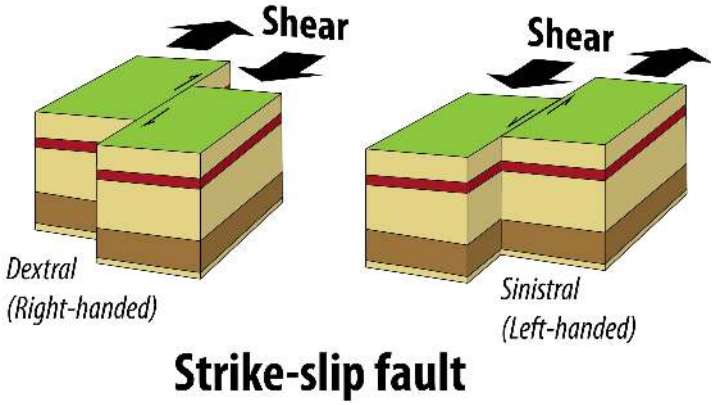


Figure 2.5.1: The two types of transform or strike-slip faults. (By Karla Panchuk; CC BY 4.0 via Physical Geology.)

Most transform boundaries are found on the ocean floor, around mid-ocean ridges. These boundaries form **aseismic fracture zones**, filled with earthquake-free transform faults, to accommodate different rates of spreading occurring at the ridge.

Some transform boundaries produce significant **seismic activity**, primarily as earthquakes, with very little mountain-building or volcanism. This type of transform boundary may contain a single fault or series of faults, which develop in places where plate tectonic stresses are transferred to the surface. As with other types of active boundaries, if the plates are unable to **shear** past each other, the tectonic forces will continue to build up. If the built-up energy between the plates is suddenly released, the result is an **earthquake**.

In the eyes of humanity, the most significant **transform faults** occur within continental plates and have a shearing motion that frequently produces moderate-to-large magnitude earthquakes. Notable examples include the San Andreas Fault in California, Northern and Eastern Anatolian Faults in Turkey, Altyn Tagh Fault in Central Asia, and Alpine Fault in New Zealand.

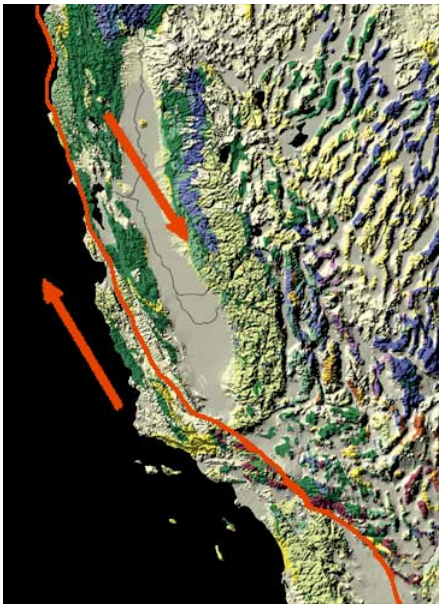


Figure 2.5.2: Map of the San Andreas fault, showing relative motion. (By Kate Barton, David Howell and Joe Vigil for USGS; public domain via Wikimedia Commons.)

Transpression and Transtension

Bends along transform faults may create compressional or extensional forces that cause secondary faulting zones. **Transpression** occurs where there is a component of compression in addition to the shearing motion. These forces build up around the area of the bend, where the opposing plates are restricted from sliding past each other. As the forces continue to build up, they create mountains in the **restraining bend** around the fault. The Big Bend area, located in the southern part of the San Andreas Fault includes a large area of transpression where many mountains have been built, moved, and even rotated [76].

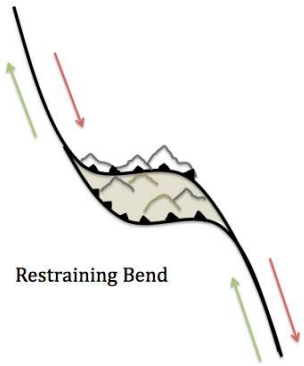


Figure 2.5.3: A transpressional strike-slip fault, causing uplift in the area called a restraining bend. (By GeoAsh; CC BY-SA 4.0 via Wikimedia Commons.)

Transtension zones require a fault that includes a **releasing bend**, where the plates are being pulled apart by extensional forces. Depressions and sometimes volcanism develop in the releasing bend along the fault. The Dead Sea found between Israel and Jordan, and the Salton Sea of California are examples of basins formed by transtensional forces.



Figure 2.5.4: A transtensional strike-slip fault, resulting in a releasing bend. In the center of the fault, the opposite sides are being pulled apart, which would create a depression. (By K. Martin; CC BY-SA 3.0 via Wikimedia Commons.)

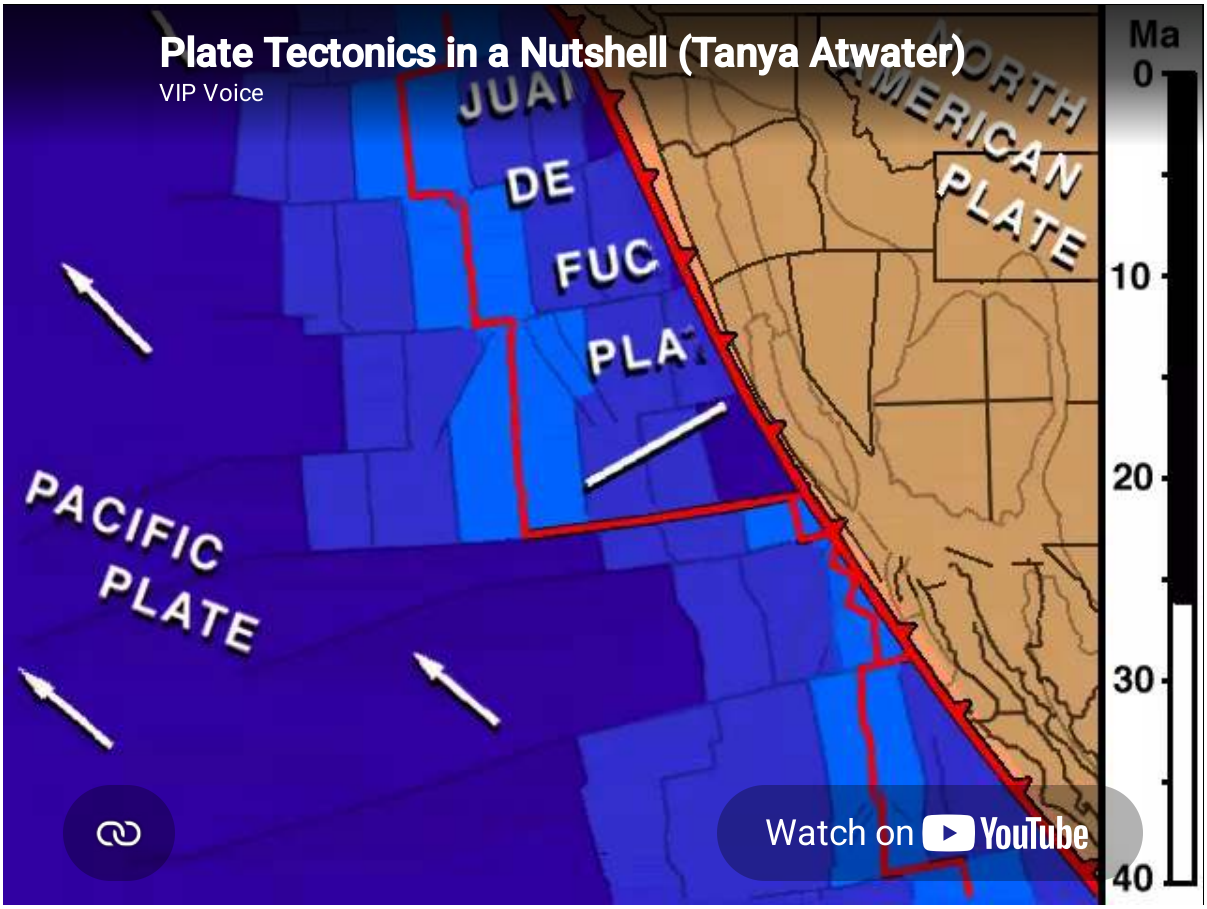
Piercing Points

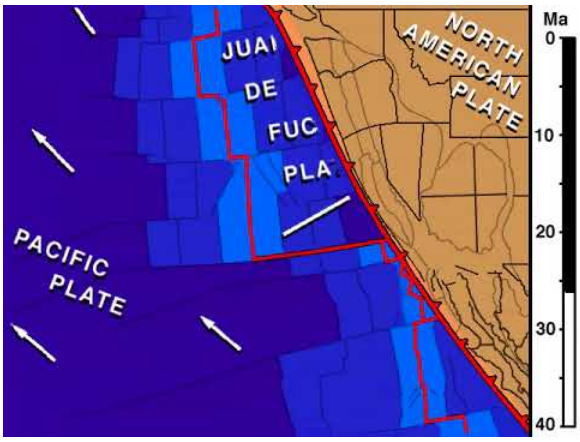
When a geological feature is cut by a fault, it is called a **piercing point**. Piercing points are very useful for recreating past fault movement, especially along transform boundaries. Transform faults are unique because their horizontal motion keeps a geological feature relatively intact, preserving the record of what happened. Other types of faults—normal and reverse —tend to be more destructive, obscuring or destroying these features. The best type of piercing point includes unique patterns that are used to match the parts of a geological feature separated by fault movement. Detailed studies of piercing points show the San Andreas Fault has experienced over 225 km of movement in the last 20 million years, and this movement occurred at three different fault traces.



Figure 2.5.5: Left: Wallace (dry) Creek on the Carrizo Plain, California. As the creek flows from the northern mountainous part of the image, it takes a sharp right (as viewed from the flow of water), then a sharp left. This is caused by the San Andreas Fault cutting roughly perpendicular to the creek, and shifting the location of the creek over time. The fault can be seen about halfway down, trending left to right, as a change in the topography. (By Robert E. Wallace for USGS; public domain via Wikimedia Commons.) Right: The same photo showing the offset in Wallace Creek by the San Andreas Fault. (By IRIS; CC BY 4.0 via EarthScope Consortium.)

Watch the video of the origin of the San Andreas fault. As the mid-ocean ridge subducts, the relative motion between the remaining plates become transform, forming the fault system. Note that because the motion of the plates is not exactly parallel to the fault, it causes divergent motion in the interior of North America.





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2.6: The Wilson Cycle

The **Wilson Cycle** is named for J. Tuzo Wilson who first described it in 1966 [24], and it outlines the ongoing origin and breakup of supercontinents, such as Pangea and Rodinia [77]. Scientists have determined this cycle has been operating for at least three billion years and possibly earlier.

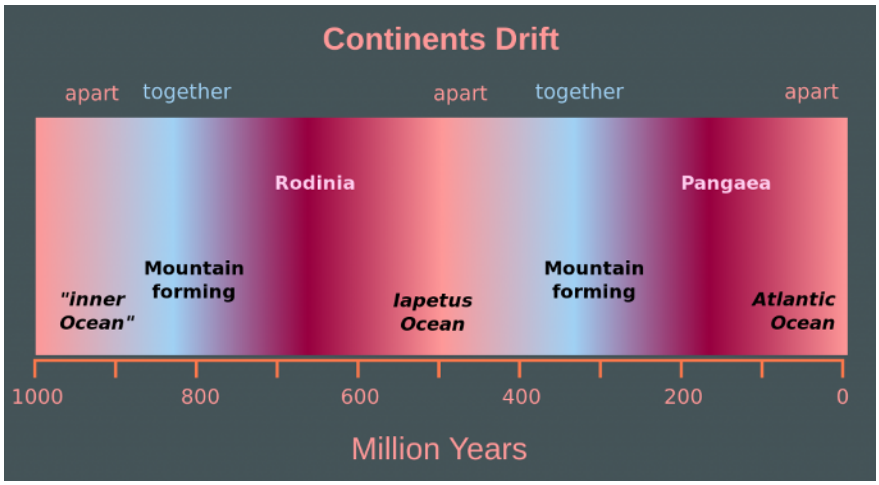


Figure 2.6.1: Diagram of the Wilson Cycle, showing rifting and collision phases. (By Hannes Grobe; CC BY-SA 2.5 via Wikimedia Commons.)

There are a number of hypotheses about how the Wilson Cycle works. One mechanism proposes that rifting happens because continental plates reflect the heat much better than oceanic plates [78]. When continents congregate together, they reflect more of the Earth's heat back into the mantle, generating more vigorous convection currents that then start the continental rifting process [79]. Some geologists believe **mantle plumes** are remnants of these periods of increased mantle temperature and convection upwelling and study them for clues about the origin of continental rifting.

The mechanism behind how supercontinents are created is still largely a mystery. There are three schools of thought about what continues to drive the continents further apart and eventually bring them together. The **ridge-push** hypothesis suggests after the initial rifting event, plates continue to be pushed apart by mid-ocean spreading centers and their underlying convection currents. **Slab-pull** proposes the plates are pulled apart by descending slabs in the subduction zones of the oceanic-continental margins [80]. A third idea, **gravitational sliding**, attributes the movement to gravitational forces pulling the lithospheric plates down from the elevated mid-ocean ridges and across the underlying asthenosphere [81]. Current evidence seems to support slab pull more than ridge push or gravitational sliding.

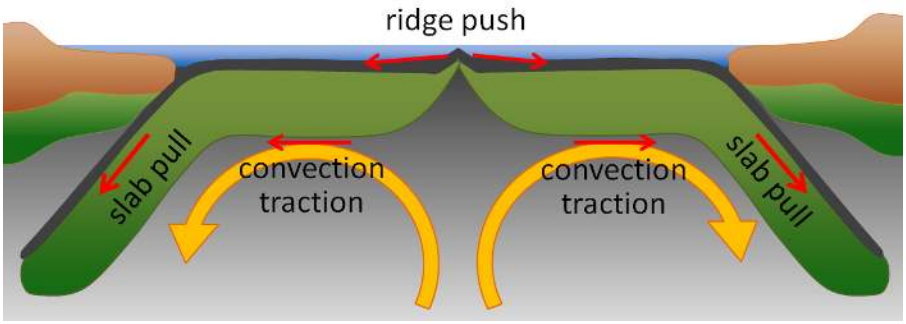


Figure 2.6.2: In this model, there are three forces working to move the plates. Ridge-push forces cause two plates to pull apart on the surface. Slab-pull forces pull the plates down. This movement of out-and-down is also encouraged by convection traction, or clockwise and counterclockwise currents that are present beneath the plates. (By Steven Earle; CC BY 4.0 via Physical Geology.)

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2.7: Hotspots

The Wilson Cycle provides a broad overview of tectonic plate movement. To analyze plate movement more precisely, scientists study **hotspots**. First postulated by J. Tuzo Wilson in 1963, a hotspot is an area in the lithospheric plate where molten magma breaks through and creates a volcanic center, such as volcanic islands in the ocean and volcanoes on land [82]. As the plate moves across the hotspot, the volcanic center becomes extinct because it is no longer over an active magma source. Instead, the magma emerges through another area in the plate to create a new active volcano. Over time, the combination of a moving plate and a stationary hotspot creates a chain of islands or mountains. The classic definition of hotspots states they do not move, although recent evidence suggests that there may be exceptions [83].

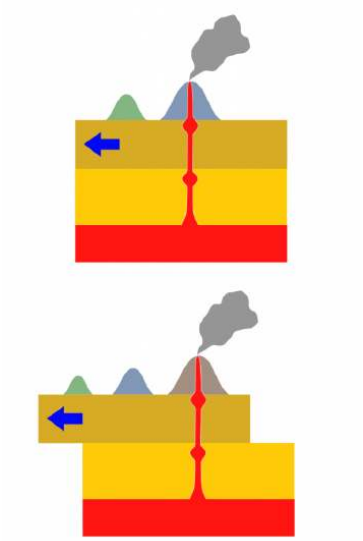


Figure 2.7.1: Diagram showing a non-moving source of magma (mantle plume) and a moving overriding plate. (By Los688; public domain via Wikimedia Commons.)

Hotspots are the only types of volcanism not associated with subduction or rifting zones at plate boundaries; they seem totally disconnected from any plate tectonics processes, such as tectonic earthquakes. However, there are relationships between hotspots and plate tectonics. There are several hotspots, current and former, that are believed to have begun at the time of rifting. Also, scientists use the age of volcanic eruptions and shape of the chain to quantify the rate and direction of plate movements relative to the hotspot.

Scientists are divided over how magma is generated in hotspots. Some suggest that hotspots originate from super-heated material from as deep as the core that reaches the Earth's crust as a mantle plume [84]. Others argue the molten material that feeds hotspots is sourced from the mantle [85]. Of course, it is difficult to collect data from these deep-Earth features due to the extremely high pressure and temperature [86].

How hotspots are initiated is another highly debated subject. The prevailing mechanism has hotspots starting in divergent boundaries during supercontinent rifting [87]. Scientists have identified a number of current and past hotspots believed to have begun this way. Subducting slabs have also been named as causing mantle plumes and hotspot volcanism [88]. Some geologists have suggested another geological process not involving plate tectonics may be involved, such as large space objects crashing into the Earth [89]. Regardless of how they are formed, dozens of hotspots are on the Earth. Some well-known hotspot locations include the Tahiti Islands, Afar Triangle, Easter Island, Iceland, Galapagos Islands, and the Samoan Islands. The United States is home to two of the largest and best-studied hotspots: Hawaii and Yellowstone.

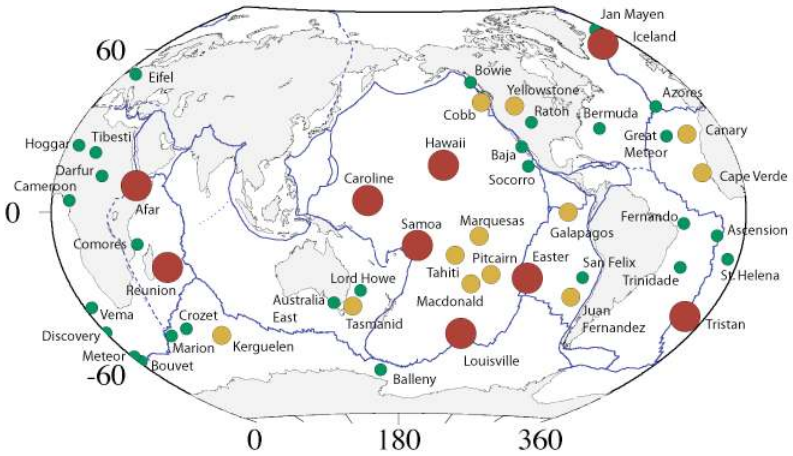


Figure 2.7.2: Map of world hotspots. Large red circles represent possible core-mantle-boundary origin hotspots. Medium yellow circles represent possible upper-mantle origin hotspots. Small green circles represent possible lithospheric origin hotspots. (By Foulger; public domain via Wikimedia Commons.)

Hawaiian Hotspot

The active volcanoes in Hawaii represent one of the most active hotspot sites on Earth. Scientific evidence indicates the **Hawaiian hotspot** is at least 80 million years old [90]. Geologists believe it is actually much older. However, any rocks with proof of this have been subducted under the ocean floor. The big island of Hawaii sits atop a large mantle plume that marks the active hotspot. Kilauea volcano is the main vent for this hotspot and has been actively erupting since 1983.

This enormous volcanic island chain, much of which is underwater, stretches across the Pacific for almost 6000 km. The seamount chain's most striking feature is a sharp 60° bend located at the midpoint, which marks a significant change in plate movement direction that occurred 50 million years ago. The change in direction has been more often linked to a plate reconfiguration [91], but also to other things like plume migration [83].

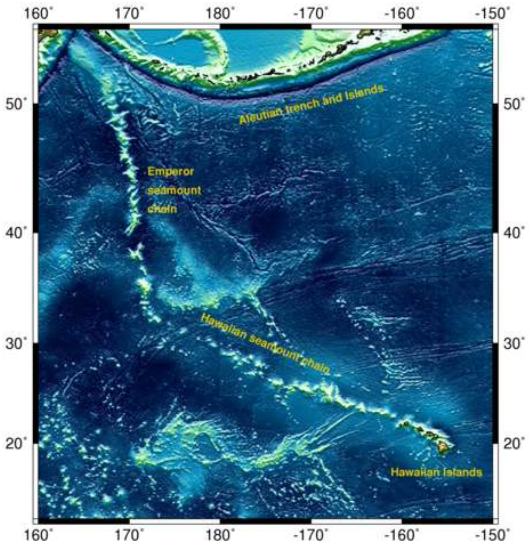


Figure 2.7.3: The Hawaii-Emperor seamount and island chain. (By Ingo Wolbern; public domain via Wikimedia Commons.)

In an attempt to map the Hawaiian mantle plume as far down as the lower mantle [92], scientists have used **tomography**, a type of three-dimensional seismic imaging. This information—along with other evidence gathered from rock ages, vegetation types, and island size—indicates the oldest islands in the chain are located the furthest away from the active hotspot.

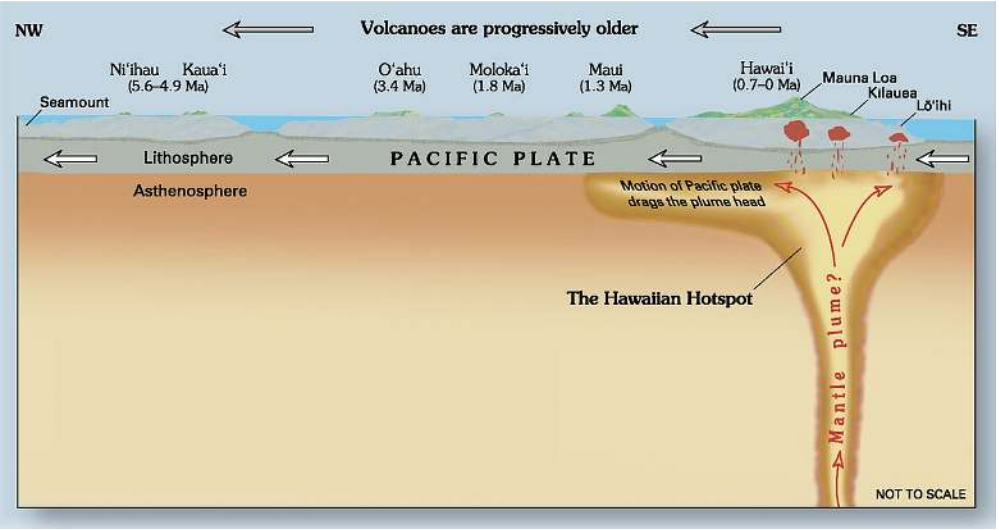


Figure 2.7.4:: Diagram of the Hawaiian hotspot and islands that it formed. (By Joel E Robinson for USGS; public domain via Wikimedia Commons.)

Yellowstone Hotspot

Like the Hawaiian version, the **Yellowstone hotspot** is formed by magma rising through the lithosphere. What makes it different is this hotspot is located under a thick, continental plate. Hawaii sits on a thin oceanic plate, which is easily breached by magma coming to the surface. At Yellowstone, the thick continental plate presents a much more difficult barrier for magma to penetrate. When it does emerge, the eruptions are generally much more violent. Thankfully they are also less frequent.

Over 15 million years of eruptions by this hotspot have carved a curved path across the western United States. It has been suggested the Yellowstone hotspot is connected to the much older Columbia River flood basalts [93] and even to 70 million-year-old volcanism found in the Yukon region of Canada [94].

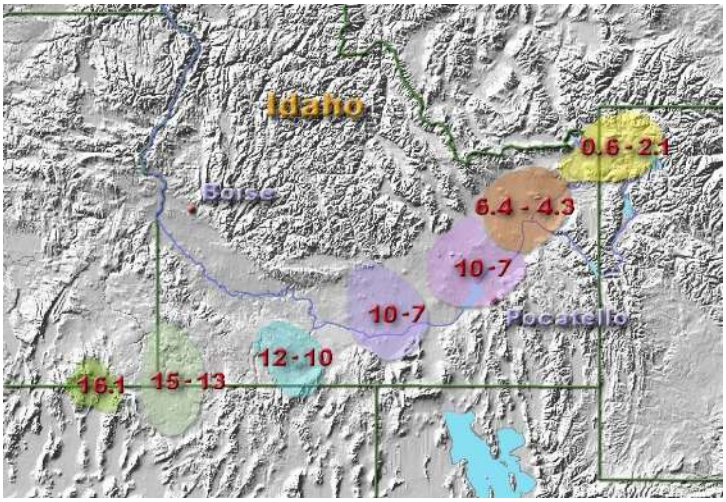


Figure 2.7.5:: The track of the Yellowstone hotspot which shows the age of different eruptions millions of years ago. (By Kelvin Case at English Wikipedia; CC BY 3.0 via Wikimedia Commons.)

The most recent major eruption of this hotspot created the Yellowstone Caldera and Lava Creek tuff formation approximately 631,000 years ago [95]. The eruption threw 1000 cubic kilometers of ash and magma into the atmosphere, some of which was found as far away as Mississippi. Should the hotspot erupt again, scientists predict it will be another massive event. This would be a calamity reaching far beyond the western United States. These super volcanic eruptions fill the Earth's atmosphere with so much gas and ash, they block sunlight from reaching the Earth. Not only would this drastically alter climates and environments around the globe, but it could also affect worldwide food production.

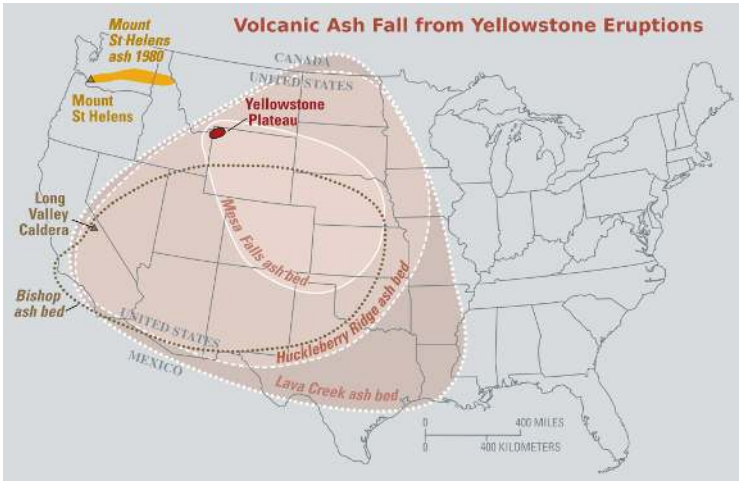


Figure 2.7.6:: Several prominent ash beds found in North America, including three Yellowstone eruptions which are shaded pink (Mesa Falls, Huckleberry Ridge, and Lava Creek), the Bishop Tuff ash bed (brown dashed line), and the modern May 18, 1980 Mt. St. Helens ash fall (yellow). (By USGS; public domain via Wikimedia Commons.)

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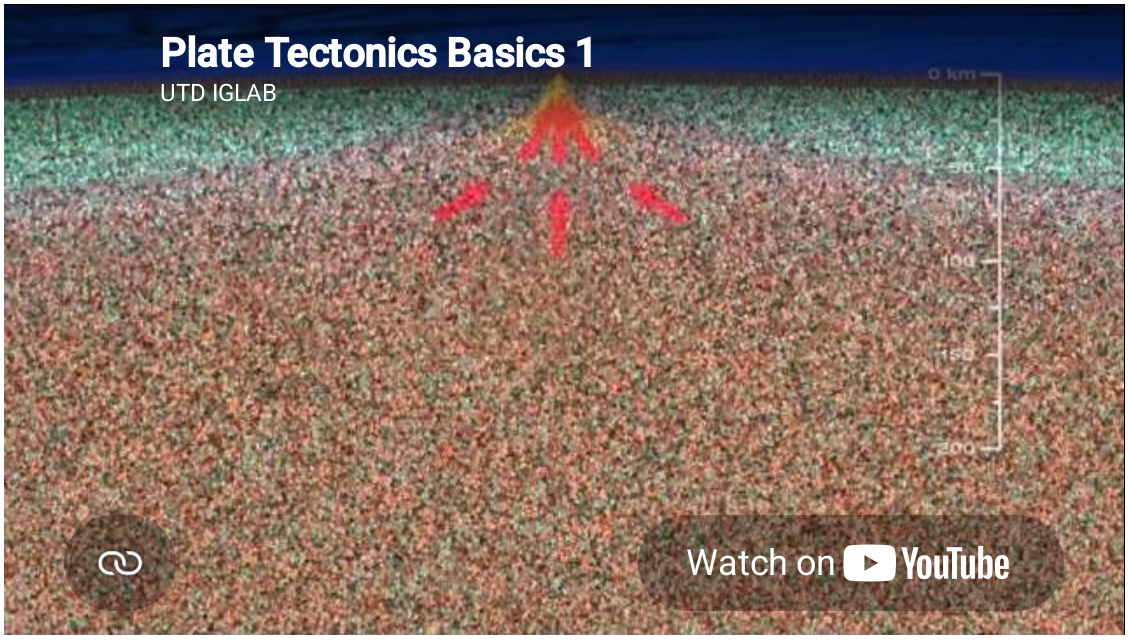
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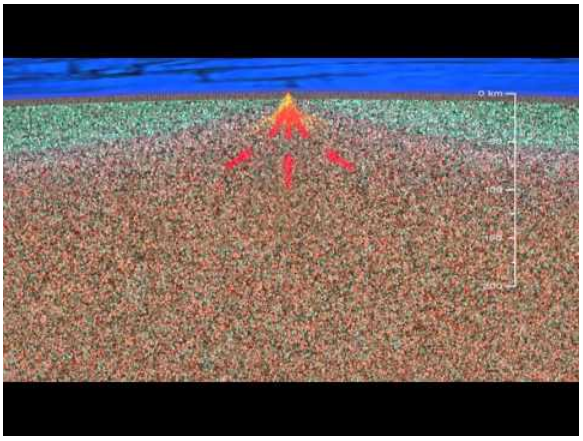
2.S: Summary

Summary

Plate tectonics is a unifying theory; it explains nearly all of the major geologic processes on Earth. Since its early inception in the 1950s and 1960s, geologists have been guided by this revolutionary perception of the world. The theory of plate tectonics states the surface layer of the Earth is broken into a network of solid, relatively brittle plates. Underneath the plates is a much hotter and more ductile layer that contains zones of convective upwelling generated by the interior heat of Earth. These convection currents move the surface plates around—bringing them together, pulling them apart, and shearing them side-by-side. Earthquakes and volcanoes form at the boundaries where the plates interact, with the exception of volcanic hotspots, which are not caused by plate movement.

Watch this video for the basics on plate tectonics.





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CHAPTER OVERVIEW

3: Minerals

Learning Objectives

- Define mineral.
- Describe the basic structure of the atom.
- Derive basic atomic information from the Periodic Table of Elements.
- Describe chemical bonding related to minerals.
- Describe the main ways minerals form.
- Describe the silicon-oxygen tetrahedron and how it forms common silicate minerals.
- List common non-silicate minerals in oxide, sulfide, sulfate, and carbonate groups.
- Identify minerals using physical properties and identification tables.



Figure 3.1: These selenite (gypsum) crystals, found in The Cave of the Crystals in Naica, Mexico, are some of the largest minerals ever found. The largest crystal found here is 39 feet (12 meters) and 55 tons. Note the person for scale. (By Alexander Van Driessche; CC BY 3.0 via Wikimedia Commons.)

The term “minerals” as used in nutrition labels and pharmaceutical products is not the same as a mineral in a geological sense. In geology, the classic definition of a **mineral** is:

1. Naturally occurring
2. Inorganic
3. Solid at room temperature
4. Regular crystal structure
5. Defined chemical composition

Some natural substances technically should not be considered minerals, but are included by exception. For example, water and mercury are liquid at room temperature. Both are considered minerals because they were classified before the room-temperature rule was accepted as part of the definition. Calcite is quite often formed by organic processes but is considered a mineral because it is widely found and geologically important. Because of these discrepancies, the International Mineralogical Association in 1985 amended the definition to: “A mineral is an element or chemical compound that is normally crystalline and that has been formed as a result of geological processes.” This means that the calcite in the shell of a clam is not considered a mineral. But once that clamshell undergoes burial, diagenesis, or other geological processes, then the calcite is considered a mineral. Typically, substances like coal, pearl, opal, or obsidian that do not fit the definition of a mineral are called **mineraloids**.

A **rock** is a substance that contains one or more minerals or mineraloids. As is discussed in later chapters, there are three types of rocks composed of minerals: igneous (rocks crystallizing from molten material), sedimentary (rocks made of products of

mechanical weathering (sand, gravel, etc.), or chemical weathering (things precipitated from solution)), and metamorphic (rocks produced by alteration of other rocks by heat and pressure).

3.1: Chemistry of Minerals
3.2: Formation of Minerals
3.3: Silicate Minerals
3.4: Non-Silicate Minerals
3.5: Identifying Minerals
3.S: Summary

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3.1: Chemistry of Minerals

Rocks are composed of minerals that have a specific chemical composition. To understand mineral chemistry, it is essential to examine the fundamental unit of all matter, the **atom**.

The Atom

Matter is made of atoms. Atoms consist of subatomic particles—protons, neutrons, and electrons. A simple model of the atom has a central **nucleus** composed of **protons**, which have positive charges, and **neutrons** which have no charge. A cloud of negatively charged **electrons** surrounds the nucleus, the number of electrons equaling the number of protons thus balancing the positive charge of the protons for a neutral atom. Protons and neutrons each have a **mass number** of 1. The mass of an electron is less than $1/1000^{th}$ that of a proton or neutron, meaning most of the atom's mass is in the nucleus.

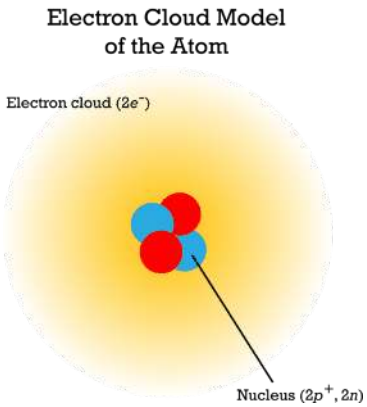


Figure 3.1.1: Electron cloud model of the atom. (By Christopher AuYeung; CC BY-NC 3.0 via CK-12 Foundation.)

Periodic Table of the Elements

Matter is composed of **elements** which are atoms that have a specific number of protons in the nucleus. This number of protons is called the **atomic number** for the element. For example, an oxygen atom has 8 protons and an iron atom has 26 protons. An element cannot be broken down chemically into a simpler form and retains unique chemical and physical properties. Each element behaves in a unique manner in nature. This uniqueness led scientists to develop a **periodic table of the elements**, a tabular arrangement of all known elements listed in order of their atomic number.

Periodic Table of Elements

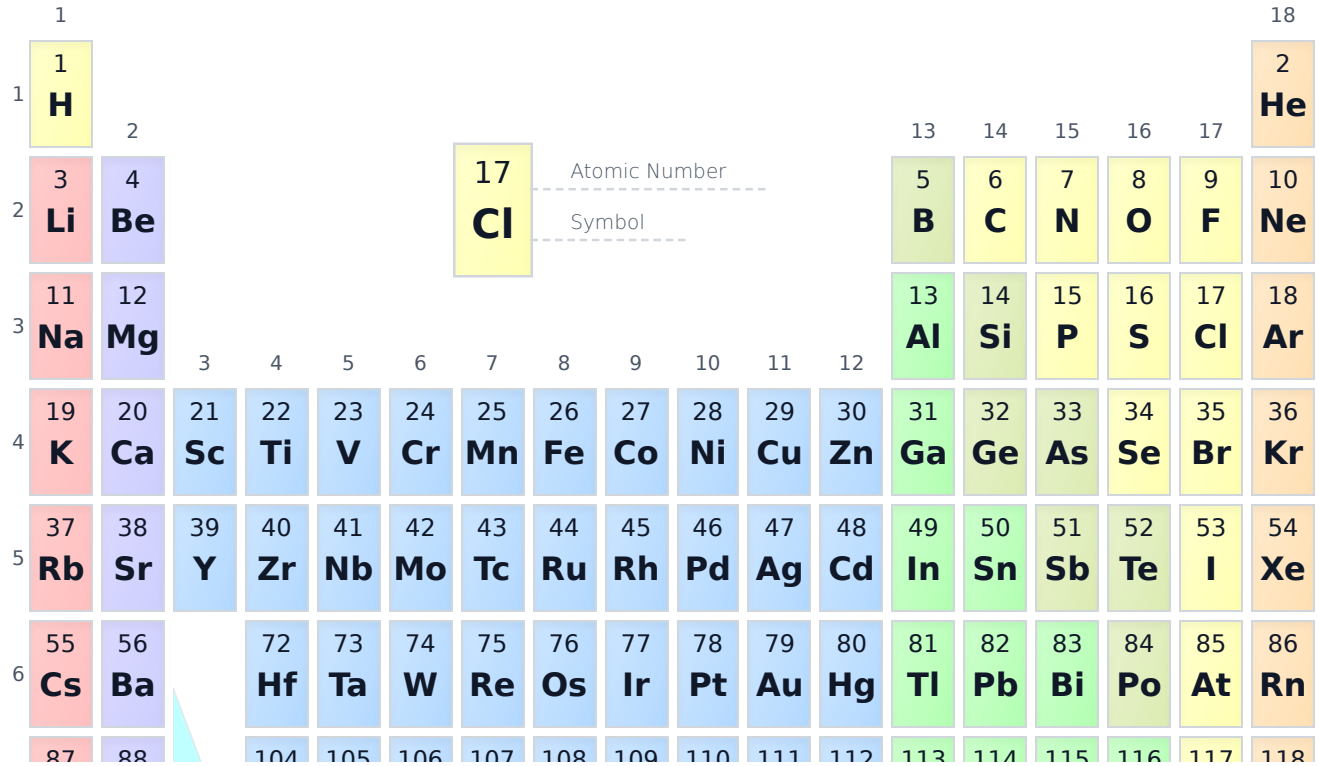


Figure 3.1.2: The Periodic Table of the Elements

The first arrangement of elements into a periodic table was done by Dmitri Mendeleev in 1869 using the elements known at the time [1]. In the periodic table, each element has a chemical symbol, name, atomic number, and atomic mass. The **chemical symbol** is an abbreviation for the element, often derived from a Latin or Greek name for the substance [2]. The atomic number is the number of protons in the nucleus. The **atomic mass** is the number of protons and neutrons in the nucleus, each with a mass number of one. Since the mass of electrons is so much less than the protons and neutrons, the atomic mass is effectively the number of protons plus neutrons.

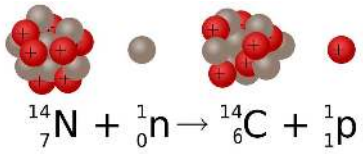


Figure 3.1.3: Formation of carbon-14 from nitrogen-14. (By NikNaks from Sgbeer; CC BY-SA 3.0 via Wikimedia Commons.)

The atomic mass of natural elements represents an average mass of the atoms comprising that substance in nature and is usually not a whole number as seen on the periodic table, meaning that an element exists in nature with atoms having different numbers of neutrons. The differing number of neutrons affects the mass of an element in nature and the atomic mass number represents this average. This gives rise to the concept of isotope. **Isotopes** are forms of an element with the same number of protons but different numbers of neutrons. There are usually several isotopes for a particular element. For example, 98.9% of carbon atoms have 6 protons and 6 neutrons. This isotope of carbon is called carbon-12 (^{12}C). A few carbon atoms, carbon-13 (^{13}C), have 6 protons and 7 neutrons. A trace amount of carbon atoms, carbon-14 (^{14}C), has 6 protons and 8 neutrons.

Among the 118 known elements, the heaviest are fleeting human creations known only in high energy particle accelerators, and they decay rapidly. The heaviest naturally occurring element is uranium, with atomic number 92. The eight most abundant elements in Earth's continental crust are shown in Table 3.1.1: [3; 4]. These elements are found in the most common rock-forming minerals.

Table 3.1.1:: Eight Most Abundant Elements in the Earth's Continental Crust % by weight (Source: USGS). All other elements are less than 1%..

Element	Symbol	Abundance %
Oxygen	O	47%
Silicon	Si	28%
Aluminum	Al	8%
Iron	Fe	5%
Calcium	Ca	4%
Sodium	Na	3%
Potassium	K	3%
Magnesium	Mg	2%

Chemical Bonding

Most substances on Earth are compounds containing multiple elements. **Chemical bonding** describes how these atoms attach with each other to form compounds, such as sodium and chlorine combining to form NaCl, common table salt. Compounds that are held together by chemical bonds are called **molecules**. Water is a compound of hydrogen and oxygen in which two hydrogen atoms are covalently bonded with one oxygen making the water molecule. The oxygen we breathe is formed when one oxygen atom covalently bonds with another oxygen atom to make the molecule O_2. The subscript 2 in the chemical formula indicates the molecule contains two atoms of oxygen.

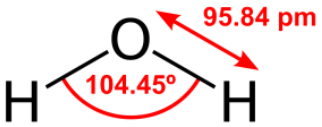


Figure 3.1.4: A model of a water molecule, showing the bonds between the hydrogen and oxygen. (By Dan Craggs; public domain via Wikimedia Commons.)

Most minerals are also compounds of more than one element. The common mineral calcite has the chemical formula $CaCO_3$ indicating the molecule consists of one calcium, one carbon, and three oxygen atoms. In calcite, one carbon and three oxygen atoms are held together by covalent bonds to form a molecular ion, called carbonate, which has a negative charge. Calcium as an ion has a positive charge of plus two. The two oppositely charged ions attract each other and combine to form the mineral calcite, $CaCO_3$. The name of the **chemical compound** is calcium carbonate, where calcium is Ca and carbonate refers to the molecular ion CO_3^{2-}.

The mineral olivine has the chemical formula $(Mg,Fe)_2SiO_4$, in which one silicon and four oxygen atoms are bonded with two atoms of either magnesium or iron. The comma between iron (Fe) and magnesium (Mg) indicates the two elements can occupy the same location in the crystal structure and substitute for one another.

Valence and Charge

The electrons around the atom's nucleus are located in **shells** representing different energy levels. The outermost shell is called the **valence shell**. Electrons in the valence shell are involved in chemical bonding. In 1913, Niels Bohr proposed a simple model of the atom that states atoms are more stable when their outermost shell is full [5; 6]. Atoms of most elements thus tend to gain or lose electrons so the outermost or valence shell is full. In Bohr's model, the innermost shell can have a maximum of two electrons and the second and third shells can have a maximum of eight electrons. When the innermost shell is the valence shell, as in the case of

hydrogen and helium, it obeys the **octet rule** when it is full with two electrons. For elements in higher rows, the octet rule of eight electrons in the valence shell applies.

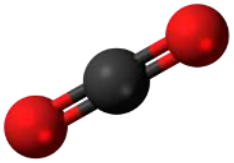


Figure 3.1.5: The carbon dioxide molecule. Since Oxygen is -2 and Carbon is +4, the two oxygens bond to the carbon to form a neutral molecule. (By Jynto; public domain via Wikimedia Commons.)

The rows in the periodic table present the elements in order of atomic number and the columns organize elements with similar characteristics, such as the same number of electrons in their valence shells. Columns are often labeled from left to right with Roman numerals I to VIII, and Arabic numerals 1 through 18. The elements in columns I and II have 1 and 2 electrons in their respective valence shells and the elements in columns VI and VII have 6 and 7 electrons in their respective valence shells.

In row 3 and column I, sodium (Na) has 11 protons in the nucleus and 11 electrons in three shells—2 electrons in the inner shell, 8 electrons in the second shell, and 1 electron in the valence shell. To maintain a full outer shell of 8 electrons per the octet rule, sodium readily gives up that 1 electron so there are 10 total electrons. With 11 positively charged protons in the nucleus and 10 negatively charged electrons in two shells, sodium when forming chemical bonds is an ion with an overall net charge of +1.

All elements in column I have a single electron in their valence shell and a valence of 1. The other column I elements also readily give up this single valence electron and thus become ions with a +1 charge. Elements in column II readily give up 2 electrons and end up as ions with a charge of +2. Note that elements in columns I and II which readily give up their valence electrons, often form bonds with elements in columns VI and VII which readily take up these electrons. Elements in columns 3 through 15 are usually involved in covalent bonding. The last column 18 (VIII) contains the **noble gases**. These elements are chemically inert because their valence shell is already full with 8 electrons, so they do not gain or lose electrons. An example is the noble gas helium which has 2 valence electrons in the first shell. Its valence shell is therefore full. All elements in column VIII possess full valence shells and do not form bonds with other elements.

As seen above, an atom with a net positive or negative charge as a result of gaining or losing electrons is called an **ion**. In general, the elements on the left side of the table lose electrons and become positive ions, called **cations** because they are attracted to the cathode in an electrical device. The elements on the right side tend to gain electrons. These are called **anions** because they are attracted to the anode in an electrical device. The elements in the center of the periodic table, columns 3 through 15, do not consistently follow the octet rule. These are called **transition elements**. A common example is iron, which has a +2 or +3 charge depending on the oxidation state of the element. Oxidized Fe^{3+} carries a +3 charge and reduced Fe^{2+} is +2. These two different oxidation states of iron often impart dramatic colors to rocks containing their minerals—the oxidized form producing red colors and the reduced form producing green.

Ionic Bonding

Ionic bonds, also called **electron-transfer bonds**, are formed by the electrostatic attraction between atoms having opposite charges. Atoms of two opposite charges attract each other electrostatically and form an ionic bond in which the positive ion transfers its electron (or electrons) to the negative ion which takes them up. Through this transfer both atoms thus achieve a full valence shell. For example one atom of sodium (Na^+) and one atom of chlorine (Cl^-) form an ionic bond to make the compound sodium chloride (NaCl). This is also known as the mineral halite or common table salt. Another example is calcium (Ca^{2+}) and chlorine (Cl^-) combining to make the compound calcium chloride ($CaCl_2$). The subscript 2 indicates two atoms of chlorine are ionically bonded to one atom of calcium.

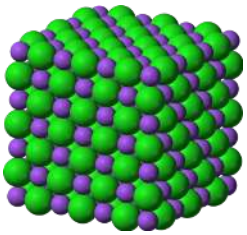


Figure 3.1.6: Cubic arrangement of Na and Cl ions in halite. (By Benjah-bmm27; public domain via Wikimedia Commons.)

Covalent Bonding

Ionic bonds are usually formed between a metal and a nonmetal. Another type, called a **covalent** or **electron-sharing bond**, commonly occurs between nonmetals. Covalent bonds share electrons between ions to complete their valence shells. For example, oxygen (atomic number 8) has 8 electrons—2 in the inner shell and 6 in the valence shell. Gases like oxygen often form diatomic molecules by sharing valence electrons. In the case of oxygen, two atoms attach to each other and share 2 electrons to fill their valence shells to become the common oxygen molecule we breathe (O_2). Methane (CH_4) is another covalently bonded gas. The carbon atom needs 4 electrons and each hydrogen needs 1. Each hydrogen shares its electron with the carbon to form a molecule as shown in the figure.

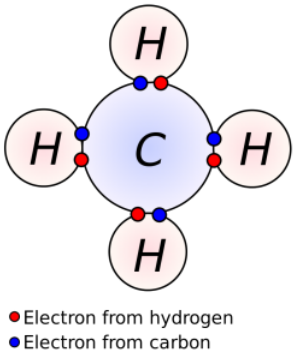


Figure 3.1.7: Methane molecule. (By DynaBlast; CC BY-SA 2.5 via Wikimedia Commons.)

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3.2: Formation of Minerals

Minerals form when atoms bond together in a crystalline arrangement. Three main ways this occurs in nature are:

1. Precipitation directly from an aqueous (water) solution with a temperature change
2. Crystallization from a magma with a temperature change
3. Biological precipitation by the action of organisms

Precipitation from Aqueous Solution

Solutions consist of ions or molecules, known as **solutes**, dissolved in a medium or **solvent**. In nature, this solvent is usually water. Many minerals can be dissolved in water, such as halite or table salt, which has the composition sodium chloride, NaCl. The Na^+ and Cl^- ions separate and disperse into the solution.

Precipitation is the reverse process, in which ions in solution come together to form solid minerals. Precipitation is dependent on the concentration of ions in solution and other factors such as temperature and pressure. The point at which a solvent cannot hold any more solute is called **saturation**. Precipitation can occur when the temperature of the solution falls, when the solvent evaporates, or with changing chemical conditions in the solution. An example of precipitation in our homes is when water evaporates and leaves behind a rind of minerals on faucets, shower heads, and drinking glasses.



Figure 3.2.1: Calcium carbonate deposits from hard water on a faucet. (By Bbypnda; CC BY-SA 3.0 via Wikimedia Commons.)

In nature, changes in environmental conditions may cause the minerals dissolved in water to form bonds and grow into crystals or may cement grains of sediment together. In Utah, deposits of tufa formed from mineral-rich springs that emerged into the ice age Lake Bonneville. Now exposed in dry valleys, this porous tufa was used by pioneers when building their homes as natural insulation to protect against summer heat and winter cold. The travertine terraces at Mammoth Hot Springs in Yellowstone National Park were similarly formed by calcite precipitation at the edges of the shallow spring-fed ponds.



Figure 3.2.2: The Bonneville Salt Flats of Utah. (By BLM; CC BY 2.0 via Wikimedia Commons.)

Another example of precipitation occurs in the Great Salt Lake, Utah, where the concentration of sodium chloride and other salts is nearly eight times greater than in the world's oceans [7]. Streams carry salt ions into the lake from the surrounding mountains. With no other outlet, the water in the lake evaporates and the concentration of salt increases until saturation is reached and the minerals precipitate out as sediments. Similar salt deposits include halite and other precipitates, and occur in other lakes like Mono Lake in California and the Dead Sea.

Crystallization from Magma

Heat is energy that causes atoms in substances to vibrate. **Temperature** is a measure of the intensity of the vibration. If the vibrations are violent enough, chemical bonds are broken and the crystals melt, releasing the ions into the melt. **Magma** is molten rock with freely moving ions. When magma is emplaced at depth or extruded onto the surface (then called **lava**), it starts to cool and mineral crystals can form.



Figure 3.2.3: Lava, which is magma at the Earth's surface. (By USGS; public domain via Wikimedia Commons.)

Precipitation by Organisms

Many organisms build bones, shells, and body coverings by extracting ions from water and precipitating minerals biologically. The most common mineral precipitated by organisms is **calcite**, or calcium carbonate ($CaCO_3$). Calcite is often precipitated by organisms as a polymorph called **aragonite**. **Polymorphs** are crystals with the same chemical formula but different crystal structures. Marine invertebrates such as corals and clams precipitate aragonite or calcite for their shells and structures. Upon death, their hard parts accumulate on the ocean floor as sediments and eventually may become the sedimentary rock **limestone**. Though limestone can form inorganically, the vast majority is formed by this biological process.



Figure 3.2.4: Fossilized ammonite shell made of calcium carbonate. (By ; CC BY-SA 3.0, via Wikimedia Commons.)

Another example is marine organisms called **radiolaria**, which are zooplankton that precipitate silica for their microscopic external shells. When the organisms die, the shells accumulate on the ocean floor and can form the sedimentary rock **chert**. An example of biologic precipitation from the vertebrate world is bone, which is composed mostly of a type of **apatite**, a mineral in the phosphate group. The apatite found in bones contains calcium and water in its structure and is called hydroxycarbonate apatite, $Ca_5(PO_4, CO_3)_3(OH)$. As mentioned above, such substances are not technically minerals until the organism dies and these hard parts become fossils.

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3.3: Silicate Minerals

Minerals are categorized based on their composition and structure. **Silicate minerals** are built around a molecular ion called the **silicon-oxygen tetrahedron**. A tetrahedron has a pyramid-like shape with four sides and four corners. Silicate minerals form the largest group of minerals on Earth, comprising the vast majority of the Earth's mantle and crust. Of the nearly 4,000 known minerals on Earth, most are rare. There are only a few that make up most of the rocks likely to be encountered by surface-dwelling creatures like us. These are generally called the **rock-forming minerals**.

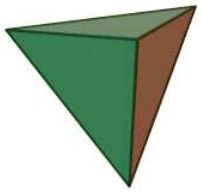


Figure 3.3.1: Rotating animation of a tetrahedron. (By Cyp; CC BY-SA 3.0 via Wikimedia Commons.)

The silicon-oxygen tetrahedron (SiO_4) consists of a single silicon atom and four oxygen atoms located at the four corners of the tetrahedron. The silicon ion is much smaller than the oxygen ions and fits in the space in the center of the oxygen ions. Each oxygen ion has a -2 charge and the silicon ion has a +4 charge. The silicon ion shares one of its four valence electrons with each of the four oxygen ions in a covalent bond to create a symmetrical four-sided pyramid figure. Only half of the oxygen's valence electrons are shared, giving the silicon-oxygen tetrahedron an ionic charge of -4. Because only one of the valence electrons of the corner oxygens is shared, the silicon-oxygen tetrahedron has chemically active corners available to form bonds with other silicate tetrahedra or other positively charged ions such as Al^{3+}, $Fe^{2+,3+}$, Mg^{2+}, K^{+}, Na^{+}, and Ca^{2+}. Depending on many factors, such as the original magma chemistry, silicon-oxygen tetrahedra can combine with other tetrahedra in several different configurations. For example, tetrahedra can be isolated, attached in chains, sheets, or three-dimensional structures. These combinations and others create the unique chemical compositions forming the large group of silicate minerals.

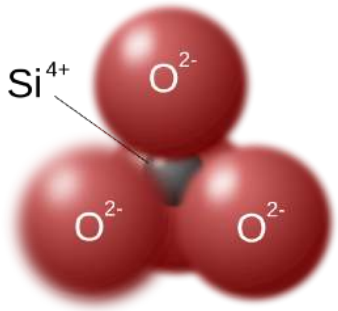


Figure 3.3.2: Silicon-oxygen tetrahedron. (By Helgi; CC BY-SA 3.0 via Wikimedia Commons.)

The Dark Ferromagnesian Silicates

Olivine Family

Olivine is the primary mineral component in mantle rock such as peridotite and basalt. It is characteristically green when not weathered. The chemical formula is $(Fe,Mg)_2SiO_4$. As previously described, the comma between iron (Fe) and magnesium (Mg) indicates these two elements occur in a **solid solution**. Not to be confused with a liquid solution, a solid solution occurs when two or more elements have similar properties and can freely substitute for each other in the same location in the crystal structure. Chemically, olivine is mostly silicate, iron, and magnesium and therefore is grouped among the **dark-colored ferromagnesian** (iron=ferro, magnesium=magnesian) or **mafic minerals**, a contraction of their chemical symbols Ma and Fe. Ferromagnesian silicates tend to be more dense than non-ferromagnesian silicates. This difference in density ends up being important in controlling the behavior of the igneous rocks that are built from these minerals: whether a tectonic plate subducts or not is largely governed by the density of its rocks, which are in turn controlled by the density of the minerals that comprise them.



Figure 3.3.3: Green olivine in basalt. (By Vsmith; CC BY-SA 3.0, via Wikimedia Commons.)

Olivine is referred to as a **mineral family** because of the ability of iron and magnesium to substitute for each other. The iron and magnesium ions in the solid solution are about the same size and charge, so either atom can fit into the same location in the growing crystals. Within the cooling magma, the mineral crystals continue to grow until they solidify into igneous rock. The relative amounts of iron and magnesium in the parent magma determine which minerals in the series form. Olivine can form crystals of all iron as one end member (called **fayalite**), all magnesium as the other end member (called **forsterite**), and mixtures of iron and magnesium in between. Different mineral names are applied to compositions between the end members. Other rarer elements with similar properties to iron or magnesium, like manganese (Mn), can substitute into the olivine crystalline structure in small amounts. Such ionic substitutions in mineral crystals give rise to the great variety of minerals and are often responsible for differences in color and other properties within a group or family of minerals.

The crystal structure of olivine is built from independent silicate tetrahedra. Minerals with independent tetrahedral structures are called **neosilicates** (or **orthosilicates**). In addition to olivine, other common neosilicate minerals include garnet, topaz, kyanite, and zircon.

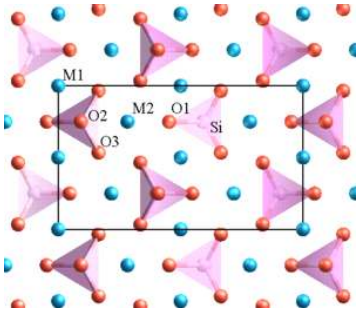


Figure 3.3.4: Tetrahedral structure of olivine. (By Matanya; public domain via Wikimedia Commons.)

Two other similar arrangements of tetrahedra are close in structure to the neosilicates and grade toward the next group of minerals, the pyroxenes. In a variation on independent tetrahedra called **sorosilicates**, there are minerals that share one oxygen between two tetrahedra and include minerals like pistachio-green epidote, a gemstone. Another variation are the **cyclosilicates**, which as the name suggests, consist of tetrahedral rings, and include gemstones such as beryl, emerald, aquamarine, and tourmaline.

Pyroxene Family

Pyroxene is another family of dark ferromagnesian minerals, typically black or dark green in color. Members of the pyroxene family have a complex chemical composition that includes iron, magnesium, aluminum, and other elements bonded to polymerized silicate tetrahedra. **Polymers** are chains, sheets, or three-dimensional structures, and are formed by multiple tetrahedra covalently bonded via their corner oxygen atoms. Pyroxenes are commonly found in mafic igneous rocks such as peridotite, basalt, and gabbro, as well as metamorphic rocks like eclogite and blue-schist.



Figure 3.3.5: Crystals of diopside, a member of the pyroxene family. (By Robert M. Lavinsky, iRocks.com; CC BY-SA 3.0, via Wikimedia Commons.)

Pyroxenes are built from long, single chains of polymerized silicate tetrahedra in which tetrahedra share two corner oxygens. The silicate chains are bonded together into the crystal structures by metal cations. A common member of the pyroxene family is **augite**, which forms in several solid solution series with a complex chemical formula $(Ca,Na)(Mg,Fe,Al,Ti)(Al,Si)_2O_6$ that gives rise to a number of individual mineral names.

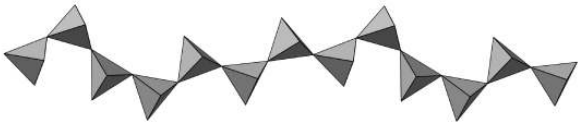


Figure 3.3.6: Single chain tetrahedral structure in pyroxene. (By Bubenik; CC BY-SA 3.0 via Wikimedia Commons.)

This single-chain crystalline structure bonds with many elements, which can also freely substitute for each other. The generalized chemical composition for pyroxene is $XZ(Al,Si)_2O_6$. X represents the ions Na, Ca, Mg, or Fe, and Z represents Mg, Fe, or Al. These ions have similar ionic sizes, which allows many possible substitutions among them. Although the cations may freely substitute for each other in the crystal, they carry different ionic charges that must be balanced out in the final crystalline structure. For example, Na has a charge of +1, but Ca has a charge of +2. If a Na^+ ion substitutes for a Ca^{2+} ion, it creates an unequal charge that must be balanced by other ionic substitutions elsewhere in the crystal. Note that ionic size is more important than ionic charge for substitutions to occur in solid solution series in crystals.

Amphibole Family

Amphibole minerals are built from polymerized double silicate chains and they are also referred to as **inosilicates**. Imagine two pyroxene chains that connect together by sharing the third oxygen on each tetrahedron. Amphiboles are usually found in igneous and metamorphic rocks and typically have a long-bladed crystal habit. The most common amphibole, **hornblende**, is usually black; however, they come in a variety of colors depending on their chemical composition. The metamorphic rock, amphibolite, is primarily composed of amphibole minerals.

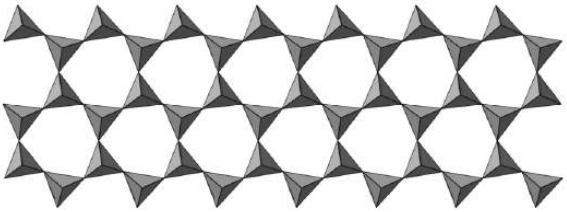


Figure 3.3.7: Double chain structure of inosilicates. (By Bubenik; CC BY-SA 3.0 via Wikimedia Commons.)

Amphiboles are composed of iron, magnesium, aluminum, and other cations bonded with silicate tetrahedra. These dark ferromagnesian minerals are commonly found in gabbro, basalt, diorite, and often form the black specks in granite. Their chemical formula is very complex and generally written as $(RSi_4O_{11})_2$, where R represents many different cations. For example, it can also be written more exactly as $AX_2Z_5((Si,Al,Ti)_8O_{22})(OH,F,Cl,O)_2$. In this formula A may be Ca, Na, K, Pb, or blank; X can be Li, Na, Mg, Fe, Mn, or Ca; and Z can be Li, Na, Mg, Fe, Mn, Zn, Co, Ni, Al, Cr, Mn, V, Ti, or Zr. The substitutions create a wide variety of colors such as green, black, colorless, white, yellow, blue, or brown. Amphibole crystals can also include **hydroxide ions** (OH^-), which occur from an interaction between the growing minerals and water dissolved in the magma.



Figure 3.3.8: Left: Hornblende crystals. (By USGS; public domain via Wikimedia Commons.) Right: Elongated crystals of hornblende in orthoclase. (By Dave Dyet; public domain via Wikimedia Commons.)

Sheet Silicates

Sheet silicates are built from tetrahedra which share all three of their bottom corner oxygens thus forming sheets of tetrahedra with their top corners available for bonding with other atoms. **Micas** and **clays** are common types of sheet silicates, also known as **phyllosilicates**. Mica minerals are usually found in igneous and metamorphic rocks, while clay minerals are more often found in sedimentary rocks. Two frequently found micas are dark-colored **biotite**, frequently found in granite, and light-colored **muscovite**, found in the metamorphic rock called schist.



Figure 3.3.9: Left: Crystal of biotite mica. (By Fred Kruijen; CC BY-SA 3.0 NL via Wikimedia Commons.) Right: A stack of sheets of muscovite mica. (By USGS; public domain via Wikimedia Commons.)

Chemically, sheet silicates usually contain silicon and oxygen in a 2:5 ratio (Si_4O_{10}). Micas contain mostly silicate, aluminum, and potassium. Biotite mica has more iron and magnesium and is considered a ferromagnesian silicate mineral. Muscovite micas belong to the felsic silicate minerals. **Felsic** is a contraction formed from feldspar, the dominant mineral in felsic rocks, and silicate.

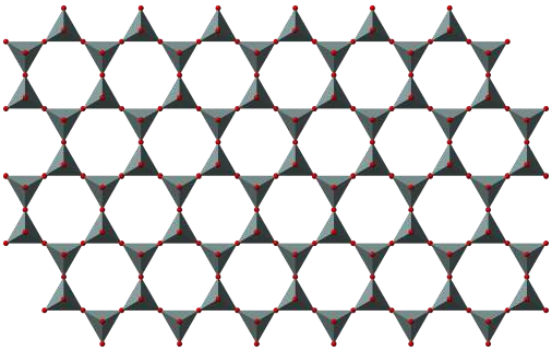
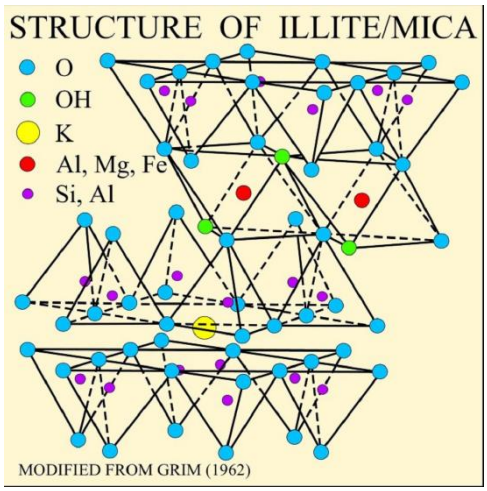


Figure 3.3.10: Sheet structure of mica, view perpendicular to the sheets. (By Benjah-bmm27; public domain via Wikimedia Commons.)

The illustration of the crystalline structure of mica shows the corner oxygen atoms (O) bonded with K, Al, Mg, Fe, and Si atoms, forming polymerized sheets of linked tetrahedra, with an octahedral layer of Fe, Mg, or Al, between them. The yellow potassium ions form **Van der Waals bonds** (attraction and repulsion between atoms, molecules, and surfaces) and hold the sheets together. Van der Waals bonds differ from covalent and ionic bonds, and exist here between the sandwiches, holding them together into a stack of sandwiches. The Van der Waals bonds are weak compared to the bonds within the sheets, allowing the sandwiches to be separated along the potassium layers. This gives mica its characteristic property of easily cleaving into sheets.



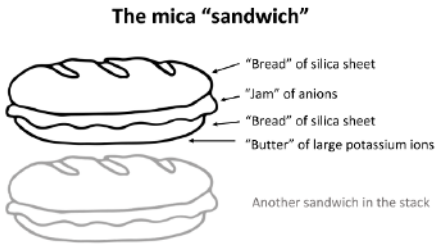


Figure 3.3.11: Left: Crystal structure of a mica, view parallel to the sheets. (By USGS; public domain via Wikimedia Commons.) Right: Mica silicate "sandwich" structure. (By Kindred Grey from Sandwich by Alex Muravev; CC BY 4.0 via Introduction to Earth Science.)

Clays minerals occur in sediments formed by the weathering of rocks and are another family of silicate minerals with a tetrahedral sheet structure. Clay minerals form a complex family and are an important component of many sedimentary rocks. Other sheet silicates include **serpentine** and **chlorite**, found in metamorphic rocks.

Clay minerals are composed of hydrous aluminum silicates. One type of clay, **kaolinite**, has a structure like an open-faced sandwich, with the bread being a single layer of silicon-oxygen tetrahedra and a layer of aluminum as the spread in an octahedral configuration with the top oxygens of the sheets.

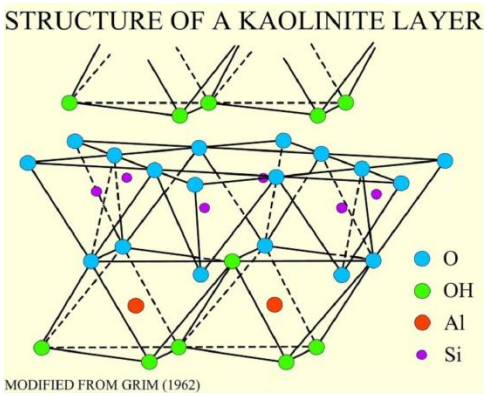


Figure 3.3.12: Structure of kaolinite. (By USGS; public domain)

Framework Silicates

Quartz and **feldspar** are the two most abundant minerals in the continental crust. There are two types of feldspar: one containing potassium and abundant in felsic rocks of the continental crust, and the other with sodium and calcium and abundant in the mafic rocks of oceanic crust. Together with quartz, these minerals are classified as **framework silicates**. They are built with a three-dimensional framework of silicate tetrahedra in which all four corner oxygens are shared with adjacent tetrahedra. Framework silicates are called **tectosilicates** and also include the alkali metal-rich feldspathoids and zeolites.



Figure 3.3.13: Pink orthoclase feldspar crystals. (By Didier Descouens; CC BY-SA 4.0, via Wikimedia Commons.)

Quartz is a felsic mineral composed of pure **silica**, SiO_2 with the tetrahedra arranged in a three-dimensional framework. Impurities consisting of atoms within this framework give rise to many varieties of quartz among which are gemstones like amethyst, rose quartz, and citrine.



Figure 3.3.14: Freely growing quartz crystals showing crystal faces. (By Didier Descouens; CC BY-SA 4.0 via Wikimedia Commons.)

Feldspars are mostly silica with aluminum, potassium, sodium, and calcium. The silica tetrahedra framework has holes and spaces into which other ions like aluminum, potassium, sodium, and calcium can fit, giving rise to a variety of mineral compositions and mineral names. Aluminum, which has a similar ionic size to silicon, can substitute for silicon inside the tetrahedra. Because potassium ions are so much larger than sodium and calcium ions, which are very similar in size, the inability of the crystal lattice to accommodate both potassium and sodium/calcium gives rise to the two families of feldspar: **orthoclase** and **plagioclase** respectively. Orthoclase feldspar ($KAlSi_3O_8$), also called **potassium feldspar** or **K-spar,** is made of silica, aluminum, and potassium. In plagioclase feldspar, $(Ca,Na)AlSi_3O_8$, the solid solution (Ca,Na) indicates a series of minerals, one end with calcium $CaAl_2Si_2O_8$, called **anorthite**, and the other end with sodium $NaAlSi_3O_8$, called **albite**.

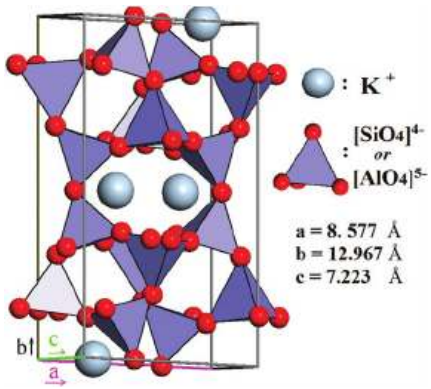


Figure 3.3.15: Crystal structure of potassium feldspar. (By Taisiya Skorina, Antoine Allanore; CC BY 3.0 via Research Gate.)

Feldspars are usually found in igneous rocks, such as granite, rhyolite, and basalt as well as metamorphic rocks and detrital sedimentary rocks. Detrital sedimentary rocks are composed of mechanically weathered rock particles, like sand and gravel. Quartz is especially abundant in detrital sedimentary rocks because it is very resistant to disintegration by weathering. While quartz is the

most abundant mineral on the Earth's surface due to its durability, the feldspar minerals are the most abundant minerals in the Earth's crust, comprising roughly 50% of the total minerals that make up the crust.

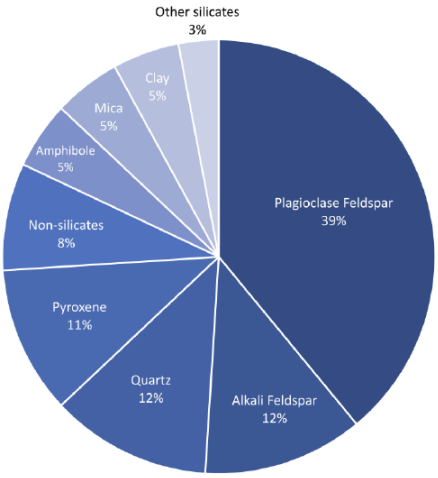


Figure 3.3.16: Pie chart of mineral abundance in Earth's crust. (By Kindred Grey; CC BY 4.0 via Introduction to Earth Science.)

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3.4: Non-Silicate Minerals

The crystal structure of **non-silicate minerals** (see table) does not contain silicon-oxygen tetrahedra. Many non-silicate minerals are economically important and provide **metallic** resources such as copper, lead, and iron. They also include valuable **non-metallic** products such as salt, construction materials, and fertilizer.

Table 3.4.1: Common non-silicate mineral groups.

Mineral Group	Examples	Formula	Uses
Native elements	gold, silver, copper	Au, Ag, Cu	Jewelry, coins, industry
Carbonates	calcite, dolomite	$CaCO_3$, $CaMg(CO_3)_2$	Lime, Portland cement
Oxides	hematite, magnetite, bauxite	Fe_2O_3, Fe_3O_4, a mixture of aluminum oxides	Ores of iron & aluminum, pigments
Halides	halite, sylvite	NaCl, KCl	Table salt, fertilizer
Sulfides	galena, chalcopyrite, cinnabar	PbS, $CuFeS_2$, HgS	Ores of lead, copper, mercury
Sulphates	gypsum, epsom salts	$CaSo_4{\cdot}2H_2O$, $MgSO_4{\cdot}7H_2O$	Sheetrock, therapeutic soak
Phosphates	apatite	$Ca_5(PO_4)_3(F,Cl,OH)$	Fertilizer, teeth, bones

Carbonates

Calcite ($CaCO_3$) and **dolomite** ($CaMg(CO_3)_2$) are the two most frequently occurring **carbonate minerals**, and usually occur in sedimentary rocks, such as limestone and dolostone rocks, respectively. Some carbonate minerals, such as calcite and dolomite, are formed via evaporation and precipitation. However, most carbonate-rich rocks, such as limestone, are created by the lithification of fossilized marine organisms. These organisms, including those we can see and many microscopic organisms, have shells or exoskeletons consisting of calcium carbonate ($CaCO_3$). When these organisms die, their remains accumulate on the floor of the water body in which they lived and the soft body parts decompose and dissolve away. The calcium carbonate hard parts become included in the sediments, eventually becoming the sedimentary rock called **limestone**. While limestone may contain large, easy to see fossils, most limestones contain the remains of microscopic creatures and thus originate from biological processes.



Figure 3.4.1: Limestone full of small fossils. (By Jim Stuby; public domain via Wikimedia Commons.)

Calcite crystals show an interesting property called **birefringence**, meaning they polarize light into two wave components vibrating at right angles to each other. As the two light waves pass through the crystal, they travel at different velocities and are separated by refraction into two different travel paths. In other words, the crystal produces a double image of objects viewed through it. Because they polarize light, calcite crystals are used in special petrographic microscopes for studying minerals and rocks.



Figure 3.4.2: Calcite crystal in the shape of a rhombohedron. Note the double-refracted word “calcite” in the center of the figure due to birefringence. (By Alkivar; public domain via Wikimedia Commons.)

The figure shows the crystal structure of calcite ($CaCO_3$). Like silicon, carbon has four valence electrons. The carbonate unit consists of carbon atoms (tiny white spheres) covalently bonded to three oxygen atoms (red), one oxygen sharing two valence electrons with the carbon and the other two sharing one valence electron each with the carbon, thus creating triangular units with a charge of -2. The negatively charged carbonate unit forms an ionic bond with the Ca ion (blue), which as a charge of +2.

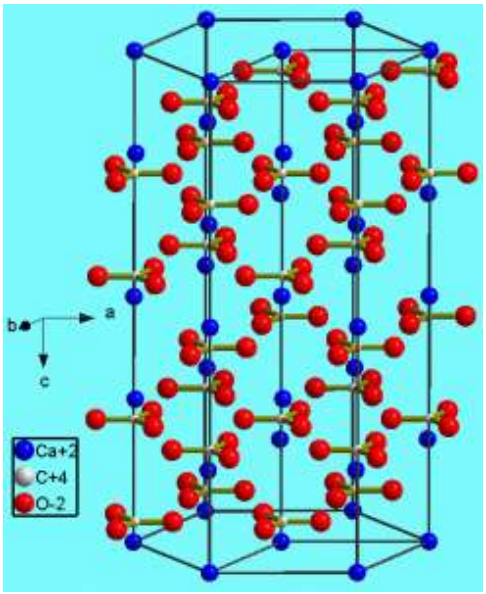


Figure 3.4.3: Crystal structure of calcite. (By Materialscientist at English Wikipedia; CC BY-SA 3.0 via Wikimedia Commons.)

Many non-silicate minerals are referred to as salts. The term **salts** used here refers to compounds made by replacing the hydrogen in natural acids. The most abundant natural acid is **carbonic acid** that forms by the solution of carbon dioxide in water ($CO_2 + H_2O \rightarrow H_2CO_3$). Carbonate minerals are salts built around the carbonate ion (CO_3^{2-}) where calcium and/or magnesium replace the hydrogen in carbonic acid (such as calcite, $CaCO_3$). Calcite and closely-related polymorph **aragonite** are secreted by organisms to form shells and physical structures like corals. Many such creatures draw both calcium and carbonate from dissolved bicarbonate ions (HCO_3^-) in ocean water. As seen in the mineral identification section, calcite is easily dissolved in acid and thus effervesces in dilute **hydrochloric acid** (HCl). Small dropper bottles of dilute hydrochloric acid are often carried by geologists in the field as well as used in mineral identification labs.

Other salts include halite (NaCl) in which sodium replaces the hydrogen in hydrochloric acid, and gypsum ($Ca[SO_4] \cdot 2\ H_2O$) in which calcium replaces the hydrogen in sulfuric acid (H_2SO_4). Note that some water molecules are also included in the gypsum crystal. Salts are often formed by evaporation and are called **evaporite minerals**.

Oxides, Halides, and Sulfides

After carbonates, the next most common non-silicate minerals are the oxides, halides, and sulfides.

Oxides consist of metal ions covalently bonded with oxygen. The most familiar oxide is **rust**, which is a combination of iron oxides (Fe_2O_3) and hydrated oxides. **Hydrated oxides** form when the iron is exposed to oxygen and water. Iron oxides are important for producing metallic iron. When iron oxide or ore is smelted, it produces carbon dioxide (CO_2) and metallic iron.



Figure 3.4.4: Limonite, a hydrated oxide of iron. (By USGS; public domain via Wikimedia Commons.)

The red color in rocks is usually due to the presence of iron oxides. For example, the red sandstone cliffs in Zion National Park and throughout southern Utah consist of white or colorless grains of quartz coated with iron oxide which serve as cementing agents holding the grains together.

Other iron oxides include limonite, magnetite, and hematite. **Hematite** occurs in many different crystal forms. The massive form shows no external structure. Botryoidal hematite shows large concentric blobs. Specular hematite looks like a mass of shiny metallic crystals. Oolitic hematite looks like a mass of dull red fish eggs. These different forms of hematite are polymorphs and all have the same formula, Fe_2O_3.



Figure 3.4.5: Oolitic hematite. (By Dave Dyet; public domain via Wikimedia Commons.)

Other common oxide minerals include:

- Ice (H_2O), an oxide of hydrogen
- Bauxite ($Al_2H_2O_4$), hydrated oxides of aluminum, an ore for producing metallic aluminum
- Corundum (Al_2O_3), which includes ruby and sapphire gemstones

The **halides** consist of halogens in column VII, usually fluorine or chlorine, ionically bonded with sodium or other cations. These include **halite** or sodium chloride (NaCl), common table salt; **sylvite** or potassium chloride (KCl); and **fluorite** or calcium fluoride (CaF_2).

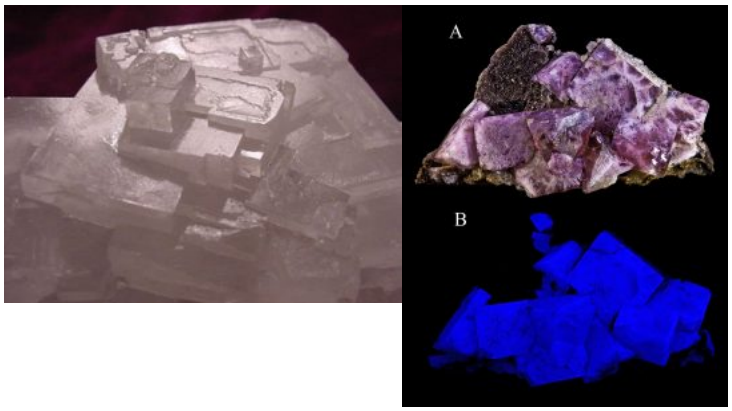


Figure 3.4.6: Left: Halite crystal showing cubic habit. (By USGS; public domain via Wikimedia Commons.) Right: Fluorite. B shows fluorescence of fluorite under UV light. (By Didier Descouens; CC BY-SA 4.0 via Wikimedia Commons.)

Halide minerals usually form from the evaporation of seawater or other isolated bodies of water. A well-known example of halide mineral deposits created by evaporation is the Bonneville Salt Flats, located west of the Great Salt Lake in Utah (see figure).



Figure 3.4.7: Salt crystals at the Bonneville Salt Flats. (By Famartin; CC BY-SA 4.0 via Wikimedia Commons.)

Many important metal ores are **sulfides,** in which metals are bonded to sulfur. Significant examples include **galena** (lead sulfide, PbS), **sphalerite** (zinc sulfide, ZnS), **pyrite** (iron sulfide, FeS_2, sometimes called "fool's gold"), and **chalcopyrite** (copper-iron sulfide, $CuFeS_2$). Sulfides are well known for being important ore minerals. For example, galena is the main source of lead, sphalerite is the main source of zinc, and chalcopyrite is the main copper ore mineral mined in porphyry deposits like the Bingham mine. The largest sources of nickel, antimony, molybdenum, arsenic, and mercury are also sulfides.



Figure 3.4.8: Cubic crystals of pyrite. (By CarlesMillan; CC BY-SA 3.0, via Wikimedia Commons.)

Sulfates

Sulfate minerals contain a metal ion, such as calcium, bonded to a sulfate ion. The **sulfate ion** is a combination of sulfur and oxygen (SO_4^{2-}). The sulfate mineral **gypsum** ($CaSO_4 \cdot 2H_2O$) is used in construction materials such as plaster and drywall. Gypsum is often formed from evaporating water and usually contains water molecules in its crystalline structure. The $\cdot 2H_2O$ in the formula indicates the water molecules are whole H_2O. This is different from minerals like amphibole, which contain a hydroxide ion (OH^-)

that is derived from water but is missing a hydrogen ion (H^+). The calcium sulfate without water is a different mineral than gypsum called **anhydrite** ($CaSO_4$).



Figure 3.4.9: Gypsum crystal. (By USGS; public domain via Wikimedia Commons.)

Phosphates

Phosphate minerals have a tetrahedral **phosphate ion** (PO_4^{3-}) combined with various anions and cations. In some cases arsenic or vanadium can substitute for phosphorus. Phosphates are an important ingredient of fertilizers as well as detergents, paint, and other products. The best known phosphate mineral is **apatite**, $Ca_5(PO_4)_3(F,Cl,OH)$, variations of which are found in teeth and bones. The gemstone **turquoise** [$CuAl_6(PO_4)_4(OH)_8{\cdot}4H_2O$] is a copper-rich phosphate mineral that, like gypsum, contains water molecules.



Figure 3.4.10: Apatite crystal on calcite. (By Didier Descouens; CC BY-SA 4.0 via Wikimedia Commons.)

Native Element Minerals

Native element minerals, usually metals, occur in nature in a pure or nearly pure state. **Gold** is an example of a native element mineral; it is not very reactive and rarely bonds with other elements so it is usually found in an isolated or pure state. The non-metallic and poorly-reactive mineral carbon is often found as a native element, such as **graphite** and **diamonds**. Mildly reactive metals like silver, copper, platinum, mercury, and sulfur sometimes occur as native element minerals. Reactive metals such as iron, lead, and aluminum almost always bond to other elements and are rarely found in a native state.



Figure 3.4.11: Left: Native copper. (By Jonathan Zander; CC BY-SA 3.0 via Wikimedia Commons.) Right: Native sulfur deposited around a volcanic fumarole. (By Brisk g; public domain via Wikimedia Commons.)

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3.5: Identifying Minerals

Geologists identify minerals by their **physical properties**. In the field, where geologists may have limited access to advanced technology and powerful machines, they can still identify minerals by testing several physical properties: luster and color, streak, hardness, crystal habit, cleavage and fracture, and some special properties. Only a few common minerals make up the majority of Earth's rocks and are usually seen as small grains in rocks. Of the several properties used for identifying minerals, it is good to consider which will be most useful for identifying them in small grains surrounded by other minerals.

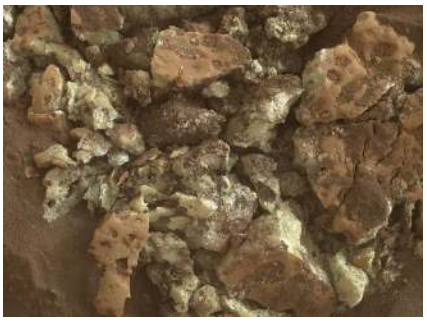


Figure 3.5.1: The rover Curiosity ran over this rock on Mars, and discovered the presence of mineral crystals of elemental sulfur on Mars. (By NASA/JPL-Caltech/MSSS; public domain via JPL.)

Luster and Color

The first thing to notice about a mineral is its surface appearance, specifically luster and color. **Luster** describes how the mineral looks. **Metallic luster** looks like a shiny metal such as chrome, steel, silver, or gold. **Submetallic luster** has a duller appearance. Pewter, for example, shows submetallic luster.

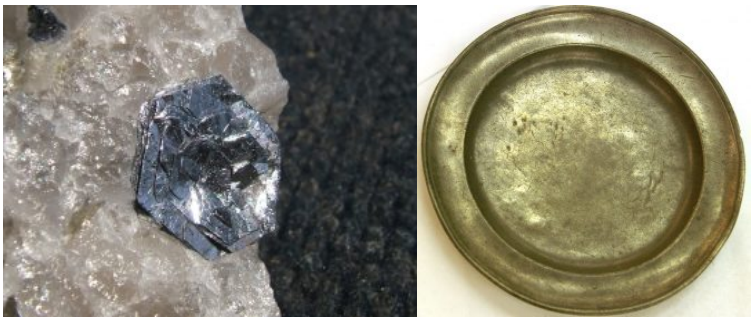


Figure 3.5.2: Left: 15 mm metallic luster hexagonal molybdenite crystal from Quebec. (By John Chapman; CC BY-SA 4.0 via Wikimedia Commons.) Right: Submetallic luster shown on an antique pewter plate. (By unknown; public domain via Wikimedia Commons.)

Nonmetallic luster doesn't look like metal and may be described as **vitreous** (glassy), **earthy**, **silky**, **pearly**, and other surface qualities. Nonmetallic minerals may be shiny, although their vitreous shine is different from metallic luster. See the table for descriptions and examples of nonmetallic luster.

Table 3.5.1: Nonmetallic luster descriptions and examples.

Luster	Image	Description
Vitreous/glassy	 Quartz crystals showing vitreous luster (By Didier Descouens; CC BY-SA 4.0, via Wikimedia Commons.)	Surface is shiny like glass
Earthy/dull	 Kaolin specimen showing dull or earthy luster (By USGS; public domain via Wikimedia Commons.)	Dull, like dried mud or clay
Silky	 Gypsum specimen showing silky luster (By Ra'ike; CC-BY-SA-3.0 via Wikimedia Commons.)	Soft shine like silk fabric
Pearly	 Mica specimen showing pearly luster (By Luis Miguel Bugallo Sanchez; CC-BY-SA-3.0 via Wikimedia Commons.)	Like the inside of a clam shell or mother-of-pearl
Submetallic		Has the appearance of dull metal, like pewter. These minerals would usually still be considered metallic. Submetallic appearance can occur in metallic minerals because of weathering.

	Submetallic luster on sphalerite (By Andreas Früh; CC-BY-SA-3.0 via Wikimedia Commons.)	

Surface **color** may be helpful in identifying minerals, although it can be quite variable within the same mineral family. Mineral colors are affected by the main elements as well as impurities in the crystals. These impurities may be rare elements—like manganese, titanium, chromium, or lithium—even other molecules that are not normally part of the mineral formula. For example, the incorporation of water molecules gives quartz, which is normally clear, a milky color.

Some minerals predominantly show a single color. Malachite and azurite are green and blue, respectively, because of their copper content. Other minerals have a predictable range of colors due to elemental substitutions, usually via a solid solution. Feldspars, the most abundant minerals in the Earth's crust, are complex, have solid solution series, and are present in several colors including pink, white, green, gray and others. Other minerals also come in several colors, influenced by trace amounts of several elements. The same element may show up as different colors in different minerals. With notable exceptions, color is usually not a definitive property of minerals. For identifying many minerals, a more reliable indicator is streak, which is the color of the powdered mineral.



Figure 3.5.3: Azurite is ALWAYS a dark blue color and has been used for centuries for blue pigment. (By Graeme Churchard; CC BY 2.0 via Wikimedia Commons.)

Streak

Streak is the color of a powdered mineral and can be seen when a mineral sample is scratched or scraped on an unglazed porcelain streak plate. A paper page in a field notebook may also be used for the streak of some minerals. Minerals that are harder than the streak plate will not show streak but will scratch the porcelain. For these minerals, a streak test can be obtained by powdering the mineral with a hammer and smearing the powder across a streak plate or notebook paper.

While mineral surface colors and appearances may vary, their streak colors can be diagnostically useful. An example of this property is seen in the iron-oxide mineral hematite. Hematite occurs in a variety of forms, colors, lusters (from shiny metallic silver to earthy red-brown), and different physical appearances. A hematite streak is consistently reddish-brown, no matter what the original specimen looks like. Iron sulfide, or pyrite, is a brassy metallic yellow. Commonly named fool's gold, pyrite has a characteristic black to greenish-black streak.



Figure 3.5.4: Some minerals have different streaks than their visual color. (By ; CC BY-SA 3.0 via Wikimedia Commons.)

Hardness

Hardness measures the ability of a mineral to scratch other substances. The **Mohs Hardness Scale** gives a number showing the relative scratch-resistance of minerals when compared to a standardized set of minerals of increasing hardness. The Mohs scale was developed by German geologist Fredrick Mohs in the early 20th century, although the idea of identifying minerals by hardness goes back thousands of years. Mohs hardness values are determined by the strength of a mineral's atomic bonds.

The figure shows the minerals associated with specific hardness values, together with some common items readily available for use in field testing and mineral identification. The hardness values run from 1 to 10, with 10 being the hardest; however, the scale is not linear. Diamond defines a hardness of 10 and is actually about four times harder than corundum, which is 9. A steel pocketknife blade, which has a hardness value of 5.5, separates between hard and soft minerals on many mineral identification keys.

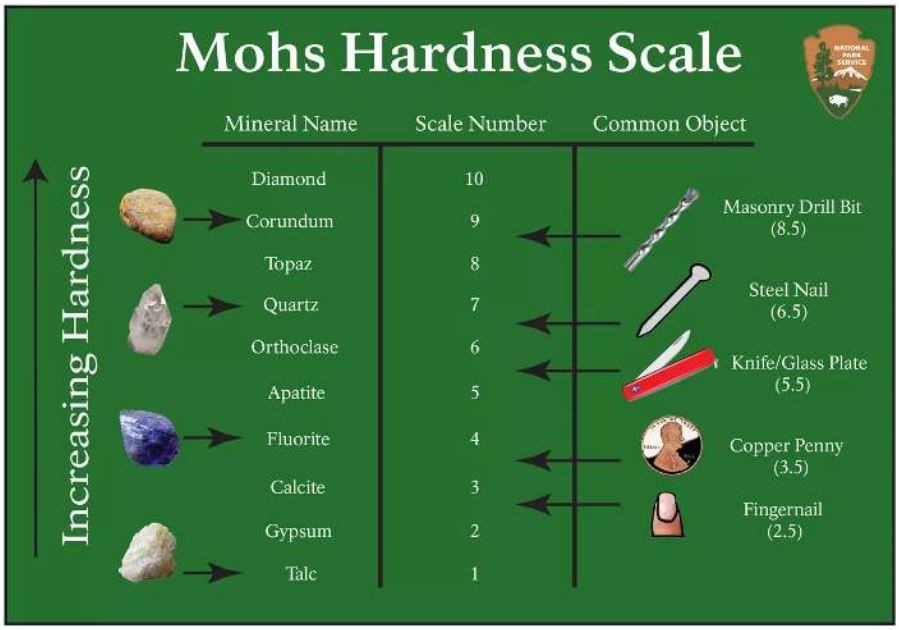
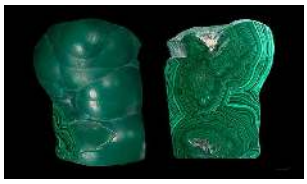


Figure 3.5.5: Mohs Hardness Scale. (By Heather Walborn; public doman via NPS.)

Crystal Habit

Minerals can be identified by **crystal habit**, how their crystals grow and appear in rocks. Crystal shapes are determined by the arrangement of the atoms within the crystal structure. For example, a cubic arrangement of atoms gives rise to a cubic-shaped mineral crystal. Crystal habit refers to typically observed shapes and characteristics; however, they can be affected by other minerals crystallizing in the same rock. When minerals are constrained so they do not develop their typical crystal habit, they are called **anhedral**. **Subhedral** crystals are partially formed shapes. For some minerals, characteristic crystal habit is to grow crystal faces even when surrounded by other crystals in the rock. An example is garnet. Minerals grow freely where the crystals are unconstrained and can take characteristic shapes often form crystal faces. A **euhedral** crystal has a perfectly formed, unconstrained shape. Some minerals crystallize in such tiny crystals, they do not show a specific crystal habit to the naked eye. Other minerals, like pyrite, can have an array of different crystal habits, including cubic, dodecahedral, octahedral, and massive. The table lists typical crystal habits of various minerals.

Table 3.5.2: Typical crystal habits of various minerals.

Habit	Image	Examples
Bladed long and flat crystals	 Bladed kyanite (By Aelwyn; CC-BY-SA-3.0 via Wikimedia Commons.)	kyanite, amphibole, gypsum
Botryoidal/mammillary blobby, circular crystals	 Malachite from the Congo (By Didier Descouens; CC BY-SA 3.0 via Wikimedia Commons.)	hematite, malachite, smithsonite
Coating/laminae/druse crystals that are small and coat surfaces	 Quartz (var. amethyst) in a geode (By Juppi66; public domain via Wikimedia Commons.)	quartz, calcite, malachite, azurite
Cubic cube-shaped crystals	 Cubic crystals of halite (By Robert M. Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	pyrite, galena, halite
Dodecahedral 12-sided polygon shapes	 Pyrite crystals with dodecahedral habit (By Didier Descouens; CC BY-SA 4.0 via Wikimedia Commons.)	garnet, pyrite

Dendritic branching crystals	 Manganese dendrites, scale in mm (By Mark A. Wilson; public domain via Wikimedia Commons.)	Mn-oxides, copper, gold
Equant crystals that do not have a long direction	 Olivine crystal (By S kitahashi; CC-BY-SA-3.0 via Wikimedia Commons.)	olivine, garnet, pyroxene
Fibrous thin, very long crystals	 Tremolite, a type of amphibole (By Didier Descouens; CC BY-SA 4.0, via Wikimedia Commons.)	serpentine, amphibole, zeolite
Layered, sheets stacked, very thin, flat crystals	 Sheet crystals of muscovite (By Rob Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	mica (biotite, muscovite, etc.)
Lenticular/platy crystals that are plate-like	 Orange wulfenite on calcite (By Rob Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	selenite roses, wulfenite, calcite

Hexagonal crystals with six sides	 Hexagonal hanksite (By Matt Affolter; CC BY-SA 3.0 via Wikimedia Commons.)	quartz, hanksite, corundum
Massive/granular Crystals with no obvious shape, microscopic crystals	 Limonite, a hydrated oxide of iron (By USGS; public domain via Wikimedia Commons.)	limonite, pyrite, azurite, bornite
Octahedral 4-sided double pyramid crystals	 Octahedral fluorite (By Ryan Salsbury; CC-BY-SA-3.0 via Wikimedia Commons.)	diamond, fluorite, magnetite, pyrite
Prismatic/columnar very long, cylindrical crystals	 Columnar tourmaline (By Rob Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	tourmaline, beryl, barite
Radiating crystals that grow from a point and fan-out		pyrite “suns”, pyrophyllite

	Pyrophyllite (By Rob Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	
Rhombohedral crystals shaped like slanted cubes	 Calcite crystal in shape of rhombohedron (By Alkivar; public domain via Wikimedia Commons.)	calcite, dolomite
Tabular/blocky/stubby sharp-sided crystals with no long direction	 Crystals of diopside, a member of the pyroxene family (By Rob Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	feldspar, pyroxene, calcite
Tetrahedral three-sided, pyramid-shaped crystals	 Tetrahedrite (By Rob Lavinsky, iRocks.com; CC BY-SA 3.0 via Wikimedia Commons.)	magnetite, spinel, tetrahedrite

Another crystal habit that may be used to identify minerals is **twinning,** which occurs when the crystal structure replicates in mirror images along certain directions in the crystal. **Striations,** which are dark and light parallel lines on a crystal face are also used in identifying minerals. Striations and twinning are related properties in some minerals including plagioclase feldspar. Striations are optical lines on a cleavage surface. Because of twinning in the crystal, striations show up on one of the two cleavage faces of the plagioclase crystal.



Figure 3.5.6: Left: Twinned staurolite (By Helgi; CC BY-SA 3.0 via Wikimedia Commons.) Right: Striations on plagioclase due to twinning. (By Mike Beauregard; CC BY 2.0, via Wikimedia Commons.)

Cleavage and Fracture

Minerals often show characteristic patterns of breaking along specific cleavage planes or show characteristic fracture patterns. **Cleavage planes** are smooth, flat, parallel planes within the crystal. The cleavage planes may show as reflective surfaces on the crystal, as parallel cracks that penetrate into the crystal, or show on the edge or side of the crystal as a series of steps. Cleavage arises in crystals where the atomic bonds between atomic layers are weaker along some directions than others, meaning they will break preferentially along these planes. Because they develop on atomic surfaces in the crystal, cleavage planes are optically smooth and reflect light, although the actual break on the crystal may appear jagged or uneven. In such cleavages, the cleavage surface may appear like rice terraces on a mountainside that all reflect sunlight from a particular sun angle.

Graphite has its carbon atoms arranged into layers with relatively strong bonds within the layer and very weak bonds between the layers. Thus graphite cleaves readily between the layers and the layers slide easily over one another giving graphite its lubricating quality. Graphite is also used to make pencil lead.

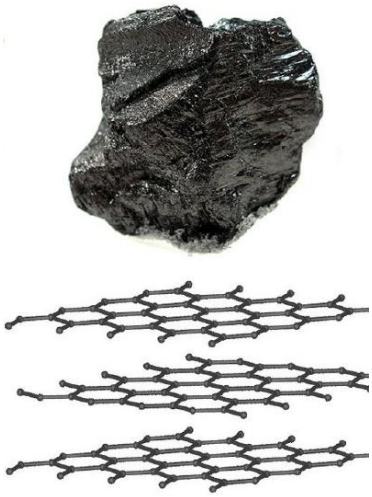


Figure 3.5.7: Graphite and atomic structure showing layers of carbon atoms separated by a gap with weak bonds holding the layers together. (By ltub; CC BY-SA 3.0 via Wikimedia Commons.)

Some minerals have strong cleavage, some minerals only have weak cleavage and some do not typically demonstrate cleavage. For example, quartz and olivine rarely show cleavage and typically break into conchoidal **fracture** patterns. Mineral fracture surfaces may be rough, uneven, or show a conchoidal fracture. Uneven fracture patterns are described as **irregular**, **splintery**, or **fibrous**. A **conchoidal fracture** has a smooth, curved surface like a shallow bowl or conch shell, often with curved ridges. Natural volcanic glass, called obsidian, breaks with this characteristic conchoidal pattern.



Figure 3.5.8: Obsidian showing conchoidal fracture. (By Siim Sepp; CC BY-SA 2.5 via Wikimedia Commons.)

To work with cleavage, it is important to remember that cleavage is a result of bonds separating along planes of atoms in the crystal structure. On some minerals, cleavage planes may be confused with crystal faces. This will usually not be an issue for crystals of minerals that grew together within rocks. The act of breaking the rock to expose a fresh face will most likely break the crystals along cleavage planes. Some cleavage planes are parallel with crystal faces but many are not. Cleavage planes are smooth, flat, parallel planes within the crystal. The cleavage planes may show as parallel cracks that penetrate into the crystal (see amphibole below), or show on the edge or side of the crystal as a series of steps like rice terraces. For some minerals, characteristic crystal habit is to grow crystal faces even when surrounded by other crystals in rock. An example is garnet. Minerals grow freely where the crystals are unconstrained and can take characteristic shapes often form crystal faces (see quartz below).



Figure 3.5.9: Freely growing quartz crystals showing crystal faces, not to be mistaken as cleavage. (By JJ Harrison; CC BY-SA 2.5 via Wikimedia Commons.)

In some minerals, distinguishing cleavage planes from crystal faces may be challenging for the student. Understanding the nature of cleavage and referring to the number of cleavage planes and cleavage angles on identification keys should provide the student with enough information to distinguish cleavages from crystal faces. Cleavage planes may show as multiple parallel cracks or flat surfaces on the crystal. Cleavage planes may be expressed as a series of steps like terraced rice paddies. See the cleavage surfaces on galena or wollastonite below. Cleavage planes arise from the tendency of mineral crystals to break along specific planes of weakness within the crystal favored by atomic arrangements. The number of cleavage planes, the quality of the cleavage surfaces, and the angles between them are diagnostic for many minerals and cleavage is one of the most useful properties for identifying minerals. Learning to recognize cleavage is an especially important and useful skill in studying minerals.

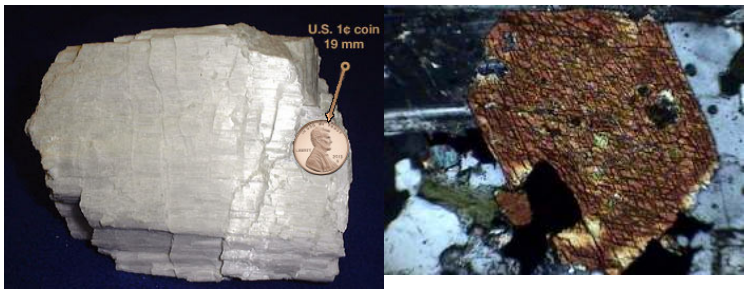


Figure 3.5.10: Left: Steps of cleavage along the same cleavage direction on wollastonite. (By USGS; public domain via Wikimedia Commons.) Right: Photomicrograph showing 120°/60° cleavage within a grain of amphibole. (By Miguel Vera Leon; CC BY 2.0 via Wikimedia Commons.)

As an identification property of minerals, cleavage is usually given in terms of the quality of the cleavage (**perfect**, **imperfect**, or **none**), the number of cleavage surfaces, and the angles between the surfaces. The most common number of cleavage plane directions in the common rock-forming minerals are **one perfect cleavage** (as in mica), **two cleavage planes** (as in feldspar, pyroxene, and amphibole), and **three cleavage planes** (as in halite, calcite, and galena). One perfect cleavage (as in mica) develops on the top and bottom of the mineral specimen with many parallel cracks showing on the sides but no angle of intersection. Two cleavage planes intersect at an angle. Common cleavage angles are 60°, 75°, 90°, and 120°. Amphibole has two cleavage planes at 60° and 120°. Galena and halite have three cleavage planes at 90° (cubic cleavage). Calcite cleaves readily in three directions producing a cleavage figure called a rhomb that looks like a cube squashed over toward one corner giving rise to the approximately 75° cleavage angles. Pyroxene has an imperfect cleavage with two planes at nearly 90°.

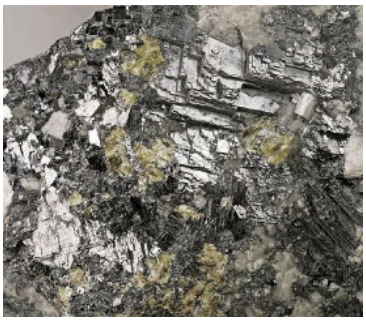


Figure 3.5.11: Cubic cleavage of galena; note how the cleavage surfaces show up as different but parallel layers in the crystal. (By Modris Baum; public domain via Wikimedia Commons.)

Cleavages on Common Rock-Forming Minerals

- Quartz—none (conchoidal fracture)
- Olivine—none (conchoidal fracture)
- Mica—1 perfect
- Feldspar—2 perfect at 90°
- Pyroxene—2 imperfect at nearly 90°
- Amphibole—2 perfect at 60°/120°
- Calcite—3 perfect at approximately 75°
- Halite, galena, pyrite—3 perfect at 90°

Special Properties

Special properties are unique and identifiable characteristics used to identify minerals or that allow some minerals to be used for special purposes. Ulexite has a fiber-optic property that can project images through the crystal like a high-definition television screen. A simple identifying special property is **taste**, such as the salty flavor of halite or common table salt (NaCl). Sylvite is potassium chloride (KCl) and has a more bitter taste.



Figure 3.5.12: A demonstration of ulexite's image projection property. (By Zeichner; public domain via Wikimedia Commons.)

Another property geologists may use to identify minerals is a property related to density called **specific gravity**. Specific gravity measures the weight of a mineral specimen relative to the weight of an equal volume of water. The value is expressed as a ratio between the mineral and water weights. To measure specific gravity, a mineral specimen is first weighed in grams then submerged in a graduated cylinder filled with pure water at room temperature. The rise in water level is noted using the cylinder's graduated

scale. Since the weight of water at room temperature is 1 gram per cubic centimeter, the ratio of the two weight numbers gives the specific gravity. Specific gravity is easy to measure in the laboratory but is less useful for mineral identification in the field than other more easily observed properties, except in a few rare cases such as the very dense galena or native gold. The high density of these minerals gives rise to a qualitative property called "**heft**." Experienced geologists can roughly assess specific gravity by heft, a subjective quality of how heavy the specimen feels in one's hand relative to its size.



Figure 3.5.13: Native gold has one of the highest specific gravities. (By Gump Stump; CC BY-SA 3.0, via Wikimedia Commons.)

A simple test for identifying calcite and dolomite is to drop a bit of dilute hydrochloric acid (10-15% HCl) on the specimen. If the acid drop **effervesces** or fizzes on the surface of the rock, the specimen is calcite. If it does not, the specimen is scratched to produce a small amount of powder and test with acid again. If the acid drop fizzes slowly on the powdered mineral, the specimen is dolomite. The difference between these two minerals can be seen in the video (https://youtu.be/DX6ZMPbA09U). Geologists who work with carbonate rocks carry a small dropper bottle of dilute HCl in their field kit. Vinegar, which contains acetic acid, can be used for this test and is used to distinguish non-calcite fossils from limestone. While acidic, vinegar produces less of a fizzing reaction because acetic acid is a weaker acid.

Some iron-oxide minerals are **magnetic** and are attracted to magnets. A common name for naturally magnetic iron oxide is **lodestone**. Others include magnetite (Fe_3O_4) and ilmenite ($FeTiO_3$). Magnetite is strongly attracted to magnets and can be magnetized. Ilmenite and some types of hematite are weakly magnetic.

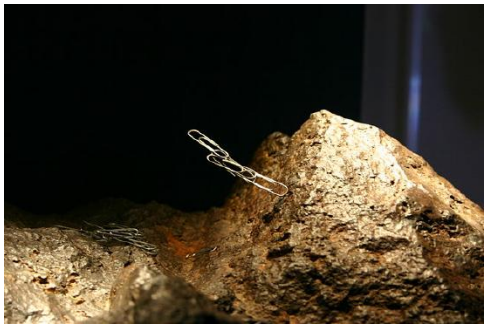


Figure 3.5.14: Paperclips attracted to lodestone (magnetite). (By Ryan Somma; CC BY-SA 2.0, via Wikimedia Commons.)

Some minerals and mineraloids scatter light via a phenomenon called **iridescence**. This property occurs in labradorite (a variety of plagioclase) and opal. It is also seen in biologically created substances like pearls and seashells. Cut diamonds show iridescence and the jeweler's diamond cut is designed to maximize this property.

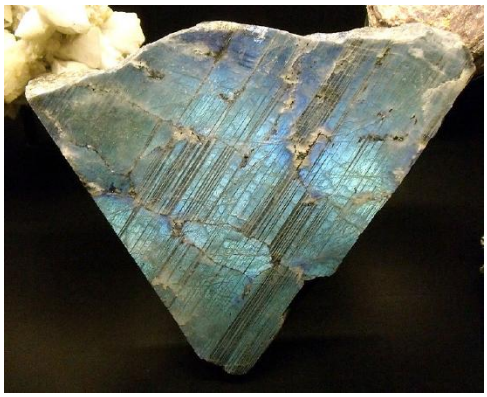


Figure 3.5.15: Iridescence on plagioclase; also showing striations on the cleavage surface. (By Mike Beauregard; CC BY 2.0, via Wikimedia Commons.)

Striations on mineral cleavage faces are an optical property that can be used to separate plagioclase feldspar from potassium feldspar (orthoclase). A process called twinning creates parallel zones in the crystal that are repeating mirror images. The actual cleavage angle in plagioclase is slightly different than 90° and the alternating mirror images in these twinned zones produce a series of parallel lines on one of plagioclase's two cleavage faces. Light reflects off these twinned lines at slightly different angles which then appear as light and dark lines called striations on the cleavage surface. Potassium feldspar does not exhibit twinning or striations but may show linear features called **exsolution lamellae**, also known as **perthitic lineation** or simply **perthite**. Because sodium and potassium do not fit into the same feldspar crystal structure, the lines are created by small amounts of sodium feldspar (albite) separating from the dominant potassium feldspar (orthoclase) within the crystal structure. The two different feldspars crystallize out into roughly parallel zones within the crystal, which are seen as these linear markings.

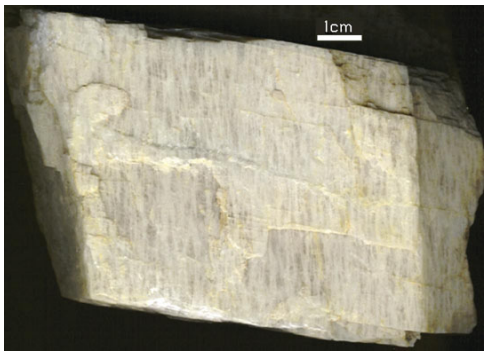


Figure 3.5.16: Exsolution lamellae or perthitic lineations within potassium feldspar. (By Jstuby; public domain via Wikimedia Commons.)

One of the most interesting special mineral properties is **fluorescence**. Certain minerals, or trace elements within them, give off visible light when exposed to ultraviolet radiation or black light. Many mineral exhibits have a fluorescence room equipped with black lights so this property can be observed. An even rarer optical property is phosphorescence. **Phosphorescent** minerals absorb light and then slowly release it, much like a glow-in-the-dark sticker.

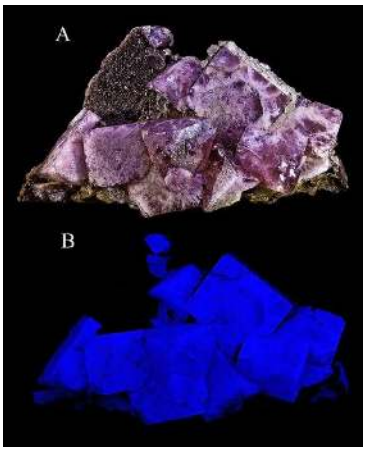


Figure 3.5.17: Fluorite. Lower image shows fluorescence of fluorite under UV light. (By Didier Descouens; CC BY-SA 4.0 via Wikimedia Commons.)

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3.S: Summary

Minerals are the building blocks of rocks and essential to understanding geology. Mineral properties are determined by their atomic bonds. Most minerals begin in a fluid, and either crystallize out of cooling magma or precipitate as ions and molecules out of a saturated solution. The silicates are largest group of minerals on Earth, by number of varieties and relative quantity, making up a large portion of the crust and mantle. Based on the silicon-oxygen tetrahedra, the crystal structure of silicates reflects the fact that silicon and oxygen are the top two of Earth's most abundant elements. Non-silicate minerals are also economically important, providing many types of construction and manufacturing materials. Minerals are identified by their unique physical properties, including luster, color, streak, hardness, crystal habit, fracture, cleavage, and special properties.

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CHAPTER OVERVIEW

4: Igneous Processes and Volcanoes

Learning Objectives

- Explain the origin of magma it relates to plate tectonics.
- Describe how the Bowen's Reaction Series relates mineral crystallization and melting temperatures.
- Explain how cooling of magma leads to rock compositions and textures, and how these are used to classify igneous rocks.
- Analyze the features of common igneous landforms and how they relate to their origin.
- Describe how silica content affects magma viscosity and eruptive style of volcanoes.
- Describe volcano types, eruptive styles, composition, and their plate tectonic settings.
- Describe volcanic hazards.

Igneous rock is formed when liquid rock freezes into solid rock. This molten material is called **magma** when it is in the ground and **lava** when it is on the surface. Only the Earth's outer core is liquid; the Earth's mantle and crust are naturally solid. However, there are a few minor pockets of magma that form near the surface where geologic processes cause melting. It is this magma that becomes the source for volcanoes and igneous rocks. This chapter will describe the classification of igneous rocks, the unique processes that form magmas, types of volcanoes and volcanic processes, volcanic hazards, and igneous landforms.



Figure 4.1: Lava flow in Hawaii

Lava cools quickly on the surface of the Earth and forms tiny microscopic crystals. These are known as fine-grained **extrusive**, or **volcanic**, igneous rocks. Extrusive rocks are often **vesicular**, filled with holes from escaping gas bubbles. **Volcanism** is the process in which lava has erupted. Depending on the properties of the lava that is erupted, the volcanism can be drastically different, from smooth and gentle to dangerous and explosive. This leads to different types of volcanoes and different volcanic hazards.



Figure 4.2: Half Dome, a massive intrusive igneous rock in Yosemite National Park, now exposed by erosion.

In contrast, magma that cools slowly below the Earth's surface forms larger crystals which can be seen with the naked eye. These are known as coarse-grained **intrusive**, or **plutonic**, igneous rocks. This relationship between cooling rates and grain sizes of the solidified minerals in igneous rocks is important for interpreting the rock's geologic history.

4.1: Classification of Igneous Rocks

4.2: Bowen’s Reaction Series
4.3: Magma Generation
4.4: Volcanism
4.5: Volcanic Structures
4.6: Volcanic Hazards and Monitoring
4.S: Summary

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4.1: Classification of Igneous Rocks

Igneous rocks are classified based on texture and composition. **Texture** describes the physical characteristics of the minerals, such as grain size. This relates to the cooling history of the molten magma from which it came. Composition refers to the rock's specific mineralogy and chemical composition. Cooling history is also related to changes that can occur to the composition of igneous rocks.

Texture

If magma cools slowly, deep within the crust, the resulting rock is called **intrusive** or **plutonic**. The slow cooling process allows crystals to grow large, giving the intrusive igneous rock a **coarse-grained** or **phaneritic** texture. The individual crystals in phaneritic texture are readily visible to the unaided eye.

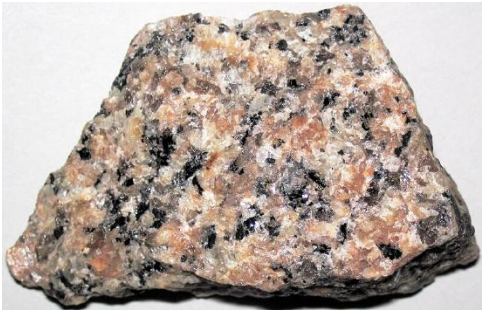


Figure 4.1.1: Granite is a classic coarse-grained (phaneritic) intrusive igneous rock. The different colors are unique minerals. The black colors are likely two or three different minerals. (By James St. John; CC BY 2.0 via Wikimedia Commons.)

When lava is extruded onto the surface, or intruded into shallow fissures near the surface and cools, the resulting igneous rock is called **extrusive** or **volcanic**. Extrusive igneous rocks have a **fine-grained** or **aphanitic** texture, in which the grains are too small to see with the unaided eye. The fine-grained texture indicates the quickly cooling lava did not have time to grow large crystals. These tiny crystals can be viewed under a petrographic microscope [1]. In some cases, extrusive lava cools so rapidly it does not develop crystals at all. This non-crystalline material is not classified as minerals but as **volcanic glass**. This is a common component of volcanic ash and rocks like obsidian.



Figure 4.1.2: Basalt is a classic fine-grained (aphanitic) extrusive igneous rock. (By James St. John; CC BY 2.0 via Wikimedia Commons.)

Some igneous rocks have a mix of coarse-grained minerals surrounded by a **matrix** of fine-grained material in a texture called **porphyritic**. The large crystals are called **phenocrysts** and the fine-grained matrix is called the **groundmass**. Porphyritic texture indicates the magma body underwent a multi-stage cooling history, cooling slowly while deep under the surface and later rising to a shallower depth or the surface where it cooled more quickly.

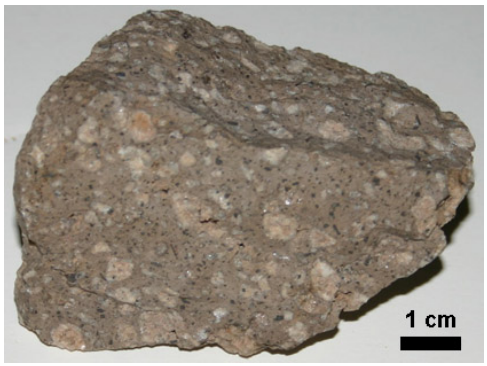


Figure 4.1.3: Porphyritic texture.

Residual molten material expelled from igneous intrusions may form veins or masses containing very large crystals of minerals like feldspar, quartz, beryl, tourmaline, and mica. This texture, which indicates a very slow crystallization, is called **pegmatitic**. A rock that chiefly consists of pegmatitic texture is known as a **pegmatite**. To give an example of how large these crystals can get, transparent cleavage sheets of pegmatitic muscovite mica were used as windows during the Middle Ages.



Figure 4.1.4: Pegmatitic texture.

All magmas contain **volatiles**, which are gases dissolved in a solution. As the magma rises to the surface, the drop in pressure causes the dissolved volatiles to come bubbling out of solution, like the fizz in an opened bottle of soda. The gas bubbles become trapped in the solidifying lava to create a **vesicular** texture, with the holes specifically called **vesicles**. A volcanic rock with vesicles is called **scoria.**



Figure 4.1.5: Scoria, a vesicular extrusive igneous rock.

An extreme version of scoria occurs when volatile-rich lava is very quickly quenched and becomes a meringue-like froth of glass called **pumice**. Some pumice is so full of vesicles that the density of the rock drops low enough that it will float.



Figure 4.1.6: Pumice.

Lava that cools extremely quickly may not form crystals at all, even microscopic ones. The resulting rock is called volcanic glass. **Obsidian** is a rock consisting of volcanic glass. Obsidian as a glassy rock often shows conchoidal fracture similar to the mineral quartz.

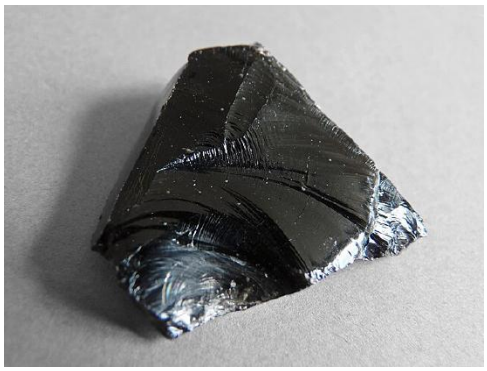


Figure 4.1.7: Obsidian (volcanic glass). Note the conchoidal fracture.

When volcanoes erupt explosively, vast amounts of lava, rock, ash, and gases are thrown into the atmosphere. The solid parts, called **tephra**, settle back to Earth and cool into rocks with **pyroclastic** textures. *Pyro*, meaning fire, refers to the igneous source of the tephra and *clastic* refers to the rock fragments. Tephra fragments are named based on size—**ash** (<2 mm), **lapilli** (2-64 mm), and **bombs or blocks** (>64 mm). Pyroclastic texture is usually recognized by the chaotic mix of crystals, angular glass shards, and rock fragments. Rock formed from large deposits of tephra fragments is called **tuff**. If the fragments accumulate while still hot, the heat may deform the crystals and weld the mass together, forming a **welded tuff**.



Figure 4.1.8: This tuff has crystals, rock fragments, and bedrock mixed together.

Composition

Composition refers to a rock's chemical and mineral make-up. For igneous rocks, composition is divided into four groups: felsic, intermediate, mafic, and ultramafic. These groups refer to differing amounts of silica, iron, and magnesium found in the minerals that make up the rocks. It is important to realize these groups do not have sharp boundaries in nature, but rather lie on a continuous spectrum with many transitional compositions and names that refer to specific quantities of minerals. As an example, granite is a commonly-used term but has a very specific definition which includes exact quantities of minerals like feldspar and quartz. Rocks labeled as 'granite' in laymen applications can be several other rocks, including syenite, tonalite, and monzonite. To avoid these complications, the following figure presents a simplified version of igneous rock nomenclature focusing on the four main groups, which is adequate for an introductory student.

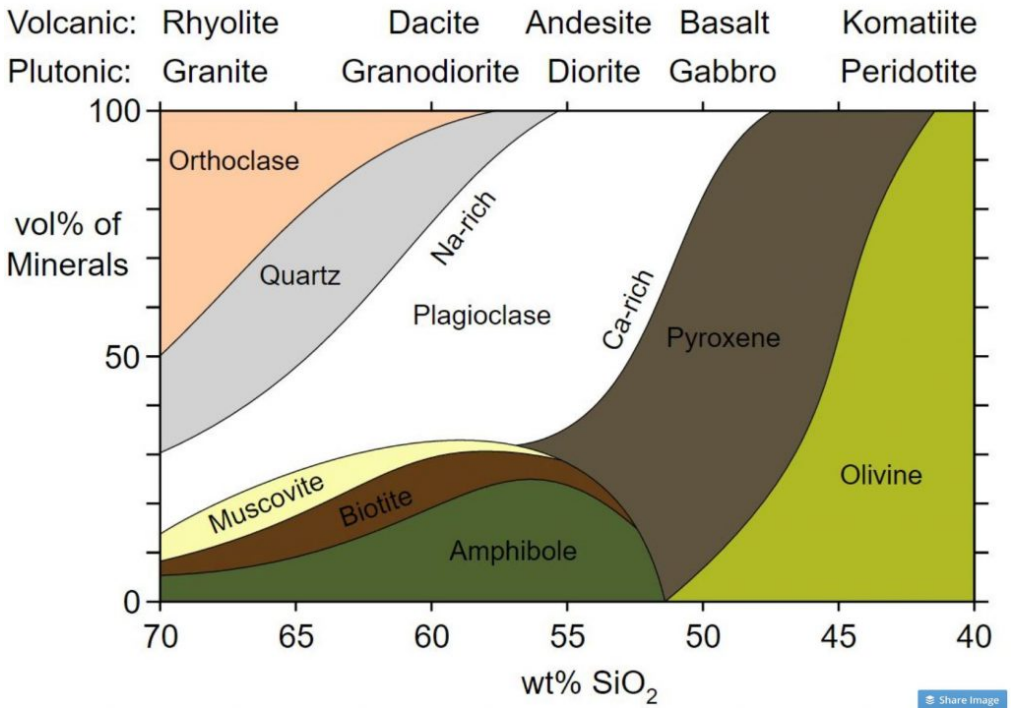


Figure 4.1.9: Mineral composition of common igneous rocks. The percentage of minerals is shown on the vertical axis. The percentage of silica is shown on the horizontal axis. Rock names at the top include a continuous spectrum of compositions grading from one into another.

- **Felsic** refers to a predominance of the light-colored (felsic) minerals feldspar and silica in the form of quartz. These light-colored minerals have more silica as a proportion of their overall chemical formula. Minor amounts of dark-colored (mafic) minerals like amphibole and biotite mica may be present as well. Felsic igneous rocks are rich in silica (in the 65-75% range, meaning the rock would be 65-75% weight percent SiO_2) and poor in iron and magnesium.
- **Intermediate** is a composition between felsic and mafic. It usually contains roughly equal amounts of light and dark minerals, including light grains of plagioclase feldspar and dark minerals like amphibole. It is intermediate in silica in the 55-60% range.
- **Mafic** refers to an abundance of ferromagnesian minerals (with magnesium and iron, chemical symbols Mg and Fe) plus plagioclase feldspar. It is mostly made of dark minerals like pyroxene and olivine, which are rich in iron and magnesium and relatively poor in silica. Mafic rocks are low in silica, in the 45-50% range.
- **Ultramafic** refers to the extremely mafic rocks composed of mostly olivine and some pyroxene which have even more magnesium and iron and even less silica. These rocks are rare on the surface, but make up peridotite, the rock of the upper mantle. It is poor in silica, in the 40% or less range.

On the figure above, the top row has both plutonic and volcanic igneous rocks arranged in a continuous spectrum from felsic on the left to intermediate, mafic, and ultramafic toward the right. **Rhyolite** refers to the volcanic and felsic igneous rocks and **granite** refer to intrusive and felsic igneous rocks. **Andesite** and **diorite** likewise refer to extrusive and intrusive intermediate rocks (with dacite and granodiorite applying to those rocks with composition between felsic and intermediate). **Basalt** and **gabbro** are the extrusive and intrusive names for mafic igneous rocks, and **peridotite** is ultramafic, with **komatiite** as the fine-grained extrusive

equivalent. Komatiite is a rare rock because volcanic material that comes directly from the mantle is not common, although some examples can be found in ancient Archean rocks [2]. Nature rarely has sharp boundaries and the classification and naming of rocks often impose what appears to be sharp boundary names onto a continuous spectrum.

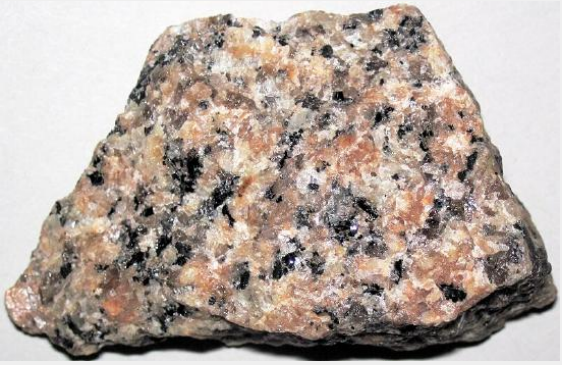
To identify a rock, match the texture (below) with the composition (right)

Texture	Composition: Felsic • >10% quartz • >50% feldspar • <15% mafic minerals	Intermediate • >10% quartz • >50% plagioclase • >10% orthoclase • <50% mafic minerals	Mafic • 20–85% plagioclase • 15–50% pyroxene • >35% olivine	Ultramafic Olivine & pyroxene
Phaneritic Coarse grained	**Granite**	**Diorite**	**Gabbro**	**Peridotite**
Aphanitic Fine grained	**Rhyolite**	**Andesite**	**Basalt**	X
Porphyritic Large crystals in a fine matrix	**Porphyritic rhyolite**	**Porphyritic andesite**	**Porphyritic basalt**	X
Vesicular Bubbly or frothy	**Pumice**	X	**Scoria**	X
Glassy	**Obsidian**	X	X	X
Pyroclastic Fragmental	**Rhyolitic tuff or volcanic breccia**	**Andesitic tuff or volcanic breccia**	**Basaltic tuff or volcanic breccia**	X

Figure 4.1.10: Igneous rock classification table with composition as vertical columns and texture as horizontal rows. (By Kindred Grey; CC BY-NC-SA 4.0 via Introduction to Earth Science.)

Aphanitic/Phaneritic Rock Types with Images

Felsic Composition



Granite

Granite is a coarse-crystalline felsic intrusive rock. The presence of quartz is a good indicator of granite. Granite commonly has large amounts of salmon pink potassium feldspar and white plagioclase crystals that have visible cleavage planes. Granite is a good approximation for the continental crust, both in density and composition.



Rhyolite

Rhyolite is a fine-crystalline felsic extrusive rock. Rhyolite is commonly pink and will often have glassy quartz phenocrysts. Because felsic lavas are thick and less mobile, rhyolite is less common than granite since the magma may cool and crystallize before reaching the surface and erupting. Examples of rhyolite include several lava flows in Yellowstone National Park and the altered rhyolite that makes up the Grand Canyon of the Yellowstone.

Intermediate Composition



Diorite

Diorite is a coarse-crystalline intermediate intrusive igneous rock. Diorite is identifiable by its Dalmatian-like appearance of black hornblende and biotite and white plagioclase feldspar. It is found in the Andes Mountains as well as the Henry and Abajo mountains of Utah.



Andesite

Andesite is a fine crystalline intermediate extrusive rock. It is commonly gray and porphyritic. It can be found in the Andes Mountains and in some island arcs. It is the fine-grained compositional equivalent of diorite.

Mafic Composition



Gabbro



Vesicular Basalt

Gabbro is a coarse-grained mafic igneous rock, made with mainly mafic minerals like pyroxene and only minor plagioclase. Because mafic lava is more mobile and can more easily reach the surface and erupt (instead of cooling below the surface), gabbro is less common than basalt. Gabbro is a major component of the lower oceanic crust.	Basalt is a fine-grained mafic igneous rock. It is commonly vesicular and aphanitic. When porphyritic, it often has either olivine or plagioclase phenocrysts. Basalt is the main rock which is formed at mid-ocean ridges, and is therefore the most common rock on the Earth's surface, making up the entirety of the ocean floor (except where covered by sediment).

Igneous Rock Bodies

Igneous rocks are common in the geologic record, but surprisingly, it is the intrusive rocks that are more common. Extrusive rocks, because of their small crystals and glass, are less durable. Plus, they are, by definition, exposed to the elements of erosion immediately. Intrusive rocks, forming underground with larger, stronger crystals, are more likely to last. Therefore, most landforms and rock groups that owe their origin to igneous rocks were formed as intrusive bodies. A significant exception to this is active volcanoes, which are discussed in a later section on volcanism. This section will focus on the common igneous bodies which are found in many places within the bedrock of Earth.

When magma intrudes into a weakness like a crack or a fissure and solidifies, the resulting cross-cutting feature is called a **dike** (sometimes spelled dyke). Because of this, dikes are often vertical or at an angle relative to the pre-existing rock layers that they intersect. Dikes are therefore **discordant intrusions**, not following any layering that was present. Dikes are important to geologists, not only for the study of igneous rocks themselves but also for dating rock sequences and interpreting the geologic history of an area. The dike is younger than the rocks it cuts across and may be used to assign actual numeric ages to sedimentary sequences, which are notoriously difficult to age date.



Figure 4.1.11: Dike of olivine gabbro cuts across Baffin Island in the Canadian Arctic.

Sills are another type of intrusive structure. A sill is a **concordant intrusion** that runs parallel to the sedimentary layers in the country rock. They are formed when magma exploits a weakness between these layers, shouldering them apart and squeezing

between them. As with dikes, sills are younger than the surrounding layers and may be radioactively dated to study the age of sedimentary strata.



Figure 4.1.12: Igneous sill intruding between Paleozoic strata in Nova Scotia.

A **magma chamber** is a large underground reservoir of molten rock. A body of rising magma is called a **diapir**. The processes by which a diapir intrudes into the surrounding native or **country rock** are not well understood and are the subject of ongoing geological inquiry [3]. For example, it is not known what happens to the pre-existing country rock as the diapir intrudes. One theory is the overriding rock gets shouldered aside, displaced by the increased volume of magma. Another is the native rock is melted and consumed into the rising magma or broken into pieces that settle into the magma, a process known as **stoping**. It has also been proposed that diapirs are not a real phenomenon, but just a series of dikes that blend into each other. The dikes may be intruding over millions of years, but since they may be made of similar material, they would be appearing to be formed at the same time. Regardless, when a diapir cools, it forms a mass of intrusive rock called a **pluton**. Plutons can have irregular shapes, but can often be somewhat round.



Figure 4.1.13: Butte Quartz Monzonite in the Boulder Batholith in Montana. (By James St. John; CC BY 2.0 via Wikimedia Commons.)

When many plutons merge together in an extensive single feature, it is called a **batholith**. Batholiths are found in the cores of many mountain ranges, including the granite formations of Yosemite National Park in the Sierra Nevada of California. They are typically more than 100 km^2 in area, associated with subduction zones, and mostly felsic in composition. A **stock** is a type of pluton with less surface exposure than a batholith and may represent a narrower neck of material emerging from the top of a batholith. Batholiths and stocks are discordant intrusions that cut across and through surrounding country rock.

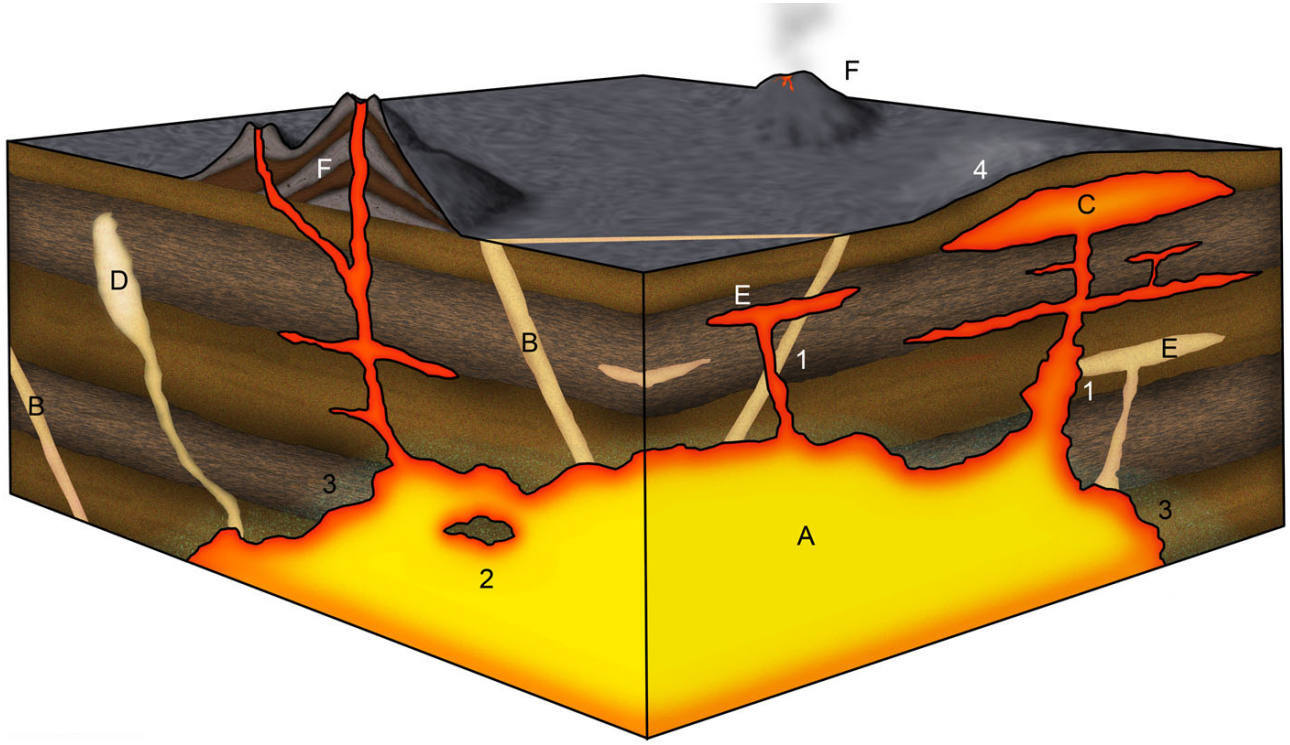


Figure 4.1.14: Half Dome in Yosemite National Park, California, is a part of the Sierra Nevada batholith which is mostly made of granite.

Laccoliths are blister-like, concordant intrusions of magma that form between sedimentary layers. The Henry Mountains of Utah are a famous topographic landform formed by this process. Laccoliths bulge upwards; a similar downward-bulging intrusion is called a **lopolith**.



Figure 4.1.15: Left: A laccolith forms as a blister in between sedimentary strata. Right: The Henry Mountains in Utah are interpreted to be a laccolith, exposed by erosion of the overlying layers.



Guide for Igneous Structures Image (shown above)

Number/Letter	Description
1	Young, emerging subvolcanic intrusion cutting through older one
2	**Xenolith** (solid rock of high melting temperature which has been transported within the magma from deep below) or **roof pendant** (fragment of the roof of the magma chamber that has detached from the roof and sunk into the melt)
3	**Contact metamorphism** in the country rock adjacent to the magma chamber (caused by the heat of the magma)
4	Uplift at the surface due to laccolith emplacement in the near sub-ground
A	Active magma chamber (called pluton when cooled and entirely crystallized; a batholith is a large rock body composed of several plutonic intrusions)
B	Old magmatic dikes
C	Emerging laccolith
D	Old pegmatite (late-magmatic dike formed by aggressive and highly mobile residual melts of a magma chamber)
E	Old and emerging magmatic sills
F	Stratovolcano

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3. Glazner, A. F., Bartley, J. M., Coleman, D. S., Gray, W. & Taylor, R. Z. Are plutons assembled over

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4.2: Bowen's Reaction Series

Bowen's Reaction Series describes the temperature at which minerals crystallize when cooled, or melt when heated. The low end of the temperature scale where all minerals crystallize into solid rock is approximately 700°C (1292°F). The upper end of the range where all minerals exist in a molten state is approximately 1250°C (2282°F) [4]. These numbers reference minerals that crystallize at standard sea-level pressure, 1 bar. The values will be different for minerals located deep below the Earth's surface due to the increased pressure, which affects crystallization and melting temperatures. However, the order and relationships are maintained.

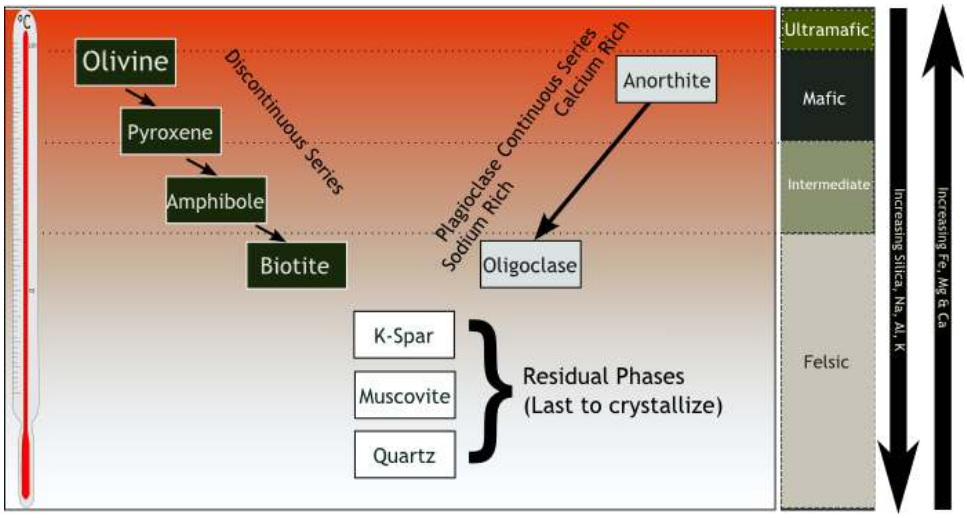


Figure 4.2.1: Bowen's Reaction Series. Minerals that crystallize at higher temperatures are at the top (olivine) and minerals that crystallize at lower temperatures are at the bottom (quartz). (Source Colivine, modified from Bowen, 1922)

In the figure, the righthand column lists the four groups of igneous rock from top to bottom: ultramafic, mafic, intermediate, and felsic. The down-pointing arrow on the far right shows increasing amounts of silica, sodium, aluminum, and potassium as the mineral composition goes from ultramafic to felsic. The up-pointing arrow shows increasing ferromagnesian components, specifically iron, magnesium, and calcium. To the far left of the diagram is a temperature scale. Minerals near the top of the diagram, such as olivine and anorthite (a type of plagioclase), crystallize at higher temperatures. Minerals near the bottom, such as quartz and muscovite, crystallize at lower temperatures.



Figure 4.2.2: Olivine, the first mineral to crystallize in a melt.

The most important aspect of Bowen's Reaction Series is to notice the relationships between minerals and temperature. Norman L. Bowen (1887-1956) was an early 20th century geologist who studied igneous rocks. He noticed that in igneous rocks, certain minerals always occur together and these mineral assemblages exclude other minerals. Curious as to why, and with the hypothesis in mind that it had to do with the temperature at which the rocks cooled, he set about conducting experiments on igneous rocks in the early 1900s. He conducted experiments on igneous rocks—grinding combinations of rocks into powder, sealing the powders into metal capsules, heating them to various temperatures, and then cooling them.



Figure 4.2.3: Normal L. Bowen

When he opened the quenched capsules, he found a glass surrounding mineral crystals that he could identify under his petrographic microscope. The results of many of these experiments, conducted at different temperatures over a period of several years, showed that the common igneous minerals crystallize from magma at different temperatures. He also saw that minerals occur together in rocks with others that crystallize within similar temperature ranges, and never crystallize with other minerals. This relationship can explain the main difference between mafic and felsic igneous rocks. Mafic igneous rocks contain more mafic minerals, and therefore, crystallize at higher temperatures than felsic igneous rocks. This is even seen in lava flows, with felsic lavas erupting hundreds of degrees cooler than their mafic counterparts. Bowen's work laid the foundation for understanding igneous **petrology** (the study of rocks) and resulted in his book, *The Evolution of the Igneous Rocks* in 1928 [5].



Figure 4.2.4: Norman L. Bowen working with his petrographic microscope.

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4.3: Magma Generation

Magma and lava contain three components: melt, solids, and volatiles. The **melt** is made of ions from minerals that have liquefied. The **solids** are made of crystallized minerals floating in the liquid melt. These may be minerals that have already cooled. **Volatiles** are gaseous components—such as water vapor, carbon dioxide, sulfur, and chlorine—dissolved in the magma [6]. The presence and amount of these three components affect the physical behavior of the magma and will be discussed more below.

Geothermal Gradient

Below the surface, the temperature of the Earth rises. This heat is caused by residual heat left from the formation of Earth and ongoing radioactive decay. The rate at which temperature increases with depth is called the **geothermal gradient**. The average geothermal gradient in the upper 100 km (62 mi) of the crust is about 25°C per kilometer of depth. So for every kilometer of depth, the temperature increases by about 25°C.

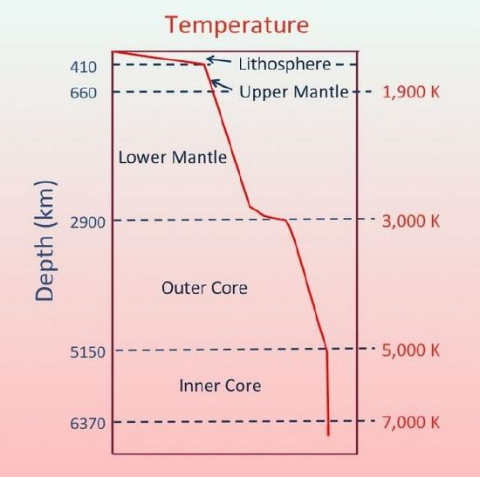


Figure 4.3.1: Geothermal gradient.

The depth-temperature graph (see figure below) illustrates the relationship between the geothermal gradient (geotherm, red line) and the start of rock melting (**solidus**, green line). The geothermal gradient changes with depth (which has a direct relationship to pressure) through the crust into the upper mantle. The area to the left of the green line includes solid components; to the right is where liquid components start to form. The increasing temperature with depth makes the depth of about 125 kilometers (78 miles) where the natural geothermal gradient is closest to the solidus.

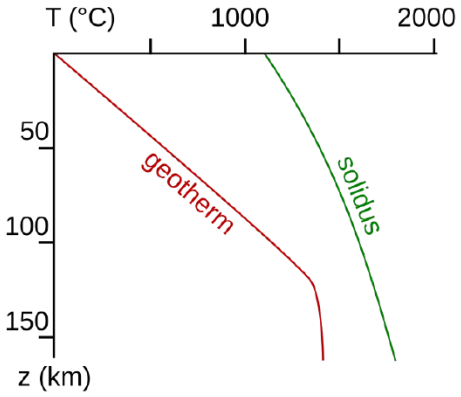


Figure 4.3.2: Pressure-temperature diagram showing temperature in degrees Celsius on the x-axis and depth below the surface in kilometers (km) on the y-axis. The red line is the geothermal gradient and the green solidus line represents the temperature and pressure regime at which melting begins. Rocks at pressures and temperatures left of the green line are solid. If pressure/temperature conditions change so that rocks pass to the right of the green line, then they will start to melt. (Source: Woudloper)

The temperature at 100 km (62 mi) deep is about 1200°C (2192°F). At the bottom of the crust, 35 km (22 mi) deep, the pressure is about 10,000 bars [7]. A **bar** is a measure of pressure, with 1 bar being normal atmospheric pressure at sea level. At these pressures and temperatures, the crust and mantle are solid. To a depth of 150 km (93 mi), the geothermal gradient line stays to the left of the solidus line. This relationship continues through the mantle to the core-mantle boundary, at 2880 km (1790 mi).

The solidus line slopes to the right because the melting temperature of any substance depends on pressure. The higher pressure created at greater depth increases the temperature needed to melt rock. In another example, at sea level with an atmospheric pressure close to 1 bar, water boils at 100°C. But if the pressure is lowered, as shown in the video, water boils at a much lower temperature.





There are three principal ways rock behavior crosses to the right of the green solidus line to create molten magma: 1) **decompression melting** caused by lowering the pressure, 2) **flux melting** caused by adding volatiles, and 3) heat-induced melting caused by increasing the temperature. The Bowen's Reaction Series shows that minerals melt at different temperatures. Since magma is a mixture of different minerals, the solidus boundary is more of a fuzzy zone rather than a well-defined line; some minerals are melted and some remain solid. This type of rock behavior is called **partial melting** and represents real-world magmas, which typically contain solid, liquid, and volatile components.

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4.4: Volcanism

When magma emerges onto the Earth's surface, the molten rock is called lava. A **volcano** is a type of land formation created when lava solidifies into rock. Volcanoes have been an important part of human society for centuries, and studying them has greatly increased as our understanding of plate tectonics and has made them less mysterious. The next sections describe volcano location, type, hazards, and monitoring.

Distribution and Tectonics

Most volcanoes are interplate volcanoes. **Interplate volcanoes** are located at active plate boundaries created by volcanism at mid-ocean ridges, subduction zones, and continental rifts. The prefix "*inter-*" means between, such as between tectonic plates. Some volcanoes are **intraplate volcanoes**. The prefix "*intra-*" means within, and intraplate volcanoes are located within tectonic plates, far removed from plate boundaries. Many intraplate volcanoes are formed by hotspots.

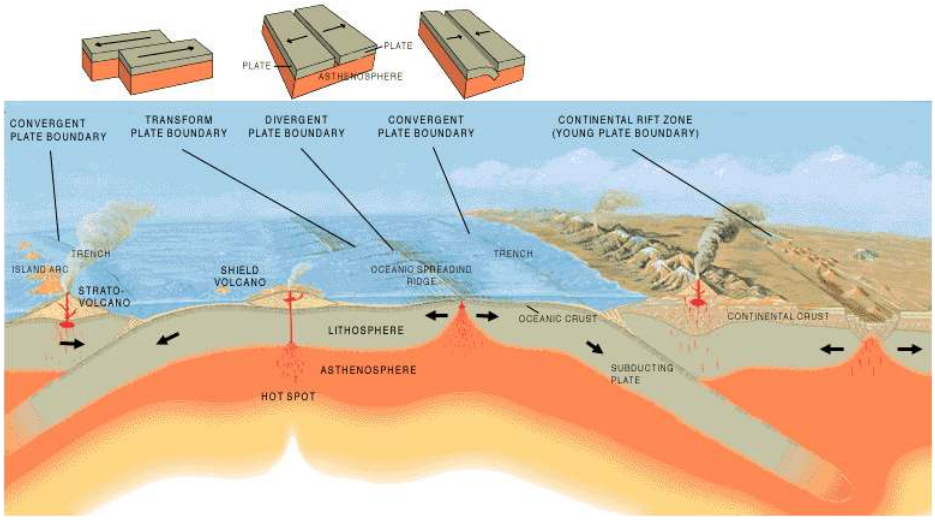


Figure 4.4.1: Association of volcanoes with plate boundaries

Volcanoes at Mid-Ocean Ridges

Most volcanism on Earth occurs on the ocean floor along **mid-ocean ridges**, a type of divergent plate boundary. These interplate volcanoes are also the least observed and famous since most of them are located under 3000-4500 m (10,000-15,000 ft) of the ocean and the eruptions are slow, gentle, and oozing. One exception is the interplate volcanoes of Iceland. The diverging and thinning oceanic plates allow hot mantle rock to rise, releasing pressure and causing decompression melting. Ultramafic mantle rock, consisting largely of peridotite, partially melts and generates magma that is basaltic. Because of this, almost all volcanoes on the ocean floor are basaltic. In fact, most oceanic lithosphere is basaltic near the surface, with phaneritic gabbro and ultramafic peridotite underneath [10].

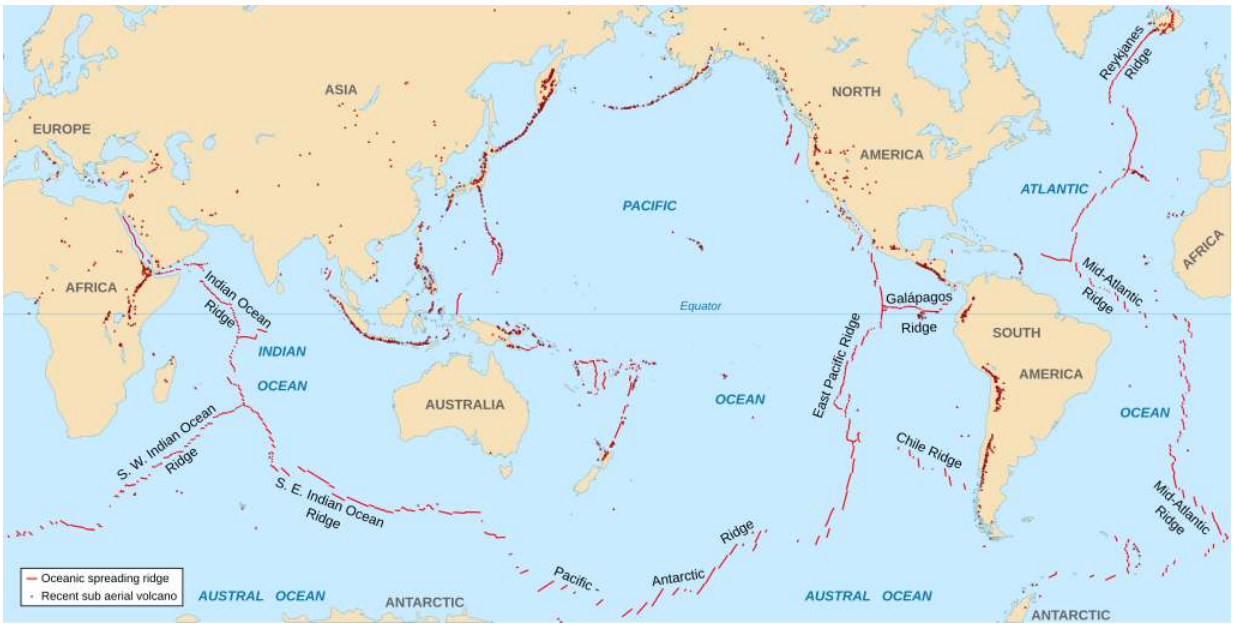


Figure 4.4.2: Map of mid-ocean ridges throughout the world.

When basaltic lava erupts underwater it emerges in small explosions and/or forms pillow-shaped structures called **pillow basalts**. These seafloor eruptions enable entire underwater ecosystems to thrive in the deep ocean around mid-ocean ridges. This ecosystem exists around tall vents emitting black, hot mineral-rich water called deep-sea **hydrothermal vents**, also known as **black smokers**.

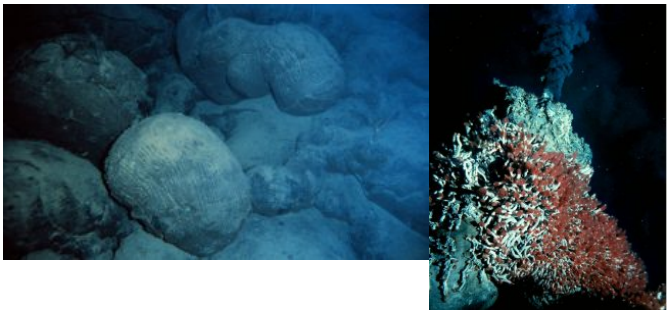


Figure 4.4.3: Left: Pillow basalt on the seafloor near Hawaii. Right: Black smoker hydrothermal vent with a colony of giant (6 feet) tube worms.

Without sunlight to support photosynthesis, these organisms instead utilize a process called **chemosynthesis**. Certain bacteria are able to turn hydrogen sulfide (H_2S), a gas that smells like rotten eggs, into life-supporting nutrients and water. Larger organisms may eat these bacteria or absorb nutrients and water produced by bacteria living symbiotically inside their bodies [11].

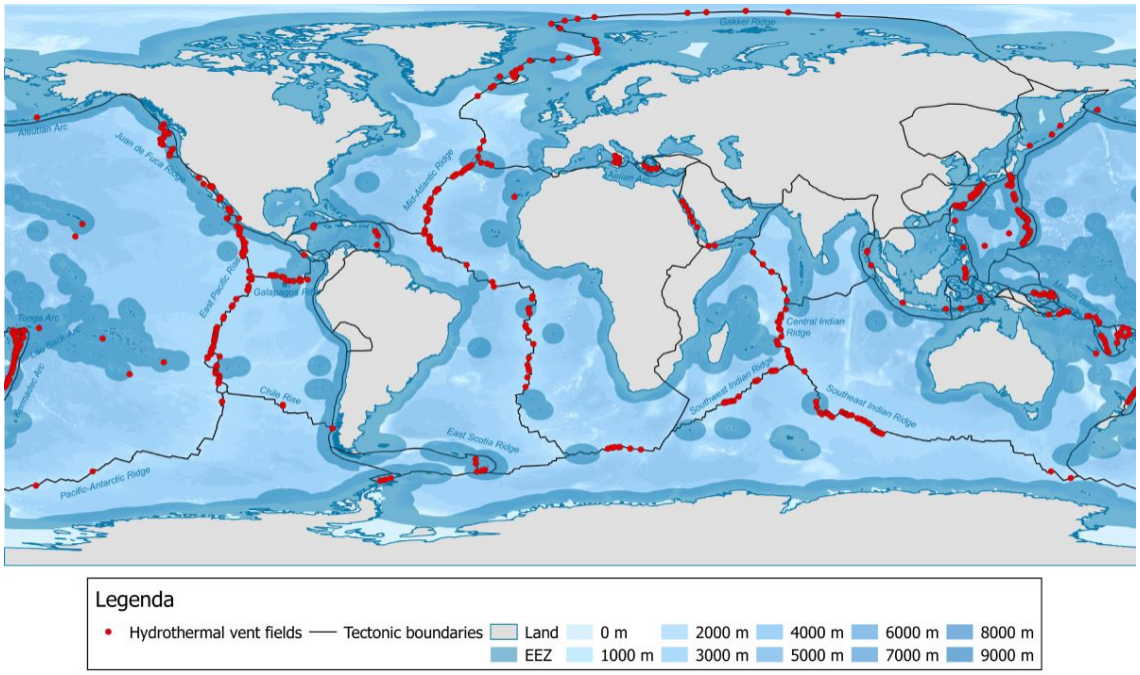
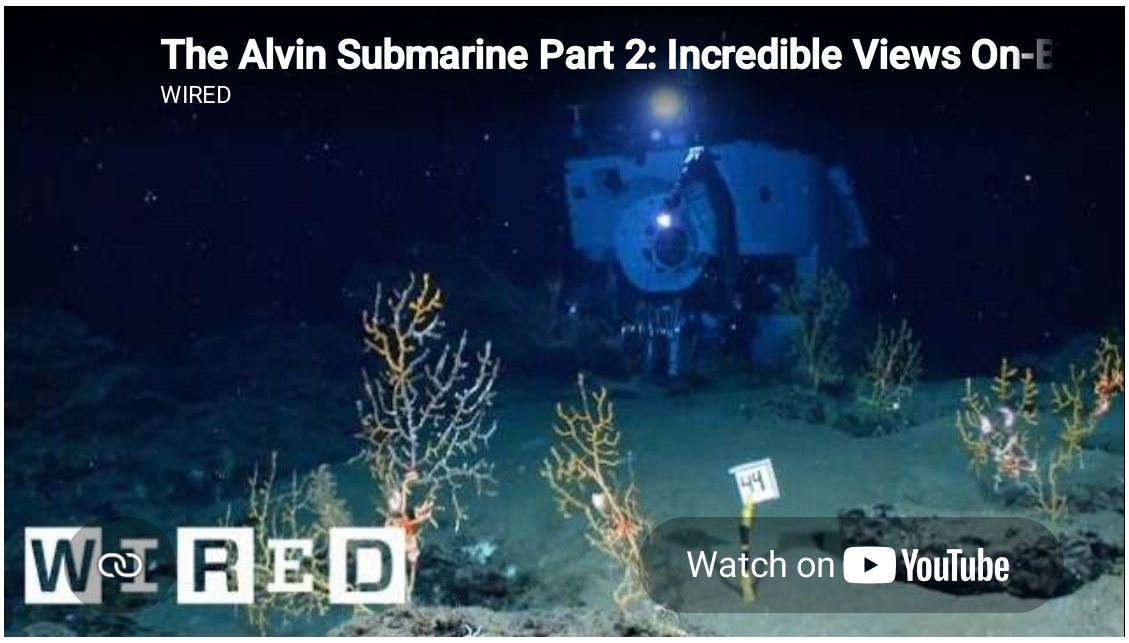


Figure 4.4.4: Distribution of hydrothermal vent fields.

The video shows some of the ecosystems found around deep-sea hydrothermal vents.





Volcanoes at Subduction Zones

The second most commonly found location for volcanism is adjacent to **subduction zones**, a type of convergent plate boundary. The process of subduction expels water from hydrated minerals in the descending slab, which causes flux melting in the overlying mantle rock. Because subduction volcanism occurs in a volcanic arc, the thickened crust promotes partial melting and magma differentiation. These evolve the mafic magma from the mantle into more silica-rich magma. The **Ring of Fire** surrounding the Pacific Ocean, for example, is dominated by subduction-generated eruptions of mostly silica-rich lava; the volcanoes and plutons consist largely of intermediate-to-felsic rock such as andesite, rhyolite, pumice, and tuff.

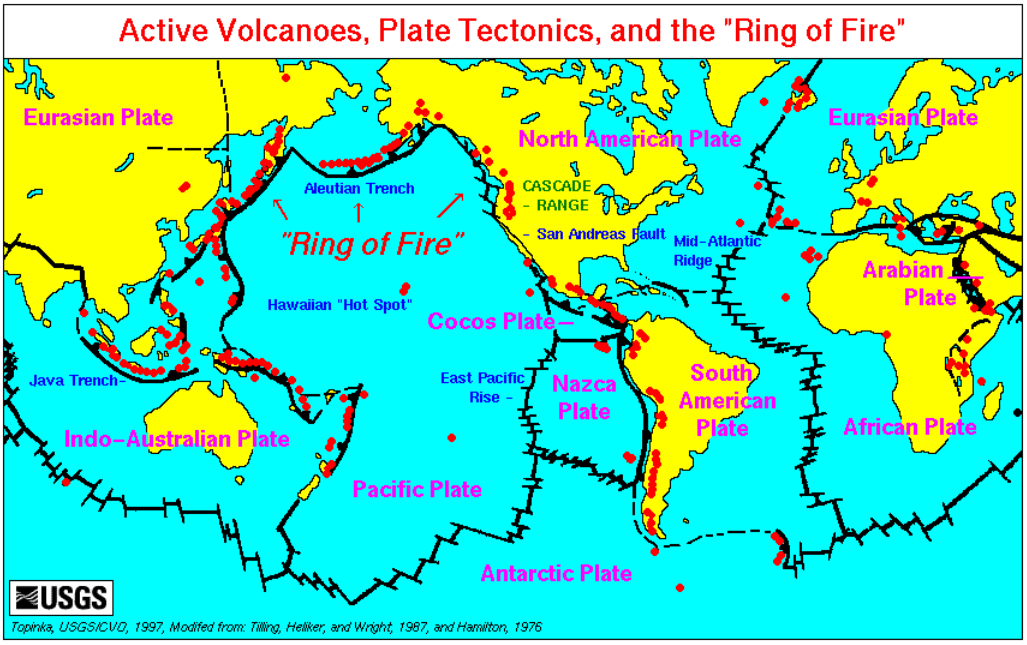


Figure 4.4.5: Distribution of volcanoes on the planet. Click here for an interactive map of volcano distributions.

Volcanoes at Continental Rifts

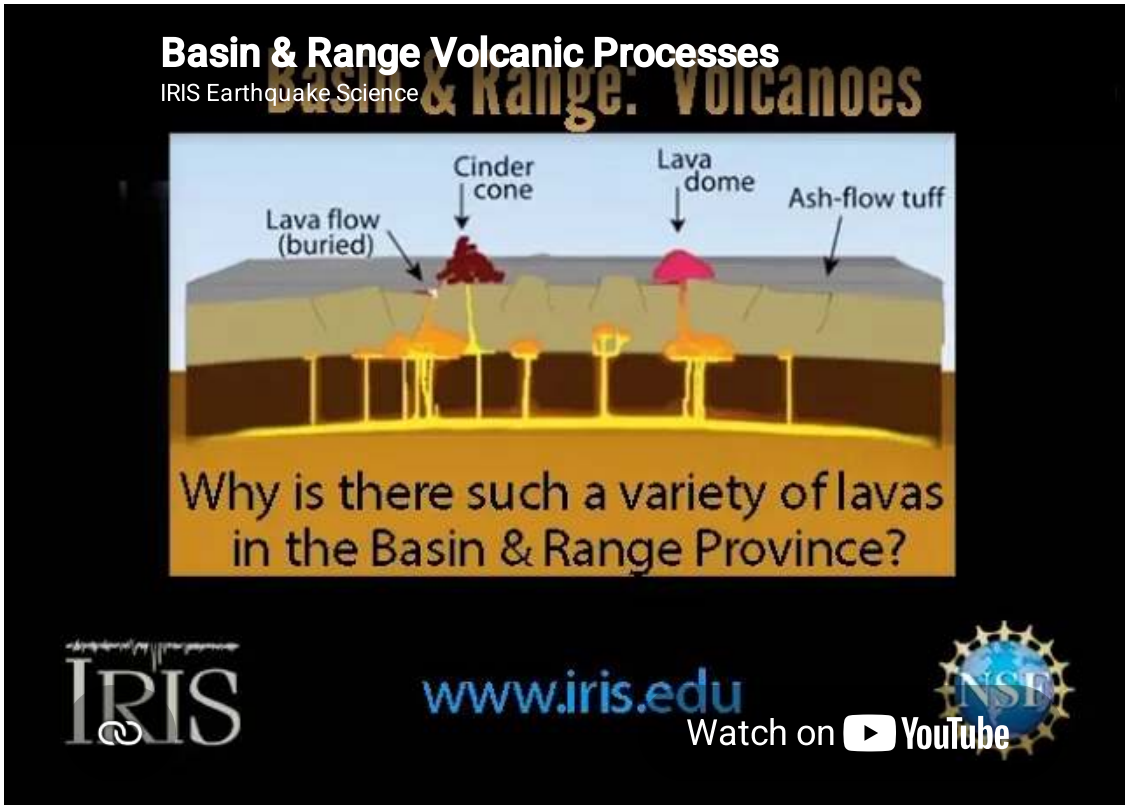
Some volcanoes are created at **continental rifts**, where crustal thinning is caused by diverging lithospheric plates, such as the East African Rift Basin in Africa. Volcanism caused by crustal thinning without continental rifting is found in the Basin and Range Province in North America. In this location, volcanic activity is produced by rising magma that stretches the overlying crust. Lower crust or upper mantle material rises through the thinned crust, releases pressure, and undergoes decompression-induced partial melting. This magma is less dense than the surrounding rock and continues to rise through the crust to the surface, erupting as basaltic lava. These eruptions usually result in flood basalts, cinder cones, and basaltic lava flows (see video). Relatively young cinder cones of basaltic lava can be found in south-central Utah, in the Black Rock Desert Volcanic Field, which is part of the zone

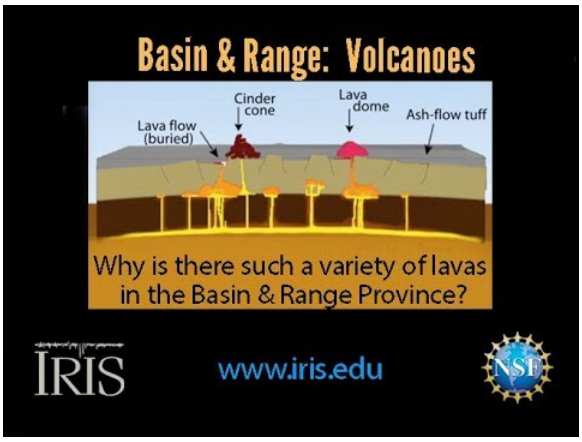
of Basin and Range crustal extension. These Utah cinder cones and lava flows started erupting around 6 million years ago, with the last eruption occurring 720 years ago [12].



Figure 4.4.6: Basaltic cinder cones of the Black Rock Desert near Beaver, Utah.

The video shows the Basin and Range volcanic processes.





Hotspots

Hotspots are the main source of intraplate volcanism. **Hotspots** occur when lithospheric plates glide over a hot **mantle plume**, an ascending column of solid heated rock originating from deep within the mantle. The mantle plume generates melts as material rises, with the magma rising even more. When the ascending magma reaches the lithospheric crust, it spreads out into a mushroom-shaped head that is tens to hundreds of kilometers across.

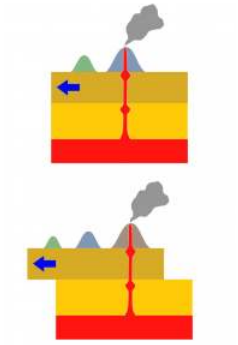


Figure 4.4.7: Diagram showing a non-moving source of magma (mantle plume) and a moving overriding plate.

Since most mantle plumes are located beneath the oceanic lithosphere, the early stages of intraplate volcanism typically take place underwater. Over time, basaltic volcanoes may build up from the seafloor into islands, such as the Hawaiian Islands [13]. Where a hotspot is found under a continental plate, contact with the hot mafic magma may cause the overlying felsic rock to melt and mix with the mafic material below, forming intermediate magma. Or the felsic magma may continue to rise, and cool into a granitic batholith or erupt as a felsic volcano. The Yellowstone caldera is an example of hotspot volcanism that resulted in an explosive eruption.

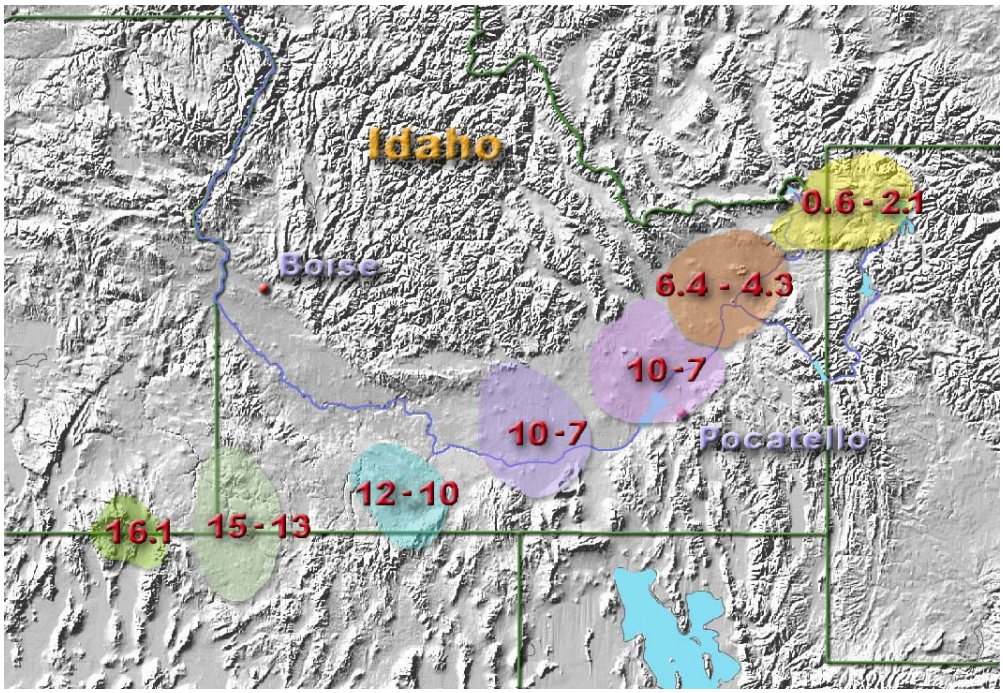


Figure 4.4.8: The track of the Yellowstone hotspot, which shows the age of different eruptions millions of years ago.

A zone of actively erupting volcanism connected to a chain of extinct volcanoes indicates intraplate volcanism located over a hotspot. These volcanic chains are created by the overriding oceanic plate slowly moving over a hotspot mantle plume. These chains are seen on the seafloor and continents and include volcanoes that have been inactive for millions of years. The Hawaiian Islands on the Pacific oceanic plate are the active end of a long volcanic chain that extends from the northwest Pacific Ocean to the Emperor Seamounts, all the way to the subduction zone beneath the Kamchatka Peninsula. The adjacent overriding North American continental plate moved across a mantle plume hotspot for several million years, creating a chain of volcanic calderas that extends from Southwestern Idaho to the presently active Yellowstone caldera in Wyoming.

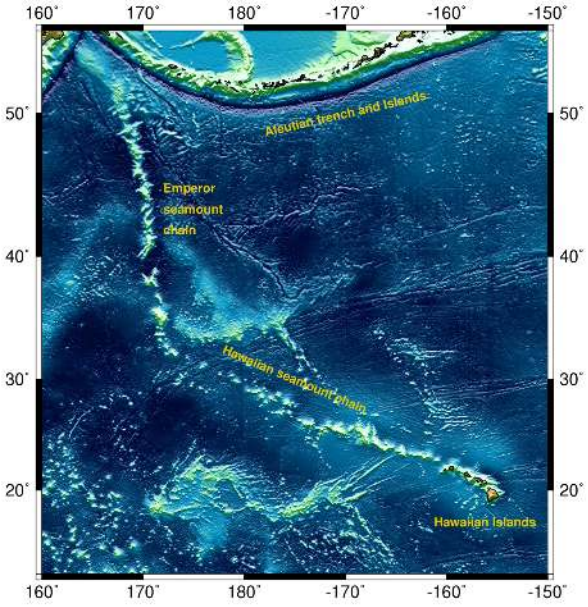
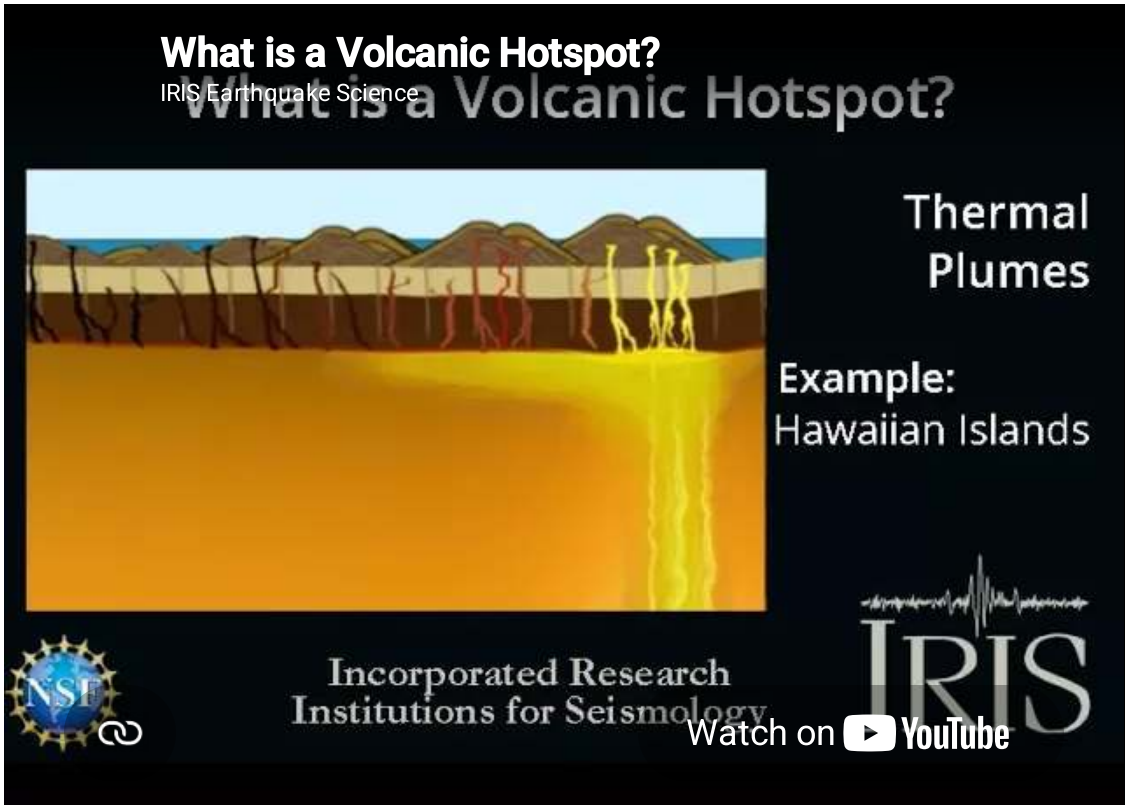
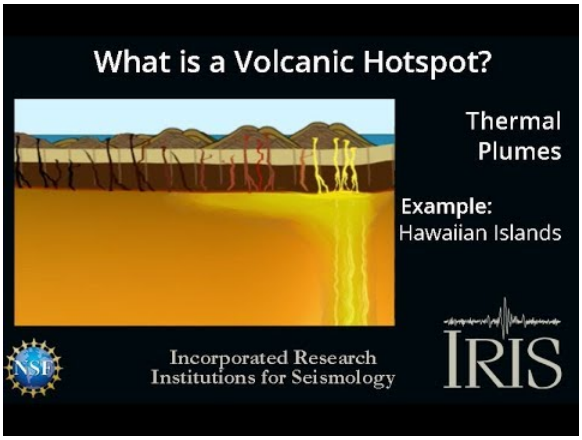


Figure 4.4.9: The Hawaii-Emperor seamount and island chain.

The three-minute video illustrates hotspot volcanoes.





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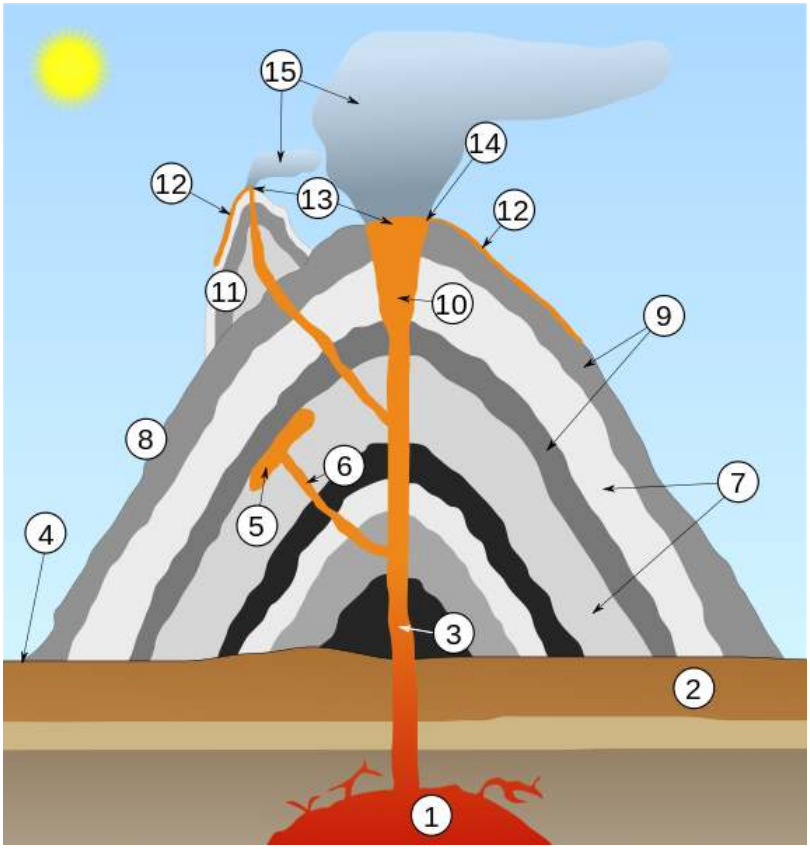
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4.5: Volcanic Structures

Volcano Features and Types

There are several different types of volcanoes based on their shape, eruption style, magmatic composition, and other aspects.



The figure shows the main features of a typical stratovolcano:

1. magma chamber
2. bedrock or upper layers of lithosphere
3. the **conduit** or narrow pipe through which the lava erupts
4. the **base** or edge of the volcano
5. a sill of magma between layers of the volcano
6. a **branch pipe** or feeder tube to the sill
7. layers of tephra (ash) from previous eruptions
8. **flank** (side) of the volcano
9. layers of lava erupted from the vent that flowed down the sides of the volcano
10. the **throat** of the volcano
11. a **parasitic cone,** a small volcano located on the flank of a larger volcano such as Shastina on Mount Shasta (Kilauea sitting on the flank of Mauna Loa is not considered a parasitic cone because it has its own separate magma chamber [13].)
12. layers of lava on the main summit and parasitic cone
13. the **vents** of the parasitic cone and the main volcano
14. the **crater**
15. clouds of ash blown into the sky by the eruption; this settles back onto the volcano and surrounding land.



Figure 4.5.1: Mt. Shasta in California with Shastina, its parasitic cone

The largest craters are called calderas, such as the Crater Lake Caldera in Oregon.



Figure 4.5.2: Oregon's Crater Lake was formed about 7700 years ago after the eruption of Mount Mazama.

Many volcanic features are produced by **viscosity**, a basic property of lava. Viscosity is the resistance to flow by a fluid. Low viscosity magma flows easily more like syrup such as the basaltic volcanism that occurs in Hawaii on shield volcanoes. High viscosity magma is thick and sticky, typically felsic or intermediate, that flows slowly, similar to toothpaste.

Shield Volcano

The largest volcanoes are **shield volcanoes**. They are characterized by broad low-angle flanks, small vents at the top, and mafic magma chambers. The name comes from the side view, which resembles a medieval warrior's shield. They are typically associated with hotspots, mid-ocean ridges, or continental rifts with rising upper mantle material. The low-angle flanks are built up slowly from numerous low-viscosity basaltic lava flows that spread out over long distances. The basaltic lava erupts **effusively**, meaning the eruptions are small, localized, and predictable.



Figure 4.5.3: Mauna Kea, a shield volcano in Hawaii. (By Nula666; CC BY-SA 3.0 via Wikimedia Commons.)

Typically, shield volcano eruptions are not much of a hazard to human life—although non-explosive eruptions of Kilauea (Hawaii) in 2018 produced uncharacteristically large lavas that damaged roads and structures. Mauna Loa (see USGS page) and Kilauea (see USGS page) in Hawaii are examples of shield volcanoes. Shield volcanoes are also found in Iceland, the Galapagos Islands, Northern California, Oregon, and the East African Rift [14].



Figure 4.5.4: Eruption of Kilauea in 2018 produced relatively high viscosity a'a lava shown here crossing and destroying a road. This eruption caused much property damage in Hawaii.

The largest volcanic edifice in the solar system is Olympus Mons on Mars. This (possibly extinct) shield volcano covers an area the size of the state of Arizona. This may indicate the volcano erupted over a hotspot for millions of years, which means Mars had little, if any, plate tectonic activity. [15] [16].

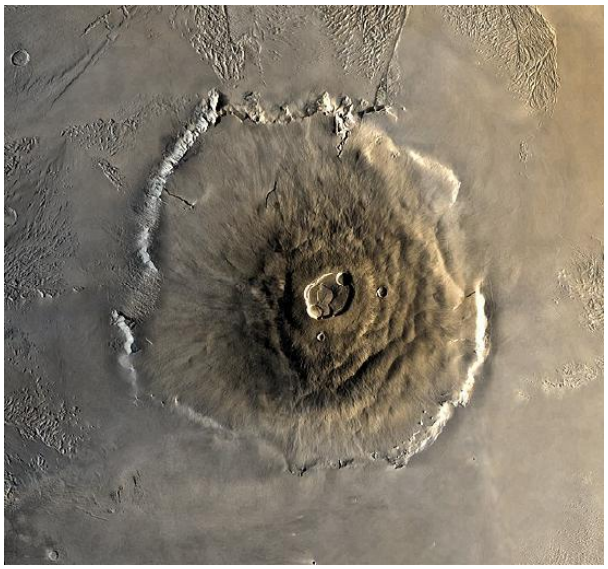


Figure 4.5.5: Olympus Mons, an enormous shield volcano on Mars, the largest volcano in the solar system.

Basaltic lava forms special landforms based on magma temperature, composition, and content of dissolved gases and water vapor. The two main types of basaltic volcanic rock have Hawaiian names—*pahoehoe* and *aa*. **Pahoehoe** might come from low-viscosity lava that flows easily into ropey strands.



Figure 4.5.6: Ropey pahoehoe lava.

Aa (sometimes spelled a'a or ʻaʻā and pronounced "ah-ah") is more viscous and has a crumbly blocky appearance [17]. The exact details of what forms the two types of flows are still up for debate. Felsic lavas have lower temperatures and more silica and thus are higher viscosity. These also form aa-style flows.



Figure 4.5.7: Blocky a'a lava. (By Librex; CC BY 2.0 via Wikimedia Commons.)

Low-viscosity, fast-flowing basaltic lava tends to harden on the outside into a tube and continue to flow internally. Once lava flow subsides, the empty outer shell may be left as a **lava tube**. Lava tubes, with or without collapsed roofs, make famous caves in Hawaii, Northern California, the Columbia River Basalt Plateau of Washington and Oregon, El Malpais National Monument in New Mexico, and Craters of the Moon National Monument in Idaho.

Fissures are cracks that commonly originate from shield-style eruptions. Lava emerging from fissures is typically mafic and very fluid. The 2018 Kilauea eruption included fissures associated with the lava flows. Some fissures are caused by volcanic seismic activity rather than lava flows. Some fissures are influenced by plate tectonics, such as the common fissures located parallel to the divergent boundary in Iceland.



Figure 4.5.8: Volcanic fissure and flow, which could eventually form a lava tube. (By M. Patrick; public domain via USGS.)

Cooling lava can contract into columns with semi-hexagonal cross sections called **columnar jointing**. This feature forms the famous Devils Tower in Wyoming, possibly an ancient volcanic vent from which the surrounding layers of lava and ash have been removed by erosion. Another well-known exposed example of columnar jointing is the Giant's Causeway in Ireland.



Figure 4.5.9: Left: Devils Tower in Wyoming has columnar jointing. Right: Columnar jointing on Giant's Causeway in Ireland.

Stratovolcano

A **stratovolcano**, also called a **composite cone volcano**, has steep flanks, a symmetrical cone shape, a distinct crater, and rises prominently above the surrounding landscape. The term composite refers to the alternating layers of pyroclastic fragments like ash and bombs, and solidified lava flows of varying composition. Examples include Mount Rainier in Washington state and Mount Fuji in Japan.



Figure 4.5.10: Left: Mount Rainier towers over Tacoma, Washington. Right: Mt. Fuji in Japan, a typical stratovolcano: symmetrical, increasing slope, and a visible crater at the top.

Stratovolcanoes usually have felsic to intermediate magma chambers, but can even produce mafic lavas. Stratovolcanoes have viscous lava flows and domes, punctuated by explosive eruptions. This produces volcanoes with steep flanks [14].

Lava Domes

Lava domes are accumulations of silica-rich volcanic rock, such as rhyolite and obsidian. Too viscous to flow easily, the felsic lava tends to pile up near the vent in blocky masses. Lava domes often form in a vent within the collapsed crater of a stratovolcano and grow by internal expansion. As the dome expands, the outer surface cools, hardens, and shatters, and spills loose fragments down the sides. Mount St. Helens has a good example of a lava dome inside of a collapsed stratovolcano crater. Examples of stand-alone lava domes are Chaiten in Chile and Mammoth Mountain in California [18][14].

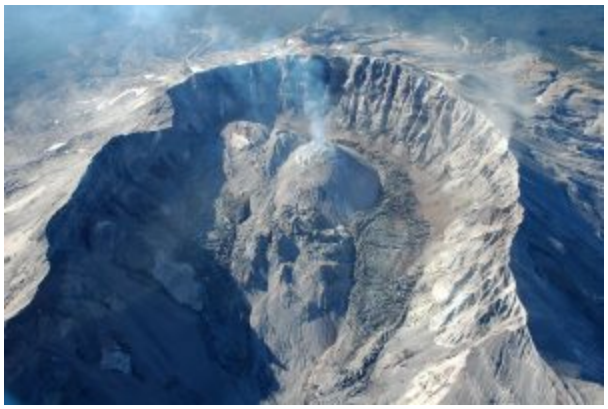


Figure 4.5.11: Lava domes have started the rebuilding process at Mount St. Helens, Washington.

Caldera

Calderas are steep-walled, basin-shaped depressions formed by the collapse of a volcanic edifice into an empty magma chamber. Calderas are generally very large, with diameters of up to 25 km (15.5 mi). The term caldera specifically refers to a volcanic vent but it is frequently used to describe a volcano type. Caldera volcanoes are typically formed by eruptions of high-viscosity felsic lava having high content of volatiles.

Crater Lake, Yellowstone, and the Long Valley Caldera are good examples of this type of volcanism. The caldera at Crater Lake National Park in Oregon was created about 6800 years ago when Mount Mazama, a composite volcano, erupted in a huge explosive blast. The volcano ejected large amounts of volcanic ash and rapidly drained the magma chamber, causing the top to collapse into a large depression that later filled with water. Wizard Island in the middle of the lake is a later resurgent lava dome that formed within the caldera basin [14].

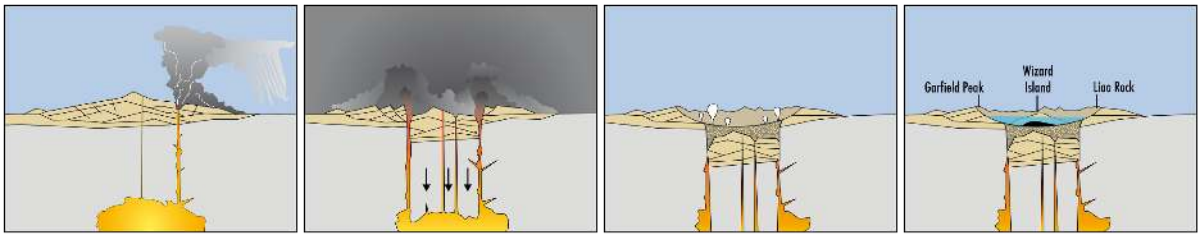


Figure 4.5.12: Timeline of events at Mount Mazama. 1. Eruptions of ash and pumice. The cataclysmic eruption started from a vent on the northeast side of the volcano as a towering column of ash, with pyroclastic flows spreading to the northeast. 2. Caldera collapse. As more magma was erupted, cracks opened up around the summit, which began to collapse. Fountains of pumice and ash surrounded the collapsing summit, and pyroclastic flows raced down all sides of the volcano. 3. Steam explosions. When the dust had settled, the new caldera was 5 miles (8 km) in diameter and 1 mile (1.6 km) deep. Ground water interacted with hot deposits causing explosions of steam and ash. 4. Today. In the first few hundred years after the cataclysmic eruption, renewed eruptions built Wizard Island, Merriam Cone, and the central platform. Water filled the new caldera to form the deepest lake in the United States. (By USGS and NPS; public domain via Wikimedia Commons.)



Figure 4.5.13: Wizard Island sits in the caldera at Crater Lake.

The Yellowstone volcanic system erupted three times in the recent geologic past—2.1, 1.3, and 0.64 million years ago—leaving behind three caldera basins. Each eruption created large rhyolite lava flows as well as pyroclastic flows that solidified into tuff formations. These extra-large eruptions rapidly emptied the magma chamber, causing the roof to collapse and form a caldera. The youngest of the three calderas contains most of Yellowstone National Park, as well as two resurgent lava domes. The calderas are difficult to see today due to the amount of time since their eruptions and subsequent erosion and glaciation.

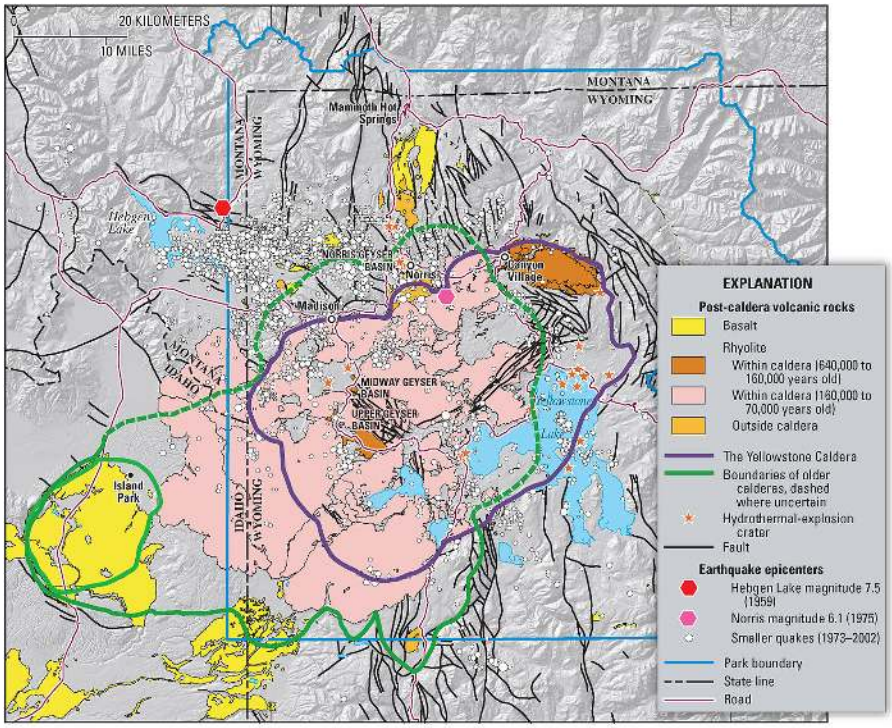


Figure 4.5.14: Map of calderas and related rocks around Yellowstone.

Yellowstone volcanism started about 17 million years ago as a hotspot under the North American lithospheric plate near the Oregon/Nevada border. As the plate moved to the southwest over the stationary hotspot, it left behind a track of past volcanic activities. Idaho's Snake River Plain was created from volcanism that produced a series of calderas and lava flows. The plate eventually arrived at its current location in northwestern Wyoming, where hotspot volcanism formed the Yellowstone calderas [19].



Figure 4.5.15: Several prominent ash beds found in North America, including three Yellowstone eruptions shaded pink (Mesa Falls, Huckleberry Ridge, and Lava Creek), the Bishop Tuff ash bed (brown dashed line), and the modern May 18, 1980 ash fall from Mt. St. Helens (yellow).

The Long Valley Caldera near Mammoth, California, is the result of a large volcanic eruption that occurred 760,000 years ago. The explosive eruption dumped enormous amounts of ash across the United States, in a manner similar to the Yellowstone eruptions. The Bishop Tuff deposit near Bishop, California, is made of ash from this eruption. The current caldera basin is 17 km by 32 km (10 mi by 20 mi), large enough to contain the town of Mammoth Lakes, major ski resort, airport, major highway, **resurgent dome**, and several hot springs [20].

Cinder Cone

Cinder cones are small volcanoes with steep sides and made of pyroclastic fragments that have been ejected from a pronounced central vent. The small fragments are called **cinders** and the largest are **volcanic bombs**. The eruptions are usually short-lived events, typically consisting of mafic lavas with a high content of volatiles. Hot lava is ejected into the air, cooling and solidifying into fragments that accumulate on the flank of the volcano. Cinder cones are found throughout western North America [14].



Figure 4.5.16: Sunset Crater, Arizona is a cinder cone.

A recent and striking example of a cinder cone is the eruption near the village of Parícutin, Mexico that started in 1943 [21]. The cinder cone started explosively shooting cinders out of the vent in the middle of a farmer's field. The volcanism quickly built up the cone to a height of over 90 m (300 ft) within a week and 365 m (1200 ft) within the first 8 months. After the initial explosive eruption of gases and cinders, basaltic lava poured out from the base of the cone. This is a common order of events for cinder cones: violent eruption, cone, and crater formation, low-viscosity lava flow from the base. The cinder cone is not strong enough to support a column of lava rising to the top of the crater, so the lava breaks through and emerges near the bottom of the volcano. During nine years of eruption activity, the ashfall covered about 260 km^2 (100 mi^2) and destroyed the nearby town of San Juan [14].

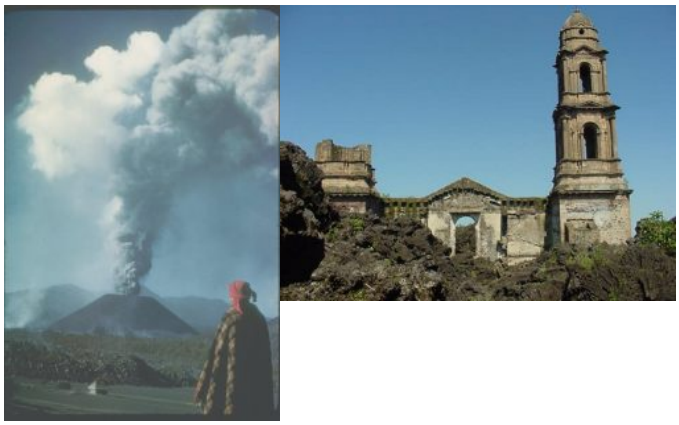


Figure 4.5.17: Left: Soon after the birth of Parícutin in 1943. Right: Lava from Parícutin covered the local church and destroyed the town of San Juan, Mexico.

Flood Basalts

A rare volcanic eruption type, unobserved in modern times, is the **flood basalt**. Flood basalts are some of the largest and lowest viscosity types of eruptions known. They are not known from any eruption in human history, so the exact mechanisms of eruption are still mysterious. Some famous examples include the Columbia River Flood Basalts in Washington, Oregon, and Idaho, the Deccan Traps, which cover about 1/3 of the country of India, and the Siberian Traps, which may have been involved in the Earth's largest mass extinction.

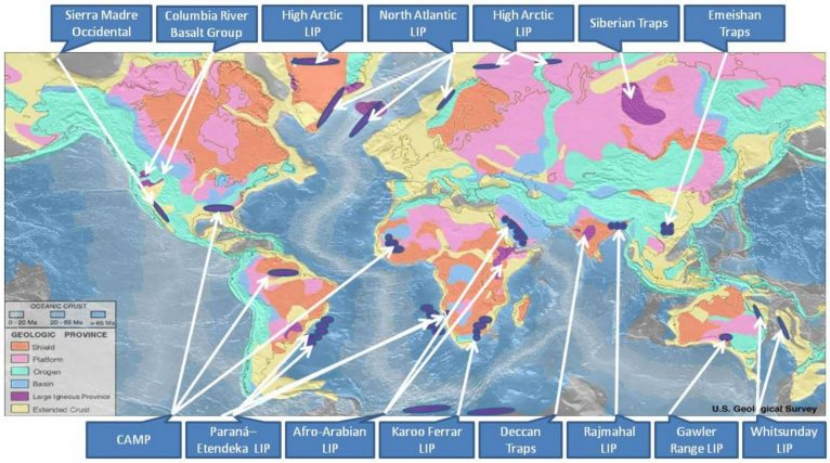


Figure 4.5.18: World map of flood basalts. Note the largest is the Siberian Traps.

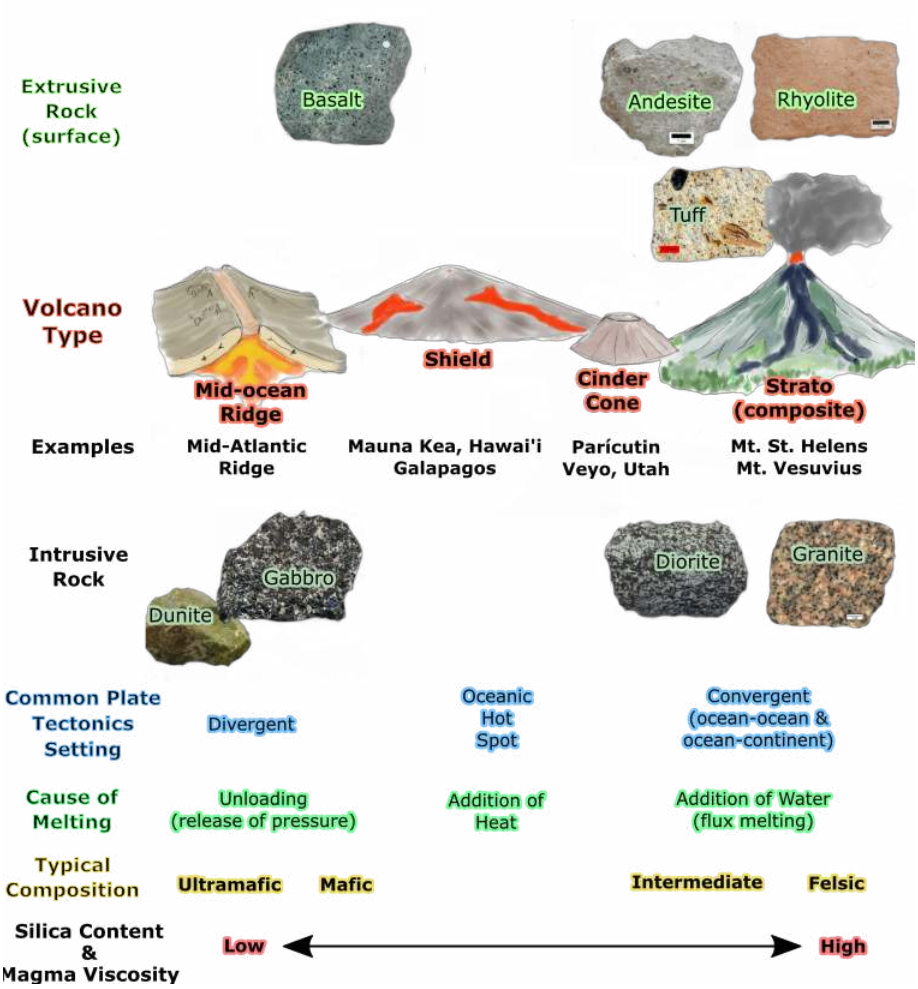
Carbonatites

Arguably the most unusual volcanic activity is a **carbonatite** eruption. Only one actively erupting carbonatite volcano exists on Earth today, Ol Doinyo Lengai, in the East African Rift Zone of Tanzania. While all other volcanism on Earth originates from silicate-based magma, carbonatites are a product of carbonate-based magma and produce volcanic rocks containing greater than 50% carbonate minerals. Carbonatite lavas are very low viscosity and relatively cold for lava. The erupting lava is black and solidifies to brown/grey rock that eventually turns white. These rocks are occasionally found in the geologic record and require special study to distinguish them from metamorphic marbles. They are mostly associated with continental rifting [22].



Figure 4.5.19: Crater of Ol Doinyo Lengai in 2011. Note the white carbonatite in the walls of the crater.

A summary of types of volcanoes with associated rock types and other characteristics is shown in the figure below.



4.5.20: Igneous rock types and related volcano types. Mid-ocean ridges and shield volcanoes represent more mafic compositions, and stratovolcanoes (composite volcanoes) generally represent a more intermediate or felsic composition and a convergent plate tectonic boundary. Note that there are exceptions to this generalized layout of volcano types and igneous rock composition.

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4.6: Volcanic Hazards and Monitoring

Volcanic Hazards

While the most obvious volcanic hazard is lava, the dangers posed by volcanoes go far beyond lava flows. For example, on May 18, 1980, Mount St. Helens (Washington, United States) erupted with an explosion and landslide that removed the upper 400 m (1300 ft) of the mountain. The initial explosion was immediately followed by a lateral blast, which produced a pyroclastic flow that covered nearly 600 km^2 (230 mi^2) of the forest with hot ash and debris [23]. The pyroclastic flow moved at speeds of 80-130 km per hour (50-80 mph), flattening trees and ejecting clouds of ash into the air.

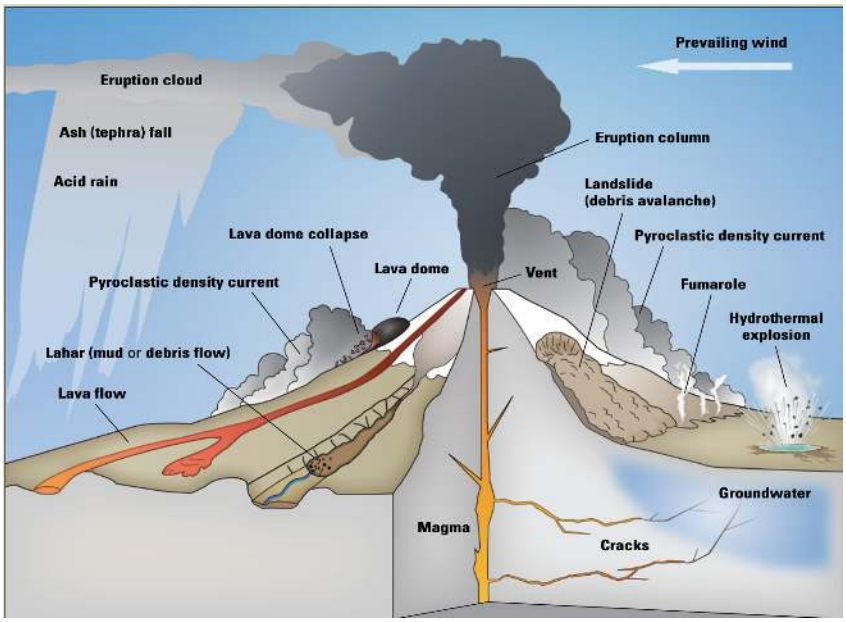


Figure 4.6.1: General diagram of volcanic hazards around a stratovolcano. (By USGS; public domain.)

The USGS video provides an account of this explosive eruption that killed 57 people [24].







Figure 4.6.2: Left: Mount St. Helens, the day before the May 18, 1980 eruption. Right: 4 months after the major eruption of Mount St. Helens.

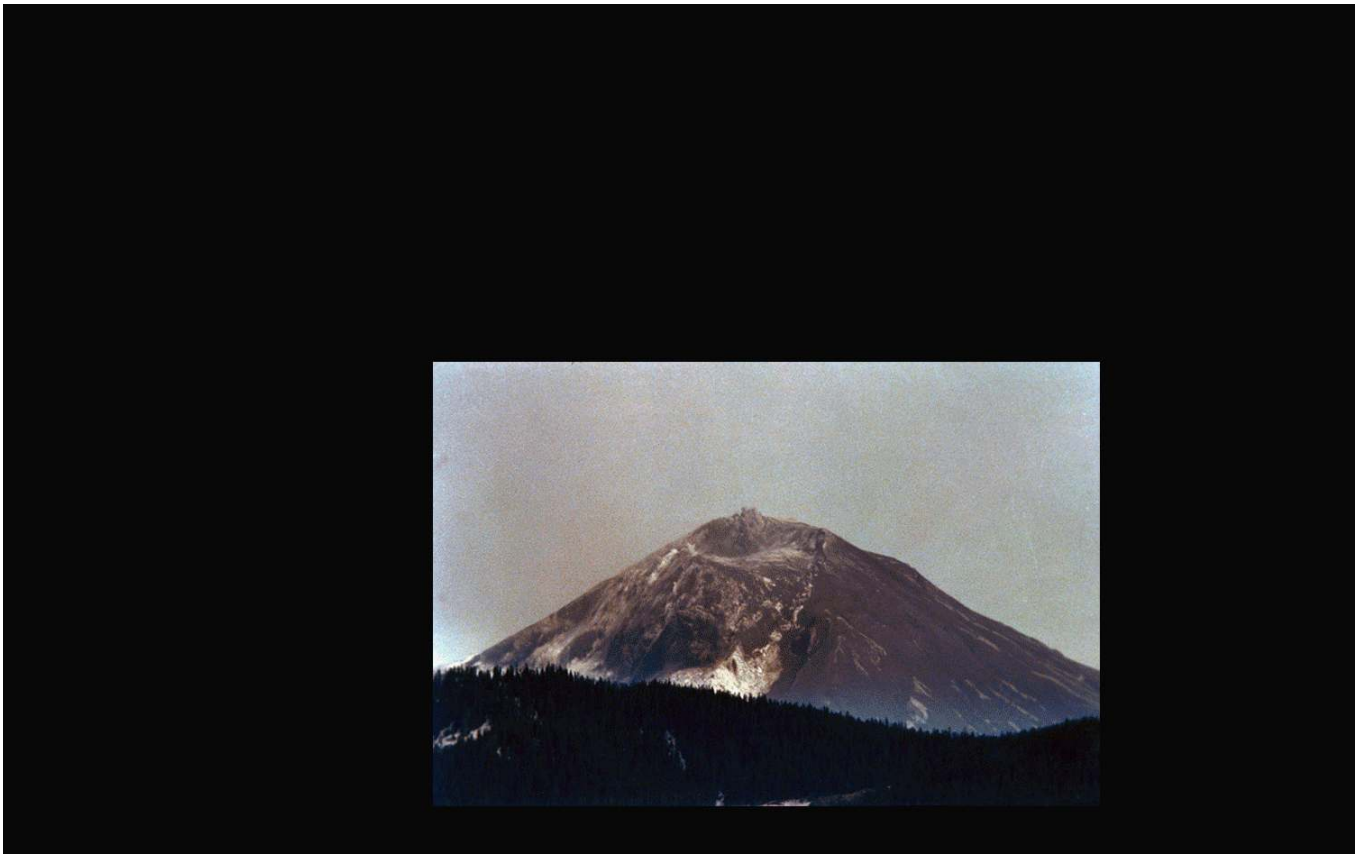


Figure 4.6.3: Series of still images of the May 18, 1980, eruption of Mount St. Helens, Washington showing largest recorded landslide in history and subsequent eruption and pyroclastic flow (By The Associated Press via The Atlantic)

In 79 CE, Mount Vesuvius, located near Naples, Italy, violently erupted sending a pyroclastic flow over the Roman countryside, including the cities of Herculaneum and Pompeii. The buried towns were discovered in an archeological expedition in the 18th century [25]. Pompeii famously contains the remains (casts) of people suffocated by ash and covered by 10 feet (3 m) of ash, pumice lapilli, and collapsed roofs [26].



Figure 4.6.4: Human remains from the 79 CE eruption of Vesuvius. (By Gary Todd; public domain via Wikimedia Commons.)

Pyroclastic Flows

The most dangerous volcanic hazard is a **pyroclastic flow**. These flows are a mix of lava blocks, pumice, ash, and hot gases between 200°C-700°C (400°F-1300°F). The turbulent cloud of ash and gas races down the steep flanks at high speeds up to 193 km per hour (120 mph) into the valleys around composite volcanoes [24]. Most explosive, silica-rich, high viscosity magma volcanoes such as composite cones usually have pyroclastic flows. The rocks tuff and welded tuff are often formed from these pyroclastic flows.



Figure 4.6.5: The material coming down from the eruption column onto the flanks is a pyroclastic flow.

There are numerous examples of deadly pyroclastic flows. In 2014, the Mount Ontake pyroclastic flow in Japan killed 47 people. The flow was caused by magma heating groundwater into steam, which then rapidly ejected with ash and volcanic bombs. Some were killed by inhalation of toxic gases and hot ash, while others were struck by volcanic bombs [27]. This video shows an eye-witness account of the Ontake pyroclastic flow.





In the early 1990s, Mount Unzen erupted several times with pyroclastic flows. The pyroclastic flow shown in this short video killed 41 people.





In 1902, on the Caribbean Island Martinique, Mount Pelee erupted with a violent pyroclastic flow that destroyed the entire town of St. Pierre and killing 28,000 people in moments.



Figure 4.6.6: The remains of St. Pierre after the eruption of Mount Pelee.

Landslides and Landslide-Generated Tsunamis

The steep and unstable flanks of a volcano can lead to slope failure and dangerous **landslides**. These landslides can be triggered by magma movement, explosive eruptions, large earthquakes, and/or heavy rainfall. During the 1980 Mount St. Helens eruption, the entire north flank of the volcano collapsed and released a huge landslide that moved at speeds of 160-290 km per hour (100-180 mph).

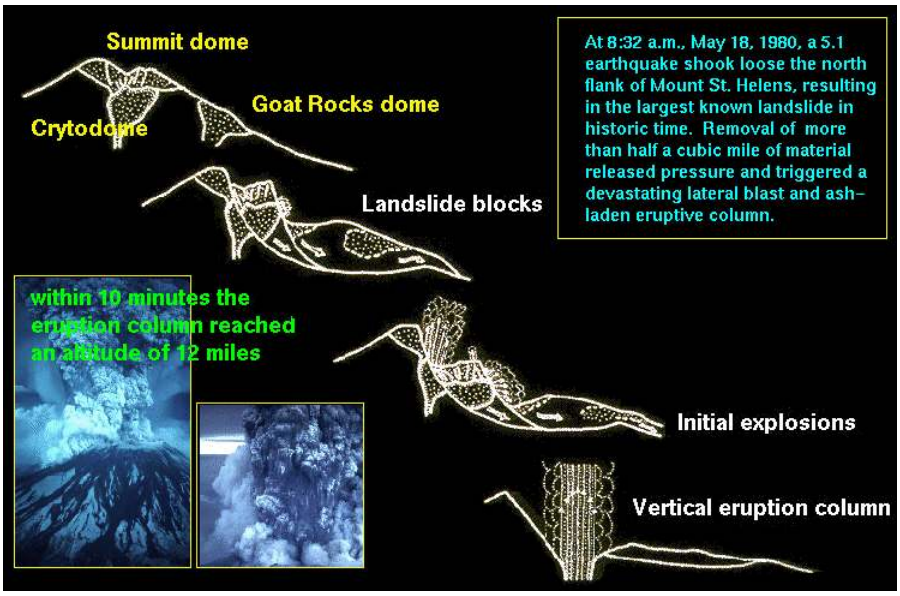


Figure 4.6.7: Sequence of events for Mount St. Helens, May 18, 1980. Note that an earthquake caused a landslide, which caused the "uncorking" of the mountain and started the eruption.

If enough landslide material reaches the ocean, it may cause a **tsunami**. In 1792, a landslide caused by the Mount Unzen eruption reached the Ariaka Sea, generating a tsunami that killed 15,000 people. When Mount Krakatau in Indonesia erupted in 1883, it generated ocean waves that towered 40 m (131 ft) above sea level. The tsunami killed 36,000 people and destroyed 165 villages [24].

Tephra

Volcanoes, especially composite volcanoes, eject large amounts of tephra (ejected rock materials), most notably ash (tephra fragments less 2 mm). Larger tephra is heavier and falls closer to the vent. Larger blocks and bombs pose hazards to those close to the eruption such as at the 2014 Mount Ontake disaster in Japan discussed earlier.



Figure 4.6.8: A man sweeps ash from an eruption of Kelud, Indonesia.

Hot ash poses an immediate danger to people, animals, plants, machines, roads, and buildings located close to the eruption. Ash is fine-grained (< 2mm) and can travel airborne long distances away from the eruption site. Heavy accumulations of ash can cause buildings to collapse. In people, it may cause respiratory issues like silicosis. Ash is destructive to aircraft and automobile engines, which can disrupt transportation and shipping services [24]. In 2010, the Eyjafjallajökull volcano in Iceland emitted a large ash cloud into the upper atmosphere, causing the largest air-travel disruption in northern Europe since World War II. No one was injured, but the service disruption was estimated to have cost the world economy billions of dollars [29].

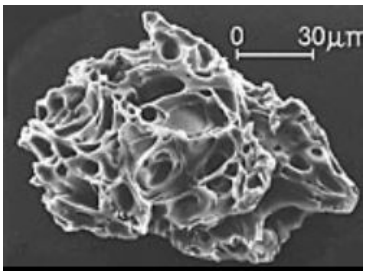


Figure 4.6.9: Micrograph of silica particle in volcanic ash. A cloud of these is capable of destroying an aircraft or automobile engine.

Volcanic Gases

As magma rises to the surface, the confining pressure decreases and allows dissolved gases to escape into the atmosphere. Even volcanoes that are not actively erupting may emit hazardous gases, such as carbon dioxide (CO_2), sulfur dioxide (SO_2), hydrogen sulfide (H_2S), and hydrogen halides (HF, HCl, or HBr).

Carbon dioxide tends to sink and accumulate in depressions and basins. In volcanic areas known to emit carbon dioxide, low-lying areas may trap hazardous concentrations of this colorless and odorless gas. The Mammoth Mountain Ski Resort in California, is located within the Long Valley Caldera, is one such area of carbon dioxide-producing volcanism. In 2006, three ski patrol members died of suffocation caused by carbon dioxide after falling into a snow depression near a fumarole (article).

In rare cases, volcanism may create a sudden emission of gases without warning. **Limnic eruptions** (*limne* is Greek for lake), occur in crater lakes associated with active volcanism. The water in these lakes is supercharged with high concentrations of dissolved gases. If the water is physically jolted by a landslide or earthquake, it may trigger an immediate and massive release of gases out of solution. An analogous example would be what happens to a vigorously shaken bottle of carbonated soda when the cap is opened. An infamous limnic eruption occurred in 1986 at Lake Nyos, Cameroon. Almost 2000 people were killed by a massive release of carbon dioxide [24].

Lahars

Lahar is an Indonesian word and is used to describe a volcanic mudflow that forms from rapidly melting snow or glaciers. Lahars are slurries resembling wet concrete and consist of water, ash, rock fragments, and other debris. These mudflows flow down the flanks of volcanoes or mountains covered with freshly-erupted ash, and on steep slopes can reach speeds of up to 80 km per hour (50 mph).



Figure 4.6.10: Mud line shows the extent of lahars around Mount St. Helens.

Several major cities, including Tacoma, are located on prehistoric lahar flows that extend for many kilometers across the flood plains surrounding Mount Rainier in Washington (see map). A map of Mount Baker in Oregon shows a similar potential hazard for lahar flows [24]. A tragic scenario played out in 1985, when a lahar from the Nevado del Ruiz volcano in Colombia buried the town of Armero and killed an estimated 23,000 people.

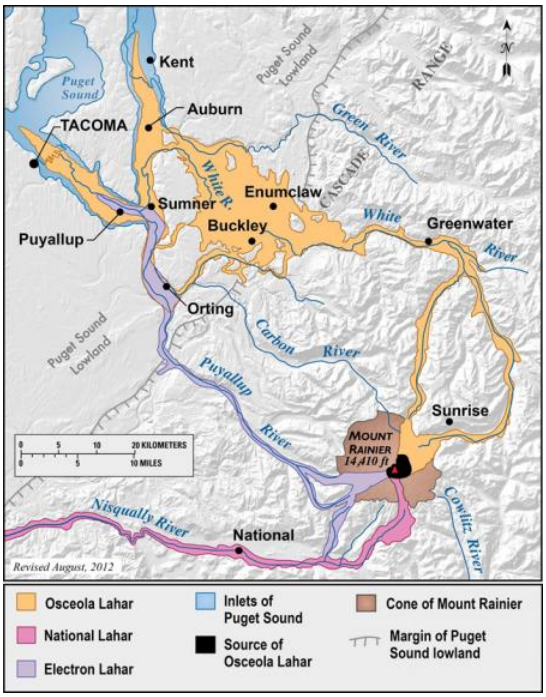
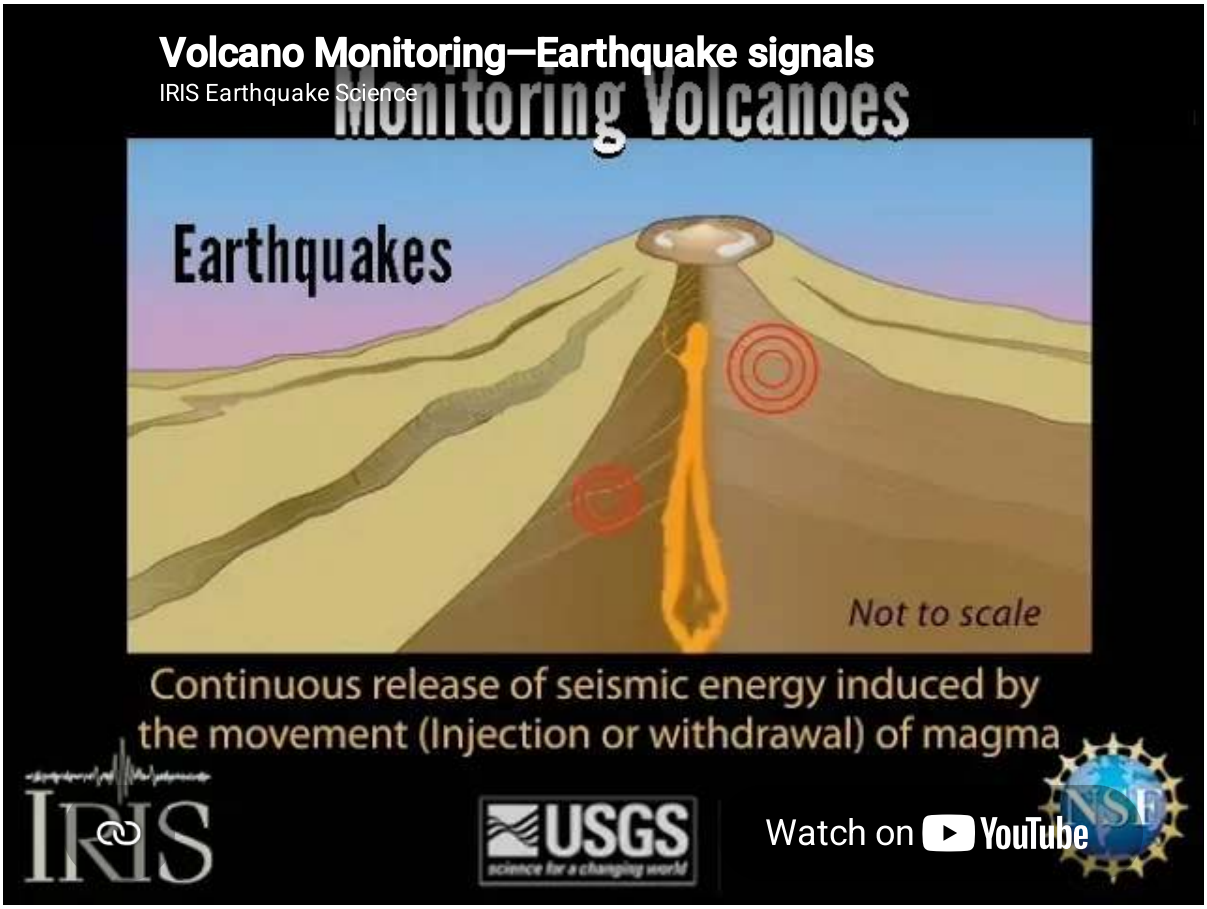


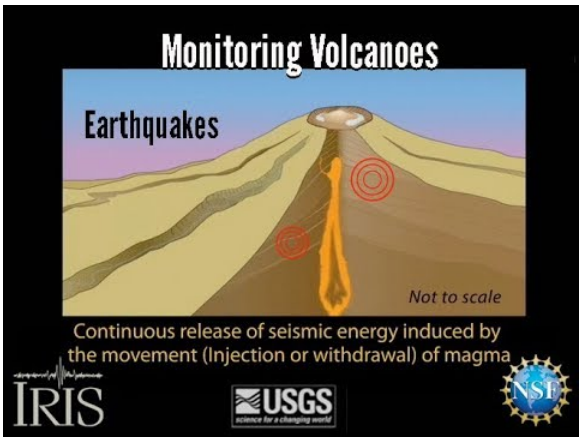
Figure 4.6.11: Old lahars around Tacoma, Washington.

Monitoring

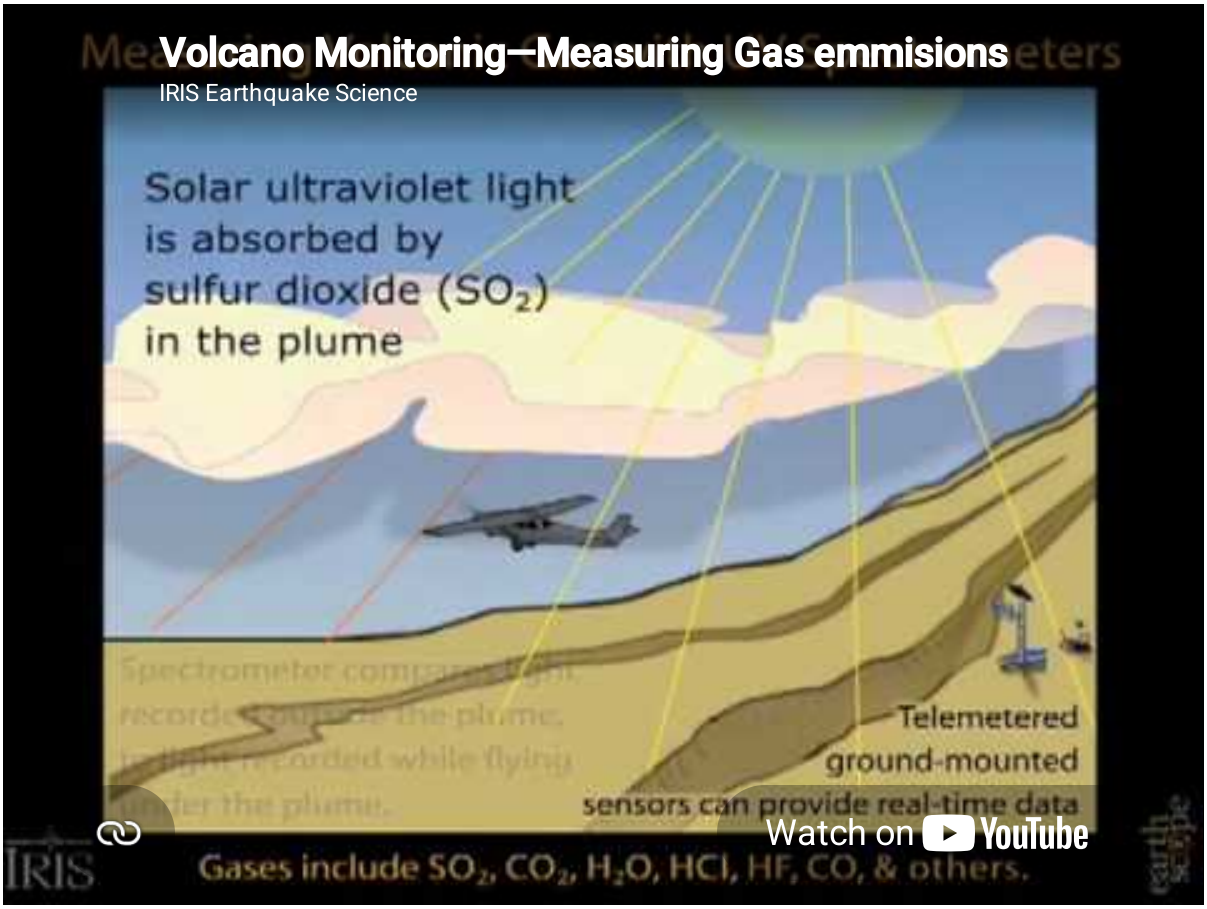
Geologists use various instruments to detect changes or indications that an eruption is imminent [30][31]. The three videos show different types of volcanic monitoring used to predict eruptions 1) earthquake activity, 2) increases in gas emission, and 3) changes in land surface orientation and elevation.

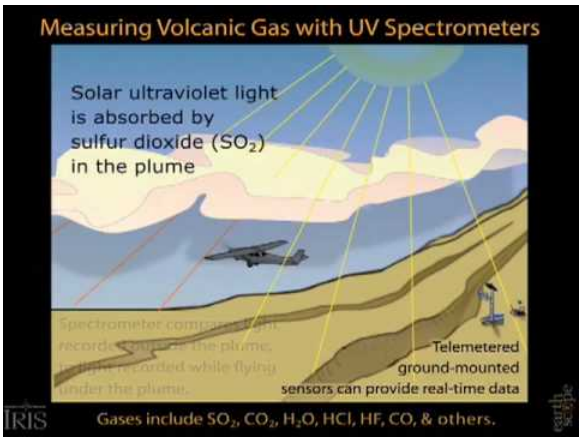
This video shows how monitoring earthquake frequency, especially special vibrational earthquakes called **harmonic tremors**, can detect magma movement and possible eruption.



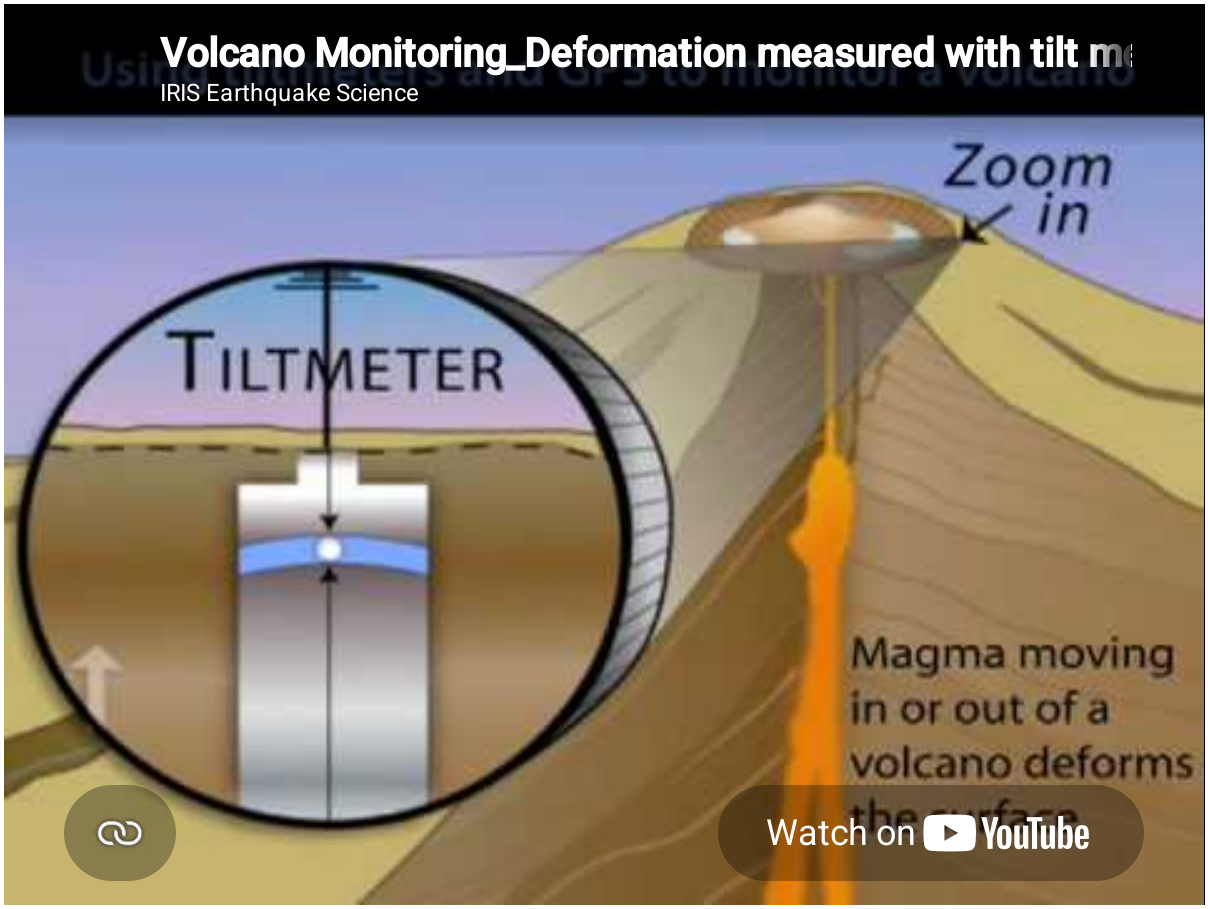


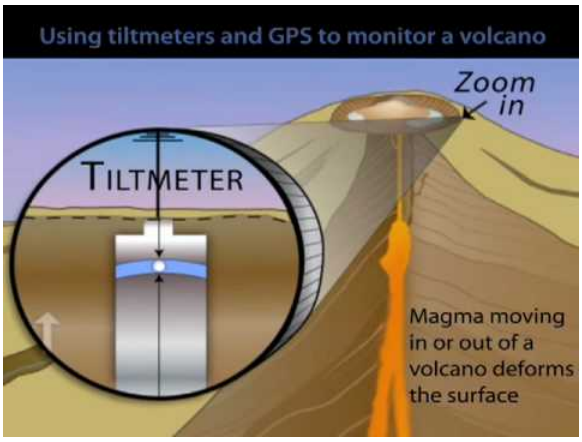
This video shows how gas monitoring may be used to predict an eruption. A rapid increase of gas emission may indicate magma that is actively rising to the surface and releasing dissolved gases out of solution, and that an eruption is imminent.





The last video shows how a GPS unit and **tiltmeter** can detect land surface changes, indicating the magma is moving underneath it.





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4.S: Summary

Igneous rock is divided into two major groups: intrusive rock that solidifies from underground magma, and extrusive rock formed from lava that erupts and cools on the surface. Magma is generated from mantle material at several plate tectonics situations by three types of melting: decompression melting, flux melting, or heat-induced melting. Magma composition is determined by differences in the melting temperatures of the mineral components (Bowen's Reaction Series). Volcanoes come in a wide variety of shapes and sizes, and are classified by multiple factors, including magma composition, and plate tectonic activity. Because volcanism presents serious hazards to human civilization, geologists carefully monitor volcanic activity to mitigate or avoid the dangers it presents.

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CHAPTER OVERVIEW

5: Weathering, Erosion, and Sedimentary Rocks

Learning Objectives

- Describe how water is an integral part of all sedimentary rock formation.
- Explain how chemical and mechanical weathering turn bedrock into sediment.
- Differentiate the two main categories of sedimentary rocks : clastic rock formed from pieces of weathered bedrock; and chemical rock that precipitates out of solution by organic or inorganic means.
- Explain the importance of sedimentary structures and analysis of depositional environments, and how they provide insight into the Earth's history.



Figure 5.1: Horseshoe Bend of the Colorado River near the Grand Canyon. (By Christian Mehlführer; CC BY 2.5 via Wikimedia Commons.)

Sedimentary rock and the processes that create it, which include weathering, erosion, and lithification, are an integral part of understanding Earth Science. This is because the majority of the Earth's surface is made up of sedimentary rocks and their common predecessor, sediments. Even though sedimentary rocks can form in drastically different ways, their origin and creation have one thing in common, water.

5.1: The Unique Properties of Water

5.2: Weathering and Erosion

5.3: Sedimentary Rocks

5.4: Sedimentary Structures

5.5: Depositional Environments

5.S: Summary

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5.1: The Unique Properties of Water

Water plays a role in the formation of most sedimentary rocks. It is one of the main agents involved in creating the minerals in chemical sedimentary rock. It also is a weathering and erosion agent, producing the grains that become detrital sedimentary rock. Several special properties make water an especially unique substance, and integral to the production of sediments and sedimentary rock.

The water molecule consists of two hydrogen atoms covalently bonded to one oxygen atom arranged in a specific and important geometry. The two hydrogen atoms are separated by an angle of about 105°, and both are located to one side of the oxygen atom [1]. This atomic arrangement, with the positively charged hydrogens on one side and negatively charged oxygen on the other side, gives the water molecule a property called **polarity**. Resembling a battery or a magnet, the molecule's positive-negative architecture leads to a whole suite of unique properties.

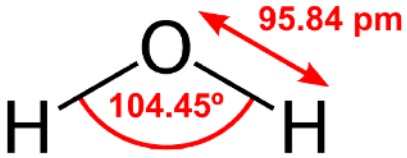


Figure 5.1.1: A model of a water molecule, showing the bonds between the hydrogen and oxygen.

Polarity allows water molecules to stick to other substances. This is called **adhesion**. Water is also attracted to itself, a property called **cohesion**, which leads to water's most common form in the air, a droplet. Cohesion is responsible for creating **surface tension**, which various insects use to walk on water by distributing their weight across the surface.



Figure 5.1.2: Dew on a spider's web.

The fact that water is attracted to itself leads to another important property, one that is extremely rare in the natural world—the liquid form is denser than the solid form. The polarity of water creates a special type of weak bonding called **hydrogen bonds**. Hydrogen bonds allow the molecules in liquid water to sit close together. Water is densest at 4°C and is less dense above and below that temperature. As water solidifies into ice, the molecules must move apart in order to fit into the crystal lattice, causing water to expand and become less dense as it freezes. Because of this, ice floats and water at 4°C sinks, which keeps the oceans liquid and prevents them from freezing solid from the bottom up. This unique property of water keeps Earth, the water planet, habitable.

Even more critical for supporting life, water remains **liquid** over a very large range of temperatures, which is also a result of cohesion. Hydrogen bonding allows liquid water to absorb high amounts of energy before turning into **vapor** or gas. The wide range across which water remains a liquid, 0°C-100°C (32°F-212°F), is rarely exhibited in other substances. Without this high **boiling point**, liquid water as we know it would be constricted to narrow temperature zones on Earth, instead of being found from pole to pole.

Water is a **universal solvent**, meaning it dissolves more substances than any other commonly found, naturally occurring liquid. The water molecules use polarity and hydrogen bonds to pry ions away from the crystal lattice. Water is such a powerful solvent, it can dissolve even the strongest rocks and minerals given enough time.

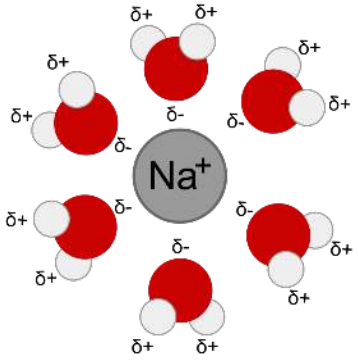


Figure 5.1.3: A sodium (Na) ion in solution.

Reference

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5.2: Weathering and Erosion

Bedrock refers to the solid crystalline rock that makes up the Earth's outer crust. **Weathering** is a process that turns bedrock into smaller particles, called **sediment** or soil. Mechanical weathering includes pressure expansion, frost wedging, root wedging, and salt expansion. Chemical weathering includes carbonic acid and hydrolysis, dissolution, and oxidation.

Erosion is a mechanical process, usually driven by water, wind, gravity, or ice, which transports sediment and soil from the place of weathering. Liquid water is the main agent of erosion. Gravity and mass wasting processes move rocks and sediment to new locations. Gravity and ice, in the form of glaciers, move large rock fragments as well as fine sediment.

Mechanical Weathering

Mechanical weathering physically breaks bedrock into smaller pieces. The usual agents of mechanical weathering are pressure, temperature, freezing/thawing cycle of water, plant or animal activity, and salt evaporation.

Pressure Expansion

Bedrock buried deep within the Earth is under high pressure and temperature. When uplift and erosion brings bedrock to the surface, its temperature drops slowly, while its pressure drops immediately. The sudden pressure drop causes the rock to rapidly expand and crack; this is called **pressure expansion**. **Sheeting** or **exfoliation** is when the rock surface spalls off in layers. **Spheroidal weathering** is a type of exfoliation that produces rounded features and is caused when chemical weathering moves along joints in the bedrock.



Figure 5.2.1: The outer layer of this granite is fractured and eroding away, known as exfoliation.

Frost Wedging

Frost wedging, also called **ice wedging**, uses the power of expanding ice to break apart rocks. Water works its way into various cracks, voids, and crevices. As the water freezes, it expands with great force, exploiting any weaknesses. When ice melts, the liquid water moves further into the widened spaces. Repeated cycles of freezing and melting eventually pry the rocks apart. The cycles can occur daily when fluctuations of temperature between day and night go from freezing to melting.

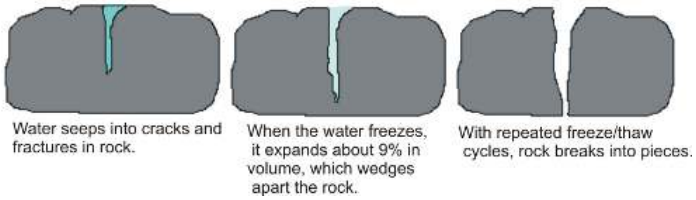


Figure 5.2.2: The process of frost wedging.

Root Wedging

Like frost wedging, **root wedging** happens when plant roots work themselves into cracks, prying the bedrock apart as they grow. Occasionally these roots may become fossilized. **Rhizolith** is the term for these roots preserved in the rock record [2]. Tunneling organisms such as earthworms, termites, and ants are biological agents that induce weathering similar to root wedging.



Figure 5.2.3: The roots of this tree are demonstrating the destructive power of root wedging. Though this picture is a man-made rock (asphalt), it works on a typical rock as well.

Salt Expansion

Salt expansion, which works similarly to frost wedging, occurs in areas of high evaporation or near-marine environments. Evaporation causes salts to precipitate out of solution and grow and expand into cracks in the rock. Salt expansion is one of the causes of **tafoni**, a series of holes in a rock. Tafoni, cracks, and holes are weak points that become susceptible to increased weathering. Another phenomenon that occurs is when salt water evaporates and leaves behind a **hopper crystal**, a square-shaped crystal, commonly made of salt, preserved in the rock.



Figure 5.2.4: Tafoni from Salt Point, California.

Chemical Weathering

Chemical weathering is the dominant weathering process in warm, humid environments. It happens when water, oxygen, and other reactants chemically degrade the mineral components of bedrock and turn them into water-soluble ions which can then be transported by water. Higher temperatures accelerate chemical weathering rates.

Chemical and mechanical weathering work hand-in-hand via a fundamental concept called **surface-area-to-volume ratio**. Chemical weathering only occurs on rock surfaces because water and reactants cannot penetrate the solid rock. Mechanical weathering penetrates the bedrock, breaking large rocks into smaller pieces and creating new rock surfaces. This exposes more surface area to chemical weathering, enhancing its effects. In other words, higher surface-area-to-volume ratios produce higher rates of overall weathering.

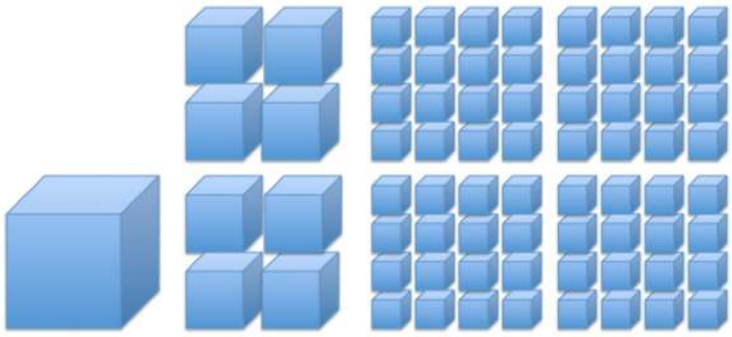


Figure 5.2.5: Each of these three groups of cubes has an equal volume. However, their surface areas are vastly different. On the left, the single cube has a length, width, and height of 4 units, giving it a surface area of $6(4 \times 4) = 96$ and a volume of $4^3 = 64$. The middle eight cubes have a length, width, and height of 2, meaning a surface area of $8(6(2 \times 2)) = 8 \times 24 = 192$. They also have a volume of $8(2^3) = 8 \times 8 = 64$. The 64 cubes on the right have a length, width, and height of 1, leading to a surface area of $64(6(1 \times 1)) = 64 \times 6 = 384$. The volume remains unchanged because $64(1^3) = 64 \times 1 = 64$. The surface area to volume ratio (SA:V), which is related to the amount of material available for reactions, changes for each as well. On the left, it is $\frac{96}{64} = 0.75$ or 3:4. The center has a SA:V of $\frac{192}{64} = 1.5$, or 3:1. On the right, the SA:V is $\frac{384}{64} = 6$, or 6:1.

Carbonic Acid and Hydrolysis

Carbonic acid (H_2CO_3) forms when carbon dioxide, the fifth-most abundant gas in the atmosphere, dissolves in water. This happens naturally in clouds, which is why precipitation is normally slightly acidic. Carbonic acid is an important agent in two chemical weathering reactions, hydrolysis, and dissolution.

Hydrolysis occurs via two types of reactions. In one reaction, water molecules ionize into positively charged H^+ and OH^- ions and replace mineral cations in the crystal lattice. In another type of hydrolysis, carbonic acid molecules react directly with minerals, especially those containing silicon and aluminum (i.e. feldspars), to form molecules of clay minerals.

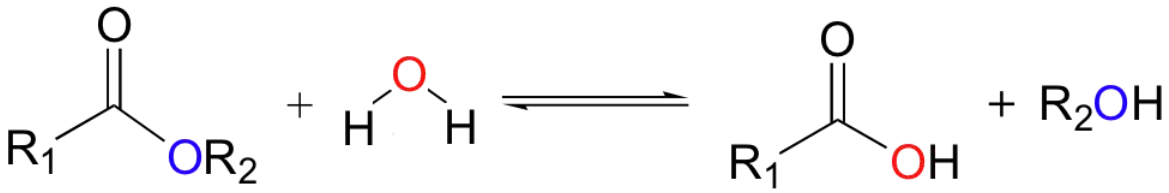


Figure 5.2.6: Generic hydrolysis diagram, where the bonds in the mineral in question would represent the left side of the diagram.

Hydrolysis is the main process that breaks down silicate rock and creates clay minerals. The following is a hydrolysis reaction that occurs when silica-rich feldspar encounters carbonic acid to produce water-soluble clay and other ions:

feldspar + carbonic acid (in water) → clay + metal cations (Fe^{++}, Mg^{++}, Ca^{++}, Na^+, etc.) + bicarbonate anions (HCO_3^-) + silica (SiO_2)

Clay minerals are platy silicates or phyllosilicates similar to micas, and are the main components of very fine-grained sediment. The dissolved substances may later precipitate into chemical sedimentary rocks like evaporite and limestone, as well as amorphous silica or chert nodules.

Dissolution

Dissolution is a hydrolysis reaction that dissolves minerals in bedrock and leaves the ions in solution, usually in water. Some evaporites and carbonates, like salt and calcite, are more prone to this reaction; however, all minerals can be dissolved. Non-acidic water, having a neutral pH of 7, will dissolve any mineral, although it may happen very slowly. Water with higher levels of acid, natural or man-made, dissolves rocks at a higher rate. Liquid water is normally slightly acidic due to the presence of carbonic acid and free H^+ ions. Natural rainwater can be highly acidic, with pH levels as low as 2 [3]. Dissolution can be enhanced by a biological agent, such as when organisms like lichen and bacteria release organic acids onto the rocks they are attached to. Regions with high humidity (airborne moisture) and precipitation experience more dissolution due to greater contact time between rocks and water.



Figure 5.2.7: In this rock, a pyrite cube has dissolved (as seen with the negative “corner” impression in the rock), leaving behind small specks of gold coloring from the pyrite.

The **Goldich Dissolution Series** shows chemical weathering rates are associated with crystallization rankings in the Bowen’s Reaction Series [4]. Minerals at the top of Bowen's series crystallize under high temperatures and pressures, and chemically weather at a faster rate than minerals ranked at the bottom. Quartz, a felsic mineral that crystallizes at 700°C, is very resistant to chemical weathering. High crystallization-point mafic minerals, such as olivine and pyroxene (1250°C), weather relatively rapidly and more completely. Olivine and pyroxene are rarely found as end products of weathering because they tend to break down into elemental ions.



Figure 5.2.8: This mantle xenolith containing olivine (green) is chemically weathering by hydrolysis and oxidation into the pseudo-mineral iddingsite, which is a complex of water, clay, and iron oxides. The more altered side of the rock has been exposed to the environment longer.

Dissolution is also noteworthy for the special geological features it creates. In places with abundant carbonate bedrock, dissolution weathering can produce a **karst topography** characterized by sinkholes or caves.



Figure 5.2.9: Eroded karst topography in Minevre, France.

Timpanogos Cave National Monument in Northern Utah is a well-known dissolution feature. The figure shows a cave formation created from dissolution followed by precipitation—groundwater saturated with calcite seeped into the cavern, where evaporation caused the dissolved minerals to precipitate out.



Figure 5.2.10: A formation called The Great Heart of Timpanogos in Timpanogos Cave National Monument.

Oxidation

Oxidation, the chemical reaction that causes rust in metallic iron, occurs geologically when iron atoms in a mineral bond with oxygen. Any minerals containing iron can be oxidized. The resultant iron oxides may permeate a rock if it is rich in iron minerals. Oxides may also form a coating that covers rocks and grains of sediment, or they can line rock cavities and fractures. If the oxides are more susceptible to weathering than the original bedrock, they may create void spaces inside the rock mass, weakness throughout the rock mass, or hollows on exposed surfaces.

Three commonly found minerals are produced by iron-oxidation reactions: red or gray hematite, brown goethite (pronounced “GUR-tite”), and yellow limonite. These iron oxides coat and bind mineral grains together into sedimentary rocks in a process called **cementation** and often give these rocks a dominant color. They color the rock layers of the Colorado Plateau, as well as Zion, Arches, and Grand Canyon National Parks. When the oxide minerals replacing existing minerals in bedrock are resistant to weathering, iron **concretions** may occur in the rock. Concretions are round, hard, cemented masses in a sedimentary rock.



Figure 5.2.11: Pyrite cubes are oxidized, becoming new mineral goethite. In this case, goethite is a pseudomorph after pyrite, meaning it has taken the form of another mineral.

Erosion

Erosion is a mechanical process, usually driven by water, gravity, wind, or ice that removes sediment from the place of weathering. Liquid water is the main agent of erosion.



Figure 5.2.12: Grand Canyon from Mather Point. The canyon was formed by erosion of the bedrock by the Colorado River.

Erosion resistance is important in the creation of distinctive geological features. This is well demonstrated in the cliffs of the Grand Canyon. The cliffs are made of rock left standing after less resistant materials have weathered and eroded away. Rocks with

different levels of erosion resistance also create unique-looking features called **hoodoos** in Bryce Canyon National Park and Goblin Valley State Park in Utah.



Figure 5.2.13: A hoodoo near Moab, Utah. The more resistant cap has protected the less resistant underlying layers.

Soil

Soil is a combination of air, water, minerals, and organic matter that forms at the transition between the biosphere and geosphere. Soil is made when weathering breaks down the bedrock and turns it into sediment. If erosion does not remove the sediment significantly, organisms can access the mineral content of the sediments. These organisms turn minerals, water, and atmospheric gases into organic substances that contribute to the soil.

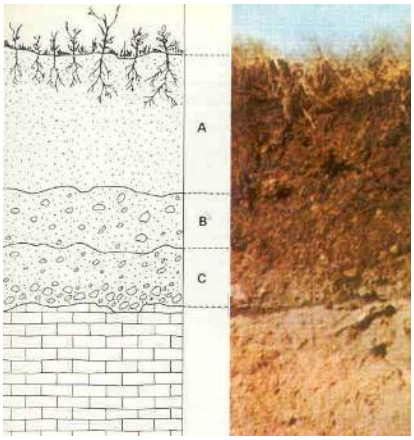


Figure 5.2.14: A sketch and picture of soil.

Soil is an important reservoir for organic components necessary for plants, animals, and microorganisms to live. The organic component of soil, called **humus**, is a rich source of bioavailable nitrogen. Nitrogen is the most common element in the atmosphere, but it exists in a form most life forms are unable to use. Special bacteria found only in the soil provide most nitrogen compounds that are usable, **bioavailable**, by life forms. These **nitrogen-fixing bacteria** absorb nitrogen from the atmosphere and convert it into nitrogen compounds. These compounds are absorbed by plants and used to make DNA, amino acids for protein, and enzymes. Animals obtain bioavailable nitrogen by eating plants, and this is the source of most of the nitrogen used by life.

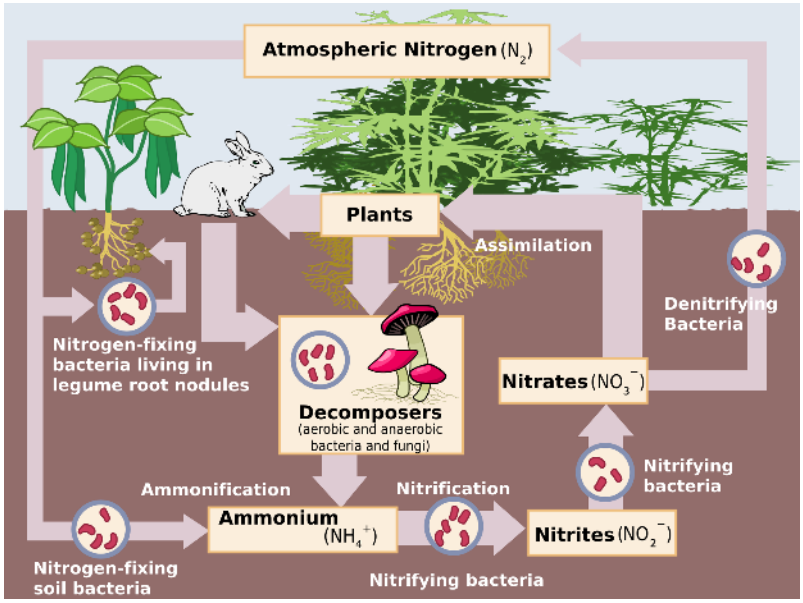


Figure 5.2.15: Schematic of the nitrogen cycle.

Soils range from poor to rich, depending on the amount of humus they contain. Soil productivity is determined by water and nutrient content. Freshly created volcanic soils, called andisols, and clay-rich soils that hold nutrients and water are examples of productive soils.

The nature of the soil, meaning its characteristics, is determined primarily by five components:

1. mineralogy of the parent material
2. topography
3. weathering
4. climate
5. organisms that inhabit the soil.

For example, soil tends to erode more rapidly on steep slopes so soil layers in these areas may be thinner than in flood plains, where it tends to accumulate. The quantity and chemistry of the organic matter of soil affect how much and what varieties of life it can sustain. Temperature and precipitation, two major weathering agents, are dependent on climate. Fungi and bacteria contribute organic matter and the ability of soil to sustain life, interacting with plant roots to exchange nitrogen and other nutrients [5].



Figure 5.2.16: Agricultural terracing, as made by the Inca culture from the Andes, helps reduce erosion and promote soil formation, leading to better farming practices.

In well-formed soils, there is a discernable arrangement of distinct layers called **soil horizons** [6]. These soil horizons can be seen in road cuts that expose the layers at the edge of the cut. Soil horizons make up the **soil profile**. Each soil horizon reflects climate, topography, and other soil-development factors, as well as its organic material and mineral sediment composition. The horizons are assigned names and letters. Differences in naming schemes depend on the area, soil type or research topic. The figure shows a simplified soil profile that uses commonly designated names and letters.

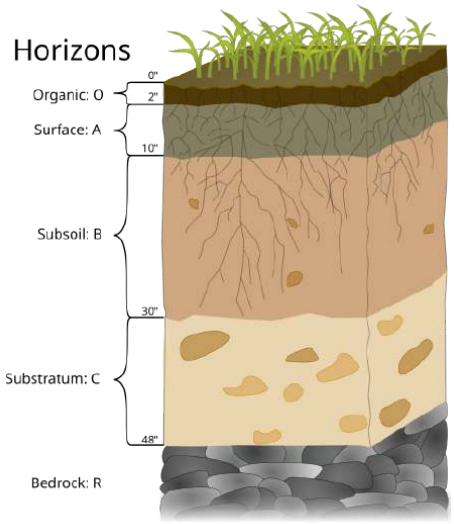


Figure 5.2.17: A simplified soil profile, showing labeled layers. (By Wilsonbiggs; CC BY-SA 4.0 via Wikimedia Commons.)

O Horizon: The top horizon is a thin layer of predominantly organic material, such as leaves, twigs, and other plant parts that are actively decaying into humus.

A Horizon: The next layer, called **topsoil**, consists of humus mixed with mineral sediment. As precipitation soaks down through this layer, it leaches out soluble chemicals. In wet climates with heavy precipitation, this leaching out produces a separate layer called **horizon E**, the **leaching** or **eluviation zone**.

B Horizon: Also called **subsoil**, this layer consists of sediment mixed with humus removed from the upper layers. The subsoil is where mineral sediment is chemically weathered. The amount of organic material and the degree of weathering decrease with depth. The upper subsoil zone, called **regolith**, is a porous mixture of humus and highly weathered sediment. In the lower zone, **saprolite**, scant organic material is mixed with largely unaltered parent rock.

C Horizon: This is called **substratum** and is a zone of mechanical weathering. Here, bedrock fragments are physically broken but not chemically altered. This layer contains no organic material.

R Horizon: The final layer consists of unweathered, parent bedrock and fragments.

The United States governing body for agriculture, the USDA, uses a taxonomic classification to identify soil types, called soil orders. **Xoxisols** or **laterite soils** are nutrient-poor soils found in tropical regions. While poorly suited for growing crops, xosisols are home to most of the world's mineable aluminum ore (bauxite). **Ardisol** forms in dry climates and can develop layers of hardened calcite, called **caliche**. **Andisols** originate from volcanic ash deposits. **Alfisols** contain silicate clay minerals. These two soil orders are productive for farming due to their high content of mineral nutrients. In general, color can be an important factor in understanding soil conditions. Black soils tend to be **anoxic**, red soils are oxygen-rich, and green soils are oxygen-poor (i.e. reduced). This is true for many sedimentary rocks as well.



Figure 5.2.18: A sample of bauxite. Note the unweathered igneous rock in the center.

Not only is soil essential to terrestrial life in nature, but also human civilization via agriculture. Careless or uninformed human activity can seriously damage soil's life-supporting properties. A prime example is the famous Dust Bowl disaster of the 1930s, which affected the midwestern United States. The damage occurred because of large-scale attempts to develop prairie land in southern Kansas, Colorado, western Texas, and Oklahoma into farmland [7]. Poor understanding of the region's geology, ecology, and climate led to farming practices that ruined the soil profile.



Figure 5.2.19: A dust storm approaches Stratford, Texas in 1935.

The prairie soils and native plants are well adapted to a relatively dry climate. With government encouragement, settlers moved in to homestead the region. They plowed vast areas of prairie into long, straight rows and planted grain. The plowing broke up the stable soil profile and destroyed the natural grasses and plants, which had long roots that anchored the soil layers. The grains they planted had shallower root systems and were plowed up every year, which made the soil prone to erosion. The plowed furrows were aligned in straight rows running downhill, which favored erosion and loss of topsoil.

The local climate does not produce sufficient precipitation to support non-native grain crops, so the farmers drilled wells and over-pumped water from the underground aquifers. The grain crops failed due to lack of water, leaving bare soil that was stripped from the ground by the prairie winds. Particles of midwestern prairie soil were deposited along the east coast and as far away as Europe. Huge dust storms called black blizzards made life unbearable, and the once-hopeful homesteaders left in droves. The setting for John Steinbeck's famous novel and John Ford's film, *The Grapes of Wrath*, is Oklahoma during this time. The lingering question is whether we have learned the lessons of the dust bowl, to avoid creating it again [8].

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5.3: Sedimentary Rocks

Sedimentary rock is classified into two main categories: clastic and chemical. Clastic or detrital sedimentary rocks are made from pieces of bedrock, sediment, derived primarily by mechanical weathering. Clastic rocks may also include chemically weathered sediment. Clastic rocks are classified by grain shape, grain size, and sorting. Chemical sedimentary rocks are precipitated from water saturated with dissolved minerals. Chemical rocks are classified mainly by the composition of minerals in the rock.



Figure 5.3.1: Sedimentary rock at Siccar Point on the east coast of Scotland.

Lithification and Diagenesis

Lithification turns loose sediment grains, created by weathering and transported by erosion, into clastic sedimentary rock via three interconnected steps. **Deposition** happens when friction and gravity overcome the forces driving sediment transport, allowing sediment to accumulate. **Compaction** occurs when material continues to accumulate on top of the sediment layer, squeezing the grains together and driving out the water. The mechanical compaction is aided by weak attractive forces between the smaller grains of sediment. Groundwater typically carries cementing agents into the sediment. These minerals, such as calcite, amorphous silica, or oxides, may have a different composition than the sediment grains. **Cementation** is the process of cementing minerals coating the sediment grains and gluing them together into a fused rock.

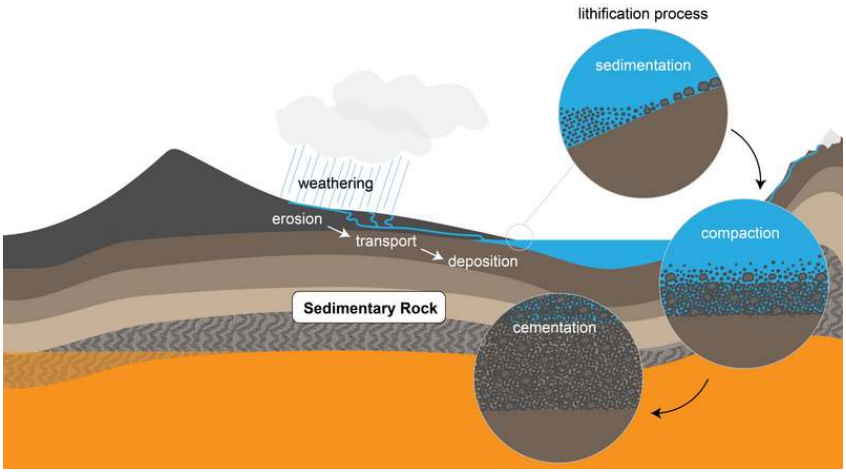


Figure 5.3.2: The process of forming sedimentary rocks from weathering, erosion, transport, deposition, sedimentation, compaction, and cementation. (By Carie Frantz; CC BY-SA 4.0 via Wikimedia Commons.)

Diagenesis is a process accompanying lithification, and is a low-temperature form of rock metamorphism. During diagenesis, sediments are chemically altered by heat and pressure. A classic example is aragonite ($CaCO_3$), a form of calcium carbonate that makes up most organic shells. When lithified aragonite undergoes diagenesis, the aragonite reverts to calcite ($CaCO_3$), which has the same chemical formula but a different crystalline structure. In sedimentary rock containing calcite and magnesium (Mg), diagenesis may transform the two minerals into dolomite ($CaMg(CO_3)_2$). Diagenesis may also reduce the **pore space**, which is the

open space between sedimentary rock grains. The processes of cementation, compaction, and ultimately lithification occur within the realm of diagenesis, which includes the processes that turn organic material into fossils.



Figure 5.3.3: Permineralization is a form of diagenesis that fossilizes wood into a rock, known as petrified wood. (By Moondigger; CC BY-SA 2.5 via Wikimedia Commons.)

Detrital Sedimentary Rocks (Clastic)

Detrital or **clastic** sedimentary rocks consist of preexisting sediment pieces that come from weathered bedrock. Most of this is mechanically weathered sediment, although some clasts may be pieces of chemical rocks. This creates some overlap between the two categories, since clastic sedimentary rocks may include chemical sediments. Detrital or clastic rocks are classified and named based on their grain size.

Grain Size

Detrital sedimentary rock is classified according to sediment grain size, which is graded from large to small on the Wentworth scale (see figure). **Grain size** is the average diameter of sediment fragments in sediment or rock. Grain sizes are delineated using a log base 2 scale [9; 10]. For example, the grain sizes in the **pebble** class are 2.52, 1.26, 0.63, 0.32, 0.16, and 0.08 inches, which correlate respectively to very coarse, **coarse**, **medium**, **fine**, and very fine granules. Large fragments, or clasts, include all grain sizes larger than 2 mm (5/64 in). These include **boulders**, **cobbles**, granules, and gravel. **Sand** has a grain size between 2 mm and 0.0625 mm, about the lower limit of the naked eye's resolution. Sediment grains smaller than sand are called **silt**. Silt is unique; the grains can be felt with a finger or as grit between your teeth, but are too small to see with the naked eye.

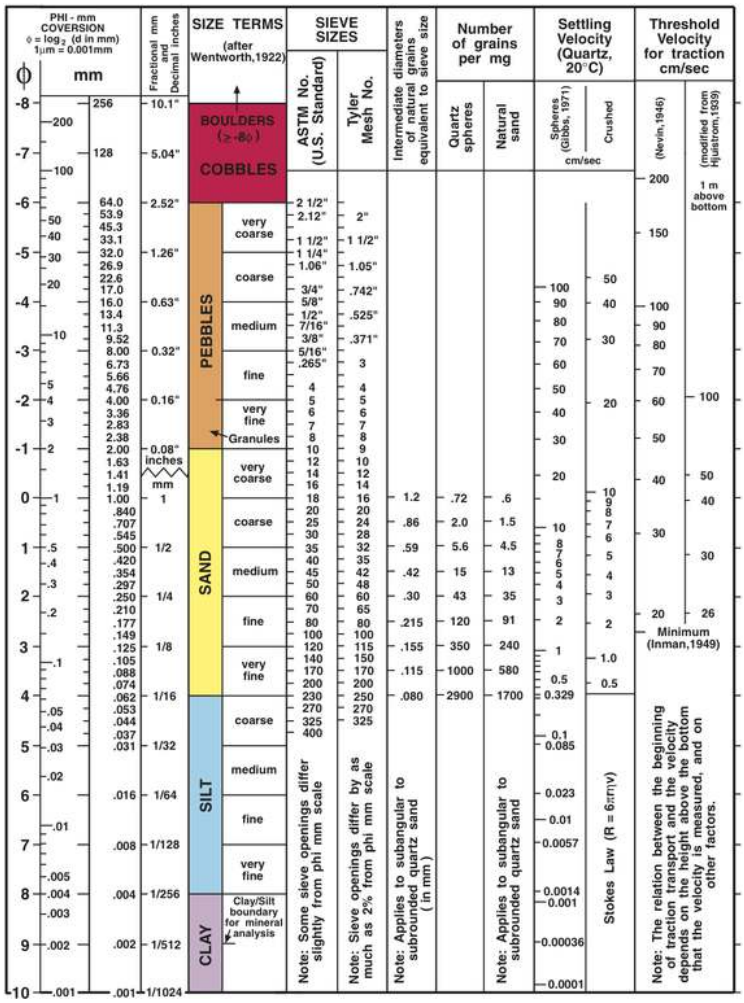


Figure 5.3.4: Size categories of sediments, known as the Wentworth scale.

Sorting and Rounding

Sorting describes the range of grain sizes within sediment or sedimentary rock. Geologists use the term "**well-sorted**" to describe a narrow range of grain sizes, and "**poorly-sorted**" for a wide range of grain sizes [11]. It is important to note that soil engineers use similar terms with opposite definitions; well-graded sediment consists of a variety of grain sizes, and poorly graded sediment has roughly the same grain sizes.

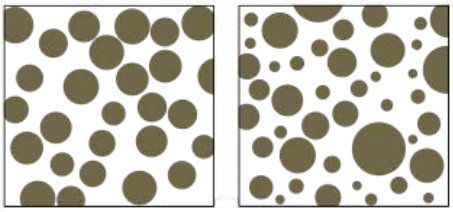


Figure 5.3.5: Left: Well-sorted sediment. Right: poorly-sorted sediment.

When reading the story told by rocks, geologists use sorting to interpret erosion or transport processes, as well as **deposition energy**. For example, wind-blown sands are typically extremely well sorted, while glacial deposits are typically poorly sorted. These characteristics help identify the type of erosion process that occurred. Coarse-grained sediment and poorly-sorted rocks are usually found nearer to the source of sediment, while fine sediments are carried farther away. In a rapidly flowing mountain stream, you would expect to see boulders and pebbles. In a lake fed by the stream, there should be sand and silt deposits. If you also find large boulders in the lake, this may indicate the involvement of another sediment transport process, such as rockfall caused by ice- or root-wedging.

Rounding is created when angular corners of rock fragments are removed from a piece of sediment due to abrasion during transport. Well-rounded sediment grains are defined as being free of all sharp edges. Very **angular** sediment retains the sharp corners. Most clast fragments start with some sharp edges due to the bedrock's crystalline structure, and those points are worn down during transport. More rounded grains imply a longer erosion time or transport distance, or more energetic erosional process. Mineral hardness is also a factor in rounding.

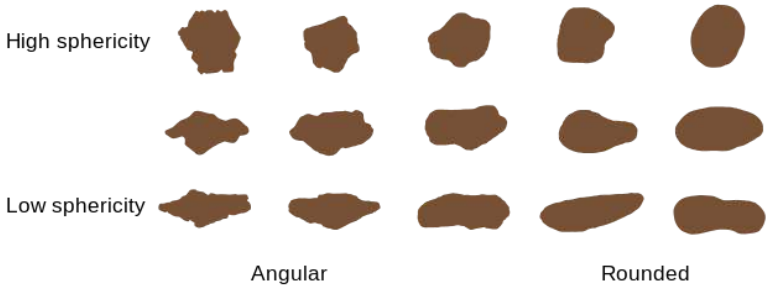


Figure 5.3.6: Degree of rounding in sediments. Sphericity refers to the spherical nature of an object, a completely different measurement unrelated to rounding.

Composition and Provenance

Composition describes the mineral components found in sediment or sedimentary rock and may be influenced by local geology, like source rock and hydrology. Other than clay, most sediment components are easily determined by visual inspection. The most commonly found sediment mineral is quartz because of its low chemical reactivity and high hardness, making it resistant to weathering, and its ubiquitous occurrence in continental bedrock. Other commonly found sediment grains include feldspar and lithic fragments. **Lithic fragments** are pieces of fine-grained bedrock [12], and include mud chips, volcanic clasts, or pieces of slate.

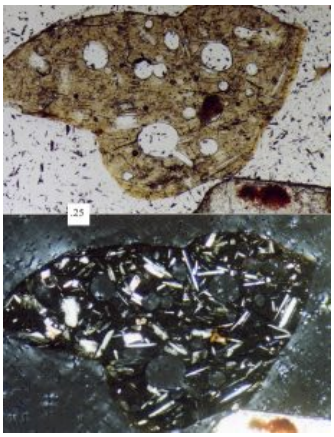


Figure 5.3.7: A microscope view of a sand grain made of basalt, known as a microlithic volcanic lithic fragment. The box is 0.25 mm. Top picture is plane-polarized light; bottom is cross-polarized light.

Weathering of volcanic rock produces Hawaii's famous black (basalt) and green (olivine) sand beaches, which are rare elsewhere on Earth. This is because the local rock is composed almost entirely of basalt and provides an abundant source of dark-colored clasts loaded with mafic minerals. According to the Goldich Dissolution Series, clasts high in mafic minerals are more easily destroyed compared to clasts composed of felsic minerals like quartz [13].



Figure 5.3.8: Left: Hawaiian beach composed of green olivine sand from weathering of nearby basaltic rock. Right: Sand sample from the green sand beach. (By Siim Sepp; CC BY-SA 3.0 via Wikimedia Commons.)

Geologists use **provenance** to discern the original source of sediment or sedimentary rock. Provenance is determined by analyzing the mineral composition and types of fossils present, as well as textural features like sorting and rounding. Provenance is important for describing tectonic history [14], visualizing paleogeographic formations [15], unraveling an area's geologic history, or reconstructing past supercontinents [16].

In quartz sandstone, sometimes called quartz arenite (SiO_2), age of the provenance may be determined using a rare, durable clast mineral called zircon ($ZrSiO_4$). Zircon, or zirconium silicate, contains traces of uranium, which can be used for age-dating the source bedrock that contributed sediment to the lithified sandstone rock.

Classification of Clastic Rocks

Clastic rocks are classified according to the grain size of their sediment [17]. Coarse-grained rocks contain clasts with a predominant grain size larger than sand. Typically, smaller sediment grains, collectively called **groundmass** or **matrix**, fill in much of the volume between the larger clasts, and hold the clasts together. **Conglomerates** are rocks containing coarse rounded clasts, and **breccias** contain angular clasts. Both conglomerates and breccias are usually poorly sorted.



Figure 5.3.9: Megabreccia in Titus Canyon, Death Valley National Park, California.

Medium-grained rocks composed mainly of sand are called **sandstone**, or sometimes **arenite** if well sorted. Sediment grains in sandstone can have a wide variety of mineral compositions, roundness, and sorting. Some sandstone names indicate the rock's mineral composition. **Quartz sandstone** contains predominantly quartz sediment grains. **Arkose** is sandstone with significant amounts of feldspar, usually greater than 25%. Sandstone that contains feldspar, which weathers more quickly than quartz, is useful for analyzing the local geologic history. **Graywacke** is a term with conflicting definitions [18]. Graywacke may refer to sandstone with a muddy matrix, or sandstone with many lithic fragments (small rock pieces).

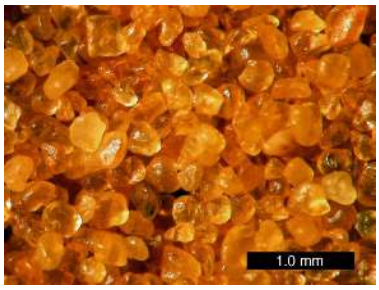


Figure 5.3.10: Enlarged image of frosted and rounded windblown sand grains.

Fine-grained rocks include mudstone, shale, siltstone, and claystone. **Mudstone** is a general term for rocks made of sediment grains smaller than sand (less than 2 mm). Rocks that are **fissile**, meaning they separate into thin sheets, are called **shale**. Rocks exclusively composed of silt or clay sediment, are called **siltstone** or **claystone**, respectively. These last two rock types are rarer than mudstone or shale.



Figure 5.3.11: The Rochester Shale, New York. Note the thin fissility in the layers.

Rock types found as a mixture between the main classifications may be named using the less-common component as a descriptor. For example, a rock containing some silt but mostly rounded sand and gravel are called silty conglomerate. Sand-rich rock containing minor amounts of clay is called clayey sandstone.



Figure 5.3.12: Claystone laminations (layers) from Glacial Lake Missoula.

Chemical, Biochemical, and Organic Sedimentary Rocks

Chemical sedimentary rocks are formed by processes that do not directly involve mechanical weathering and erosion. Chemical weathering may contribute to the dissolved materials in water that ultimately form these rocks. Biochemical and organic sediments are clastic in the sense that they are made from pieces of organic material that are deposited, buried, and lithified; however, they are usually classified as being chemically produced.

Inorganic chemical sedimentary rocks are made of minerals precipitated from ions dissolved in solution, and created without the aid of living organisms. Inorganic chemical sedimentary rocks form in environments where ion concentration, dissolved gases, temperatures, or pressures are changing, which cause minerals to crystallize.

Biochemical sedimentary rocks are formed from shells and bodies of underwater organisms. The living organisms extract chemical components from the water and use them to build shells and other body parts. The components include aragonite, a mineral similar to and commonly replaced by calcite, and silica.

Organic sedimentary rocks come from organic material that has been deposited and lithified, usually underwater. The source materials are plant and animal remains that are transformed through burial and heat, and end up as coal, oil, and methane (natural gas).

Inorganic Chemical Sedimentary Rocks

Inorganic chemical sedimentary rocks are formed when minerals precipitate out of an aqueous solution, usually due to water evaporation. The precipitate minerals form various salts known as **evaporites**. For example, the Bonneville Salt Flats in Utah flood with winter rains and dry out every summer, leaving behind salts such as **gypsum** and **halite**. The deposition order of evaporite formation is opposite to their solubility order, i.e. as water evaporates and increases the mineral concentration in solution, less soluble minerals precipitate out sooner than the highly soluble minerals. The deposition order and saturation percentages are depicted in the table, bearing in mind the process in nature may vary from laboratory-derived values [19].



Figure 5.3.13: Salt-covered plain known as the Bonneville Salt Flats, Utah.

Table 5.3.1: Deposition order and saturation percentages. Table after [20].

Mineral sequence	Percent Seawater remaining after evaporation
Calcite	50%
Gypsum/anhydrite	20%
Halite	10%
Various potassium and magnesium salts	5%

Calcium carbonate-saturated water precipitates porous masses of calcite called **tufa**. Tufa can form near degassing water and in saline lakes. Waterfalls downstream of springs often precipitate tufa as the turbulent water enhances the degassing of carbon dioxide, which makes calcite less soluble and causes it to precipitate. Saline lakes concentrate calcium carbonate from a combination of wave action causing degassing, springs in the lakebed, and evaporation. In salty Mono Lake in California, tufa towers were exposed after water was diverted and lowered the lake levels.



Figure 5.3.14: Limestone tufa towers along the shores of Mono Lake, California.

Cave deposits like **stalactites** and **stalagmites** are another form of chemical precipitation of calcite, in a form called **travertine**. Calcite slowly precipitates from water to form the travertine, which often shows banding. This process is similar to the mineral growth on faucets in your home sink or shower that comes from hard (mineral-rich) water. Travertine also forms at hot springs such as Mammoth Hot Spring in Yellowstone National Park.



Figure 5.3.15: Travertine terraces of Mammoth Hot Springs, Yellowstone National Park.

Banded iron formation deposits commonly formed early in Earth's history, but this type of chemical sedimentary rock is no longer being created. Oxygenation of the atmosphere and oceans caused free iron ions, which are water-soluble, to become oxidized and precipitate out of solution. The iron oxide was deposited, usually in bands alternating with layers of chert.

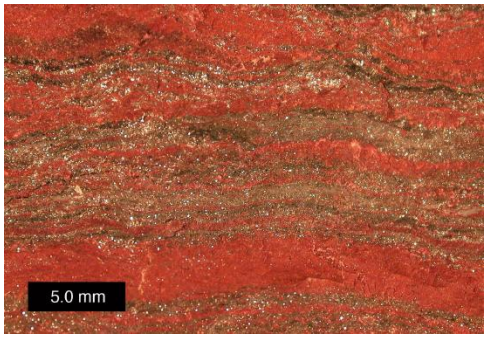


Figure 5.3.16: Alternating bands of iron-rich and silica-rich mud, formed as oxygen combined with dissolved iron.

Chert, another commonly found chemical sedimentary rock, is usually produced from silica (SiO_2) precipitated from groundwater. Silica is highly insoluble on the surface of Earth, which is why quartz is so resistant to chemical weathering. Water deep underground is subjected to higher pressures and temperatures, which helps dissolve silica into an aqueous solution. As the groundwater rises toward or emerges at the surface, the silica precipitates out, often as a cementing agent or into **nodules**. For example, the bases of the geysers in Yellowstone National Park are surrounded by silica deposits called geyserite or **sinter**. The

silica is dissolved in water that is thermally heated by a relatively deep magma source. Chert can also form biochemically and is discussed in the Biochemical subsection. Chert has many synonyms, some of which may have gem value such as jasper, flint, onyx, and agate, due to subtle differences in colors, striping, etc., but chert is the more general term used by geologists for the entire group.



Figure 5.3.17: A type of chert, flint, shown with a lighter weathered crust.

Oolites are among the few limestone forms created by an inorganic chemical process, similar to what happens in evaporite deposition. When water is oversaturated with calcite, the mineral precipitates out around a nucleus, a sand grain or shell fragment, and forms little spheres called **ooids** (see figure). As evaporation continues, the ooids continue building concentric layers of calcite as they roll around in gentle currents.



Figure 5.3.18: Ooids in an oolite.

Biochemical Sedimentary Rocks

Biochemical sedimentary rocks are not that different from chemical sedimentary rocks; they are also form from ions dissolved in solution. However, biochemical sedimentary rocks rely on biological processes to extract the dissolved materials out of the water. Most macroscopic marine organisms use dissolved minerals, primarily aragonite (calcium carbonate), to build hard parts such as shells. When organisms die, the hard parts settle as sediment, which becomes buried, compacted, and cemented into rock.

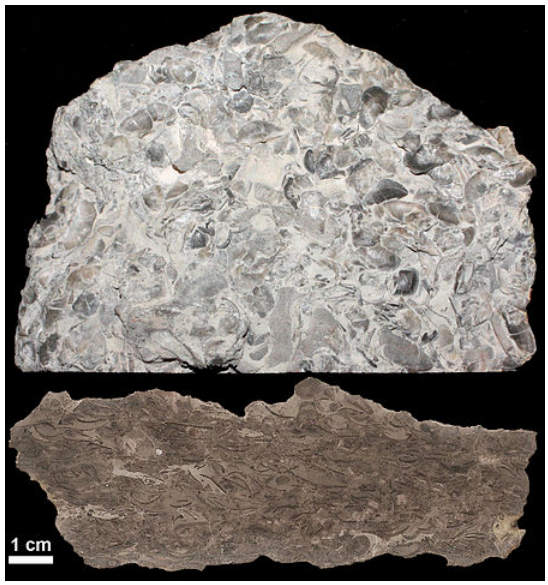


Figure 5.3.19: Fossiliferous limestone (with brachiopods and bryozoans) from the Kope Formation of Ohio. Lower image is a section of the rock that has been etched with acid to emphasize the fossils.

This biochemical extraction and secretion is the main process for forming **limestone**, the most commonly occurring, non-clastic sedimentary rock. Limestone is mostly made of calcite ($CaCO_3$) and sometimes includes dolomite ($CaMg(CO_3)_2$), a close relative. Solid calcite reacts with hydrochloric acid by effervescing or fizzing. Dolomite only reacts to hydrochloric acid when ground into a powder, which can be done by scratching the rock surface.

Limestone occurs in many forms, most of which originate from biological processes. Entire coral reefs and their ecosystems can be preserved in exquisite detail in limestone rock. **Fossiliferous limestone** contains many visible fossils. A type of limestone called **coquina** originates from beach sands made predominantly of shells that were then lithified. Coquina is composed of loosely-cemented shells and shell fragments. You can find beaches like this in modern tropical environments, such as the Bahamas. **Chalk** contains high concentrations of shells from a microorganism called a coccolithophore. **Micrite**, also known as microscopic calcite mud, is a very fine-grained limestone containing microfossils that can only be seen using a microscope.



Figure 5.3.20: Close-up of coquina, a shell hash.

Biogenetic chert forms on the deep ocean floor, created from biochemical sediment made of microscopic organic shells. This sediment, called **ooze**, may be calcareous (calcium carbonate-based) or siliceous (silica-based) depending on the type of shells deposited. For example, the shells of radiolarians (zooplankton) and diatoms (phytoplankton) are made of silica, so they produce siliceous ooze.

Organic Sedimentary Rocks

Under the right conditions, intact pieces of organic material or material derived from organic sources are preserved in the geologic record. Although not derived from sediment, this lithified organic material is associated with sedimentary strata and created by similar processes—burial, compaction, and diagenesis. Deposits of these, which are used as fuel, develop in areas where organic material collects in large quantities. Lush swamplands can create conditions conducive to **coal** formation. Shallow-water, organic

material-rich marine sediment can become highly productive **petroleum** and **natural gas** deposits. These are examples of fossil-derived energy sources.



Figure 5.3.21: Bituminous coal, frequently used for fuel. (By Amcyrus2012; CC BY 4.0 via Wikimedia Commons.)

Classification of Chemical Sedimentary Rocks

In contrast to detrital sediment, chemical, biochemical, and organic sedimentary rocks are classified based on mineral composition. Most of these are **monomineralic**, composed of a single mineral, so the rock name is usually associated with the identifying mineral. Chemical sedimentary rocks consisting of halite are called **rock salt**. Rocks made of limestone (calcite) are an exception, having elaborate subclassifications and even two competing classification methods: Folk Classification and Dunham Classification [11; 21]. The Folk Classification deals with rock grains and usually requires a specialized, petrographic microscope. The Dunham Classification is based on rock texture, which is visible to the naked eye or using a hand lens and is easier for field applications. Most carbonate geologists use the Dunham system.



Figure 5.3.22: Gyprock, a chemical sedimentary rock made of the mineral gypsum. From the Castile formation of New Mexico.

A chart of types of sedimentary rocks with associated characteristics for identification is shown in the figure below.

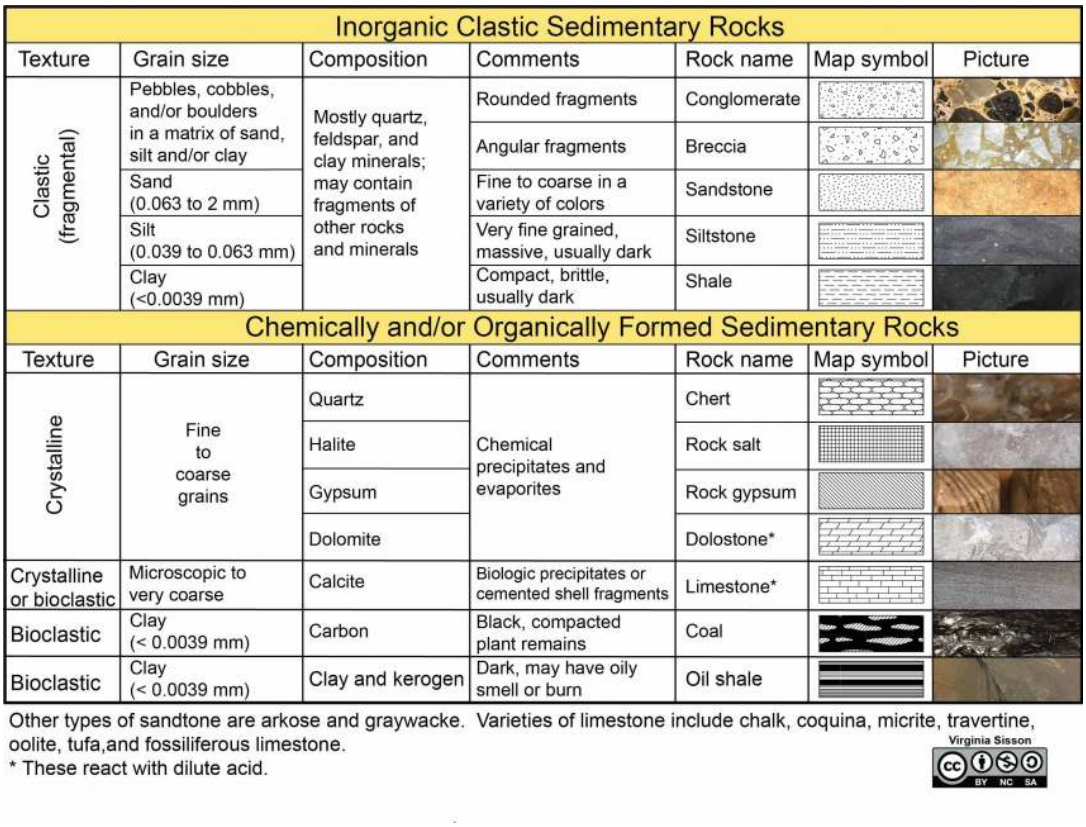


Figure 5.3.23: Sedimentary rock identification chart. (By Virginia Sisson; CC BY-NC-SA 4.0 via The Story of Earth.)

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5.4: Sedimentary Structures

Sedimentary structures are visible textures or arrangements of sediments within a rock. Geologists use these structures to interpret the processes that made the rock and the environment in which it formed. They compare sedimentary structures formed in modern environments to lithified counterparts in ancient rocks. Below is a summary discussion of common sedimentary structures that are useful for interpretations in the rock record.

Bedding Planes

The most basic sedimentary structure is **bedding planes**, the planes that separate the layers or **strata** in sedimentary and some volcanic rocks. Visible in exposed outcroppings, each bedding plane indicates a change in sediment deposition conditions. This change may be subtle. For example, if a section of underlying sediment firms up, this may be enough to create a form or a layer that is dissimilar from the overlying sediment. Each layer is called a **bed**, or **stratum**, the most basic unit of **stratigraphy**, the study of sedimentary layering.



Figure 5.4.1: Horizontal strata in southern Utah.

As would be expected, bed thickness can indicate sediment deposition quantity and timing. Technically, a bed is a bedding plane thicker than 1 cm (0.4 in) and the smallest mappable unit. A layer thinner than 1 cm (0.4 in) is called a **lamina** [22]. **Varves** are bedding planes created when laminae and beds are deposited in repetitive cycles, typically daily or seasonally [23]. Varves are valuable geologic records of climatic histories, especially those found in lakes and glacial deposits.



Figure 5.4.2: Students from the University of Wooster examine beds of Ordovician limestone in central Tennessee.

Graded Bedding

Graded bedding refers to a sequence of increasingly coarse- or fine-grained sediment layers. Graded bedding often develops when sediment deposition occurs in an environment of decreasing energy. A **Bouma sequence** is graded bedding observed in a clastic rock called **turbidite** [24]. Bouma sequence beds are formed by offshore **sediment gravity flows**, which are underwater flows of sediment. These subsea density flows begin when sediment is stirred up by an energetic process and becomes a dense slurry of

mixed grains. The sediment flow courses downward through submarine channels and canyons due to gravity acting on the density difference between the denser slurry and less dense surrounding seawater. As the flow reaches deeper ocean basins it slows down, loses energy, and deposits sediment in a Bouma sequence of coarse grains first, followed by increasingly finer grains.

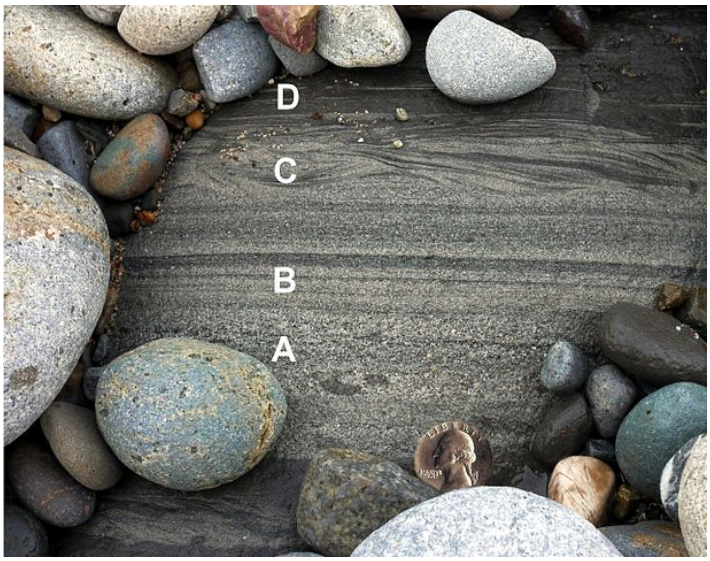


Figure 5.4.3: Image of the classic Bouma sequence. A=coarse- to fine-grained sandstone, possibly with an erosive base. B=laminated medium- to fine-grained sandstone. C=rippled fine-grained sandstone. D=laminated siltstone grading to mudstone.

Flow Regime and Bedforms

In fluid systems, such as moving water or wind, sand is the most easily transported and deposited sediment grain. Smaller particles like silt and clay are less movable by fluid systems because the tiny grains are chemically attracted to each other and stick to the underlying sediment. Under higher flow rates, the fine silt and clay sediment tend to stay in place and the larger sand grains get picked up and moved.

Bedforms are sedimentary structures created by fluid systems working on sandy sediment [25]. Grain size, flow velocity, and **flow regime** or pattern interact to produce bedforms having unique, identifiable physical characteristics. Flow regimes are divided into **upper** and **lower regimes**, which are further divided into uppermost, upper, lower, and lowermost parts. The table below shows bedforms and their associated flow regimes. For example, the dunes bedform is created in the upper part of the lower flow regime.

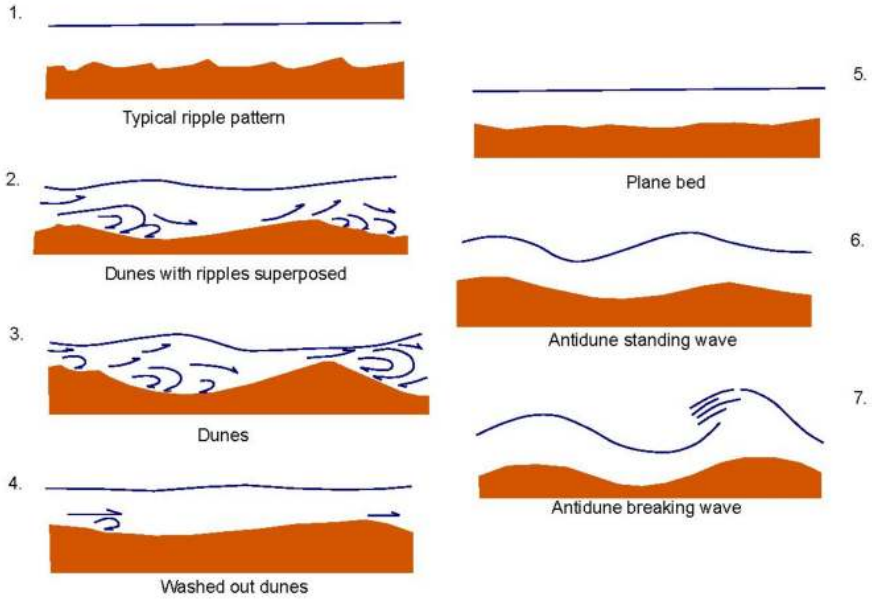


Figure 5.4.4: Bedforms from under increasing flow velocities.

Table 5.4.1: Bedforms and their associated flow regimes.

Flow Regime (part)	Bedform	Description
Lower (lowest)	Plane bed	Lower plane bed, flat laminations
Lower (lower)	Ripples	Small (with respect to flow) inclined layers dipping downflow
Lower (upper)	Dunes	Larger inclined cross beds, ±ripples, dipping downflow
Upper (lower)	Plane bed	Flat layers, can include lined-up grains (parting lineations)
Upper (upper)	Antidunes	Hard to preserve reverse dunes dipping shallowly upflow
Upper (uppermost)	Chutes/pools *(rare)*	Erosional, not really a bedform; rarely found preserved

Plane Beds

Plane beds created in the lower flow regime are like bedding planes, on a smaller scale. The flat, parallel layers form as sandy sediment piles and moves on top of layers below. Even non-flowing fluid systems, such as lakes, can produce sediment plane beds. Plane beds in the upper flow regime are created by fast-flowing fluids. They may look identical to lower-flow-regime beds; however, they typically show **parting lineations**, slight alignments of grains in rows and swaths, caused by high sediment transport rates that only occur in upper flow regimes.



Figure 5.4.5: Subtle lines across this sandstone (trending from the lower left to upper right) are parting lineations.

Ripples

Ripples are known by several names: **ripple marks**, ripple cross-beds, or ripple cross laminations. The ridges or undulations in the bed are created as sediment grains pile up on top of the plane bed. With the exception of dunes, the scale of these beds is typically measured in centimeters. Occasionally, large flows like glacial lake outbursts can produce ripples as tall as 20 m (66 ft).



Figure 5.4.6: Left: Modern current ripple in sand from the Netherlands. The flow creates a steep side down current. In this image, the flow is from right to left. Right: A bidirectional flow creates this symmetrical wave ripple in rocks from Nomgon, Mongolia. Note the crests of the ripples have been eroded away by subsequent flows in places.

First scientifically described by Hertha Ayrton [26], ripple shapes are determined by flow type and can be straight-crested, sinuous, or complex. Asymmetrical ripples form in a unidirectional flow. Symmetrical ripples are the result of an oscillating back-and-forth flow typical of intertidal swash zones. Climbing ripples are created from high sedimentation rates and appear as overlapping layers of ripple shapes.



Figure 5.4.7: Climbing ripple deposit from India.

Dunes

Dunes are very large and prominent versions of ripples and typical examples of large cross-bedding [27]. **Cross bedding** happens when ripples or dunes pile atop one another, interrupting, and/or cutting into the underlying layers. Desert sand dunes are probably the first image conjured up by this category of bedform.



Figure 5.4.8: Lithified cross-bedded dunes from the high country of Zion National Park, Utah. The complexity of bedding planes results from the three-dimensional network of ancient dune flows.

Dunes are the most common sedimentary structure found within channelized flows of air or water. The biggest difference between river dunes and air-formed (desert) dunes is the depth of the fluid system. Since the atmosphere's depth is immense when compared to a river channel, desert dunes are much taller than those found in rivers. Some famous air-formed dune landscapes include the Sahara Desert, Death Valley, and the Gobi Desert [28].



Figure 5.4.9: The modern sand dune in Morocco.

As airflow moves sediment along, the grains accumulate on the dune's **windward surface** (facing the wind). The angle of the windward side is typically shallower than the **leeward** (downwind) side, which has grains falling down over it. This difference in slopes can be seen in a bed cross-section and indicates the direction of the flow in the past. There are typically two styles of dune beds: the more common **trough cross-beds** with curved windward surfaces, and rarer **planar cross-beds** with flat windward surfaces.

In tidal locations with strong in-and-out flows, dunes can develop in opposite directions. This produces a feature called **herringbone cross-bedding**.



Figure 5.4.10: Herringbone cross-bedding from the Mazomanie Formation, upper Cambrian of Minnesota.

Another dune formation variant occurs when very strong, hurricane-strength, winds agitate parts of the usually undisturbed seafloor. These beds are called **hummocky cross-stratification** and have a 3D architecture of hills and valleys, with inclined and declined layering that matches the dune shapes.



Figure 5.4.11: Hummocky-cross stratification, seen as bedding planes within bedding planes in this rock from Australia. (By Rygel, M.C.; CC BY-SA 4.0 via Wikimedia Commons.)

Antidunes

Antidunes are so named because they share similar characteristics with dunes, but are formed by a different, opposing process [29]. While dunes form in lower flow regimes, antidunes come from fast-flowing upper flow regimes. In certain conditions of high flow rates, the sediment accumulates upstream of a subtle dip instead of traveling downstream. Antidunes form in phase with the flow; in rivers, they are marked by rapids in the current. Antidunes are rarely preserved in the rock record because the high flow rates needed to produce the beds also accelerate erosion.



Figure 5.4.12: Antidunes forming in Urdaibai, Spain.

Bioturbation

Bioturbation is the result of organisms burrowing through soft sediment, which disrupts the bedding layers. These tunnels are backfilled and eventually preserved when the sediment becomes rock. Bioturbation happens most commonly in shallow, marine environments, and can be used to indicate water depth [30].



Figure 5.4.13: Bioturbated sandstone from Australia showing tunnels created by burrowing organisms. (By Rygel, M.C.; CC BY-SA 4.0 via Wikimedia Commons.)

Mudcracks

Mudcracks occur in clay-rich sediment that is submerged underwater and later dries out. Water fills voids in the clay's crystalline structure, causing the sediment grains to swell. When this waterlogged sediment begins to dry out, the clay grains shrink. The sediment layer forms deep polygonal cracks with tapered openings toward the surface [31], which can be seen in profile. The cracks fill with new sediment and become visible veins running through the lithified rock. These dried-out clay beds are a major source of **mud chips**, small fragments of mud or shale, which commonly become inclusions in sandstone and conglomerate. What makes this sedimentary structure so important to geologists is that they only form in certain depositional environments—such as tidal flats that form underwater and are later exposed to air. **Syneresis cracks** are similar in appearance to mud cracks but much rarer; they are formed when subaqueous (underwater) clay sediment shrinks [32].

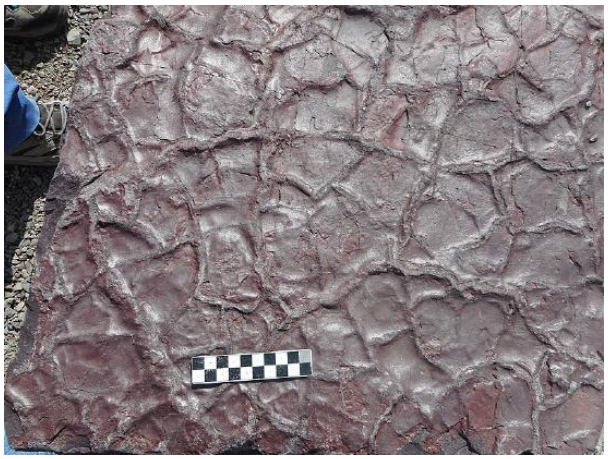


Figure 5.4.14: Mudcracks in the Grinnell Formation, Glacier National Park. (By Carpenter, Kenneth; CC BY-SA 4.0 via Wikimedia Commons.)

Sole Marks

Sole marks are small features typically found in river deposits. They form at the base of a bed, the sole, and on top of the underlying bed. They can indicate several things about the deposition conditions, such as flow direction or stratigraphic up-direction. **Flute casts** or **scour marks** are grooves carved out by the forces of fluid flow and sediment loads. The upstream part of the flow creates steep grooves and downstream the grooves are shallower. The grooves subsequently become filled by overlying sediment, creating a cast of the original hollow [33].

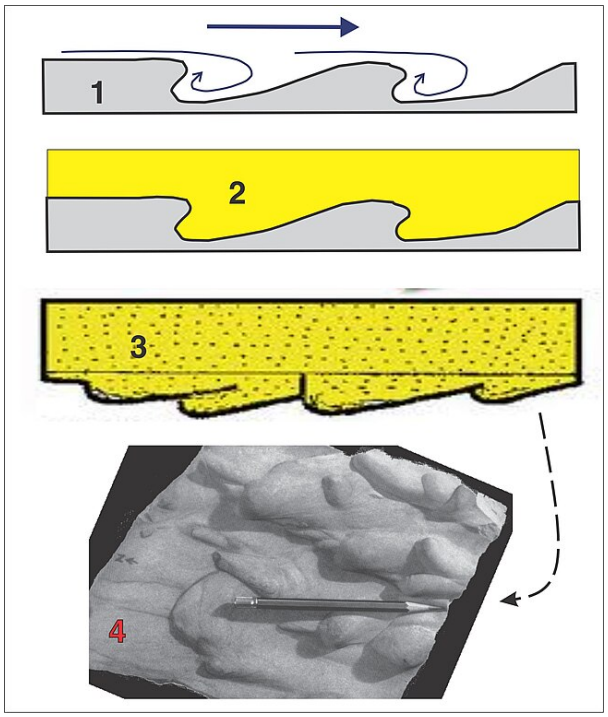


Figure 5.4.15: Formation of a flute cast. 1. Flowing water creates grooves in the sediment. 2. New sediment is deposited on top which preserves the grooves. 3. The sediment is lithified, with the mold of the grooves. 4. The rock has blob-like bulges, flute casts, and is the top layer of sediment that covered the grooves. (By Ceclre.Caballero; CC BY-SA 4.0 via Wikimedia Commons.)

Formed similarly to flute casts but with a more regular and aligned shape, **groove casts** are produced by larger clasts or debris carried along in the water that scrape across the sediment layer. **Tool marks** come from objects like sticks carried in the fluid downstream or embossed into the sediment layer, leaving a depression that later fills with new sediment.



Figure 5.4.16: Tool marks and groove casts on the base of a sandstone. (By Rygel, M.C.; CC BY-SA 4.0 via Wikimedia Commons.)

Load casts, an example of **soft-sediment deformation**, are small indentations made by an overlying layer of coarse sediment grains or clasts intruding into a softer, finer-grained sediment layer [34].



Figure 5.4.17: Load casts on the base of a bed of sandstone. (By Rygel, M.C.; CC BY-SA 4.0 via Wikimedia Commons.)

Raindrop Impressions

Like their name implies, **raindrop impressions** are small pits or bumps found in soft sediment. While they are generally believed to be created by rainfall, they may be caused by other agents such as escaping gas bubbles [35].



Figure 5.4.18: Mississippian raindrop impressions over wave ripples from Nova Scotia.

Imbrication

Imbrication is a stack of large and usually flat clasts—cobbles, gravels, mud chips, etc—that are aligned in the direction of fluid flow [36]. The clasts may be stacked in rows, with their edges dipping down and flat surfaces aligned to face the flow. Or their flat surfaces may be parallel to the layer and long axes aligned with the flow. Imbrications are useful for analyzing **paleocurrents**, or currents found in the geologic past, especially in alluvial deposits.



Figure 5.4.19: Cobbles in this conglomerate are positioned in a way that they are stacked on each other, which occurred as flow went from left to right.

Geopetal Structures

Geopetal structures [37], also called **up-direction indicators**, are used to identify which way was up when the sedimentary rock layers were originally formed. This is especially important in places where the rock layers have been deformed, tilted, or overturned. Well-preserved mud cracks, sole marks, and raindrop impressions can be used to determine up direction.



Figure 5.4.20: This bivalve (clam) fossil was partially filled with tan sediment, partially empty. Later fluids filled in the fossil with white calcite minerals. The line between the sediment and the later calcite is paleo-horizontal.

Other useful geopetal structures include:

- **Vugs**: Small voids in the rock that usually become filled during diagenesis. If the void is partially filled or filled in stages, it serves as a permanent record of a level bubble, frozen in time.
- Cross bedding – In places where ripples or dunes pile on top of one another, where one cross bed interrupts and/or cuts another below, this shows a cross-cutting relationship that indicates up direction.
- Ripples, dunes: Sometimes the ripples are preserved well enough to differentiate between the crests (top) and troughs (bottom).
- Fossils: Body fossils in life position, meaning the body parts are not scattered or broken, and trace fossils like footprints can provide an up direction. Intact fossilized coral reefs are excellent up indicators because of their large size and easily distinguishable top and bottom. Index fossils, such as ammonites, can be used to age date strata and determine up direction based on relative rock ages.
- Vesicles – Lava flows eliminate gas upwards. An increase of vesicles toward the top of the flow indicates up.



Figure 5.4.21: Trace fossil Eubrontes, a dinosaur footprint, from Utah. It shows the geopetal direction is into the image.

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5.5: Depositional Environments

Depositional Environments

The ultimate goal of many stratigraphy studies is to understand the original depositional environment. Knowing where and how a particular sedimentary rock was formed can help geologists paint a picture of past environments—such as a mountain glacier, gentle floodplain, dry desert, or deep-sea ocean floor. The setting in which sediments are accumulated is called a **depositional environment**. In the case of clastic rocks, the key environmental conditions are related to the amount of energy available to transport sediments, and how far the sediments get from their source before being deposited.

Broadly, depositional environments can be said to be terrestrial, marine, or to reflect a transitional zone between the two. **Terrestrial** refers to depositional environments on land. These may be depositional environments such as deserts found on dry land, but they could also be environments such as freshwater lakes or rivers. Marine refers to environments associated with saltwater seas and oceans. Transitional depositional environments include environments such as deltas, where freshwater rivers empty into saltwater seas or oceans.

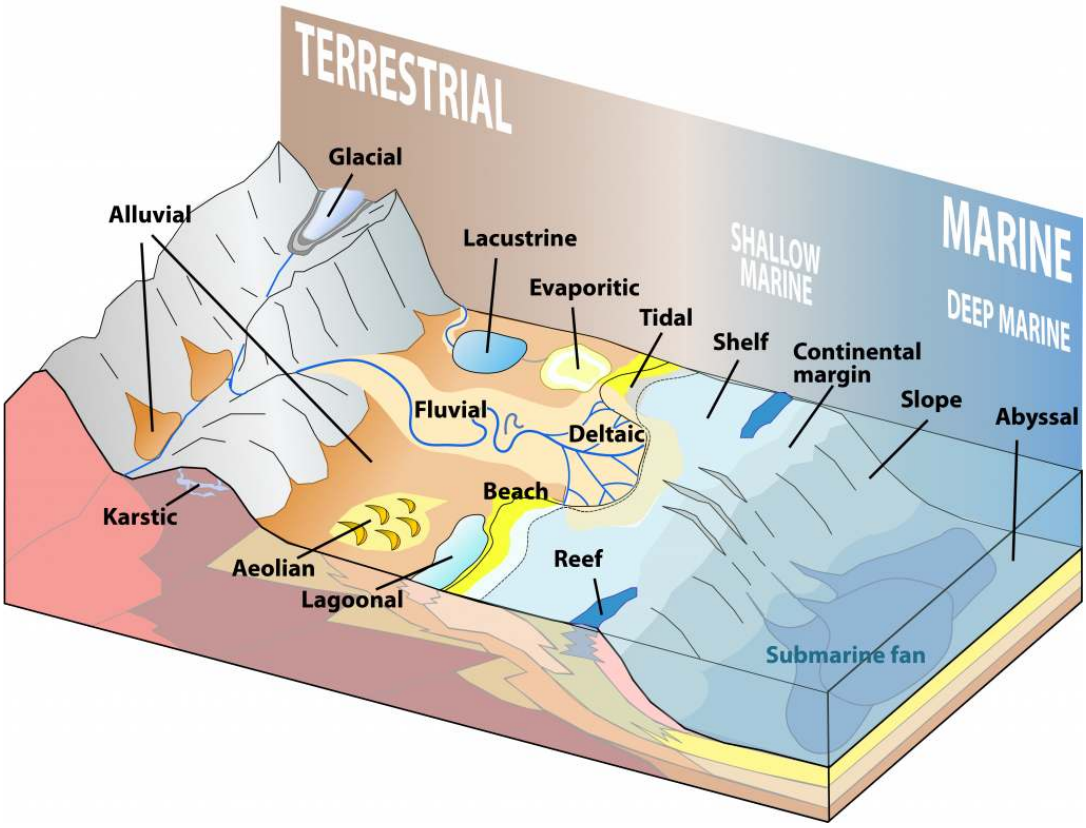


Figure 5.5.1: Some of the important depositional environments for sediments and sedimentary rocks. Source: Karla Panchuk (2021) CC BY-SA 4.0. Modified after Mike Norton (2018) CC BY-SA 3.0. View source.

Tables 5.5.1 and 5.5.2 provide a summary of the processes and sediment types that pertain to the various depositional environments illustrated in the figure above.

Table 5.5.1: Terrestrial Depositional Environments.

Environment	Key Transport Processes	Depositional Settings	Typical Sediments
Glacial	Gravity, moving ice, moving water	Valleys, plains, streams, lakes	Glacial till, gravel, sand, silt, clay
Alluvial	Gravity, moving water	Where steep-sided valleys meet plains	Coarse angular fragments
Fluvial	Moving water	Streams	Gravel, sand, silt, organic matter
Aeolian	Wind	Deserts and coastal regions	Sand, silt
Lacustrine	Moving Water	Lakes	Sand, silt, clay, organic matter
Evaporite	Still water	Lakes in arid regions	Salts, clay
Source: Karla Panchuk (2018), CC BY 4.0. Modified after Steven Earle (2015), CC BY 4.0. View source.			

Table 5.5.2: Marine and Transitional Depositional Environments.

Environment	Key Transport Processes	Depositional Settings	Typical Sediments
Deltaic	Moving water	Deltas	Sand, silt, clay, organic matter
Beach	Waves, long-shore currents	Beaches, spits, sand bars	Gravel, sand
Tidal	Tidal currents	Tidal flats	Fine-grained sand, silt, clay
Reef	Waves, tidal currents	Reefs and adjacent basins	Carbonates
Shallow marine	Waves, tidal currents	Shelves, slopes, lagoons	Carbonates in tropical climates; sand/silt/clay elsewhere
Lagoonal	Little transportation	Lagoon bottom	Carbonates in tropical climates; silt, clay elsewhere
Submarine fan	Underwater gravity flows	Continental slopes, abyssal plains	Gravel, sand, silt, clay
Deep water	Ocean currents	Deep-ocean abyssal plains	Clay, carbonate mud, silica mud
Source: Karla Panchuk (2018), CC BY 4.0. Modified after Steven Earle (2015), CC BY 4.0. View source.			

Marine and transitional depositional environments are described in the chapter on Coastlines. Terrestrial depositional environments are described next.

Terrestrial Depositional Environments

Terrestrial depositional environments are diverse. Water is a major factor in these environments, in liquid or frozen states, or even when it is lacking (arid conditions). The table shows the sediment, rocks, fossils and sedimentary structures typically found in terrestrial depositional environments.

Table 5.5.3: Rock record and terrestrial depositional environments

Location	Sediment	Common Rock Types	Typical Fossils	Sedimentary Structures
Fluvial (river)	sand and mud, can have larger sediments	sandstone, conglomerate	bone beds (rare)	cross beds, channels, asymmetric ripples
Alluvial	mud to boulders, poorly sorted	clastic rocks	rare	channels, mud cracks
Lacustrine (lake)	fine-grained laminations	shale	invertebrates, rare (deep) bone beds	laminations
Paludal (swamp)	plant material	coal	plant debris	rare
Aeolian (dunes)	very well-sorted sand and silt	sandstone	rare	cross beds (large)
Glacial	mud to boulders, poorly sorted	conglomerate (tillite)		striations, drop stones

Fluvial

Fluvial (river) systems are formed by water flowing in channels over the land. They generally come in two main varieties: meandering or braided. In **meandering** streams, the flow carries sediment grains via a single channel that wanders back and forth across the floodplain. The sediment deposited farther away from the channel is mostly fine-grained material that only gets deposited during floods.



Figure 5.5.2: The Cauto River in Cuba. Note the sinuosity in the river, which is meandering.

Braided fluvial systems generally contain coarser sediment grains, and form a complicated series of intertwined channels that flow around gravel and sand bars [49].



Figure 5.5.3: The braided Waimakariri river in New Zealand.

Alluvial

A distinctive characteristic of **alluvial** systems is the intermittent flow of water. Alluvial deposits are common in arid places with little soil development. Lithified alluvial beds are the primary basin-filling rock found throughout the Basin and Range region of the western United States. The most distinctive alluvial sedimentary deposit is the **alluvial fan**, a large cone of sediment formed by streams flowing out of dry mountain valleys into a wider and more open dry area. Alluvial sediments are typically poorly sorted and coarse-grained, and often found near playa lakes or aeolian deposits [50].



Figure 5.5.4: An alluvial fan spreads out into a broad alluvial plain. From Red Rock Canyon State Park, California.

Lacustrine

Lake systems and deposits, called **lacustrine**, often form as sediment from rivers is dumped into a lake and slowly settles to the bottom. Lacustrine deposits are found in lakes in a wide variety of locations. Lake Baikal in southeast Siberia (Russia) is in a tectonic basin. Crater Lake (Oregon) sits in a volcanic caldera. The Great Lakes (northern United States) came from glacially carved and deposited sediment. Ancient Lake Bonneville (Utah) formed in a **pluvial** (from rainfall) setting during a climate that was relatively wetter and cooler than that of modern Utah. Oxbow lakes, named for their curved shape, originated in fluvial floodplains.



Figure 5.5.5: Oregon's Crater Lake was formed about 7700 years ago after the eruption of Mount Mazama.

Lacustrine sediment tends to be very fine-grained and thinly laminated, with only minor contributions from wind-blown, current, and tidal deposits [51]. When lakes dry out or evaporation outpaces precipitation, playas form. **Playa** deposits resemble those of normal lake deposits but contain more evaporite minerals. Certain tidal flats can have playa-type deposits as well.

Paludal

Paludal systems include bogs, marshes, swamps, or other wetlands, and usually contain lots of organic matter. Paludal systems typically develop in coastal environments but are common in humid, low-lying, low-latitude, warm zones with large volumes of flowing water. A characteristic paludal deposit is a peat bog, a deposit rich in organic matter that can be converted into coal when lithified. Paludal environments may be associated with tidal, deltaic, lacustrine, and/or fluvial deposition.

Aeolian

Aeolian, sometimes spelled eolian or œolian, are deposits of windblown sediments. Since wind has a much lower carrying capacity than water, aeolian deposits typically consist of clast sizes from fine dust to sand [52]. Fine silt and clay can cross very long

distances, even entire oceans suspended in the air.

With sufficient sediment influx, aeolian systems can potentially form large dunes in dry or wet conditions. The figure shows dune features and various types. Parabolic and linear dunes grow from sand anchored by plants and are common in coastal areas.

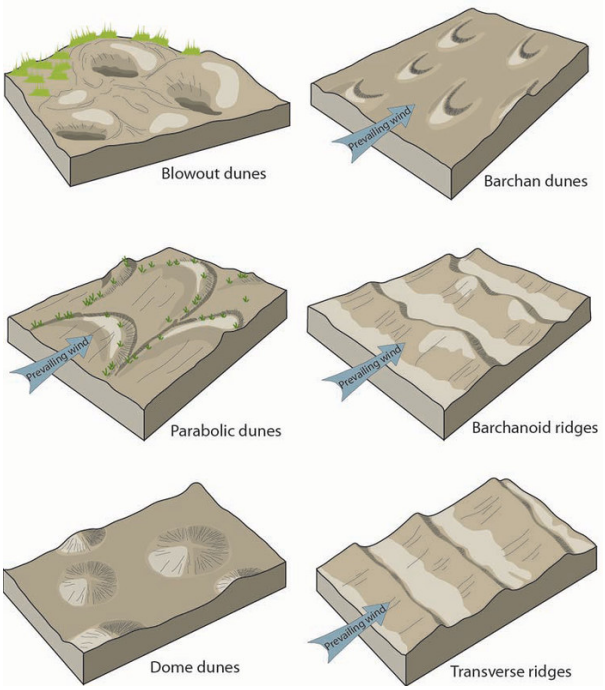


Figure 5.5.6: Formation and types of dunes. (By NPS; public domain via Flickr.)

Compacted layers of wind-blown sediment are known as **loess**. Loess commonly starts as finely ground-up rock flour created by glaciers. Such deposits cover thousands of square miles in the Midwestern United States. Loess may also form in desert regions. Silt for the Loess Plateau in China came from the Gobi Desert in China and Mongolia.

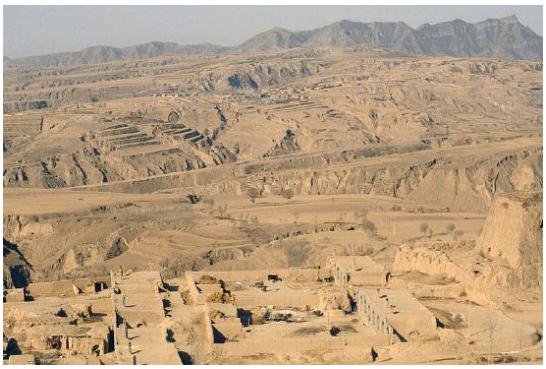


Figure 5.5.7: Loess Plateau in China. The loess is so highly compacted that buildings and homes have been carved in it.

Glacial

Glacial sedimentation is very diverse and generally consists of the most poorly-sorted sediment deposits found in nature. The main rock type is called **diamictite**, which literally means two sizes, referring to the unsorted mix of large and small rock fragments found in glacial deposits [53]. Many glacial **tills**, glacially derived sediments, include very finely-pulverized rock flour along with giant erratic boulders. The surfaces of larger clasts typically have striations from the rubbing, scraping, and polishing of surfaces by abrasion during the movement of glacial ice. Glacial systems are so large and produce so much sediment, they frequently create multiple, individualized depositional environments, such as fluvial, deltaic, lacustrine, pluvial, alluvial, and/or aeolian.



Figure 5.5.8: Wide range of sediments near Athabaska Glacier, Jasper National Park, Alberta, Canada.

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5.S: Summary

Sedimentary rocks are grouped into two main categories: clastic (detrital) and chemical. Clastic (detrital) rocks are made of mineral clasts or sediment that lithifies into solid material. Sediment is produced by the mechanical or chemical weathering of bedrock and transported away from the source via erosion. Sediment that is deposited, buried, compacted, and sometimes cemented becomes clastic rock. Clastic rocks are classified by grain size; for example sandstone is made of sand-sized particles. Chemical sedimentary rocks come from minerals precipitated out of an aqueous solution and are classified according to mineral composition. The chemical sedimentary rock limestone is made of calcium carbonate. Sedimentary structures have textures and shapes that give insight on depositional histories. Depositional environments depend mainly on fluid transport systems and encompass a wide variety of underwater and above ground conditions. Geologists analyze depositional conditions, sedimentary structures, and rock records to interpret the paleogeographic history of a region.

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CHAPTER OVERVIEW

6: Metamorphic Rocks

Learning Objectives

- Describe the temperature and pressure conditions of the metamorphic environment.
- Identify and describe the three principal metamorphic agents.
- Describe what recrystallization is and how it affects mineral crystals.
- Explain what foliation is and how it results from directed pressure and recrystallization.
- Explain the relationships among slate, phyllite, schist, and gneiss in terms of metamorphic grade.
- Define index mineral.
- Explain how metamorphic facies relate to plate tectonic processes.
- Describe what a contact aureole is and how contact metamorphism affects surrounding rock.
- Describe the role of hydrothermal metamorphism in forming mineral deposits and ore bodies.



Figure 6.1: Painted Wall of Black Canyon of the Gunnison National Park, Colorado, made of 1.7 billion-year-old gneiss intruded by younger pegmatites.

Metamorphic rock, *meta-* meaning change and *–morphos* meaning form, is one of the three rock categories in the rock cycle. Metamorphic rock material has been changed by temperature, pressure, and/or fluids. The **rock cycle** shows that both igneous and sedimentary rocks can become metamorphic rocks. And metamorphic rocks themselves can be re-metamorphosed. Because metamorphism is caused by plate tectonic motion, metamorphic rocks provide geologists with a history book of how past tectonic processes shaped our planet [1].

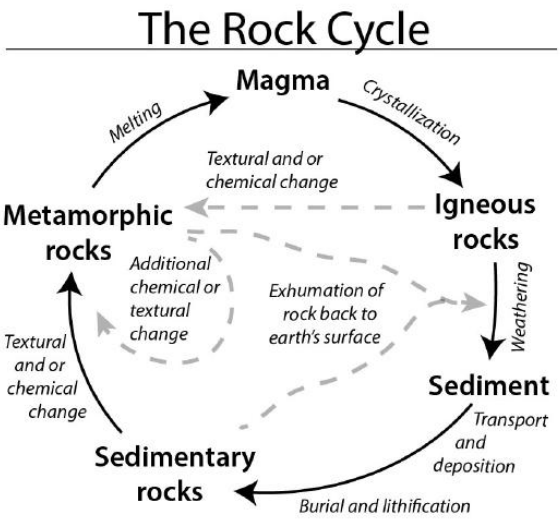


Figure 6.2: Rock cycle showing the five materials (such as igneous rocks and sediment) and the processes by which one changes into another (such as weathering). (Source: Peter Davis)

6.1: Metamorphic Processes
6.2: Metamorphic Textures
6.3: Metamorphic Grade
6.4: Metamorphic Environments
6.S: Summary

Reference

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6.1: Metamorphic Processes

Metamorphism occurs when solid rock changes in composition and/or texture without the mineral crystals melting, which is how igneous rock is generated. Metamorphic source rocks, the rocks that experience the metamorphism, are called the **parent rock** or **protolith**, from *proto–* meaning first, and *lithos-* meaning rock. Most metamorphic processes take place deep underground, inside the Earth's crust. During metamorphism, protolith chemistry is mildly changed by increased temperature (heat), a type of pressure called confining pressure, and/or chemically reactive fluids. Rock texture is changed by heat, confining pressure, and a type of pressure called directed stress.

Temperature (Heat)

Temperature measures a substance's energy—an increase in temperature represents an increase in energy [2]. Temperature changes affect the chemical equilibrium or cation balance in minerals. At high temperatures, atoms may vibrate so vigorously they jump from one position to another within the crystal lattice, which remains intact. In other words, this atom swapping can happen while the rock is still solid.

The temperatures of metamorphism lie in between surficial processes (as in sedimentary rock) and magma in the rock cycle. Heat-driven metamorphism begins at temperatures as cold as 200˚C and can continue to occur at temperatures as high as 700°C to 1100°C [3; 4; 5]. Higher temperatures would create magma, and thus, would no longer be a metamorphic process. Temperature increases with increasing depth in the Earth along the geothermal gradient and metamorphic rock records these depth-related temperature changes.

Pressure

Pressure is the force exerted over a unit area on a material. Like heat, pressure can affect the chemical equilibrium of minerals in a rock. The pressure that affects metamorphic rocks can be grouped into confining pressure and directed stress. **Stress** is a scientific term indicating a force. **Strain** is the result of this stress, including metamorphic changes within minerals.

Confining Pressure

Pressure exerted on rocks under the surface is due to the simple fact that rocks lie on top of one another. When pressure is exerted from rocks above, it is balanced from below and sides and is called **confining** or **lithostatic pressure**. Confining pressure has equal pressure on all sides (see figure) and is responsible for causing chemical reactions to occur, just like heat. These chemical reactions will cause new minerals to form.

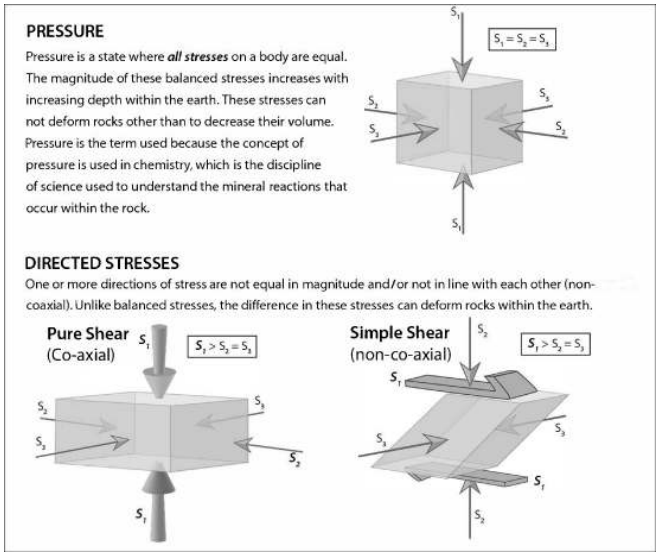


Figure 6.1.1: Difference between pressure and stress and how they deform rocks. Pressure (or confining pressure) has equal stress (forces) in all directions and increases with depth under the Earth's surface. Under directed stress, some stress directions (forces) are stronger than others, and this can deform rocks. Source: Peter Davis

Confining pressure is measured in bars and ranges from 1 bar at sea level to around 10,000 bars at the base of the crust [6]. For metamorphic rocks, pressures range from a relatively low pressure of 3000 bars [5] to around 50,000 bars [5], which occurs around 15-35 kilometers below the surface.

Directed Stress

Directed stress, also called **differential** or **tectonic stress**, is an unequal balance of forces on a rock in one or more directions (see the previous figure). Directed stresses are generated by the movement of lithospheric plates. Stress indicates a type of force acting on the rock. Strain describes the resultant processes caused by stress and includes metamorphic changes in the minerals. In contrast to confining pressure, directed stress occurs at much lower pressures and does not generate chemical reactions that change the mineral composition and atomic structure [3]. Instead, directed stress modifies the parent rock at a mechanical level, changing the arrangement, size, and/or shape of the mineral crystals. These crystalline changes create identifying textures, which is shown in the figure below comparing the phaneritic texture of igneous granite with the foliated texture of metamorphic gneiss.



Figure 6.1.2: An igneous rock granite (left) and foliated high-temperature and high-pressure metamorphic rock gneiss (right) illustrating a metamorphic texture. (Source: Peter Davis)

Directed stresses produce rock textures in many ways. Crystals are rotated, changing their orientation in space. Crystals can get fractured, reducing their grain size. Conversely, they may grow larger as atoms migrate. Crystal shapes also become deformed. These mechanical changes occur via **recrystallization**, which is when minerals dissolve from an area of rock experiencing high stress and precipitate or regrow in a location having lower stress. For example, recrystallization increases grain size much like adjacent soap bubbles coalesce to form larger ones. Recrystallization rearranges mineral crystals without fracturing the rock structure, deforming the rock like silly putty; these changes provide important clues to understanding the creation and movement of deep underground rock faults.



Figure 6.1.3: Pebbles (that used to be spherical or close to spherical) deformed by directed stress. (By Marli Miller; CC BY-NC 4.0 via Geologic Time Pics.)

Fluids

A third metamorphic agent is **chemically reactive fluids** that are expelled by crystallizing magma and created by metamorphic reactions. These reactive fluids are made of mostly water (H_2O) and carbon dioxide (CO_2), and smaller amounts of potassium (K),

sodium (Na), iron (Fe), magnesium (Mg), calcium (Ca), and aluminum (Al). These fluids react with minerals in the protolith, changing its chemical equilibrium and mineral composition, in a process similar to the reactions driven by heat and pressure. In addition to using elements found in the protolith, the chemical reaction may incorporate substances contributed by the fluids to create new minerals. In general, this style of metamorphism, in which fluids play an important role, is called **hydrothermal metamorphism** or **hydrothermal alteration**. Water actively participates in chemical reactions and allows extra mobility of the components in hydrothermal alteration.

Fluid-activated metamorphism is frequently involved in creating economically important mineral deposits that are located next to igneous intrusions or magma bodies. For example, the mining districts in the Cottonwood Canyons and Mineral Basin of northern Utah produce valuable ores such as argentite (silver sulfide), galena (lead sulfide), and chalcopyrite (copper-iron sulfide), as well as the native element gold [2]. These mineral deposits were created from the interaction between a granitic intrusion called the Little Cottonwood Stock and country rock consisting of mostly limestone and dolostone. Hot, circulating fluids expelled by the crystallizing granite reacted with and dissolved the surrounding limestone and dolostone, precipitating out new minerals created by the chemical reaction. Hydrothermal alteration of mafic mantle rock, such as olivine and basalt, creates the metamorphic rock **serpentinite**, a member of the serpentine subgroup of minerals. This metamorphic process happens at mid-ocean spreading centers where newly formed oceanic crust interacts with seawater.

Some hydrothermal alterations remove elements from the parent rock rather than deposit them. This happens when seawater circulates down through fractures in the fresh, still-hot basalt, reacting with and removing mineral ions from it. The dissolved minerals are usually ions that do not fit snugly in the silicate crystal structure, such as copper. The mineral-laden water emerges from the seafloor via hydrothermal vents called black smokers, named after the dark-colored precipitates produced when the hot vent water meets cold seawater. Ancient black smokers were an important source of copper ore for the inhabitants of Cyprus (Cypriots) as early as 4000 BCE, and later by the Romans [8].

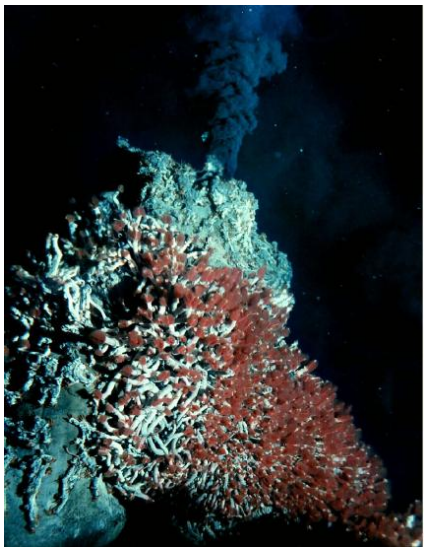


Figure 6.1.4: Black smoker hydrothermal vent with a colony of giant (6 foot) tube worms.

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6.2: Metamorphic Textures

Metamorphic texture is the description of the shape and orientation of mineral grains in a metamorphic rock. Metamorphic rock textures are foliated, non-foliated, or lineated are described below.

Table 6.2.1: Metamorphic rock identification table.

Identify rock's foliation		Textural features	Mineral composition	Rock name	Parent rock
Foliated (layered texture)	Fine-grained or no visible grains	Flat, slaty cleavage is well developed. Dense, microscopic grains may exhibit slight sheen (or dull luster). Clanky sound when struck. Breaks into hard, flat sheets.	Fine, microscopic clay or mica	Slate	Shale
		Finely crystalline; micas hardly discernible, but impart a sheen or luster. Breaks along wavy surfaces.	Dark silicates and micas	Phyllite	Siltstone or shale
	Medium- to coarse-grained	Schistose texture. Foliation formed by alignment of visible crystals. Rock breaks along scaly foliation surfaces. Medium to fine-grained. Sparkling appearance.	Common minerals include chlorite, biotite, muscovite, garnet, and hornblende. Recognizable minerals used as part of rock name. Porphyroblasts common.	Mica schist	Siltstone or shale
				Garnet schist	
		Gneissic banding. Coarse-grained. Foliation present as minerals arranged into alternating light and dark layers giving the rock a banded texture in side view. Crystalline texture. No cleavage.	Light-colored quartz and feldspar; dark ferromagnesian minerals	Gneiss	Shale or granitic rocks
Non-foliated (no layered texture)	Fine-grained or no visible grains	Medium- to coarse-grained crystalline structure.	Crystals of amphibole (hornblende) in blade-like crystals	Amphibolite	Basalt, gabbro, or ultramafic igneous rocks
		Microcrystalline texture. Glassy black sheen. Conchoidal fracture. Low density.	Fine, tar-like, organic makeup	Anthracite coal	Coal
		Dense and dark-	Microscopic dark	Hornfels	Many rock types

Identify rock's foliation		Textural features	Mineral composition	Rock name	Parent rock
		colored. Fine or microcrystalline texture. Very hard. Color can range from gray, gray-green to black.	silicates		
		Microcrystalline or no visible grains with smooth, wavy surfaces. May be dull or glossy. Usually shades of green.	Serpentine. May have fibrous asbestos visible.	Serpentinite	Ultramafic igneous rocks or peridotite
	Fine- to coarse-grained	Microcrystalline or no visible grains. Can be scratched with a fingernail. Shades of green, gray, brown, or white. Soapy feel.	Talc	Soapstone or talc schist	Ultramafic igneous rocks
		Crystalline. Hard (scratches glass). Breaks across grains. Sandy or sugary texture. Color variable; can be white, pink, buff, brown, red, purple.	Quartz grains fused together. Grains will not rub off like sandstone.	Quartzite	Quartz sandstone
		Finely crystalline (resembling a sugar cube) to medium or coarse texture. Color variable; white, pink, gray, among others. Fossils in some varieties.	Calcite or dolomite crystals tightly fused together. Calcite effervesces with HCl; dolomite effervesces only when powdered.	Marble	Limestone or dolostone
		Texture of conglomerate, but breaks across clasts as easily as around them. Pebbles may be stretched (lineated) or cut by rock cleavage.	Granules or pebbles are commonly granitic or jasper, chert, quartz, or quartzite.	Meta-conglomerate	Conglomerate

Foliation and Lineation

Foliation is a term used that describes minerals lined up in planes. Certain minerals, most notably the mica group, are mostly thin and planar by default. Foliated rocks typically appear as if the minerals are stacked like pages of a book, thus the use of the term 'folia', like a leaf. Other minerals, with hornblende being a good example, are longer in one direction, linear like a pencil or a needle, rather than a planar-shaped book. These linear objects can also be aligned within a rock. This is referred to as a **lineation**. Linear crystals, such as hornblende, tourmaline, or stretched quartz grains, can be arranged as part of a foliation, a lineation, or foliation/lineation together. If they lie on a plane with mica, but with no common or preferred direction, this is foliation. If the minerals line up and point in a common direction, but with no planar fabric, this is lineation. When minerals lie on a plane AND point in a common direction; this is both foliation and lineation.

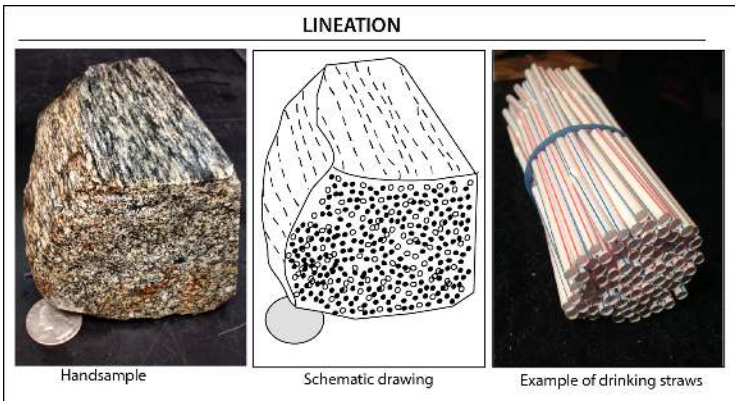


Figure 6.2.1: Example of lineation where minerals are aligned like a stack of straws or pencils. (Source: Peter Davis)

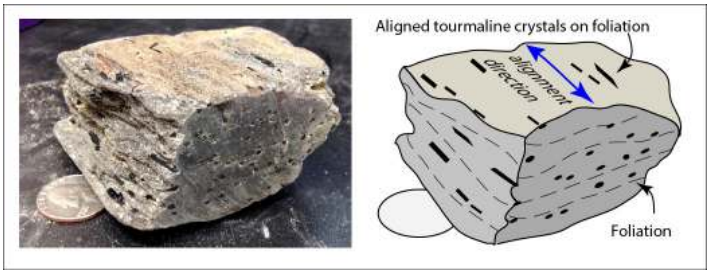


Figure 6.2.2: An example of foliation WITH lineation. (Source: Peter Davis)

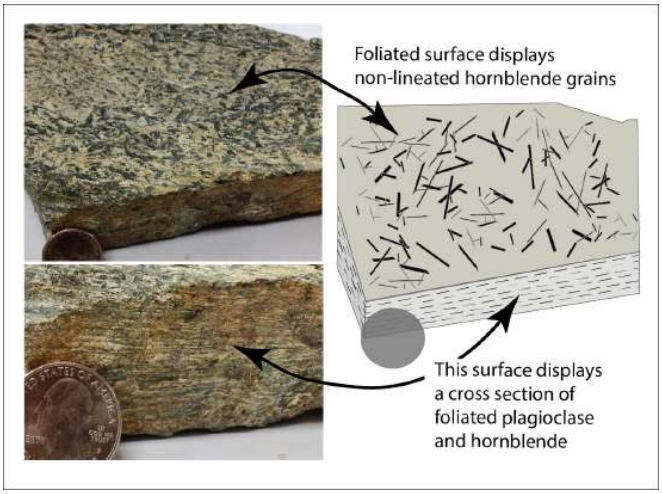


Figure 6.2.3: An example of foliation WITHOUT lineation. (Source: Peter Davis)

Foliated metamorphic rocks are named based on the style of their foliations. Each rock name has a specific texture that defines and distinguishes it, with their descriptions listed below.

Slate is a fine-grained metamorphic rock that exhibits a foliation called **slaty cleavage** that is the flat orientation of the small platy crystals of mica and chlorite forming perpendicular to the direction of stress. The minerals in slate are too small to see with the

unaided eye. The thin layers in slate may resemble sedimentary bedding, but they are a result of directed stress and may lie at angles to the original strata. In fact, original sedimentary layering may be partially or completely obscured by the foliation. Thin slabs of slate are often used as a building material for roofs and tiles.

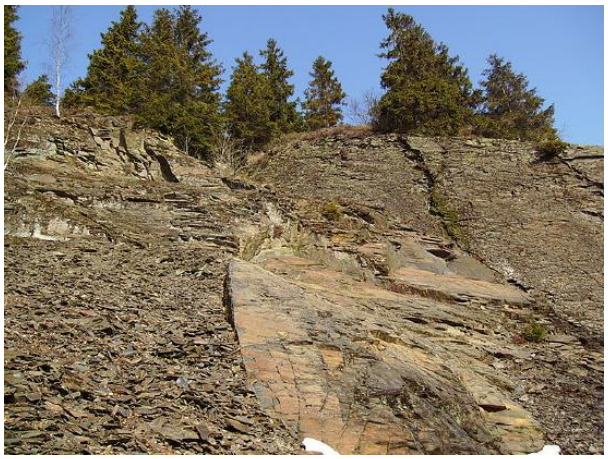


Figure 6.2.4: Slate mine in Germany. Note the rock breaks along flat, even planes, showing slaty cleavage.

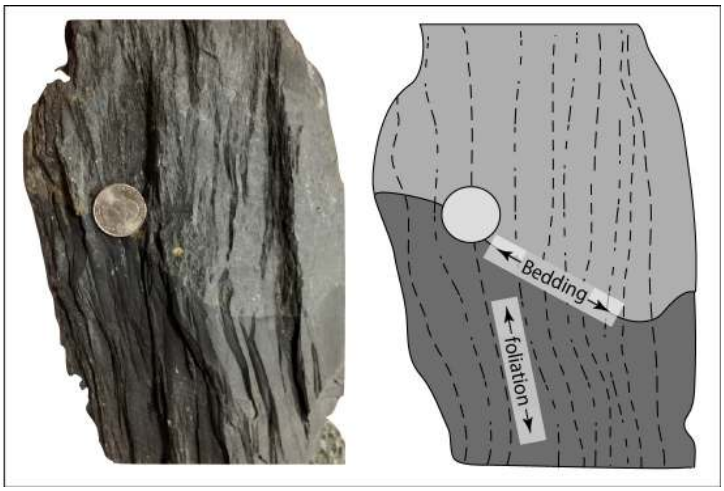


Figure 6.2.5: Foliation vs. bedding. Foliation is caused by metamorphism. Bedding is a result of sedimentary processes. They do not have to align. (Source: Peter Davis)

Phyllite is a foliated metamorphic rock in which platy minerals have grown larger and the surface of the foliation shows a sheen from light reflecting from the grains, perhaps even a wavy appearance, called **crenulations**. Similar to phyllite but with even larger grains is the foliated metamorphic rock **schist**, which has large platy grains visible as individual crystals. Common minerals are muscovite, biotite, and porphyroblasts of garnets. A **porphyroblast** is a large crystal of a particular mineral surrounded by small grains. **Schistosity** is a textural description of foliation created by the parallel alignment of platy visible grains. Some schists are named for their minerals such as mica schist (mostly micas), garnet schist (mica schist with garnets), and staurolite schist (mica schists with staurolite).



Figure 6.2.6: Phyllite with a small fold. (Source: Peter Davis)



Figure 6.2.7: Schist.



Figure 6.2.8: Garnet staurolite muscovite schist.

Gneissic banding is a metamorphic foliation in which visible silicate minerals separate into dark and light bands or lineations. These grains tend to be coarse and often folded. A rock with this texture is called **gneiss**. Since gneiss forms at the highest temperatures and pressures, some partial melting may occur. This partially melted rock is a transition between metamorphic and igneous rocks called a **migmatite** [9].



Figure 6.2.9: Gneiss.

Migmatites appear as dark and light banded gneiss that may be swirled or twisted since some minerals started to melt. Thin accumulations of light-colored rock layers can occur in a darker rock that are parallel to each other or even cut across the gneissic foliation. The lighter colored layers are interpreted to be the result of the separation of a felsic igneous melt from the adjacent highly metamorphosed darker layers, or injection of a felsic melt from some distance away.



Figure 6.2.10: Migmatite.

Non-foliated

Non-foliated textures do not have lineations, foliations, or other alignments of mineral grains. Non-foliated metamorphic rocks are typically composed of just one mineral and, therefore, usually show the effects of metamorphism with recrystallization in which crystals grow together, but with no preferred direction. The two most common examples of non-foliated rocks are quartzite and marble. **Quartzite** is a metamorphic rock from the protolith sandstone. In quartzite, the quartz grains from the original sandstone are enlarged and interlocked by recrystallization. A defining characteristic for distinguishing quartzite from sandstone is that when broken with a rock hammer, the quartz crystals break across the grains. In a sandstone, only a thin mineral cement holds the grains together, meaning that a broken piece of sandstone will leave the grains intact. Because most sandstones are rich in quartz, and quartz is a mechanically and chemically durable substance, quartzite is very hard and resistant to weathering.

Marble is metamorphosed limestone (or dolostone) composed of calcite (or dolomite). Recrystallization typically generates larger interlocking crystals of calcite or dolomite. Marble and quartzite often look similar, but these minerals are considerably softer than quartz. Another way to distinguish marble from a quartzite is with a drop of dilute hydrochloric acid. Marble will effervesce (fizz) if it is made of calcite.



Figure 6.2.11: Left: Marble. Right: Quarzite.



Figure 6.2.12: Left: Macro view of quartzite. Note the interconnectedness of the grains. Right: Unmetamorphosed, unconsolidated sand grains have space between the grains.

A third non-foliated rock is **hornfels,** identified by its dense, fine-grained, hard, blocky or splintery texture composed of several silicate minerals. Crystals in hornfels become smaller with metamorphism and get so small that specialized study is required to identify them. These are common around intrusive igneous bodies and are hard to identify. The protolith of hornfels can be even harder to distinguish, which can be anything from mudstone to basalt.

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6.3: Metamorphic Grade

Metamorphic grade refers to the range of metamorphic change a rock undergoes, progressing from **low grade** (little metamorphic change) to **high grade** (significant metamorphic change). Low-grade metamorphism begins at temperatures and pressures just above sedimentary rock conditions. The sequence *slate* → *phyllite* → *schist* → *gneiss* illustrates an increasing metamorphic grade.



Figure 6.3.1: Garnet schist.

Geologists use **index minerals** that form at certain temperatures and pressures to identify metamorphic grade. These index minerals also provide important clues to a rock's sedimentary protolith and the metamorphic conditions that created it. Chlorite, muscovite, biotite, garnet, and staurolite are index minerals representing a respective sequence of low-to-high grade rock. The figure shows a **phase diagram** of three index minerals—sillimanite, kyanite, and andalusite—with the same chemical formula (Al_2SiO_5) but having different crystal structures (**polymorphism**) created by different pressure and temperature conditions.

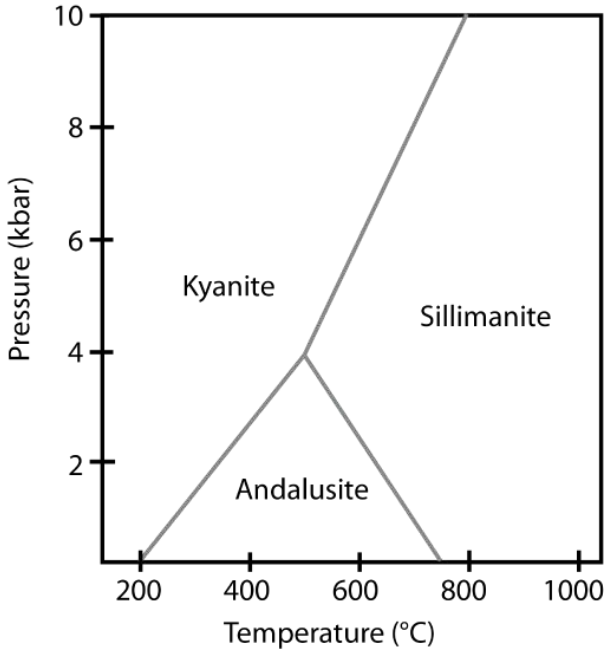


Figure 6.3.1: Phase diagram for kyanite, andalusite and sillimanite.



Figure 6.3.2: Left: Kyanite. Middle: Andalusite. Right: Sillimanite.

Some metamorphic rocks are named based on the highest grade of index mineral present. Chlorite schist includes the low-grade index mineral chlorite. Muscovite schist contains a slightly higher grade muscovite, indicating a greater degree of metamorphism. Garnet schist includes the high-grade index mineral garnet and indicating it has experienced much higher pressures and temperatures than chlorite.

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6.4: Metamorphic Environments

As with igneous processes, metamorphic rocks form at different zones of pressure (depth) and temperature as shown on the pressure-temperature (P-T) diagram. The term facies is an objective description of a rock. In metamorphic rocks, **facies** are groups of minerals called **mineral assemblages**. The names of metamorphic facies on the pressure-temperature diagram reflect minerals and mineral assemblages that are stable at these pressures and temperatures and provide information about the metamorphic processes that have affected the rocks. This is useful when interpreting the history of metamorphic rock.

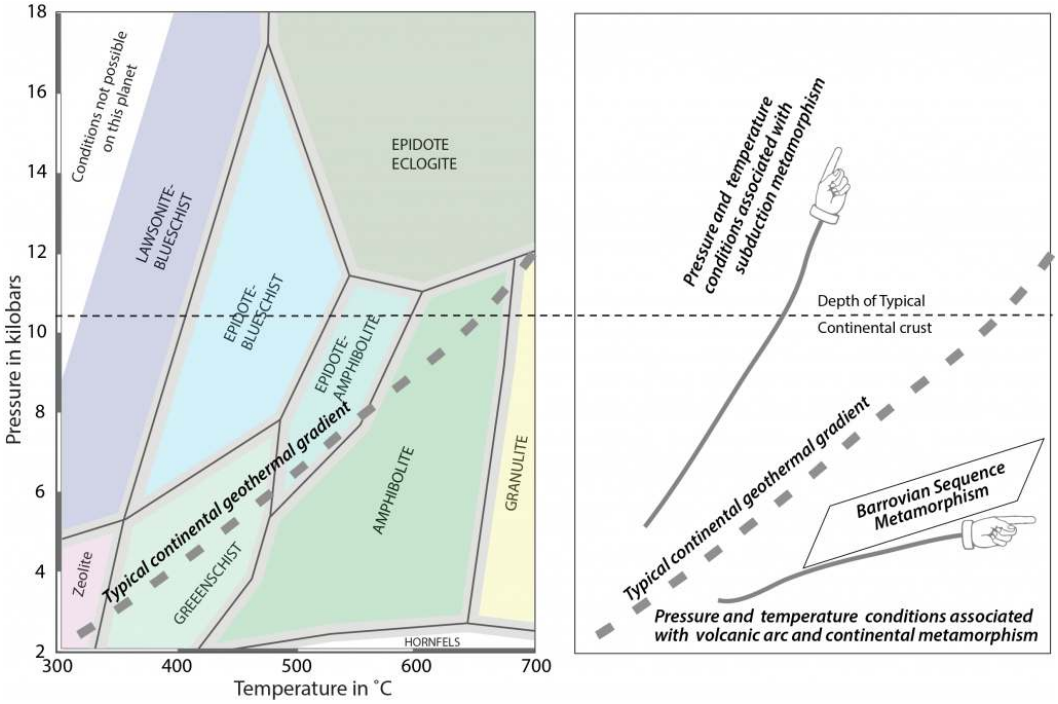


Figure 6.4.1: Pressure-temperature diagram showing metamorphic facies.

In the late 1800s, British geologist George Barrow mapped zones of index minerals in different metamorphic zones of an area that underwent regional metamorphism. Barrow outlined a progression of index minerals, named the **Barrovian Sequence**, that represents increasing metamorphic grade: chlorite (slates and phyllites) → biotite (phyllites and schists) → garnet (schists) → staurolite (schists) → kyanite (schists) → sillimanite (schists and gneisses).

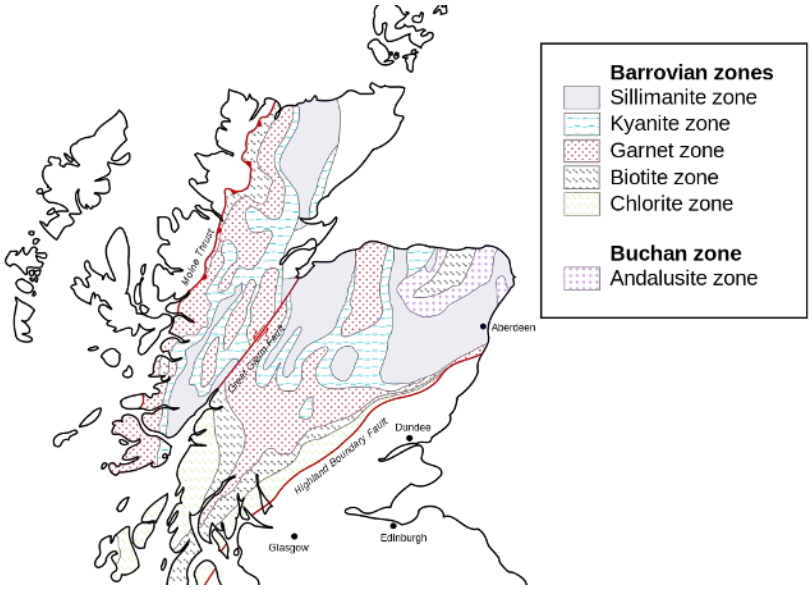


Figure 6.4.2: Barrovian sequence in Scotland.

The first of the Barrovian sequence has a mineral group that is commonly found in the metamorphic **greenschist** facies. Greenschist rocks form under relatively low pressure and temperatures and represent the fringes of regional metamorphism. The "green" part of the name is derived from green minerals like chlorite, serpentine, and epidote, and the "schist" part is applied due to the presence of platy minerals such as muscovite.

Many different styles of metamorphic facies are recognized, tied to different geologic and tectonic processes. Recognizing these facies is the most direct way to interpret the metamorphic history of rock. A simplified list of major metamorphic facies is given below.

Burial Metamorphism

Burial metamorphism occurs when rocks are deeply buried, at depths of more than 2000 meters (1.24 miles) [10]. Burial metamorphism commonly occurs in sedimentary basins, where rocks are buried deeply by overlying sediments. As an extension of diagenesis, a process that occurs during lithification, burial metamorphism can cause clay minerals, such as **smectite**, in shales to change to another clay mineral **illite**. Or it can cause quartz sandstone to metamorphose into quartzite such the Big Cottonwood Formation in the Wasatch Range of Utah. This formation was deposited as ancient near-shore sands in the late Proterozoic, deeply buried and metamorphosed to quartzite, folded, and later exposed at the surface in the Wasatch Range today. An increase of temperature with depth in combination and an increase of confining pressure produces low-grade metamorphic rocks with mineral assemblages are indicative of a **zeolite facies** [11; 12].

Contact Metamorphism

Contact metamorphism occurs in rock exposed to high temperature and low pressure, as might happen when hot magma intrudes into or lava flows over pre-existing protolith. This combination of high temperature and low pressure produces numerous metamorphic facies. The lowest pressure conditions produce **hornfels facies**, while higher pressure creates greenschist**, amphibolite**, or **granulite facies**.

As with all metamorphic rock, the parent rock texture and chemistry are major factors in determining the final outcome of the metamorphic process, including what index minerals are present. Fine-grained shale and basalt, which happen to be chemically similar, characteristically recrystallize to produce hornfels. Sandstone (silica) surrounding an igneous intrusion becomes quartzite via contact metamorphism, and limestone (carbonate) becomes marble.



Figure 6.4.3: Basaltic dike intruding into volcanic mudflow deposits in eastern Oregon. Note the baked (white) and altered (orange) edge of the country rock around the intrusion. (By Marli Miller; CC BY-NC 4.0 via Geologic Time Pics.)

When contact metamorphism occurs deeper in the Earth, metamorphism can be seen as rings of facies around the intrusion, resulting in **aureoles**. These differences in metamorphism appear as distinct bands surrounding the intrusion, as can be seen around the Alta Stock in Little Cottonwood Canyon, Utah. The Alta Stock is a granite intrusion surrounded first by rings of the index minerals amphibole (tremolite) and olivine (forsterite), with a ring of talc (dolostone) located further away [13; 14].

Regional Metamorphism

Regional metamorphism occurs when parent rock is subjected to increased temperature and pressure over a large area and is often located in mountain ranges created by converging continental crustal plates. This is the setting for the Barrovian sequence of rock

facies, with the lowest grade of metamorphism occurring on the flanks of the mountains and highest grade near the core of the mountain range, closest to the convergent boundary.

An example of an old regional metamorphic environment is visible in the northern Appalachian Mountains while driving east from New York state through Vermont and into New Hampshire. Along this route, the degree of metamorphism gradually increases from sedimentary parent rock to low-grade metamorphic rock, then higher-grade metamorphic rock, and eventually the igneous core. The rock sequence is sedimentary rock, slate, phyllite, schist, gneiss, migmatite, and granite. In fact, New Hampshire is nicknamed the Granite State. The reverse sequence can be seen heading east, from eastern New Hampshire to the coast [15].

Subduction Zone Metamorphism

Subduction zone metamorphism is a type of regional metamorphism that occurs when a slab of oceanic crust is subducted under continental crust. Because rock is a good insulator, the temperature of the descending oceanic slab increases slowly relative to the more rapidly increasing pressure, creating a metamorphic environment of high pressure and low temperature. **Glaucophane**, which has a distinctive blue color, is an index mineral found in **blueschist facies** (see metamorphic facies diagram). The California Coast Range near San Francisco has blueschist-facies rocks created by subduction-zone metamorphism, which include rocks made of **blueschist**, **greenstone**, and red chert. Greenstone, which is metamorphosed basalt, gets its color from the index mineral chlorite [16].



Figure 6.4.4: Blueschist rock with green serpentinite and mica.

Fault Metamorphism

There are a range of metamorphic rocks made along faults. Near the surface, rocks involved in repeated brittle faulting produce a material called **rock flour,** which is rock that is ground up to the particle size of flour used for food. At lower depths, faulting creates **cataclastites** [17], chaotically-crushed mixes of rock material with little internal texture. At depths below cataclastites, where strain becomes ductile, mylonites are formed. **Mylonites** are metamorphic rocks created by dynamic recrystallization through directed shear forces, generally resulting in a reduction of grain size [18]. When larger, stronger crystals (like feldspar, quartz, garnet) embedded in a metamorphic matrix are sheared into an asymmetrical eye-shaped crystal, an **augen** is formed [19; 20].

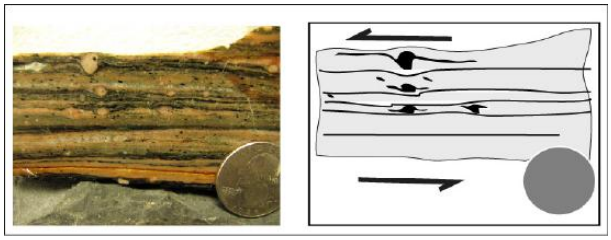


Figure 6.4.5: Left: Mylonite. Right: Diagram of the mylonite showing directed shear forces.

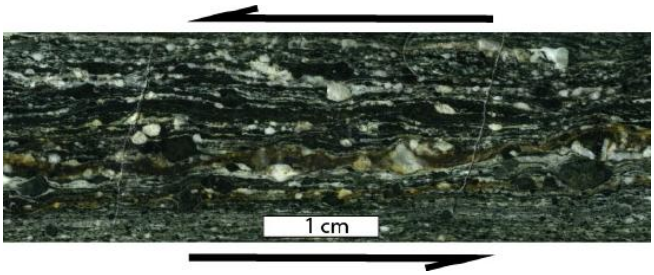


Figure 6.4.6: Examples of augens.

Shock Metamorphism

Shock (also known as **impact**) **metamorphism** is metamorphism resulting from meteorite or other bolide impacts, or from a similar high-pressure shock event. Shock metamorphism is the result of very high pressures (and higher, but less extreme temperatures) delivered relatively rapidly. Shock metamorphism produces planar deformation features, tektites, shatter cones, and quartz polymorphs. Shock metamorphism produces planar deformation features (**shock lamellae**), which are narrow planes of glassy material with distinct orientations found in silicate mineral grains. **Shocked quartz** has planar deformation features [21].

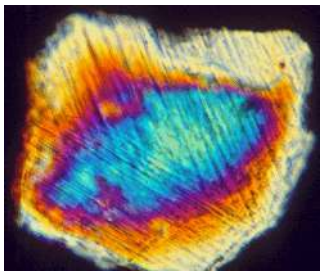


Figure 6.4.7: Close-up view of shock lamellae in a quartz grain.

Shatter cones are cone-shaped pieces of rock created by dynamic branching fractures caused by impacts [22]. While not strictly a metamorphic structure, they are common around shock metamorphism. Their diameter can range from microscopic to several meters. Fine-grained rocks with shatter cones show a distinctive horsetail pattern.



Figure 6.4.8: Shatter cone.

Shock metamorphism can also produce index minerals, though they are typically only found via microscopic analysis. The quartz polymorphs **coesite** and **stishovite** are indicative of impact metamorphism [21]. Recall that polymorphs are minerals with the same composition but different crystal structures. Intense pressure (> 10 GPa) and moderate to high temperatures (700-1200 °C) are required to form these minerals.

Shock metamorphism can also produce glass. **Tektites** are gravel-size glass grains ejected during an impact event. They resemble volcanic glass but, unlike volcanic glass, tektites contain no water or phenocrysts, and have a different bulk and isotopic chemistry. Tektites contain partially melted inclusions of shocked mineral grains [23]. Although all are melt glasses, tektites are also chemically distinct from **trinitite**, which is produced from thermonuclear detonations [24], and **fulgurites**, which are produced by lightning strikes [25]. All geologic glasses not derived from volcanoes can be called with the general term **pseudotachylyte** [26], a name that can also be applied to glasses created by faulting. The term pseudo in this context means ‘false’ or ‘in the appearance of’, a volcanic rock called tachylite because the material observed looks like volcanic rock, but is produced by significant shear heating.



Figure 6.4.9: Tektites.

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6.S: Summary

Summary

Metamorphism is the process that changes existing rocks (called protoliths) into new rocks with new minerals and new textures. Increases in temperature and pressure are the main causes of metamorphism, with fluids adding important mobilization of materials. The primary way metamorphic rocks are identified is with texture. Foliated textures come from platy minerals forming planes in a rock, while non-foliated metamorphic rocks have no internal fabric. Grade describes the amount of metamorphism in a rock, and facies are a set of minerals that can help guide an observer to an interpretation of the metamorphic history of a rock. Different tectonic or geologic environments cause metamorphism, including collisions, subduction, faulting, and even impacts from space.

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CHAPTER OVERVIEW

7: Geologic Time

Learning Objectives

- Explain the difference between relative time and numeric time.
- Describe the five principles of stratigraphy.
- Identify three types of unconformities.
- Apply relative dating principles to a block diagram and interpret the sequence of geologic events.
- Define an isotope, and explain alpha and beta decay as mechanisms of radioactive decay.
- Describe how radioisotopic dating is accomplished and list the key isotopes used.
- Explain how carbon-14 forms in the atmosphere and how it is used in dating recent events.
- Explain how scientists know the numeric age of the Earth and other events in Earth history.
- Explain how sedimentary sequences can be dated using radioisotopes and other techniques.
- Define a fossil and describe types of fossils preservation.
- Outline how natural selection takes place as a mechanism of evolution.
- Describe stratigraphic correlation.
- List the eons, eras, and periods of the geologic time scale and explain the purpose behind the divisions.
- Explain the relationship between time units and corresponding rock units—chronostratigraphy versus lithostratigraphy.

The geologic time scale and basic outline of Earth's history were worked out long before we had any scientific means of assigning numerical age units, like years, to events of Earth history. Working out Earth's history depended on realizing some key principles of relative time. Nicolaus Steno (1638-1686) introduced basic principles of stratigraphy, the study of layered rocks, in 1669 [1]. William Smith (1769-1839), working with the strata of English coal mines, noticed that strata and their sequence were consistent throughout the region. Eventually, he produced the first national geologic map of Britain [2], becoming known as "the Father of English Geology." Nineteenth-century scientists developed a relative time scale using Steno's principles, with names derived from the characteristics of the rocks in those areas. The figure of this geologic time scale shows the names of the units and subunits. Using this time scale, geologists can place all events of Earth history in order without ever knowing their numerical ages. The specific events within Earth history are discussed in tne chapter on Earth History.



Figure 7.1: Nicolas Steno, c. 1670

7.1: Relative Dating
7.2: Absolute Dating
7.3: Fossils and Evolution
7.4: Correlation
7.S: Summary

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7.1: Relative Dating

Relative dating is the process of determining if one rock or geologic event is older or younger than another, without knowing their specific ages—i.e., how many years ago the object was formed. The principles of relative time are simple, even obvious now, but were not generally accepted by scholars until the scientific revolution of the 17th and 18th centuries [3]. James Hutton realized geologic processes are slow and his ideas on **uniformitarianism** (i.e., "the present is the key to the past") provided a basis for interpreting rocks of the Earth using scientific principles.

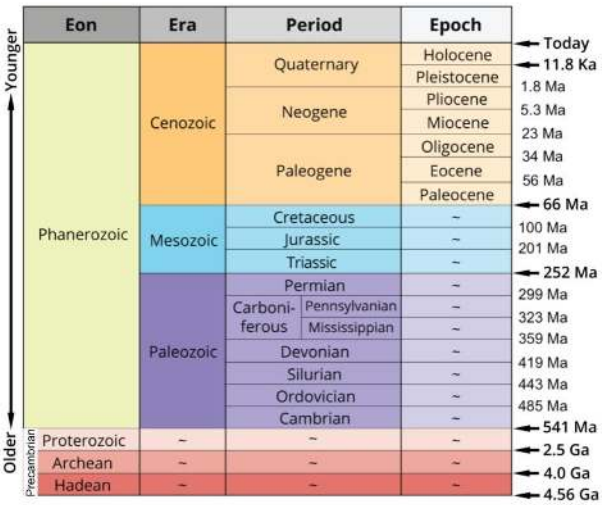


Figure 7.1.1: The geologic time scale. (By Jonathan R. Hendricks; CC BY-SA 4.0 via Historical Geology.)

Relative Dating Principles

Stratigraphy is the study of layered sedimentary rocks. This section discusses principles of relative time used in all of geology, but are especially useful in stratigraphy.

Principle of Superposition: In an otherwise undisturbed sequence of sedimentary strata, or rock layers, the layers on the bottom are the oldest and layers above them are younger.



Figure 7.1.2: Lower strata are older than those lying on top of them.

Principle of Original Horizontality: Layers of rocks deposited from above, such as sediments and lava flows, are originally laid down horizontally. The exception to this principle is at the margins of basins, where the strata can slope slightly downward into the basin.

Principle of Lateral Continuity: Within the depositional basin, strata are continuous in all directions until they thin out at the edge of that basin. Of course, all strata eventually end, either by hitting a geographic barrier, such as a ridge, or when the depositional

process extends too far from its source, either a sediment source or a volcano. Strata that are cut by a canyon later remain continuous on either side of the canyon.



Figure 7.1.3: Lateral continuity.

Principle of Cross-Cutting Relationships: Deformation events like folds, faults and igneous intrusions that cut across rocks are younger than the rocks they cut across.



Figure 7.1.4: Dark dike cutting across older rocks, the lighter of which is younger than the grey rock.

Principle of Inclusions: When one rock formation contains pieces or **inclusions** of another rock, the included rock is older than the host rock.



Figure 7.1.5: An inclusion in granite. The inclusion is older than the granite.

Principle of Fossil Succession: Evolution has produced a succession of unique fossils that correlate to the units of the geologic time scale. Assemblages of fossils contained in strata are unique to the time they lived and can be used to correlate rocks of the same age across a wide geographic distribution. Assemblages of fossils refer to groups of several unique fossils occurring together.

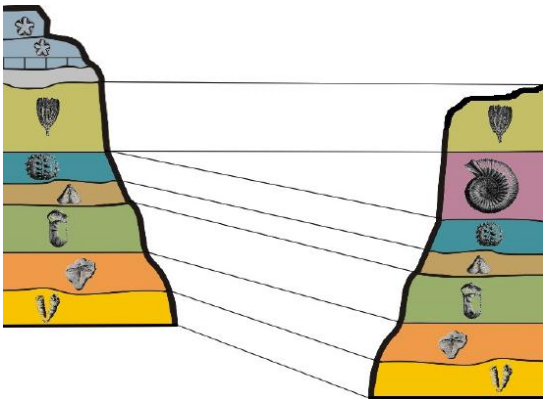


Figure 7.1.6: Fossil succession showing the correlation among strata.

Grand Canyon Example

The Grand Canyon of Arizona illustrates the stratigraphic principles. The photo shows layers of rock on top of one another in order, from the oldest at the bottom to the youngest at the top, based on the principle of superposition. The predominant white layer just below the canyon rim is the Coconino Sandstone. This layer is laterally continuous, even though the intervening canyon separates its outcrops. The rock layers exhibit the principle of lateral continuity, as they are found on both sides of the Grand Canyon which has been carved by the Colorado River.



Figure 7.1.7: The Grand Canyon of Arizona

The diagram called “Grand Canyon’s Three Sets of Rocks” shows a cross-section of the rocks exposed on the walls of the Grand Canyon, illustrating the principle of cross-cutting relationships, superposition, and original horizontality. In the lowest parts of the Grand Canyon are the oldest sedimentary formations, with igneous and metamorphic rocks at the bottom. The principle of cross-cutting relationships shows the sequence of these events. The metamorphic schist (#16) is the oldest rock formation and the cross-cutting granite intrusion (#17) is younger. As seen in the figure, the other layers on the walls of the Grand Canyon are numbered in reverse order with #15 being the oldest and #1 the youngest [4]. This illustrates the principle of superposition. The Grand Canyon region lies in Colorado Plateau, which is characterized by horizontal or nearly horizontal strata, which follows the principle of original horizontality. These rock strata have been barely disturbed from their original deposition, except by a broad regional uplift.

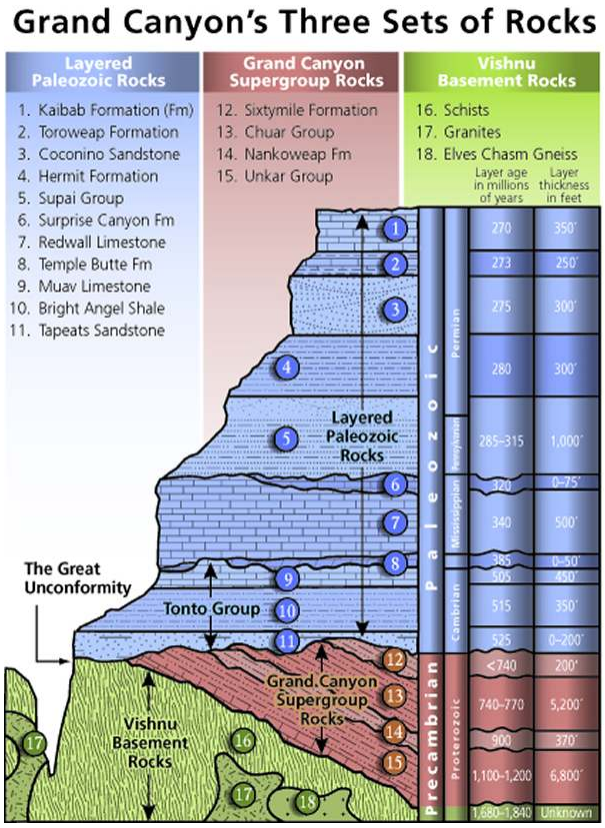


Figure 7.1.8: The rocks of the Grand Canyon

The photo of the Grand Canyon below shows strata that were originally deposited in a flat layer on top of older igneous and metamorphic "basement" rocks, per the original horizontality principle. Because the formation of the basement rocks and the deposition of the overlying strata is not continuous but broken by events of metamorphism, intrusion, and erosion, the contact between the strata and the older basement is termed an **unconformity**. An unconformity represents a period during which deposition did not occur or erosion removed rock that had been deposited, so there are no rocks that represent events of Earth history during that span of time at that place. Unconformities appear in cross-sections and stratigraphic columns as wavy lines between formations. Unconformities are discussed in the next section.



Figure 7.1.9: The red, layered rocks of the Grand Canyon Supergroup overlying the dark-colored rocks of the Vishnu schist represents a type of unconformity called a nonconformity.

Unconformities

There are three types of unconformities: nonconformity, disconformity, and angular unconformity. A **nonconformity** occurs when sedimentary rock is deposited on top of igneous and metamorphic rocks as is the case with the contact between the strata and basement rocks at the bottom of the Grand Canyon.

The strata in the Grand Canyon represent alternating marine **transgressions** and **regressions** where sea level rose and fell over millions of years. When the sea level was high, marine strata formed. When sea level fell, the land was exposed to erosion, creating an unconformity. In the Grand Canyon cross-section, this erosion is shown as heavy wavy lines between the various numbered strata. This is a type of unconformity called a **disconformity**, where either non-deposition or erosion took place. In other words, layers of rock that could have been present, are absent. The time that could have been represented by such layers is instead represented by the disconformity. Disconformities are unconformities that occur between parallel layers of strata indicating either a period of no deposition or erosion.

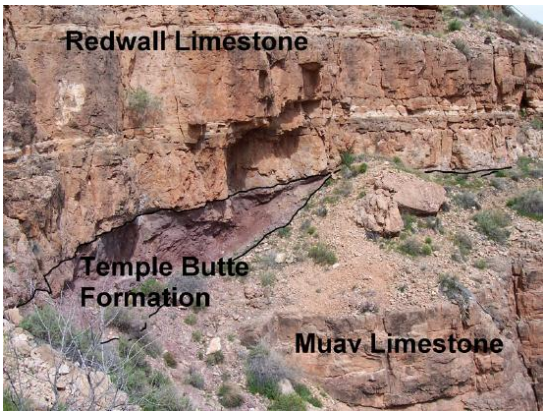


Figure 7.1.10: All three of these formations have a disconformity at the two contacts between them. The pinching Temple Butte is the easiest to see the erosion, but even between the Muav and Redwall, there is an unconformity.

The Phanerozoic strata in most of the Grand Canyon are horizontal. However, near the bottom, horizontal strata overlie tilted strata. This is known as the Great Unconformity and is an example of an **angular unconformity**. The lower strata were tilted by tectonic processes that disturbed their original horizontality and caused the strata to be eroded. Later, horizontal strata were deposited on top of the tilted strata creating an angular unconformity.



Figure 7.1.11: In the lower part of the picture is an angular unconformity in the Grand Canyon known as the Great Unconformity. Notice the flat-lying strata over dipping strata. (By James St. John; CC BY 2.0 via Wikimedia Commons.)

Here are three graphical illustrations of the three types of unconformity.

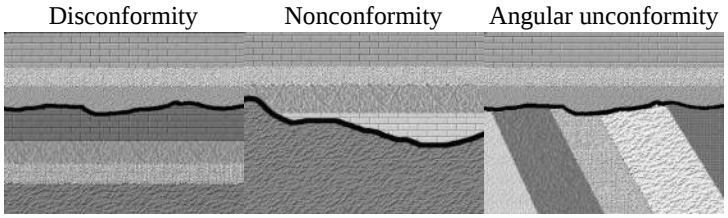


Figure 7.1.12: Left: A disconformity is a break or stratigraphic absence between strata in an otherwise parallel sequence of strata. Middle: A nonconformity is where sedimentary strata are deposited on crystalline (igneous or metamorphic) rocks. Right: An angular unconformity is where sedimentary strata are deposited on sedimentary strata that have been deformed by tilting, folding, and/or faulting so that they are no longer horizontal.

Applying Relative Dating Principles

In the block diagram, the sequence of geological events can be determined by using the relative dating principles and known properties of igneous, sedimentary, metamorphic rock. The sequence begins with the folded metamorphic gneiss on the bottom. Next, the gneiss is cut and displaced by the fault labeled A. Both the gneiss and fault A are cut by the igneous granitic intrusion called batholith B; its irregular outline suggests it is an igneous granitic intrusion emplaced as magma into the gneiss. Since batholith B cuts both the gneiss and fault A, batholith B is younger than the other two rock formations. Next, the gneiss, fault A, and batholith B were eroded forming a nonconformity as shown with the wavy line. This unconformity was actually an ancient landscape surface on which sedimentary rock C was subsequently deposited perhaps by a marine transgression. Next, igneous basaltic dike D cut through all rocks except sedimentary rock E. This shows that there is a disconformity between sedimentary rocks C and E. The top of dike D is level with the top of layer C, which establishes that erosion flattened the landscape prior to the deposition of layer E, creating a disconformity between rocks D and E. Fault F cuts across all of the older rocks B, C and E, producing a fault scarp, which is the low ridge on the upper-left side of the diagram. The final events affecting this area are current erosion processes working on the land surface, rounding off the edge of the fault scarp, and producing the modern landscape at the top of the diagram.

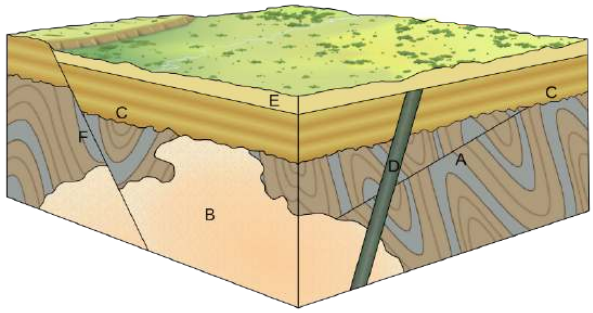


Figure 7.1.13: Block diagram to apply relative dating principles. The wavy rock is an old metamorphic gneiss, A and F are faults, B is an igneous granite, D is a basaltic dike, and C and E are sedimentary strata.

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7.2: Absolute Dating

Relative time allows scientists to tell the story of Earth events, but does not provide specific numeric ages, and thus, the rate at which geologic processes operate. Based on Hutton's principle of uniformitarianism, early geologists surmised geological processes work slowly and the Earth is very old. Relative dating principles were how scientists interpreted Earth history until the end of the 19th Century. Because science advances as technology advances, the discovery of radioactivity in the late 1800s provided scientists with a new scientific tool called radiometric dating. Using this new technology, they could assign specific time units, in this case years, to mineral grains within a rock. These numerical values are not dependent on comparisons with other rocks such as with relative dating, so this dating method is called **absolute dating** [5]. There are several types of absolute dating discussed in this section but radiometric dating is the most common and therefore is the focus on this section.



Figure 7.2.1: Canada's Nuvvuagittuq Greenstone Belt may have the oldest rocks and oldest evidence life on Earth, according to recent studies.

Radioactive Decay

Atoms contain three particles: protons, neutrons, and electrons. Protons and neutrons are located in the nucleus, while electrons orbit around the nucleus. The number of protons determines which element you're examining. For example, all atoms of carbon have six protons, all atoms of oxygen have eight protons, and all atoms of gold have 79 protons. The number of neutrons, however, is variable. An atom of an element with a different number of neutrons is an **isotope** of that element. For example, the isotope carbon-12 contains 6 neutrons in its nucleus, while the isotope carbon-13 has 7 neutrons.

Some isotopes are **radioactive**, which means they are unstable and likely to decay. This means the atom will spontaneously change from an unstable form to a stable form. There are two forms of nuclear decay that are relevant in how geologists can date rocks, shown in the table below:

Table 7.2.1: Types of nuclear decay

Particle	Composition	Effect on Nucleus
Alpha	2 protons, 2 neutrons	The nucleus contains two fewer protons and two fewer neutrons.
Beta	1 electron	One neutron decays to form a proton and an electron, which is emitted.

If an element decays by losing an **alpha particle**, it will lose 2 protons and 2 neutrons. If an atom decays by losing a **beta particle**, it loses just one electron.

Radioactive decay eventually results in the formation of stable **daughter products**. Radioactive materials decay at known rates. As time passes, the proportion of radioactive isotopes (**parent isotope**) will decrease and the proportion of daughter isotopes will increase. A rock with a relatively high proportion of radioactive isotopes is probably very young, while a rock with a high proportion of daughter products is probably very old.

Scientists measure the rate of radioactive decay with a unit called half-life. The **half-life** of a radioactive substance is the amount of time, on average, it takes for half of the parent isotope atoms to decay to daughter isotope atoms. For example, imagine a radioactive substance with a half-life of one year. When a rock is formed, it contains a certain number of radioactive parent isotope atoms. After one year (one half-life), half of the parent isotope atoms have decayed to form stable daughter products, and 50% of the radioactive parent isotope atoms remain. After another year (two half-lives), half of the remaining radioactive parent atoms have decayed, and 25% of the original radioactive parent atoms remain. After the third year (three half-lives), 12.5% of the original radioactive parent atoms remain. After four years (four half-lives), 6.25% of the original radioactive parent atoms remain, and after 5 years (five half-lives), only 3.125% of the original radioactive parent atoms remain.

If you find a rock whose radioactive material has a half-life of one year and measure 3.125% radioactive parent atoms and 96.875% daughter atoms, you can assume that the substance is 5 years old. The decay of radioactive materials is shown in the following graph. For example, a rock with 75% of the radioactive parent atoms remaining is about 0.4 years old.

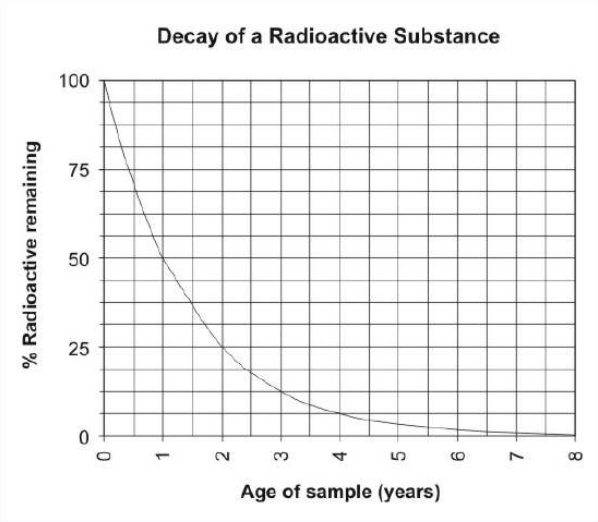


Figure 7.2.2: Decay of an imaginary radioactive substance with a half-life of one year. Note that in this example, the age also represents the number of half-lives that have elapsed. (By Kurt Rosenkrantz; CC BY-SA 3.0 via Wikimedia Commons.)

Radiometric Dating of Rocks

In the process of **radiometric dating**, several isotopes are used to date rocks and other materials. Using several different isotopes helps scientists to check the accuracy of the ages that they calculate.

Carbon Dating

Earth's atmosphere contains three isotopes of carbon. Carbon-12 (^{12}C) is stable and accounts for 98.9% of atmospheric carbon. Carbon-13 (^{13}C) is also stable and accounts for 1.1% of atmospheric carbon. Carbon-14 (^{14}C) is radioactive and is found in tiny amounts. Carbon-14 is produced naturally in the atmosphere when cosmic rays interact with nitrogen atoms. The amount of carbon-14 produced in the atmosphere at any particular time has been relatively stable through time.

Radioactive ^{14}C decays to stable nitrogen-14 (^{14}N) by releasing a beta particle. The nitrogen atoms are lost to the atmosphere, but the amount of ^{14}C decay can be estimated by measuring the proportion of radioactive ^{14}C to stable ^{12}C. As a substance ages, the relative amount of ^{14}C decreases.

Carbon is removed from the atmosphere by plants during the process of photosynthesis. Animals consume this carbon when they eat plants or other animals that have eaten plants. When an organism dies, the radiocarbon clock starts ticking as the ^{14}C decays back to ^{14}N. Therefore carbon-14 dating can be used to date plant and animal remains. Examples include timbers from an old building, bones, or ashes from a fire pit. **Carbon dating** can be effectively used to find the age of materials between 100 and 50,000 years old.

Potassium-Argon Dating

Potassium-40 (^{40}K) decays to argon-40 (^{40}Ar) with a half-life of 1.26 billion years. Because argon is a gas, it can escape from molten magma or lava. Therefore any argon that is found in a crystal probably formed as a result of the decay of potassium-40. Measuring the ratio of ^{40}K to ^{40}Ar will yield a good estimate of the age of the sample.

Potassium is a common element found in many minerals such as feldspar, mica, and amphibole. **Potassium-argon dating** can be used to date igneous rocks from 100,000 years to over a billion years old. Because it can be used to date geologically young materials, the technique has been useful in estimating the age of deposits containing the bones of human ancestors.

Uranium-Lead Dating

Two isotopes of uranium are used for radiometric dating. Uranium-238 (^{238}U) decays to form lead-206 (^{206}Pb) with a half-life of 4.47 billion years. Uranium-235 (^{235}U) decays to form lead-207 (^{207}Pb) with a half-life of 704 million years.

Uranium-lead dating is usually performed on crystals of the mineral **zircon**. When zircon forms in an igneous rock, the crystals readily accept atoms of uranium but reject atoms of lead. Therefore, if any lead is found in a zircon crystal, it can be assumed that it was produced from the decay of uranium. Zircon is resistant to weathering which makes it useful for dating geological events in ancient rocks. During metamorphic events, zircon crystals may form multiple crystal layers, with each layer recording the isotopic age of an event, thus tracing the progress of the several metamorphic events [16].



Figure 7.2.3: Zircon crystal. (By Eurico Zimbres, Tom Epaminondas; CC BY-SA 2.0 BR via Wikimedia Commons.)

Geologists have used zircon grains to do some amazing studies that illustrate how scientific conclusions can change with technological advancements. Zircon crystals from Western Australia that formed when the crust first differentiated from the mantle 4.4 billion years ago have been determined to be the oldest known rocks [6]. The zircon grains were incorporated into **metasedimentary** host rocks, sedimentary rocks showing signs of having undergone partial metamorphism. The host rocks were not very old but the embedded zircon grains were created 4.4 billion years ago and survived the subsequent processes of weathering, erosion, deposition, and metamorphism. From other properties of the zircon crystals, researchers concluded that not only were continental rocks exposed above sea level but also that conditions on the early Earth were cool enough for liquid water to exist on the surface. The presence of liquid water allowed the processes of weathering and erosion to take place [17]. Researchers at UCLA studied 4.1 billion-year-old zircon crystals and found carbon in the zircon crystals that may be biogenic in origin, meaning that life may have existed on Earth much earlier than previously thought [18].

Table 7.2.2: Common isotopes used for radiometric dating [7; 8].

Elements	Parent symbol	Daughter symbol	Half-life
Uranium-238/Lead-206	^{238}U	^{206}Pb	4.5 billion years
Uranium-235/Lead-207	^{235}U	^{207}Pb	704 million years
Potassium-40/Argon-40	^{40}K	^{40}Ar	1.25 billion years
Rubidium-87/Strontium-87	^{87}Rb	^{87}Sr	48.8 billion years
Carbon-14/Nitrogen-14	^{14}C	^{14}N	5730 years

Limitations of Radiometric Dating

The principles behind this dating method require two key assumptions. First, the mineral grains containing the isotope formed at the same time as the rock, such as minerals in an igneous rock that crystallized from magma. Second, the mineral crystals remain a closed system, meaning they are not subsequently altered by elements moving in or out of them.

These requirements place some constraints on the kinds of rock suitable for dating, with the igneous rock being the best. Metamorphic rocks are crystalline, but the processes of metamorphism may reset the clock and derived ages may represent a smear of different metamorphic events rather than the age of original crystallization. Detrital sedimentary rocks contain clasts from separate parent rocks from unknown locations and derived ages are thus meaningless. However, sedimentary rocks with precipitated minerals, such as evaporites, may contain elements suitable for radiometric dating. Igneous pyroclastic layers and lava flows within a sedimentary sequence can be used to date the sequence. Cross-cutting igneous rocks and sills can be used to bracket the ages of affected, older sedimentary rocks. The resistant mineral zircon, found as clasts in many ancient sedimentary rocks, has been successfully used for establishing very old dates, including the age of Earth's oldest known rocks [6]. Knowing that zircon minerals in metamorphosed sediments came from older rocks that are no longer available for study, scientists can date zircon to establish the age of the pre-metamorphic source rocks.

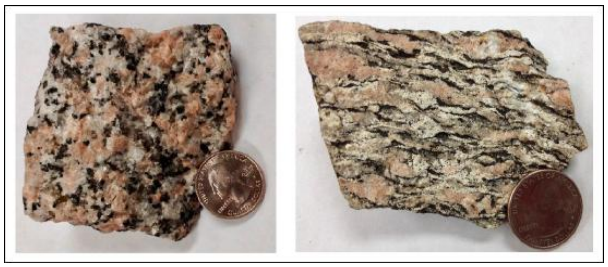


Figure 7.2.4: Granite (left) and gneiss (right). Dating a mineral within the granite would give the crystallization age of the rock while dating the gneiss might reflect the timing of metamorphism.

Ideally, several different radiometric techniques will be used to date the same rock. Agreement between these values indicates that the calculated age is accurate. Using a combination of radiometric dating, index fossils, and superposition, geologists have constructed a well-defined timeline of Earth history. For example, an overlying lava flow can give a reliable estimate of the age of a sedimentary rock formation in one location. Index fossils contained in this formation can then be matched to fossils in a different location, providing a good age measurement for that new rock formation as well. As this process has been repeated all over the world, estimates of rock and fossil ages have become more and more accurate.

Age of the Earth

The work of Hutton and other scientists gained attention after the Renaissance, spurring exploration into the idea of an ancient Earth. In the late 19th century William Thompson, a.k.a. Lord Kelvin, applied his knowledge of physics to develop the assumption that the Earth started as a hot molten sphere. He estimated the Earth is 98 million years old, but because of uncertainties in his calculations stated the age as a range of between 20 and 400 million years [12; 13]. This animation illustrates how Kelvin calculated this range and why his numbers were so far off, which has to do with unequal heat transfer within the Earth. It has also been pointed out that Kelvin failed to consider pliability and convection in the Earth's mantle as a heat transfer mechanism. Kelvin's estimate for Earth's age was considered plausible but not without challenge, and the discovery of radioactivity provided a more accurate method for determining ancient ages [14].



Figure 7.2.5: Artist's impression of Earth in the Hadean Eon, early in Earth's history.

In the 1950s, Clair Patterson (1922–1995) thought he could determine the age of the Earth using radioactive isotopes from meteorites, which he considered to be early solar system remnants that were present at the time Earth was forming. Patterson analyzed meteorite samples for uranium and lead using a mass spectrometer. He used the uranium/lead dating technique in determining the age of the Earth to be 4.55 billion years, give or take about 70 million (± 1.5%) [15]. The current estimate for the age of the Earth is 4.54 billion years, give or take 50 million (± 1.1%) [13]. It is remarkable that Patterson, who was still a graduate student at the University of Chicago, came up with a result that has been little altered in over 60 years, even as technology has improved dating methods.

Other Absolute Dating Techniques

Luminescence (aka Thermoluminescence): Radiometric dating is not the only way scientists determine numeric ages. Luminescence dating measures the time elapsed since some silicate minerals, such as coarse-sediments of silicate minerals, were last exposed to light or heat at the surface of Earth. All buried sediments are exposed to radiation from normal background radiation from the decay process described above. Some of these electrons get trapped in the crystal lattice of silicate minerals like quartz. When exposed at the surface, ultraviolet radiation and heat from the Sun release these electrons, but when the minerals are buried just a few inches below the surface, the electrons get trapped again. Samples of coarse sediments collected just a few feet below the surface are analyzed by stimulating them with light in a lab. This stimulation releases the trapped electrons as a photon of light which is called luminescence. The amount of luminescence released indicates how long the sediment has been buried. Luminescence dating is only useful for dating young sediment that is less than 1 million years old [20; 21]. In Utah, luminescence dating is used to determine when coarse-grained sediment layers were buried near a fault. This is one technique used to determine the recurrence interval of large earthquakes on faults like the Wasatch Fault that primarily cut coarse-grained material and lack buried organic soils for radiocarbon dating [22].

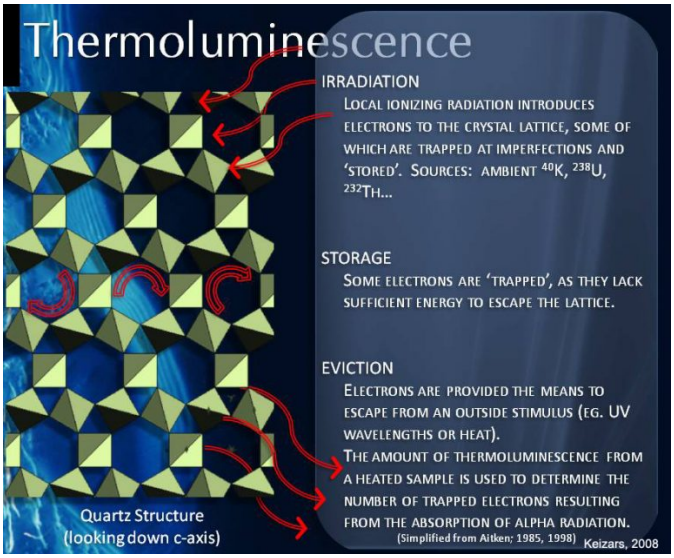


Figure 7.2.6: Thermoluminescence, a type of luminescence dating.

Fission Track: Fission track dating relies on damage to the crystal lattice produced when unstable ^{238}U decays to the daughter product ^{234}Th and releases an alpha particle. These two decay products move in opposite directions from each other through the crystal lattice leaving a visible track of damage. This is common in uranium-bearing mineral grains such as apatite. The tracks are large and can be visually counted under an optical microscope. The number of tracks corresponds to the age of the grains. Fission track dating works from about 100,000 to 2 billion (1×10^5 to 2×10^9) years ago. Fission track dating has also been used as a second clock to confirm dates obtained by other methods [23; 7].



Figure 7.2.7: Apatite from Mexico.

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7.3: Fossils and Evolution

Fossils are any evidence of past life preserved in rocks. They may be actual remains of body parts (rare), impressions of soft body parts, casts and molds of body parts (more common), body parts replaced by mineral (common) or evidence of animal behavior such as footprints and burrows. The body parts of living organisms range from the hard bones and shells of animals, soft cellulose of plants, soft bodies of jellyfish, down to single cells of bacteria and algae. Which body parts can be preserved? The vast majority of life today consists of soft-bodied and/or single-celled organisms, and will not likely be preserved in the geologic record except under unusual conditions. The best environment for preservation is the ocean, yet marine processes can dissolve hard parts and scavenging can reduce or eliminate remains. Thus, even under ideal conditions in the ocean, the likelihood of preservation is quite limited. For terrestrial life, the possibility of remains being buried and preserved is even more limited. In other words, the fossil record is incomplete and records only a small percentage of life that existed. Although incomplete, fossil records are used for stratigraphic correlation, using the **principle of faunal succession**, and provide a method used for establishing the age of a formation on the Geologic Time Scale.



Figure 7.3.1: Archaeopteryx lithographica, specimen displayed at the Museum für Naturkunde in Berlin.

Types of Preservation

Remnants or impressions of hard parts, such as a marine clamshell or dinosaur bone, are the most common types of fossils [24]. The original material has almost always been replaced with new minerals that preserve much of the shape of the original shell, bone, or cell. The common types of fossil preservation are actual preservation, permineralization, molds and casts, carbonization, and trace fossils.



Figure 7.3.2: Trilobites had a hard exoskeleton and is an early arthropod, the same group that includes modern insects, crustaceans, and arachnids.

Actual preservation is a rare form of fossilization where the original materials or hard parts of the organism are preserved. Preservation of soft-tissue is very rare since these organic materials easily disappear because of bacterial decay [24]. Examples of actual preservation are unaltered biological materials like insects in amber or original minerals like mother-of-pearl on the interior of a shell. Another example is mammoth skin and hair preserved in post-glacial deposits in the Arctic regions [25]. Rare mummification [26] have left fragments of soft tissue, skin, and sometimes even blood vessels of dinosaurs, from which proteins have been isolated and evidence for DNA fragments have been discovered [27].



Figure 7.3.3: Mosquito preserved in amber, an example of actual preservation.

Permineralization occurs when an organism is buried, and then elements in groundwater completely impregnate all spaces within the body, even cells. Soft body structures can be preserved in great detail, but stronger materials like bone and teeth are the most likely to be preserved. **Petrified wood** is an example of detailed cellulose structures in the wood being preserved. The University of California Berkeley website has more information on permineralization.



Figure 7.3.4: Permineralization in petrified wood.

Molds and **casts** form when the original material of the organism dissolves and leaves a cavity in the surrounding rock. The shape of this cavity is an external mold. If the mold is subsequently filled with sediments or a mineral precipitates, the organism's external shape is preserved as a cast. Sometimes internal cavities of organisms, such as internal casts of clams, snails, and even skulls are preserved as internal casts showing details of soft structures. If the chemistry is right, and burial is rapid, mineral nodules form around soft structures preserving the three-dimensional detail. This is called **authigenic mineralization**.

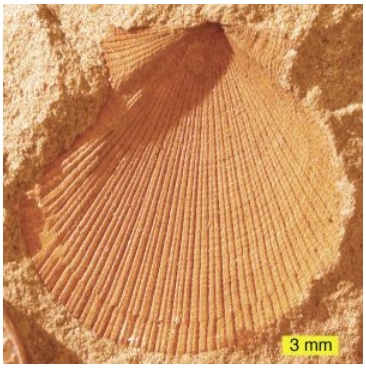


Figure 7.3.5: External mold of a clam.

Carbonization occurs when the organic tissues of an organism are compressed, the volatiles are driven out, and everything but the carbon disappears leaving a carbon silhouette of the original organism. Leaf and fern fossils are examples of carbonization [28].

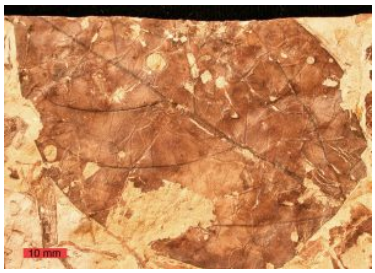


Figure 7.3.6: Carbonized leaf.

Trace fossils are indirect evidence left behind by an organism, such as burrows and footprints, as it lived its life. **Ichnology** is specifically the study of prehistoric animal tracks. Dinosaur tracks testify to their presence and movement over an area and even provide information about their size, gait, speed, and behavior [29; 30]. Burrows dug by tunneling organisms tell of their presence and mode of life [31; 32; 33]. Other trace fossils include fossilized feces called **coprolites** [34] and stomach stones called **gastroliths** [35] that provide information about diet and habitat.



Figure 7.3.7: Left: Footprints of the early crocodile Chirotherium. Right: Fossil animal droppings (coprolite).

Evolution

Evolution produced a variety of ancient fossils that are important to stratigraphic correlation. This section is a brief discussion of the process of evolution. The British naturalist Charles Darwin (1809-1882) recognized that life forms evolve into progeny life forms. He proposed **natural selection**—which operated on organisms living under environmental conditions that posed challenges to survival—was the mechanism driving the process of evolution forward.

The basic classification unit of life is the **species**: a population of organisms that exhibit shared characteristics and are capable of reproducing fertile offspring. For a species to survive, each individual within a particular population is faced with challenges posed by the environment and must survive them long enough to reproduce. Within the natural variations present in the population, there may be individuals possessing characteristics that give them some advantage in facing environmental challenges. These individuals are more likely to reproduce and pass these favored characteristics on to successive generations. If sufficient individuals in a population fail to surmount the challenges of the environment and the population cannot produce enough viable offspring, the species becomes extinct. The average lifespan of a species in the fossil record is around a million years. The fact that life still exists on Earth shows the role and importance of evolution as a natural process in meeting the continual challenges posed by our dynamic Earth. If the inheritance of certain distinctive characteristics is sufficiently favored over time, populations may become genetically isolated from one another, eventually resulting in the evolution of separate species. This genetic isolation may also be caused by a geographic barrier, such as an island surrounded by ocean. This theory of evolution by natural selection was elaborated by Darwin in his book *On the Origin of Species* [36]. Since Darwin's original ideas, technology has provided many tools and mechanisms to study how evolution and speciation take place and this arsenal of tools is growing. Evolution is well beyond the hypothesis stage and is a well-established theory of modern science.

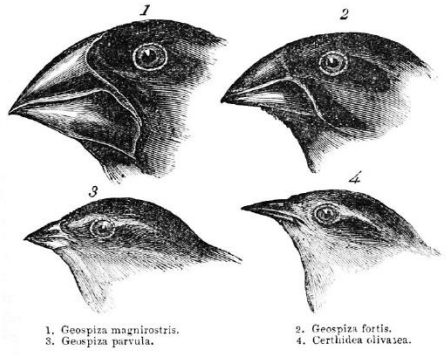


Figure 7.3.8: Illustration of Darwin's finches. Three separate species of finches evolved from the first finch, Geospiza magnirostris in the Galapagos. (By John Gould; public domain via Wikimedia Commons.)

Variation within populations occurs by the natural mixing of genes through sexual reproduction or from naturally occurring mutations. Some of this genetic variation can introduce advantageous characteristics that increase the individual's chances of survival. While some species in the fossil record show little morphological change over time, others show gradual or punctuated changes, within which intermediate forms can be seen.

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7.4: Correlation

Correlation is the process of establishing which sedimentary strata are of the same age but geographically separated. Correlation can be determined by using magnetic polarity reversals, rock types, unique rock sequences, or index fossils. There are four main types of correlation: stratigraphic, lithostratigraphic, chronostratigraphic, and biostratigraphic.

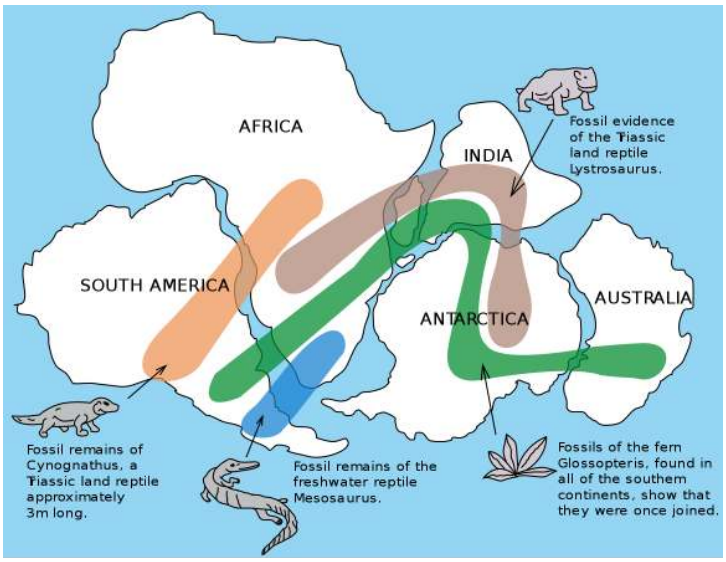


Figure 7.4.1: Image showing fossils that connect the continents of Gondwana (the southern continents of Pangea). Wegener used correlation to help develop the idea of continental drift.

Stratigraphic Correlation

Stratigraphic correlation is the process of establishing which sedimentary strata are the same age at distant geographical areas by means of their stratigraphic relationship. Geologists construct geologic histories of areas by mapping and making **stratigraphic columns**--a detailed description of the strata from bottom to top. An example of stratigraphic relationships and correlation occurs between Canyonlands National Park and Zion National Park in Utah. At Canyonlands, the Navajo Sandstone overlies the Kayenta Formation which overlies the cliff-forming Wingate Formation. In Zion, the Navajo Sandstone overlies the Kayenta formation which overlies the cliff-forming Moenave Formation. Based on the stratigraphic relationship, the Wingate and Moenave Formations correlate. These two formations have unique names because their composition and outcrop pattern are slightly different. Other strata in the Colorado Plateau and their sequence can be recognized and correlated over thousands of square miles.

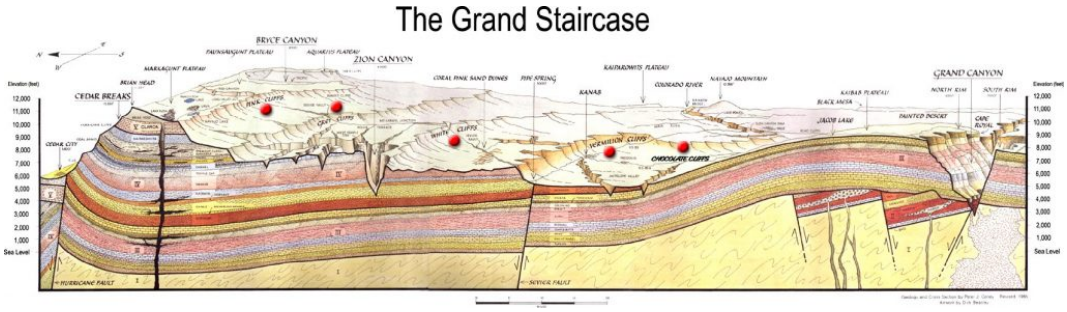


Figure 7.4.2: Correlation of Paleozoic and Mesozoic strata along the Grand Staircase from the Grand Canyon to Zion Canyon, Bryce Canyon, and Cedar Breaks. (Source: National Park Service)

Lithostratigraphic Correlation

Lithostratigraphic correlation establishes a similar age of strata based on **lithology**, which is the composition and physical properties of that strata. *Lithos* is Greek for stone and *-logy* comes from the Greek word for doctrine or science. Lithostratigraphic correlation can be used to correlate whole formations long distances or can be used to correlate smaller strata within formations to trace their extent and regional depositional environments.

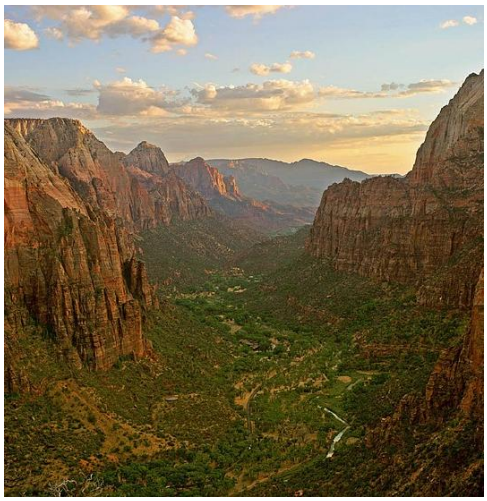


Figure 7.4.3: View of Navajo Sandstone from Angel's Landing in Zion National Park.

For example, the Navajo Sandstone, which makes up the prominent walls of Zion National Park, is the same Navajo Sandstone in Canyonlands because the lithology of the two are identical even though they are hundreds of miles apart. Extensions of the same Navajo Sandstone formation are found miles away in other parts of southern Utah, including Capitol Reef and Arches National Parks. Further, this same formation is called the Aztec Sandstone in Nevada and Nugget Sandstone near Salt Lake City because they are lithologically distinct enough to warrant new names.



Figure 7.4.4: Stevens Arch in the Navajo Sandstone at Coyote Gulch some 125 miles away from Zion National Park.

Chronostratigraphic Correlation

Chronostratigraphic correlation matches rocks of the same age, even though they are made of different lithologies. Different lithologies of sedimentary rocks can form at the same time at different geographic locations because depositional environments vary geographically. For example, at any one time in a marine setting, there could be this sequence of depositional environments from the beach to deep marine: beach, nearshore area, shallow marine lagoon, reef, slope, and deep marine. Each depositional environment will have a unique sedimentary rock formation. On the figure of the Permian El Capitan Reef at Guadalupe National Monument in West Texas, the red line shows a chronostratigraphic timeline that represents a snapshot in time. Shallow-water marine lagoon/back reef area is gray, the main Capitan reef is dark blue, and deep-water marine siltstone is yellow. All three of these unique lithologies were forming at the same time in Permian along this red timeline.

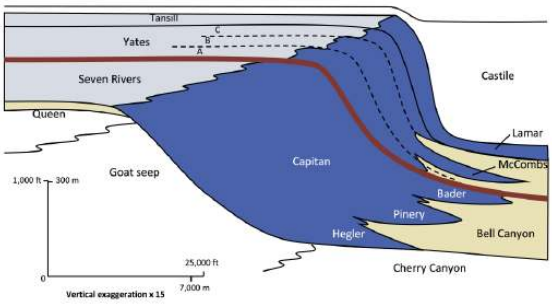


Figure 7.4.5: Cross-section of the Permian El Capitan Reef at Guadalupe National Monument, Texas. The red line shows a chronostratigraphic timeline that represents a snapshot in time in which the shallow marine lagoon/back reef area (gray), main Capitan reef (dark blue), and deep marine siltstones (yellow) were all being deposited at the same time. (By Kindred Grey; adapted under fair use from DOI:10.2110/cor.89.13.0003)

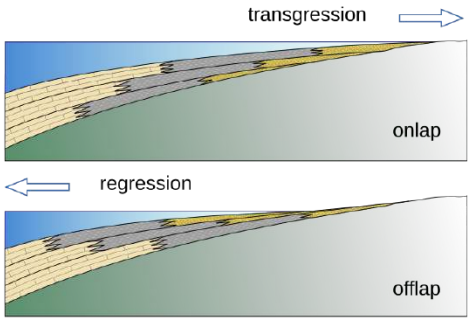


Figure 7.4.6: The rising sea levels of transgressions create onlapping sediments, regressions create offlapping. Ocean water is shown in blue so the timeline is on the surface below the water. At the same time sandstone (buff color), limestone (gray), and shale (mustard color) are all forming at different depths of water.

Biostratigraphic Correlation

Biostratigraphic correlation uses index fossils to determine strata ages. **Index fossils** represent assemblages or groups of organisms that were uniquely present during specific intervals of geologic time. **Assemblages** refer to a group of fossils. Fossils allow geologists to assign a formation to an absolute date range, such as the Jurassic Period (199 to 145 million years ago), rather than a relative time scale. In fact, most of the geologic time ranges are mapped to fossil assemblages. The most useful index fossils come from lifeforms that were geographically widespread and had a species lifespan that was limited to a narrow time interval. In other words, index fossils can be found in many places around the world, but only during a narrow time frame. Some of the best fossils for biostratigraphic correlation are **microfossils,** most of which came from single-celled organisms. As with microscopic organisms today, they were widely distributed across many environments throughout the world. Some of these microscopic organisms had hard parts, such as exoskeletons or outer shells, making them better candidates for preservation. **Foraminifera**, single-celled organisms with calcareous shells, are an example of an especially useful index fossil for the Cretaceous Period and Cenozoic Era [37].

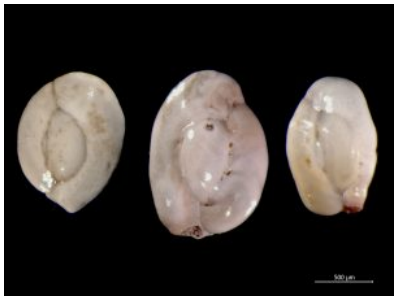


Figure 7.4.7: Foraminifera, microscopic creatures with calcareous shells.

Conodonts are another example of microfossils useful for biostratigraphic correlation of the Cambrian through Triassic Periods. **Conodonts** are tooth-like phosphatic structures of an eel-like multi-celled organism that had no other preservable hard parts. The

conodont-bearing creatures lived in shallow marine environments all over the world. Upon death, the phosphatic hard parts were scattered into the rest of the marine sediments. These distinctive tooth-like structures are easily collected and separated from limestone in the laboratory.

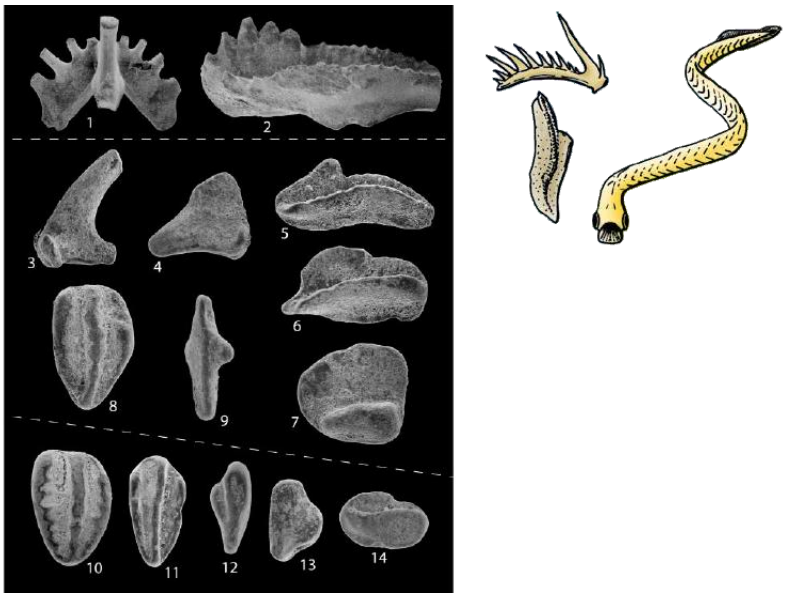


Figure 7.4.8: Left: Conodonts. (By USGS; public domain via Wikimedia Commons.) Right: Artist reconstruction of the conodont animal (right) alongside its teeth (left).

Because the conodont creatures were so widely abundant, rapidly evolving, and readily preserved in sediments, their fossils are especially useful for correlating strata, even though knowledge of the actual animal possessing them is sparse. Scientists in the 1960s carried out a fundamental biostratigraphic correlation that tied Triassic conodont zonation into **ammonoids**, which are extinct ancient cousins of the pearly nautilus. Up to that point, ammonoids were the only standard for Triassic correlation, so cross-referencing micro- and macro-index fossils enhanced the reliability of biostratigraphic correlation for either type [38]. That conodont study went on to establish the use of conodonts to internationally correlate Triassic strata located in Europe, Western North America, and the Arctic Islands of Canada [39].

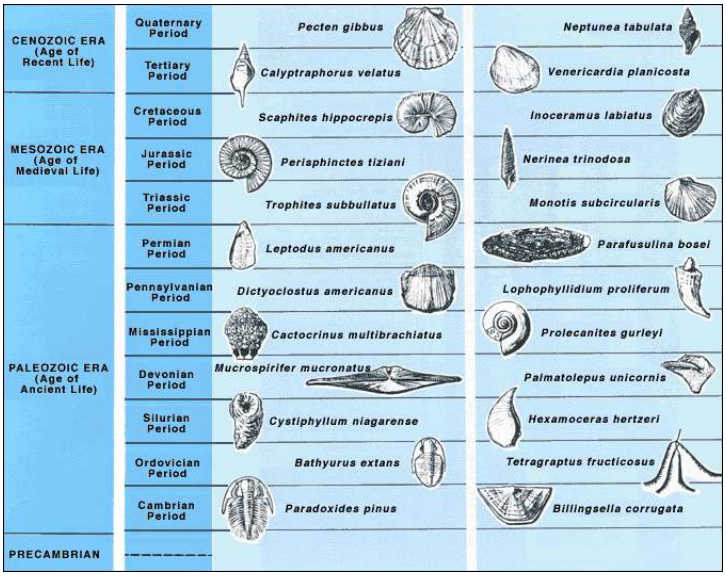


Figure 7.4.9: Index fossils used for biostratigraphic correlation.

Geologic Time Scale

Geologic time has been subdivided into a series of divisions by geologists. **Eon** is the largest division of time, followed by **era**, **period**, **epoch**, and **age**. The partitions of the **geologic time scale** are the same everywhere on Earth; however, rocks may or may not be present at a given location depending on the geologic activity going on during a particular period of time. Thus, we have the concept of time vs. rock, in which time is an unbroken continuum but rocks may be missing and/or unavailable for study. The figure of the geologic time scale represents time flowing continuously from the beginning of the Earth, with the time units presented in an unbroken sequence.

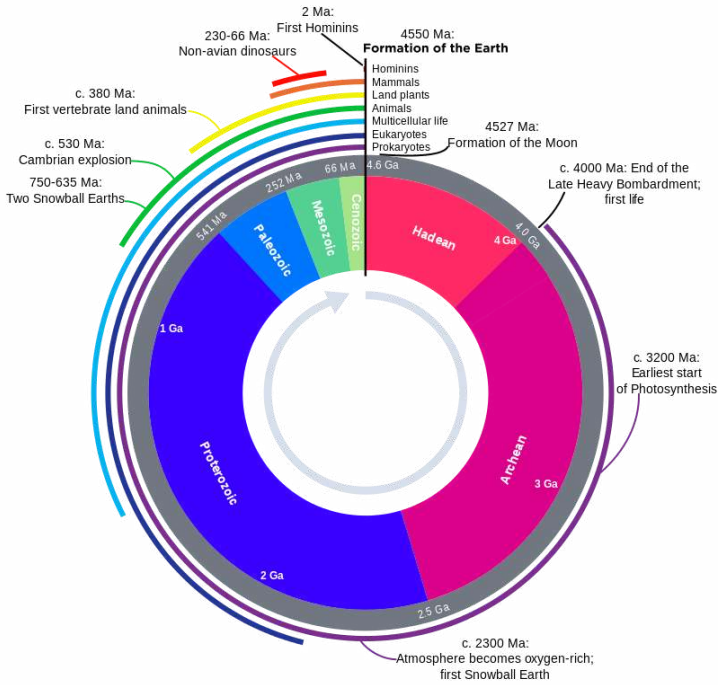


Figure 7.4.10: Geologic time on Earth, represented circularly, to show the individual time divisions and important events. Ga=billion years ago, Ma=million years ago.

The geologic time scale was developed during the 19th century using the principles of stratigraphy. The relative order of the time units was determined before geologists had the tools to assign numerical ages to periods and events. Biostratigraphic correlation used fossils to assign era and period names to sedimentary rocks on a worldwide scale [40]. With the expansion of science and technology, some geologists think the influence of humanity on natural processes has become so great they are suggesting a new geologic time period, known as the **Anthropocene**. [39; 41].

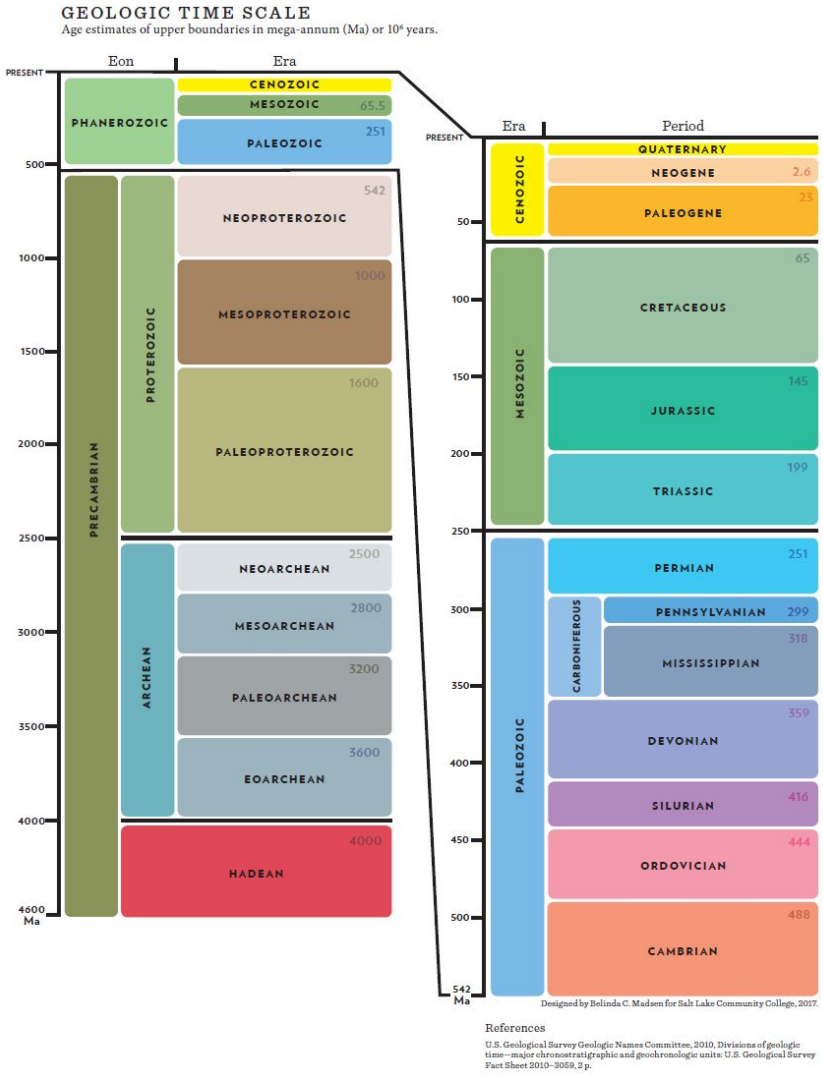


Figure 7.4.11: Geologic Time Scale with ages shown in millions of years ago (Ma).

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7.S: Summary

Summary

Events in Earth history can be placed in sequence using the five principles of relative dating. The geologic time scale was completely worked out in the 19th Century using these principles without knowing any actual numeric ages for the events. The discovery of radioactivity in the late 1800s enabled absolute dating, the assignment of numerical ages to events in the Earth's history, using decay of unstable radioactive isotopes. Accurately interpreting radiometric dating data depends on the type of rock tested and accurate assumptions about isotope baseline values. With a combination of relative and absolute dating, the history of geological events, age of Earth, and a geologic time scale have been determined with considerable accuracy. Stratigraphic correlation is an additional tool used for understanding how depositional environments change geographically. Geologic time is vast, providing plenty of time for the evolution of various lifeforms, and some of these have become preserved as fossils that can be used for biostratigraphic correlation. The geologic time scale is continuous, although the rock record may be broken because rocks representing certain time periods may be missing.

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CHAPTER OVERVIEW

8: Earth History

Learning Objectives

- Describe the turbulent beginning of Earth during the Hadean and Archean Eons.
- Identify the transition to the modern atmosphere, plate tectonics, and evolution that occurred in the Proterozoic Eon.
- Describe the Paleozoic evolution and extinction of invertebrates with hard parts, fish, amphibians, reptiles, tetrapods, and land plants; and tectonics and sedimentation associated with the supercontinent Pangea.
- Describe the Mesozoic evolution and extinction of birds, dinosaurs, and mammals; and tectonics and sedimentation associated with the breakup of Pangea.
- Describe the Cenozoic evolution of mammals and birds, paleoclimate, and tectonics that shaped the modern world.

Entire courses and careers have been based on the wide-ranging topics covering Earth's history. Throughout the long history of Earth, change has been the norm. Looking back in time, an untrained eye would see many unfamiliar life forms and terrains. The main topics studied in Earth history are paleogeography, paleontology, and paleoecology and paleoclimatology—respectively, past landscapes, past organisms, past ecosystems, and past climates. The changes that have occurred since the inception of Earth are vast and significant. From the oxygenation of the atmosphere, the progression of life forms, the assembly and deconstruction of several supercontinents, to the extinction of more life forms than exist today, having a general understanding of these changes can put present change into a more rounded perspective. This chapter will cover briefly the origin of the universe and the 4.6 billion year history of Earth. This Earth history will focus on the major physical and biological events in each eon and era.

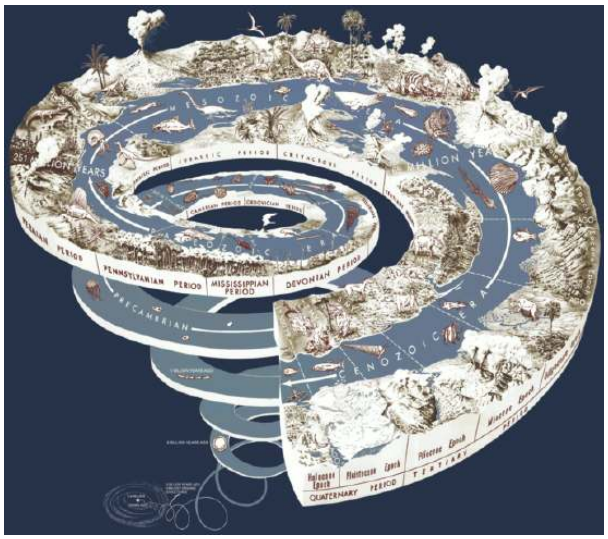


Figure 8.1: The geologic time spiral--A path to the past. (By USGS; public domain via Wikimedia Commons.)

8.1: Hadean Eon
8.2: Archean Eon
8.3: Proterozoic Eon
8.4: Paleozoic Era
8.5: Mesozoic Era
8.6: Cenozoic Era
8.S: Summary

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8.1: Hadean Eon

Geoscientists use the geologic time scale to assign relative age names to events and rocks, separating major events in Earth's history based on significant changes as recorded in rocks and fossils. This chapter summarizes the most notable events of each major time interval.

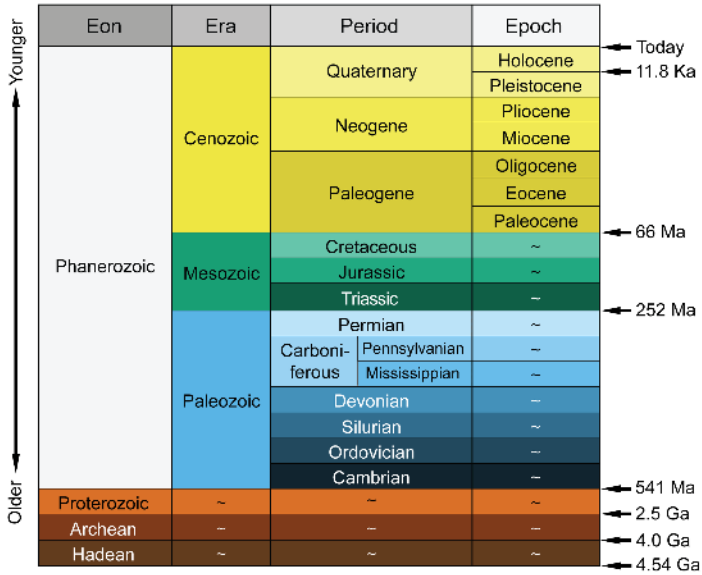


Figure 8.1.1: Geologic Time Scale. (By Jonathan R. Hendricks; CC BY-NC-SA 4.0 via Earth@Home)

The **Hadean Eon**, named after the Greek god and ruler of the underworld Hades, is the oldest eon and dates from 4.5–4.0 billion years ago. This time represents Earth's earliest history, during which the planet was characterized by a partially molten surface, volcanism, and asteroid impacts. Several mechanisms made the newly forming Earth incredibly hot: gravitational compression, radioactive decay, and asteroid impacts. Most of this initial heat still exists inside the Earth. The Hadean was originally defined as the birth of the planet occurring 4.0 billion years ago and preceding the existence of many rocks and life forms. However, geologists have dated minerals at 4.4 billion years, with evidence that liquid water was present [18]. There is possibly even evidence of life existing over 4.0 billion years ago. However, the most reliable record for early life, the microfossil record, starts at 3.5 billion years ago.



Figure 8.1.2: Artist's impression of the Earth in the Hadean.

Origin of Earth's Crust

As Earth cooled from its molten state, minerals started to crystallize and settle, resulting in a separation of minerals based on density and the creation of the crust, mantle, and core. The earliest Earth was chiefly molten material and would have been rounded by gravitational forces so it resembled a ball of lava floating in space. As the outer part of the Earth slowly cooled, the high melting-point minerals (from Bowen's Reaction Series) formed solid slabs of early crust. These slabs were probably unstable and easily reabsorbed into the liquid magma until the Earth cooled enough to allow numerous larger fragments to form a thin primitive crust. Scientists generally assume this crust was oceanic and mafic in composition and littered with impacts, much like the Moon's current crust. There is still some debate over when plate tectonics started, which would have led to the formation of the continental

and the felsic crust [23]. Regardless of this, as Earth cooled and solidified, less dense felsic minerals floated to the surface of the Earth to form the crust, while the denser mafic and ultramafic materials sank to form the mantle, and the highest-density iron and nickel sank into the core. This differentiated the Earth from a homogenous planet into a heterogeneous one with layers of felsic crust, mafic crust, ultramafic mantle, and an iron and nickel core.

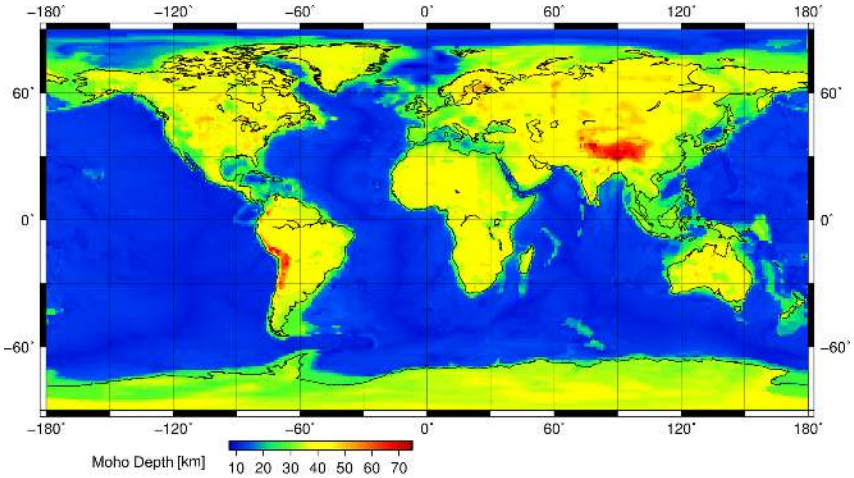


Figure 8.1.3: The global map of the depth of the Moho, or thickness of the crust.

Origin of the Moon

Several unique features of Earth's Moon have prompted scientists to develop the current hypothesis about its formation. The Earth and Moon are tidally locked, meaning that as the Moon orbits, one side always faces the Earth and the opposite side is not visible to us. Also and most importantly, the chemical compositions of the Earth and Moon show nearly identical isotope ratios [24] and volatile content. Apollo missions returned from the Moon with rocks that allowed scientists to conduct very precise comparisons between Moon and Earth rocks. Other bodies in the solar system and meteorites do not share the same degree of similarity and show much higher variability. If the Moon and Earth formed together, this would explain why they are so chemically similar.

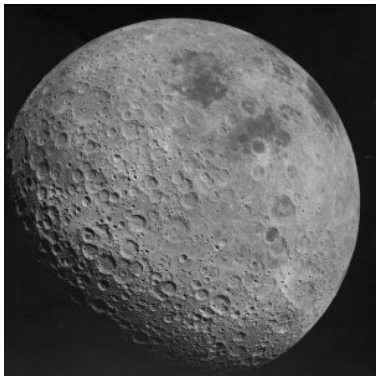


Figure 8.1.4: Far side of the Moon.

Many ideas have been proposed for the origin of the Moon: The Moon could have been captured from another part of the solar system and formed in place together with the Earth, or the Moon could have been ripped out of the early Earth. None of the proposed explanations can account for all the evidence. The currently prevailing hypothesis is the **giant-impact hypothesis**. It proposes a body about half of Earth's size must have shared at least parts of Earth's orbit and collided with it, resulting in a violent mixing and scattering of material from both objects. Both bodies would be composed of a combination of materials, with more of the lower density splatter coalescing into the Moon. This may explain why the Earth has a higher density and thicker core than the Moon. New evidence suggests that part of the object that collided with Earth is embedded in Earth's mantle (*doi: https://doi.org/10.1038/d41586-023-03385-9*).



Figure 8.1.5: Artist's concept of the giant impact from a Mars-sized object that could have formed the moon.

Origin of Earth's Water

Explanations for the origin of Earth's water include volcanic outgassing, comets, and meteorites. The **volcanic outgassing hypothesis** for the origin of Earth's water is that it originated from inside the planet, and emerged via tectonic processes as vapor associated with volcanic eruptions [27]. Since all volcanic eruptions contain some water vapor, at times more than 1% of the volume, these alone could have created Earth's surface water. Another likely source of water was from space. Comets are a mixture of dust and ice, with some or most of that ice being frozen water. Seemingly dry meteors can contain small but measurable amounts of water, usually trapped in their mineral structures [28; 29]. During heavy bombardment periods later in Earth's history, its cooled surface was pummeled by comets and meteorites, which could be why so much water exists above ground. There isn't a definitive answer to what process is the source of ocean water. Earth's water isotopically matches water found in meteorites much better than that of comets [30]. However, it is hard to know if Earth processes could have changed the water's isotopic signature over the last 4-plus billion years. It is possible that all three sources contributed to the origin of Earth's water.

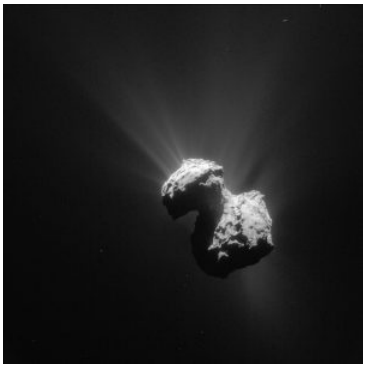


Figure 8.1.5: Water vapor leaves comet 67P/Churyumov–Gerasimenko.

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8.2: Archean Eon

The **Archean Eon**, which lasted from 4.0–2.5 billion years ago, is named after the Greek word for beginning. This eon represents the beginning of the rock record. Although there is current evidence that rocks and minerals existed during the Hadean Eon, the Archean has a much more robust rock and fossil record.



Figure 8.2.1: Artist's impression of the Archean.

Late Heavy Bombardment

Objects were chaotically flying around at the start of the solar system, building the planets and moons. There is evidence that after the planets formed, about 4.1–3.8 billion years ago, a second large spike of asteroid and comet impacted the Earth and Moon in an event called the **Late Heavy Bombardment**. Meteorites and comets in stable or semi-stable orbits became unstable and started impacting objects throughout the solar system. In addition, this event is called the **lunar cataclysm** because most of the Moon's craters are from this event. During the Late Heavy Bombardment, the Earth, Moon, and all planets in the solar system were pummeled by material from the asteroid and Kuiper belts. Evidence of this bombardment was found within samples collected from the Moon.

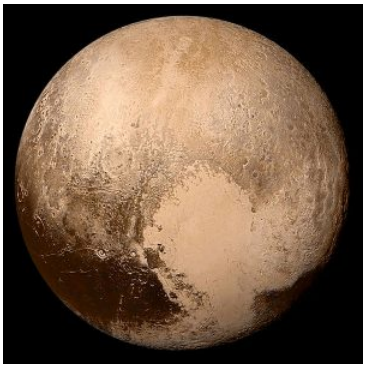


Figure 8.2.2: 2015 image from NASA's New Horizons probe of Pluto. The lack of impacts found on the Tombaugh Regio (the heart-shaped plain, lower right) has been inferred as being younger than the Late Heavy Bombardment and the surrounding surface due to its lack of impacts.

It is universally accepted that the solar system experienced extensive asteroid and comet bombardment at its start; however, some other process must have caused the second increase in impacts hundreds of millions of years later. A leading theory blames gravitational resonance between Jupiter and Saturn for disturbing orbits within the asteroid and Kuiper belts [33] based on a similar process observed in the Eta Corvi star system.

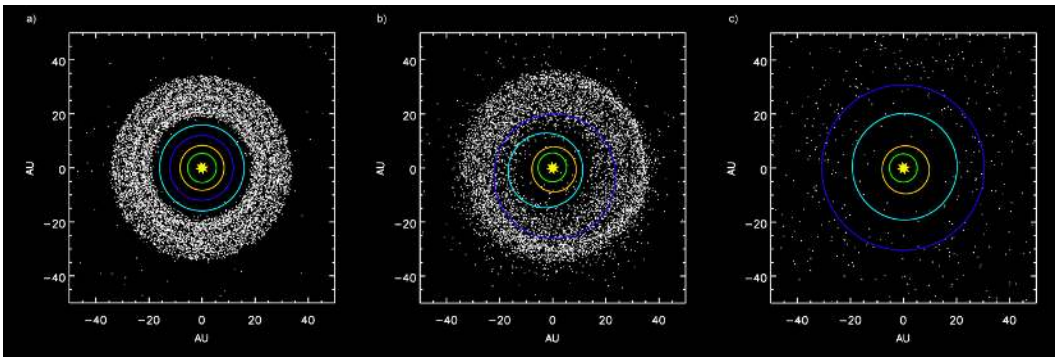


Figure 8.2.3: Simulation of before, during, and after the Late Heavy Bombardment. The orbits of the planets are denoted by the following colors: Jupiter, green; Saturn, orange; Uranus, light blue; Neptune, dark blue. Left: The solar system before the Jupiter-Saturn resonance. Middle: The scattering of Kuiper belt objects into the solar system after the orbital shift of Neptune. Right: The solar system after the ejection of Kuiper belt bodies by Jupiter.

Origin of the Continents

In order for plate tectonics to work as it does currently, it necessarily must have continents. However, the easiest way to create continental material is via assimilation and differentiation of existing continents. This chicken-and-egg quandary over how continents were made in the first place is not easily answered because of the great age of continental material and how much evidence has been lost during tectonics and erosion. While the timing and specific processes are still debated, volcanic action must have brought the first continental material to the Earth's surface during the Hadean, 4.4 billion years ago [18]. This model does not solve the problem of continent formation since magmatic differentiation seems to need a thicker crust. Nevertheless, the continents formed by some incremental process during the early history of Earth. The best idea is that density differences allowed lighter felsic materials to float upward and heavier ultramafic materials and metallic iron to sink. These density differences led to the layering of the Earth, the layers that are now detected by seismic studies. Early protocontinents accumulated felsic materials as developing plate-tectonic processes brought lighter material from the mantle to the surface [36].

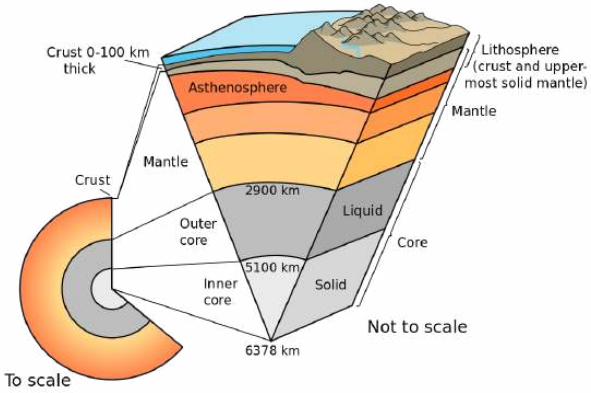


Figure 8.2.4: The layers of the Earth. Physical layers include the lithosphere and asthenosphere; chemical layers are crust, mantle, and core.

The first solid evidence of modern plate tectonics is found at the end of the Archean, indicating at least some continental lithosphere must have been in place. This evidence does not necessarily mark the starting point of plate tectonics; remnants of earlier tectonic activity could have been erased by the rock cycle.

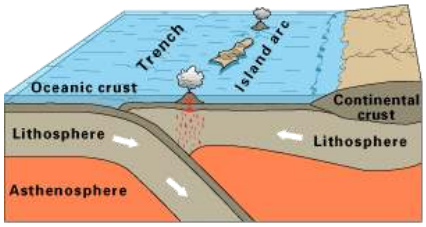


Figure 8.2.5: Subduction of an oceanic plate beneath another oceanic plate, forming a trench and an island arc. Several island arcs might combine and eventually evolve into a continent.

The stable interiors of the current continents are called **cratons** and were mostly formed in the Archean Eon. A craton has two main parts: the **shield**, which is crystalline basement rock near the surface, and the **platform** made of sedimentary rocks covering the shield. Most cratons have remained relatively unchanged with most tectonic activity having occurred around cratons instead of within them. Whether they were created by plate tectonics or another process, Archean continents gave rise to the Proterozoic continents that now dominate our planet.

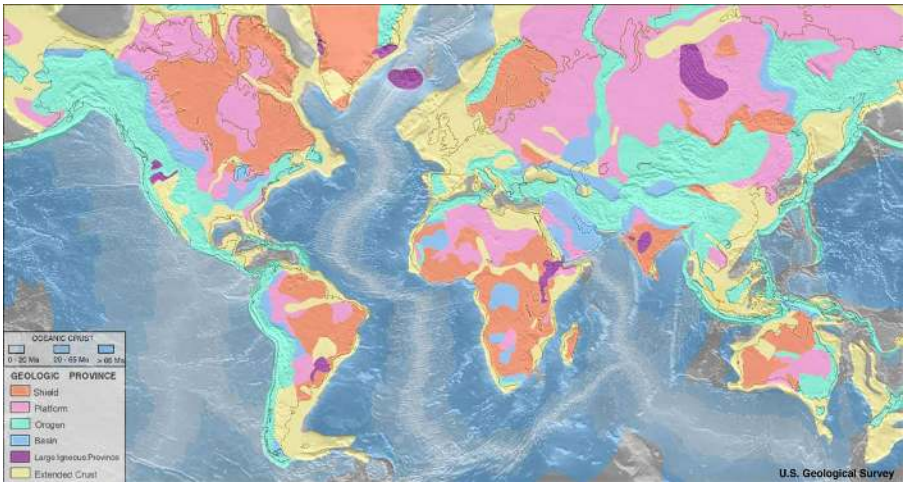


Figure 8.2.6: Geologic provinces with shield rocks (orange) and platform rocks (pink) comprising cratons, the stable interiors of continents.

The general guideline as to what constitutes a continent and differentiates oceanic from continental crust is under some debate. At passive margins, continental crust grades into the oceanic crust, making a distinction difficult. Even island-arc and hot-spot material can seem more closely related to continental crust than oceanic. Continents usually have a craton in the middle with felsic igneous rocks. There is evidence that submerged masses like Zealandia, that includes present-day New Zealand, would be considered a continent [39]. Continental crust that does not contain a craton is called a **continental fragment**, such as the island of Madagascar off the east coast of Africa.

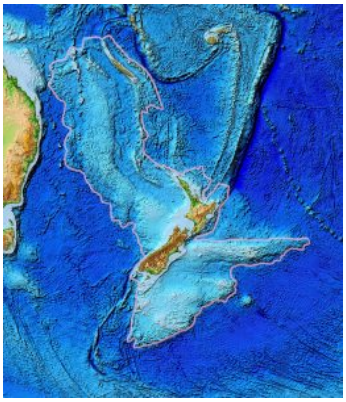


Figure 8.2.7: The continent of Zealandia.

First Life on Earth

Life most likely started during the late Hadean or early Archean Eons. The earliest evidence of life is chemical signatures, microscopic filaments, and microbial mats. Carbon found in 4.1 billion-year-old zircon grains has a chemical signature suggesting an organic origin. Other evidence of early life is the 3.8–4.3 billion-year-old microscopic filaments from a hydrothermal vent deposit in Quebec, Canada. While the chemical and microscopic filaments evidence is not as robust as fossils, there is significant fossil evidence for life at 3.5 billion years ago. These first well-preserved fossils are photosynthetic microbial mats, called **stromatolites**, found in Australia [41].



Figure 8.2.8: Fossilized microbial mats or stromatolites from Australia that are about 3.5 billion years old. (By Didier Descouens; CC BY-SA 4.0 via Wikimedia Commons.)

Although the origin of life on Earth is unknown, hypotheses include a chemical origin in the early atmosphere and ocean, deep-sea hydrothermal vents, and delivery to Earth by comets or other objects. One hypothesis is that life arose from the chemical environment of the Earth's early atmosphere and oceans, which was very different than today. The oxygen-free atmosphere produced a reducing environment with abundant methane, carbon dioxide, sulfur, and nitrogen compounds. This is what the atmosphere is like on other bodies in the solar system. In the famous Miller-Urey experiment, researchers simulated early Earth's atmosphere and lightning within a sealed vessel. After igniting sparks within the vessel, they discovered the formation of amino acids, the fundamental building blocks of proteins [42].

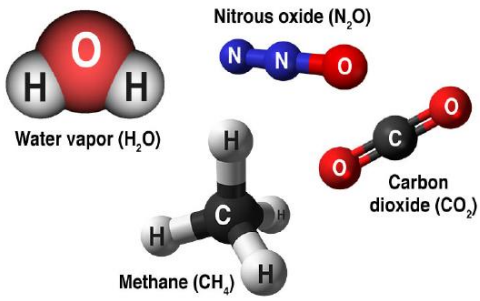
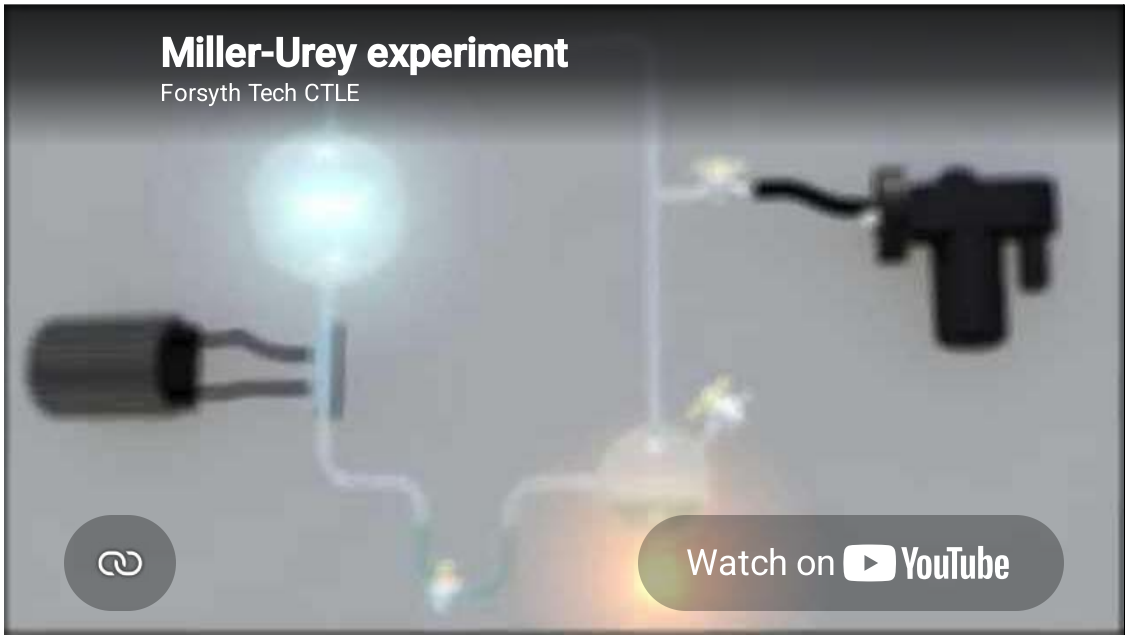
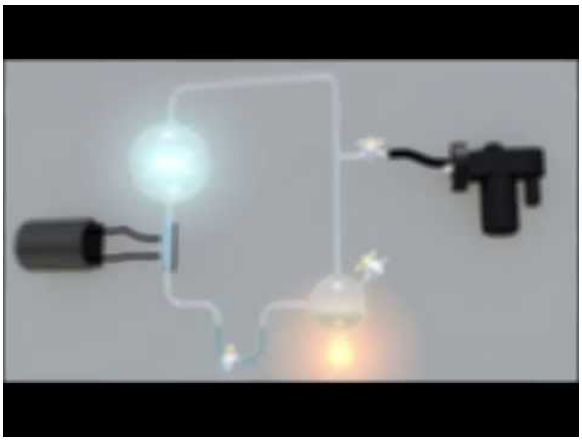


Figure 8.2.9: Greenhouse gases were more common in Earth's early atmosphere.

Animation of the original Miller-Urey 1959 experiment that simulated the early atmosphere and created amino acids from simple elements and compounds.





In 1977, when scientists discovered an isolated ecosystem around hydrothermal vents on a deep-sea mid-ocean ridge, it opened the door for another explanation of the origin of life. The hydrothermal vents have a unique ecosystem of critters with chemosynthesis as the foundation of the food chain instead of photosynthesis. The ecosystem derives its energy from hot chemical-rich waters pouring out of underground towers. This suggests that life could have started on the deep ocean floor and derived energy from the heat from the Earth's interior via chemosynthesis. Scientists have since expanded the search for life to more unconventional places, like Jupiter's icy moon Europa.

Another possibility is that life or its building blocks came to Earth from space, carried aboard comets or other objects. Amino acids, for example, have been found within comets and meteorites. This intriguing possibility also implies a high likelihood of life existing elsewhere in the cosmos.

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8.3: Proterozoic Eon

The **Proterozoic Eon**, meaning “earlier life,” is the eon of time after the Archean eon and ranges from 2.5 billion years old to 541 million years old. During this time, most of the central parts of the continents had formed and the plate tectonic process had started. Photosynthesis in organisms such as single-celled cyanobacteria (like stromatolites) had already been slowly adding oxygen to the oceans. As cyanobacteria evolved into multicellular organisms, they completely transformed the oceans and later the atmosphere by adding massive amounts of free oxygen gas (O_2) and initiated what is known as the **Great Oxygenation Event**. This drastic environmental change decimated the anaerobic (non-oxygen) bacteria, which could not survive in the presence of free oxygen [47]. In an oxygenated world, aerobic organisms could thrive in ways they could not earlier.

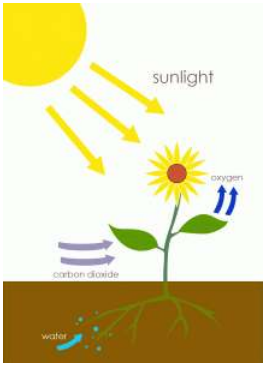


Figure 8.3.1: Diagram showing the main products and reactants in photosynthesis. The one product that is not shown is sugar, which is the chemical energy that goes into constructing the plant, and the energy that is stored in the plant which is used later by the plant or by animals that consume the plant.

Oxygen also changed the chemistry of the planet in significant ways. For example, iron can be carried in solution in a non-oxygenated environment. In chemistry, this is known as a **reducing environment**. However, once the environment was oxygenated, iron combined with free oxygen to form solid precipitates of iron oxide, such as the minerals hematite or magnetite. These precipitates accumulated into large deposits with red chert known as **banded-iron formations,** which are dated at around 2 billion years old [48].

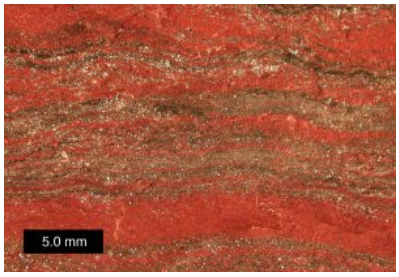


Figure 8.3.2: Alternating bands of iron-rich and silica-rich mud, formed as oxygen combined with dissolved iron.

The formation of the banded iron lasted a long time and prevented the oxygen levels from increasing significantly in the oceans since precipitation took the oxygen out of the water and formed alternating layers of iron-oxide minerals and red chert. Eventually, as oxygen continued to be made, absorption of oxygen in mineral precipitation leveled off, and dissolved oxygen gas eventually saturated the oceans and started bubbling out into the atmosphere. Oxygenation of the atmosphere is the single biggest event that distinguishes the Archean and the Proterozoic environments [49]. In addition to changing mineral and ocean chemistry, the Great Oxygenation Event is also tabbed as the likely cause of Earth’s first glaciation, the **Huron Glaciation**, that occurred around 2.1 billion years ago [50]. Free oxygen reacted with methane in the atmosphere, turning it into carbon dioxide. Carbon dioxide and methane are called **greenhouse gases** because they trap heat within the Earth's atmosphere, like the insulated glass of a greenhouse. Methane is a more effective insulator than carbon dioxide, and as CO_2 increased in the atmosphere, the greenhouse effect actually decreased, thus cooling the planet.

Rodinia

By the Proterozoic Eon, lithospheric plates had formed and started moving according to plate tectonic motions similar to today. As the moving plates collided, the ocean basins closed to form a **supercontinent** called **Rodinia**. It formed about 1 billion years ago and broke up at the end of the Proterozoic, about 750-600 million years ago. One of the resulting fragments was a continental mass called **Laurentia**, that would later become North America. The reconstruction of Rodinia has been accomplished by matching and aligning ancient mountain chains to assemble the pieces like a jigsaw puzzle, and using paleomagnetic data to orient them with respect to magnetic north.

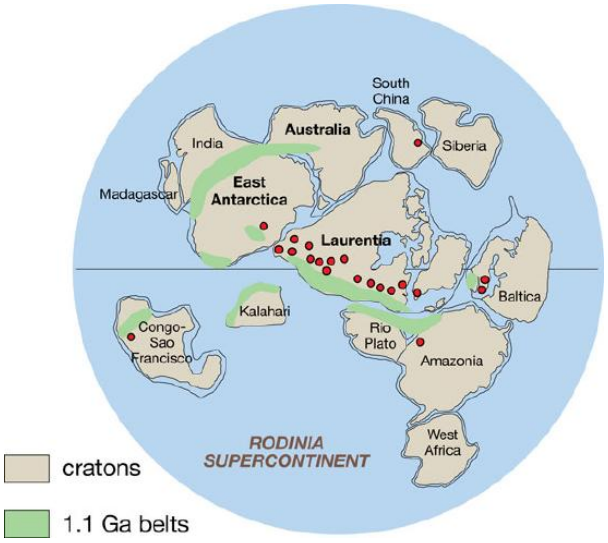


Figure 8.3.3: One possible reconstruction of Rodinia 1.1 billion years ago. Source: John Goodge, modified from Dalziel (1997).

As examples of the complexity of the issue and disagreements among geologists over the reconstructions, there are at least six different models of what broke away from Laurentia in the Panthalassa Ocean (early Pacific), including Australia, Antarctica [53], parts of China, the Tarim craton north of the Himalaya [54], Siberia [55], or the Kalahari craton of eastern Africa [56]. Regardless of the exact details, it was this breakup that created lots of biologically-favorable shallow water environments that fostered the evolutionary breakthroughs that mark the start of the next eon, the Phanerozoic.

Life Evolves

Early life in the Archean and earlier is poorly documented in the fossil record, but based on chemical evidence and evolutionary theory, scientists propose life would have been single-celled photosynthetic organisms such as cyanobacteria in stromatolites. Fossil cyanobacteria in these stromatolites produced free oxygen in the atmosphere through photosynthesis. Cyanobacteria, archaea and bacteria are **prokaryotes**, i.e. single-celled organisms with simple cells that lack a cell nucleus and other organelles.



Figure 8.3.4: Modern cyanobacteria (as stromatolites) in Shark Bay, Australia.

However, during the Proterozoic, a large evolutionary step occurred with the appearance of **eukaryotes**. Evolving around 2.1-1.6 billion years ago, eukaryotic cells are more complex, having a nucleus and cell organelles. The nuclear DNA is capable of more complex replication and regulation than that of prokaryotic cells. The organelles include mitochondria for producing energy and

chloroplasts for photosynthesis. The eukaryote branch in the tree of life gave rise to fungi, plants and animals. About 1.2 billion years ago, another important event in Earth's biological history occurred when some eukaryotes invented **sexual reproduction** [59]. By sharing genetic material between reproducing individuals (male and female), genetic variability was greatly increased in their offspring. This genetic mixing accelerated evolutionary change, allowed more complexity among individual organisms, and eventually, ecosystems.

Proterozoic land surfaces were barren of plants and animals, and geologic processes actively shaped the environment differently because land surfaces were not protected by leafy and woody vegetation. For example, rain and rivers would have caused erosion at much higher rates on land surfaces devoid of plants. This resulted in thick accumulations of pure quartz sandstone from the Proterozoic Eon such as the extensive quartzite formations in the core of the Uinta Mountains in Utah.

Fauna during the **Ediacaran Period** (635.5-541 million years ago, [60]), known as the **Ediacaran fauna**, offers the first glimpse at the diversity of ecosystems that evolved toward the end of the Proterozoic. These soft-bodied organisms were among the first multicellular life forms and may have been similar to soft jellyfish or worm-like organisms [61; 62]. Since the Ediacaran fauna did not have hard parts like shells, they are not well preserved in Proterozoic rocks. However, studies suggest that they were widespread around Earth's oceans [63]. Scientists still debate how many of these are extinct evolutionary dead-ends or the ancestors to modern biological groups [61]. The transition of life from the soft-bodied Ediacaran forms to the explosion of forms with hard parts at the end of the Proterozoic and beginning of the Phanerozoic made a dramatic difference in our ability to understand Earth history and the history of life.

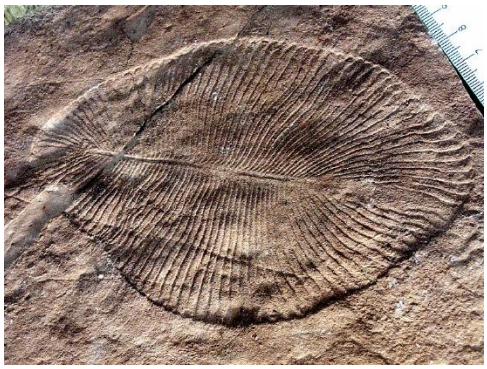


Figure 8.3.5: Dickinsonia, a typical Ediacaran fossil.

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8.4: Paleozoic Era

The **Phanerozoic Eon** is the most recent eon and represents time in which fossils are common, 541 million years ago to today. The word Phanerozoic means “visible life.” Older rocks, collectively known as the **Precambrian** (sometimes referred to as the Cryptozoic, meaning “invisible life”), are less common and have only rare fossils, and the fossils that exist represent soft-bodied life forms. The invention of hard parts like claws, scales, shells, and bones made fossils more easily preserved, and thus, easier to find. Since the younger rocks of the Phanerozoic are more common and contain the majority of fossils, the study of this eon yields much greater detail. It is further subdivided into three eras: Paleozoic (“ancient life”), Mesozoic (“middle life”), and Cenozoic (“recent life”). The next three sections are on these important eras.



Figure 8.4.1: The trilobites had a hard exoskeleton and were an early arthropod, the same group that includes modern insects, crustaceans, and arachnids.

The **Paleozoic Era** was dominated by marine organisms, but by the middle of the era, plants and animals had evolved to live and reproduce on land, including amphibians and reptiles. Fish evolved jaws and fins evolved into jointed limbs. Lungs evolved and life emerged from the sea onto land to become the first four-legged tetrapods, amphibians. Amphibians eventually evolved into reptiles once they developed hard-shelled eggs. From reptiles evolved an early ancestor to birds and mammals, and their scales became feathers and fur. The Carboniferous Period near the end of the Paleozoic had some of the most productive forests in the history of Earth and produced the coal that powered the Industrial Revolution in Europe and the United States. Tectonically, during the early Paleozoic, North America was separated from the other continents until the supercontinent Pangea formed towards the end of the era.

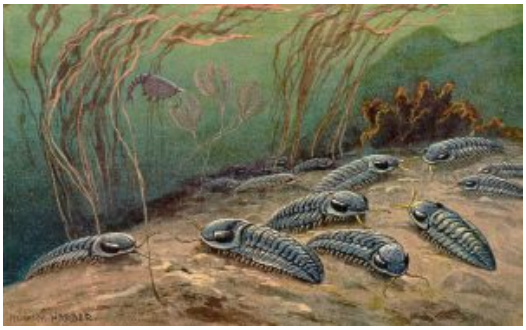


Figure 8.4.2: Trilobites, by Heinrich Harder, 1916.

Paleozoic Tectonics and Paleogeography

After the breakup of Rodinia toward the end of the Proterozoic, sea level remained high relative to land in the early Paleozoic. During the Paleozoic Era, sea levels rose and fell four times. With each sea-level rise, the majority of North America was covered by a shallow tropical ocean. Evidence of these submersions are the abundant marine sedimentary rocks such as limestone with fossils, corals and ooids. Extensive sea-level falls are documented by widespread unconformities. Today, the midcontinent has extensive marine sedimentary rocks from the Paleozoic, and western North America has thick layers of marine limestone on block-faulted mountain ranges such as Mt. Timpanogos near Provo, Utah.



Figure 8.4.3: Laurentia, which makes up the North American craton.

The assembly of **Pangea** (sometimes spelled Pangaea) was completed by the late Paleozoic Era. The name Pangea, originally coined by Alfred Wegener, means “all land.” Pangea formed a supercontinent by a series of tectonic events including subduction with island arc accretion, continental collisions and eventually ocean-basin closures. In North America, the tectonic events that occurred on the east coast and are known as the Taconic, Acadian, Caledonian, and Alleghanian **orogenies** [66; 68]. The Appalachian Mountains are the erosional remnants of these mountain-building events. Surrounding Pangea was a global ocean basin known as the **Panthalassa Ocean**. Continued plate movement extended the ocean into Pangea, forming a large bay called the **Tethys Sea** that formed between **Laurasia** (the northern continents of Laurentia and Eurasia) and **Gondwana** (the southern continents of South America, India, Australia, Antarctica, and Africa).

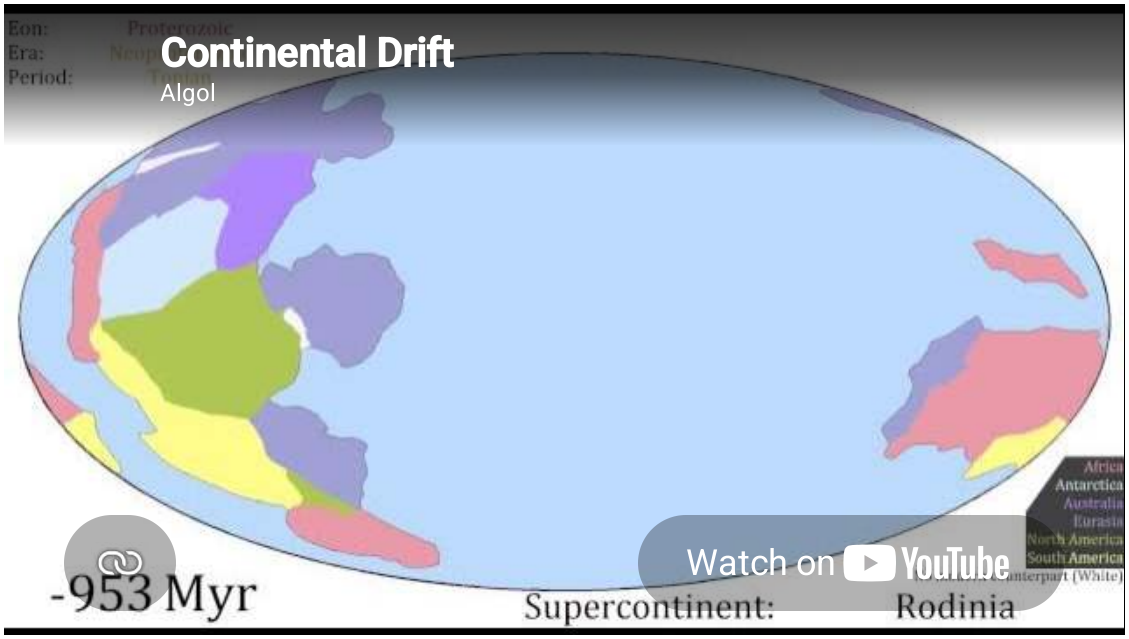


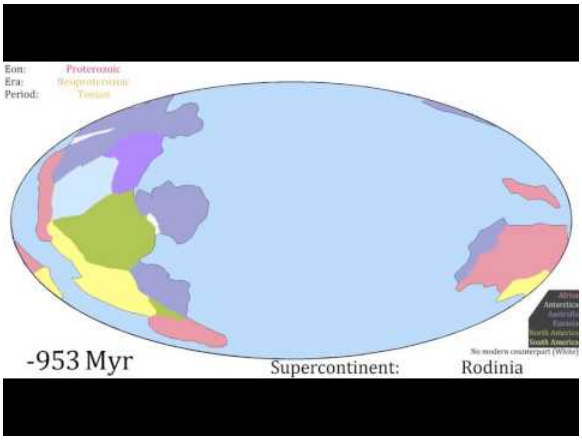
Figure 8.4.4: A reconstruction of Pangea, showing approximate positions of modern continents.

While the east coast of North America was tectonically active during the Paleozoic Era, the west coast remained mostly inactive as a passive margin during the early Paleozoic. The western edge of North American continent was near the present-day Nevada-Utah

border and was an expansive shallow continental shelf near the paleoequator. However, by the **Devonian Period**, the **Antler orogeny** started on the west coast and lasted until the **Pennsylvanian Period**. The Antler orogeny was a volcanic island arc that was accreted onto western North America with the subduction direction away from North America [71]. This created a mountain range on the west coast of North American called the Antler highlands and was the first part of building the land in the west that would eventually make most of California, Oregon, and Washington states. By the late Paleozoic, the **Sonoma orogeny** began on the west coast and was another collision of an island arc. The Sonoma orogeny marks the change in subduction direction to be toward North America with a volcanic arc along the entire west coast of North America by late Paleozoic to early Mesozoic Eras.

Animation of plate movement in the last 3.3 billion years. Pangea occurs at the 4:40 mark.





Paleozoic Evolution

The earliest Paleozoic had a significant biological explosion and contains evidence of a wide variety of evolutionary paths, including the evolutionary invention of hard parts like shells, spikes, teeth, and scales. Paleontologists refer to this event as the **Cambrian Explosion**, named after the first period in the Paleozoic. Scientists debate whether this was a manifestation of a true evolutionary pattern of diversification as a result of warmer climate following the late Proterozoic glacial environments, better preservation from easier to fossilize creatures, or simply an artifact of a more complete recent rock record. Ediacaran fauna, which lacked easily-fossilized hard parts, may have already been diverse and set the stage for the Cambrian Explosion [72]. Regardless, during the **Cambrian Period**, 541-485 million years ago, a large majority of the phyla of modern marine animals appeared [73]. These new organisms had simple cone- or tube-shaped shells that quickly became more complex. Some of these life forms have survived to today, and some were “experimental” whose lineage did not continue past the Cambrian period.

Fossil evidence of the Cambrian Explosion was first discovered by Charles Walcott in a rock layer called the **Burgess Shale** in western Canada in 1909. The Burgess Shale is a **lagerstätte**, or fossil site of exceptional preservation, including impressions of soft body parts. This allowed scientists to learn immense details of the animals that existed at the time, in addition to their tough shells, spikes, and claws. Other lagerstätte sites of similar age in China and Utah have allowed the forming of a fairly detailed picture of what the biodiversity was like in the Cambrian. The biggest mystery is the animals that do not fit existing lineages and are unique to that time. This includes famous fossil creatures like the first compound-eyed trilobites, and many other strange ones, including *Wiwaxia*, a spiked shell creature; *Hallucigenia*, a walking worm with spikes; *Opabinia*, a 5-eyed lobed arthropod with a trunk and a grappling claw at the end; and the related *Anomalocaris*, the alpha predator of the time, complete with grabbing arms and a deadly circular mouth full of teeth. Most notably at this time, an important ancestor to humans evolved. *Pikaia*, a segmented worm, is thought to be the earliest ancestor of the **Chordata** phylum (including **vertebrates**; animals with backbones [76]). These astonishing creatures offer a glimpse at evolutionary creativity.



Figure 8.4.5: Left: Original drawing from Walcott's 1912 description of Opabinia, with labels: fp = frontal appendage, e = eye, ths = thoracic somites, i = intestine, ab = abdominal segment. Top: Fossilized side view of Opabinia collected by Walcott from the Burgess Shale. (By Han Zeng, public domain via Wikimedia Commons.) Bottom: Animated illustration of Opabinia. (By Junnn11; CC BY-SA 4.0 via Wikimedia Commons.)

By the end of the Cambrian, mollusks, brachiopods, nautiloids, gastropods, graptolites, echinoderms, and trilobites covered the sea floor. Although most animal phyla appeared by the Cambrian, the biodiversity at the family, genus, and species level was low until the **Ordovician Period**. During the **Great Ordovician Biodiversification Event**, vertebrates and invertebrates (animals without backbone) became more diverse and complex at family, genus, and species level. The cause of the rapid speciation event is still debated but some likely causes are a combination of warm temperatures, expansive continental shelves near the equator, and more volcanism along the mid-ocean ridges. Some have shown evidence that an asteroid breakup event and consequent heavy meteorite impacts correlate with this diversification event. The additional volcanism added nutrients to ocean water helping support a robust ecosystem. Many life forms and ecosystems that would be recognizable in current times appeared at this time. Mollusks, corals, and arthropods in particular multiplied to dominate the oceans. [77].



Figure 8.4.6: A modern coral reef.

One important evolutionary advancement during the Ordovician Period was reef-building organisms, mostly colonial coral. Corals took advantage of the ocean chemistry, using calcite to build large structures [78] that resembled modern reefs like the Great Barrier Reef off the coast of Australia. These reefs housed thriving ecosystems of organisms that swam around, hid in, and crawled

over them. Reefs are important to paleontologists because of their preservation potential, massive size, and in-place ecosystems. Few other fossils offer more diversity and complexity than reef assemblages. Warm temperatures and high sea levels in the Ordovician most likely helped spur this diversification.

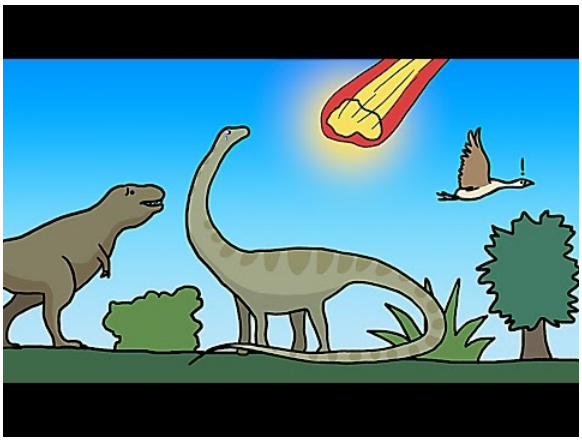


Figure 8.4.7: Guadalupe National Park is made of a giant fossil reef complex.

According to evidence from glacial deposits, a small ice age caused sea levels to drop and led to a major mass extinction by the end of the Ordovician. **Mass extinction** is when an unusually large number of species abruptly vanish and go extinct. This is the earliest of five mass extinction events documented in the fossil record. During this mass extinction, an unusually large number of species abruptly disappear in the fossil record (see video below).

3-minute video describing mass extinctions and how they are defined.





Life bounced back in the **Silurian** [78]. The period's major evolutionary event was the development of jaws from the forward pair of gill arches in bony fishes and sharks. Hinged jaws allowed fish to exploit new food sources and ecological niches. This period also included the start of armored fishes, known as the **placoderms**. In addition to fish and jaws, Silurian rocks provide the first evidence of **terrestrial** or land-dwelling plants and animals [80; 81]. The first **vascular plant,** *Cooksonia*, had woody tissues, pores for gas exchange, and veins for water and food transport. Insects, spiders, scorpions, and crustaceans began to inhabit moist, freshwater terrestrial environments.



Figure 8.4.8: The armor-plated fish (placoderm) Bothriolepis panderi from the Devonian of Russia.

The **Devonian Period**, called the Age of Fishes, saw a rise in plated fish and jawed fish [85], along with the lobe-finned fish. The lobe-finned fish (relatives of the modern lungfish and coelacanth) are important for their eventual evolution into **tetrapods**, the four-limbed vertebrate animals that can walk on land. The first evidence of land-walking fish, named *Tiktaalik* (about 385 million years ago), gave rise to early tetrapods. [87]. Though Tiktaalik was clearly a fish, it had some tetrapod structures as well. Several fossils from the Devonian are more tetrapod-like than fish-like but these weren't fully terrestrial. The first fully terrestrial tetrapod appeared in the **Mississippian** (early Carboniferous) **Period**. By the Mississippian (early Carboniferous) period, tetrapods had evolved into two main groups, **amphibians** and **amniotes**, from a common tetrapod ancestor. The amphibians were able to breathe air and live on land but still needed water to nurture their soft eggs. The first reptile (an amniote) could live and reproduce entirely on land with hard-shelled eggs that wouldn't dry out.

Land plants had also evolved into the first trees and forests [88]. Toward the end of the Devonian, another mass extinction event occurred. This extinction, while severe, is the least temporally defined, with wide variations in the timing of the event or events. Reef-building organisms were the hardest hit, leading to dramatic changes in marine ecosystems [89].

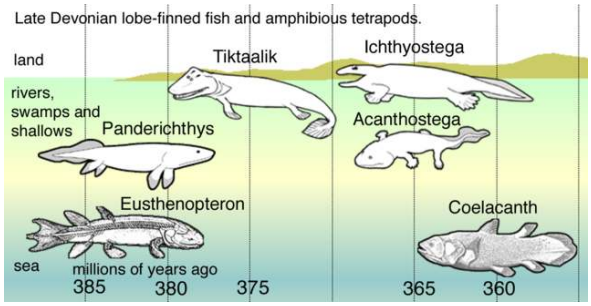


Figure 8.4.9: Several different types of fish and amphibians that led to walking on land.

The next time period called the **Carboniferous** (North American geologists have subdivided this into the Mississippian and **Pennsylvanian** periods), saw the highest levels of oxygen ever known, with forests (e.g., ferns, club mosses) and swamps dominating the landscape [90]. This helped cause the largest arthropods ever, like the millipede *Arthropleura*, at 2.5 meters (6.4 feet) long! It also saw the rise of a new group of animals, the **reptiles**. The evolutionary advantage that reptiles have over amphibians is the **amniote egg** (egg with a protective shell), which allows them to rely on non-aquatic environments for reproduction. This widened the terrestrial reach of reptiles compared to amphibians. This booming life, especially plant life, created cooling temperatures as carbon dioxide was removed from the atmosphere [92]. By the middle Carboniferous, these cooler temperatures led to an ice age (called the **Karoo Glaciation**) and less-productive forests. The reptiles fared much better than the amphibians, leading to their diversification [93]. This glacial event lasted into the early **Permian** [94].



Figure 8.4.10: A reconstruction of the giant arthropod (insects and their relatives) Arthropleura.

By the Permian, with Pangea assembled, the supercontinent led to a drier climate and even more diversification and domination by the reptiles [95]. The groups that developed in this warm-climate eventually radiated into **dinosaurs**. Another group, known as the **synapsids**, eventually evolved into **mammals**. Synapsids, including the famous sail-backed *Dimetrodon* are commonly confused with dinosaurs. **Pelycosaurs** (of the Pennsylvanian to early Permian like *Dimetrodon)* are the first group of synapsids that exhibit the beginnings of mammalian characteristics such as well-differentiated dentition: incisors, highly developed canines in lower and upper jaws and cheek teeth, premolars and molars. Starting in the late Permian, the second group of synapsids, called the **therapsids** (or mammal-like reptiles) evolve, and become the ancestors to mammals.

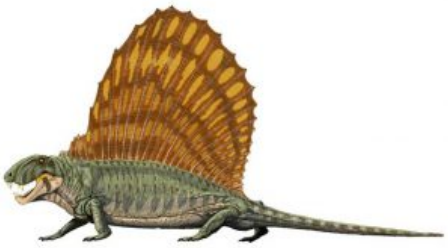


Figure 8.4.11: Reconstruction of Dimetrodon.

Permian Mass Extinction

The end of the Paleozoic era is marked by the largest mass extinction in Earth history. The Paleozoic era had two smaller mass extinctions, but these were not as large as the **Permian Mass Extinction**, also known as the **Permian-Triassic Extinction Event**. It is estimated that up to 96% of marine species and 70% of land-dwelling (terrestrial) vertebrates went extinct. Many famous organisms, like sea scorpions and trilobites, were never seen again in the fossil record. What caused such a widespread extinction event? The exact cause is still debated, though the leading idea relates to extensive volcanism associated with the **Siberian Traps**, which are one of the largest deposits of flood basalts known on Earth, dating to the time of the extinction event [100]. The eruption size is estimated at over 3 million cubic kilometers [101] that is approximately 4,000,000 times larger than the famous 1980 Mount St. Helens eruption in Washington. The unusually large volcanic eruption would have contributed a large amount of toxic gases, aerosols, and greenhouse gases into the atmosphere. Further, some evidence suggests that volcanism burned vast coal deposits releasing methane (a greenhouse gas) into the atmosphere. Greenhouse gases cause the climate to warm. This extensive addition of greenhouse gases from the Siberian Traps may have caused a runaway greenhouse effect that rapidly changed the climate, acidified the oceans, disrupted food chains, disrupted carbon cycling, and caused the largest mass extinction.

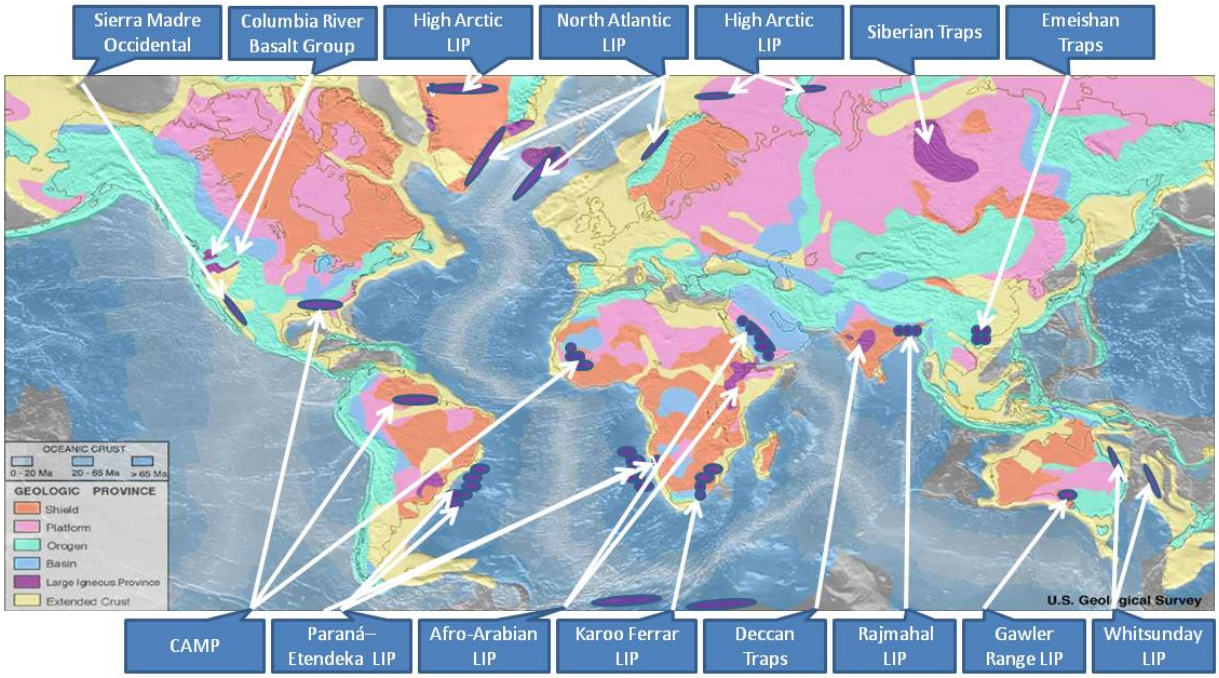


Figure 8.4.12: Map of global flood basalts. Note the largest is the Siberian Traps.

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8.5: Mesozoic Era

Following the Permian Mass Extinction, the **Mesozoic** (“middle life”) was from 252 million years ago to 66 million years ago. As Pangea started to break apart, mammals, birds, and flowering plants developed. The Mesozoic is probably best known as the age of reptiles, most notably, the dinosaurs.



Figure 8.5.1: Perhaps the greatest fossil ever found, a velociraptor attacked a protoceratops, and both were fossilized mid sequence.

Mesozoic Tectonics and Paleogeography

Pangea started breaking up (in a region that would become eastern Canada and United States) around 210 million years ago in the Late **Triassic**. Clear evidence for this includes the age of the sediments in the Newark Supergroup rift basins and the Palisades sill of the eastern part of North America and the age of the Atlantic ocean floor. Due to sea-floor spreading, the oldest rocks on the Atlantic’s floor are along the coast of northern Africa and the east coast of North America, while the youngest are along the mid-ocean ridge.

This age pattern shows how the Atlantic Ocean opened as the young Mid-Atlantic Ridge began to create the seafloor. This means the Atlantic Ocean started opening and was first formed here. The southern Atlantic opened next, with South America separating from central and southern Africa. Last (happening after the Mesozoic ended) was the northernmost Atlantic, with Greenland and Scandinavia parting ways. The breaking points of each rifted plate margin eventually turned into the passive plate boundaries of the east coast of the Americas today.

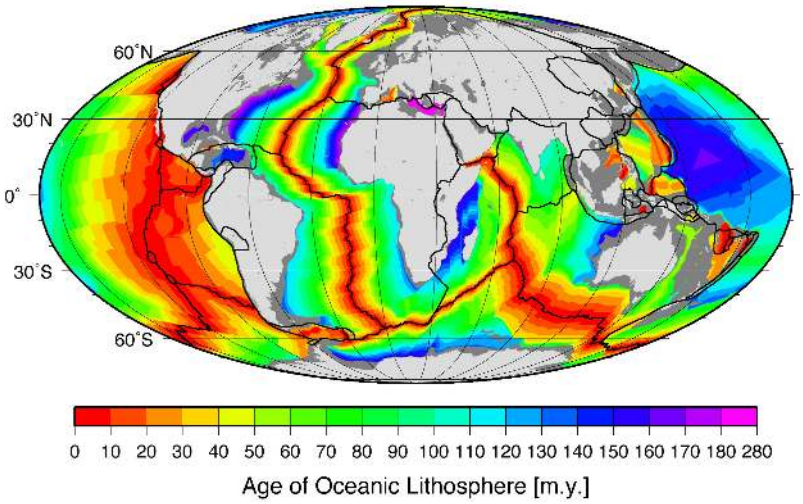


Figure 8.5.2: Age of oceanic lithosphere, in millions of years [m.y.]. Notice the differences in the Atlantic Ocean along the coasts of the continents.

Video of Pangea breaking apart and plates moving to their present locations. By Tanya Atwater.





In western North America, an active plate margin had started with subduction, controlling most of the tectonics of that region in the Mesozoic. Another possible island-arc collision created the Sonoma Orogeny in Nevada during the latest Paleozoic to the Triassic. In the **Jurassic**, another island-arc collision caused the **Nevadan orogeny**, a large Andean-style volcanic arc and thrust belt [105]. The **Sevier orogeny** followed in the **Cretaceous**, which was mainly a volcanic arc to the west and a thin-skinned fold and thrust belt to the east, meaning stacks of shallow faults and folds built up the topography. Many of the structures in the Rocky Mountains today date from this orogeny.

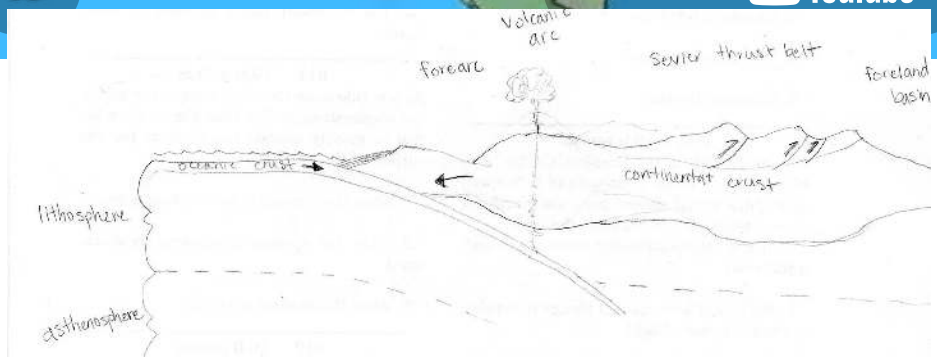


Figure 8.5.3: Sketch of the major features of the Sevier Orogeny.

Tectonics had an influence in one more important geographic feature in North America: the **Cretaceous Western Interior Foreland Basin**, which flooded during high sea levels forming the **Cretaceous Interior Seaway**. The **Farallon Plate**, an oceanic plate connected to the Pacific Plate (seen today as remnants such as the Juan de Fuca Plate, off the coast of the Pacific Northwest) was subducting in the west. Subduction was shallow at this time because a very young, hot and less dense portion of the Farallon Plate was subducted. This shallow subduction caused a downwarping in the central part of North America [107]. High sea levels due to shallow subduction, and increasing rates of seafloor spreading and subduction, high temperatures, and melted ice also contributed to the high sea levels [108]. These factors allowed a shallow epicontinental seaway that extended from the Gulf of Mexico to the Arctic Ocean to divide North America into two separate landmasses [109], **Laramidia** to the west and **Appalachia** to the east, for 25 million years [59]. Many of the coal deposits in Utah and Wyoming formed from swamps along the shores of this seaway [111]. By the end of the Cretaceous, cooling temperatures caused the seaway to regress [112].



Figure 8.5.4: The Western Interior Seaway 75 million years ago in the Cretaceous. (Modified by N. Ikeda from Gates et al; CC BY 1.0 via Wikimedia Commons.)

Mesozoic Evolution

The Mesozoic Era is dominated by reptiles, and more specifically, the dinosaurs. The Triassic had devastated ecosystems that took over 30 million years to fully re-emerge after the Permian Mass Extinction [113]. The first appearance of many modern groups of animals that would later flourish occurred at this time. This includes frogs (amphibians), turtles (reptiles), marine ichthyosaurs and plesiosaurs (marine reptiles), mammals, and the archosaurs. The **archosaurs** ("ruling reptiles") include ancestral groups that went extinct at the end of the Triassic, as well as the flying **pterosaurs**, **crocodilians**, and the dinosaurs. Archosaurs, like the placental mammals after them, occupied all major environments: terrestrial (dinosaurs), in the air (pterosaurs), aquatic (crocodilians) and even fully marine habitats (marine crocodiles). The pterosaurs, the first vertebrate group to take flight, like the dinosaurs and mammals, start small in the Triassic.



Figure 8.5.5: A Mesozoic scene from the late Jurassic.

At the end of the Triassic, another mass extinction event occurred [114], the fourth major mass extinction in the geologic record. This was perhaps caused by the **Central Atlantic Magmatic Province flood basalt** [115]. The end-Triassic extinction made certain lineages go extinct and helped spur the evolution of survivors like mammals, pterosaurs (flying reptiles), **ichthyosaurs/plesiosaurs/mosasaurs** (marine reptiles), and dinosaurs.

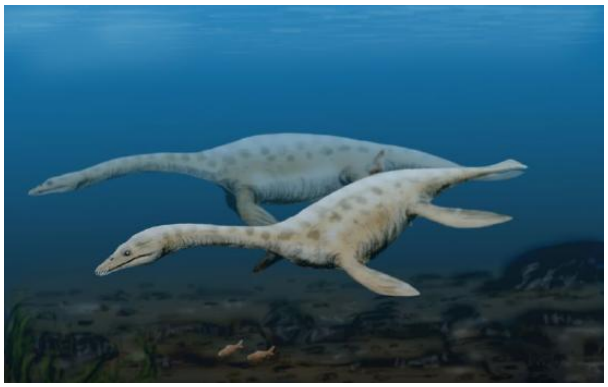


Figure 8.5.6: A drawing of the early plesiosaur Augustasaurus from the Triassic of Nevada. (By N. Tamura; CC BY-SA 3.0 via Wikimedia Commons.)

Mammals, as previously mentioned, got their start from a reptilian synapsid ancestor possibly in the late Paleozoic. Mammals stayed small, in mainly nocturnal niches, with insects being their largest prey. The development of warm-blooded circulation and fur may have been a response to this lifestyle [118].



Figure 8.5.7: Reconstruction of the small (less than 5 inches) Megazostrodon, one of the first animals considered to be a true mammal.

In the Jurassic, species that were previously common flourished due to a warmer and more tropical climate. The dinosaurs were relatively small animals in the Triassic period of the Mesozoic but became truly massive in the Jurassic. Dinosaurs are split into two groups based on their hip structure [120], i.e. orientation of the **pubis** and **ischium** bones in relationship to each other. This is referred to as the “reptile hipped” **saurischians** and the “bird-hipped” **ornithischians**. This has recently been brought into question by a new idea for dinosaur lineage [121].

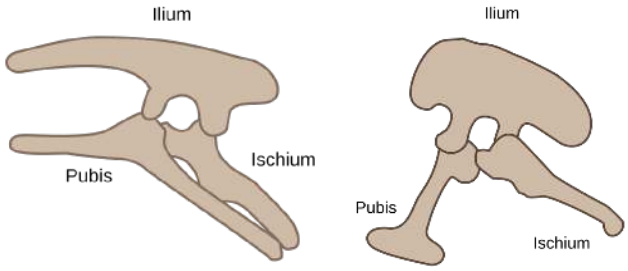


Figure 8.5.8: Left: Closed structure of an ornithischian hip, which is similar to a bird's hip. Right: Open structure of a saurischian hip, which is similar to a lizard's hip.

Most of the dinosaurs of the Triassic were saurischians, but all of them were bipedal. The major adaptive advantage dinosaurs had was changes in the hip and ankle bones, tucking the legs under the body for improved locomotion as opposed to the semi-erect gait of crocodiles or the sprawling posture of reptiles. In the Jurassic, limbs (or a lack thereof) were also important to another group of reptiles, leading to the evolution of *Eophis*, the oldest snake.

There is a paucity of dinosaur fossils from the Early and Middle Jurassic but by the Late Jurassic, they were dominating the planet. The saurischians diversified into the giant herbivorous (plant-eating) long-necked **sauropods** weighing up to 100 tons and bipedal carnivorous **theropods**, with the possible exception of the *Therizinosaurs* [125]. All of the ornithischians (e.g *Stegosaurus*, *Iguanodon*, *Triceratops*, *Ankylosaurus*, *Pachycephalosaurus*) were herbivorous with a strong tendency to have a "turtle-like" beak at the tips of their mouths.



Figure 8.5.9: Therizinosaurs, like Beipiaosaurus (shown in this restoration), are known for their enormous hand claws.

The pterosaurs grew and diversified in the Jurassic, and another notable aerial organism developed and thrived in the Jurassic: **birds**. When *Archaeopteryx* was found in the Solnhofen Lagerstätte of Germany, a seeming dinosaur-bird hybrid, it started the conversation on the origin of birds. The idea that birds evolved from dinosaurs occurred very early in the history of research into evolution, only a few years after Darwin's *On the Origin of Species* [127]. This study used a remarkable fossil of *Archeopteryx* from a transitional animal between dinosaurs and birds. Small meat-eating theropod dinosaurs were likely the branch that became birds due to their similar features [128]. A significant debate still exists over how and when powered flight evolved. Some have stated a running-start model [129], while others have favored a tree-leaping gliding model or even a semi-combination: flapping to aid in climbing.



Figure 8.5.10: Iconic "Berlin specimen" Archaeopteryx lithographica fossil from Germany.

The Cretaceous had a further diversification, specialization, and domination of the dinosaurs and other fauna. One of the biggest changes on land was the transition to angiosperm-dominated flora. **Angiosperms**, which are plants with flowers and seeds, had originated in the Cretaceous [132], switching many plains to grasslands by the end of the Mesozoic [133]. By the end of the period, they had replaced **gymnosperms** (evergreen trees) and ferns as the dominant plant in the world's forests. Haplodiploid eusocial insects (bees and ants) are descendants from Jurassic wasp-like ancestors that co-evolved with the flowering plants during this time period. The breakup of Pangea not only shaped our modern world's geography but biodiversity at the time as well. Throughout the Mesozoic, animals on the isolated, now separated island continents (formerly parts of Pangea), took strange evolutionary turns. This includes giant titanosaurian sauropods (*Argentinosaurus*) and theropods (*Giganotosaurus*) from South America.

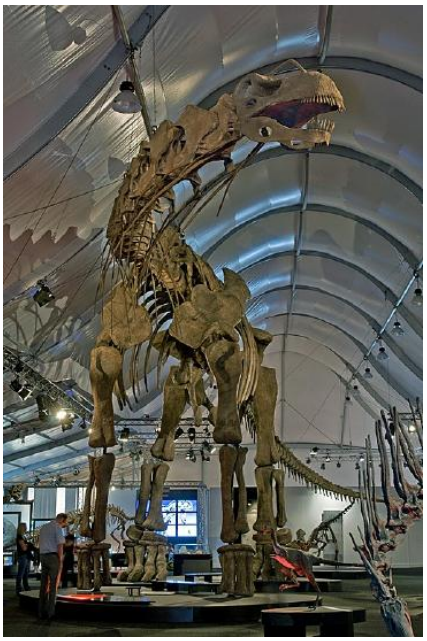


Figure 8.5.11: Reconstructed skeleton of Argentinosaurus, from Naturmuseum Senckenberg in Germany.

K-T Extinction

Similar to the end of the Paleozoic era, the Mesozoic Era ended with the **K-Pg Mass Extinction** (previously known as the **K-T Extinction**; "K" = Cretaceous and "Pg" = Paleogene) 66 million years ago [136]. This extinction event was likely caused by a large **bolide** (an extraterrestrial impactor such as an asteroid, meteoroid, or comet) that collided with Earth. Ninety percent of plankton species, 75% of plant species, and all the dinosaurs went extinct at this time.

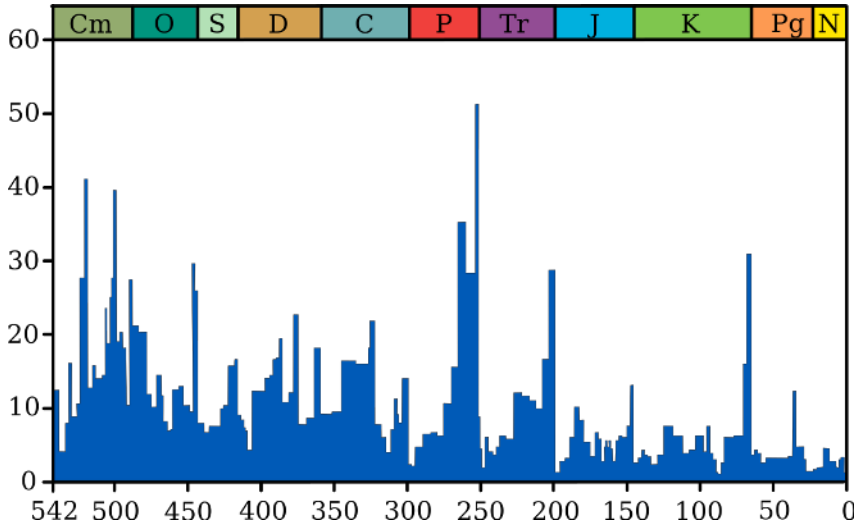


Figure 8.5.12: Graph of the rate of extinctions. Time in millions of years before present is on the horizontal axis. The percentage of marine animal genera becoming extinct during a given time interval is on the vertical axis. Across the top are the geologic time periods. Note the large spike at the end of the Cretaceous (labeled as K).

One of the strongest pieces of evidence comes from the element iridium. Quite rare on Earth, and more common in meteorites, it has been found all over the world in higher concentrations at a particular layer of rock that formed at the time of the K-Pg boundary. Soon other scientists started to find evidence to back up the claim. Melted rock spheres [138], a special type of “shocked” quartz called **stishovite**, that only is found at impact sites, was found in many places around the world. The huge impact created a strong thermal pulse that could be responsible for global forest fires [141], strong acid rains [142], a corresponding abundance of ferns, the first colonizing plants after a forest fire [143], enough debris thrown into the air to significantly cool temperatures afterward [144; 145], and a 2-km high tsunami inferred from deposits found from Texas to Alabama.



Figure 8.5.13: Artist's depiction of the impact event.

Still, with all this evidence, one large piece remained missing: the crater where the bolide impacted the planet. It was not until 1991 that the crater was confirmed using petroleum company geophysical data. Even though it is the third-largest confirmed crater on Earth at roughly 180 km wide, the **Chicxulub Crater** was hard to find due to being partially underwater and partially obscured by the dense forest canopy of the Yucatan Peninsula. Coring of the center of the impact called the **peak ring** contained granite, indicating the impact was so powerful that it lifted basement sediment from the crust several miles toward the surface [149]. In 2010, an international team of scientists reviewed 20 years of research and blamed the impact for the extinction [150].

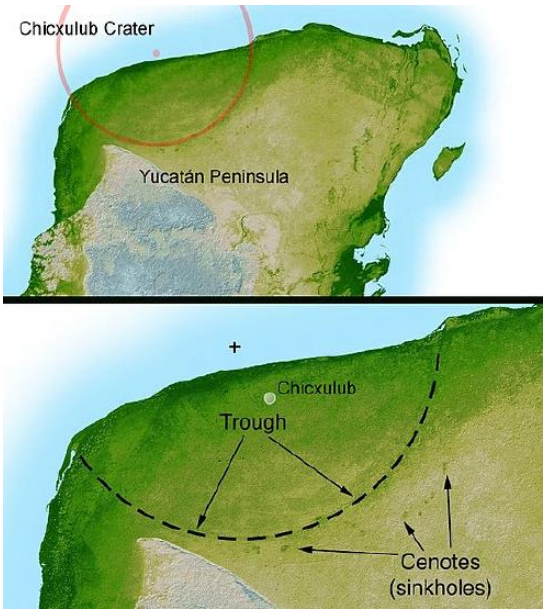


Figure 8.5.14: The land expression of the Chicxulub crater. The other side of the crater is within the Gulf of México.

With all of this information, it seems like the case would be closed. However, there are other events at this time which could have partially aided the demise of so many organisms. For example, sea levels are known to be slowly decreasing at the time of the K-Pg event, which is tied to marine extinctions [151], though any study on gradual vs. sudden changes in the fossil record is flawed due to the incomplete nature of the fossil record [152]. Another big event at this time was the **Deccan Traps** flood basalt volcanism in India. At over 1.3 million cubic kilometers of material, it was certainly a large source of material hazardous to ecosystems at the time, and it has been suggested as at least partially responsible for the extinction [153]. Some have found the impact and eruptions too much of a coincidence, and have even linked the two together [154].

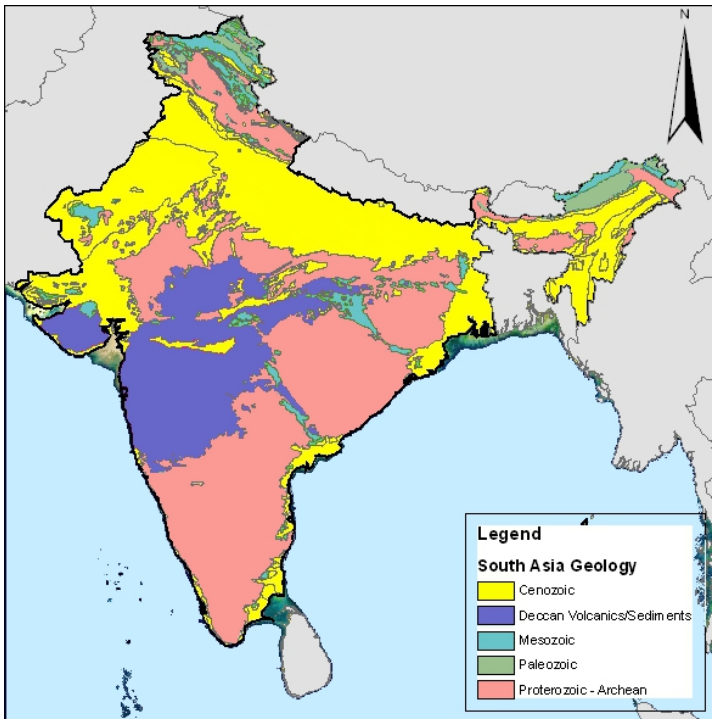


Figure 8.5.15: Geology of India, showing purple as Deccan Traps-related rocks.

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8.6: Cenozoic Era

The **Cenozoic**, meaning “new life,” is known as the age of mammals because it is in this era that mammals came to be a dominant and large life form, including human ancestors. Birds, as well, flourished in the open niches left by the dinosaur’s demise. Most of the Cenozoic has been relatively warm, with the main exception being the ice age that started about 2.558 million years ago and (despite recent warming) continues today. Tectonic shifts in the west caused volcanism but eventually changed the long-standing subduction zone into a transform boundary.

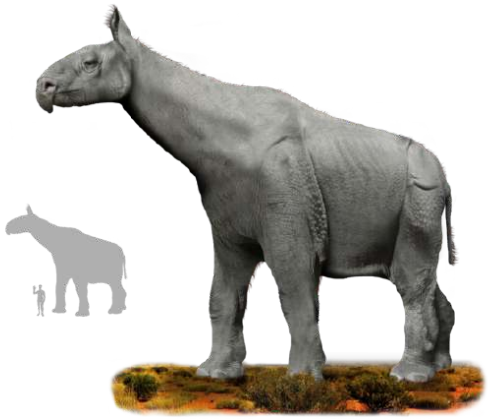
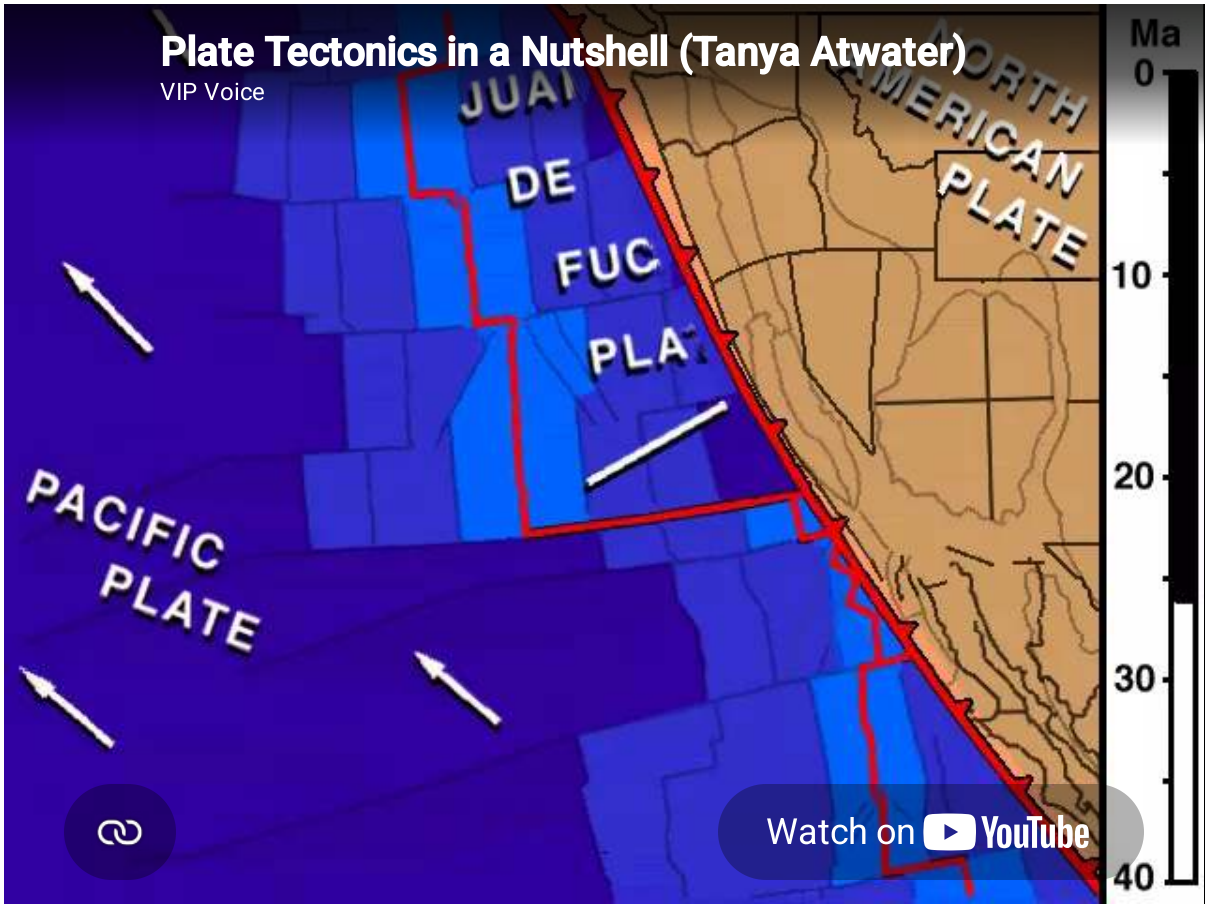
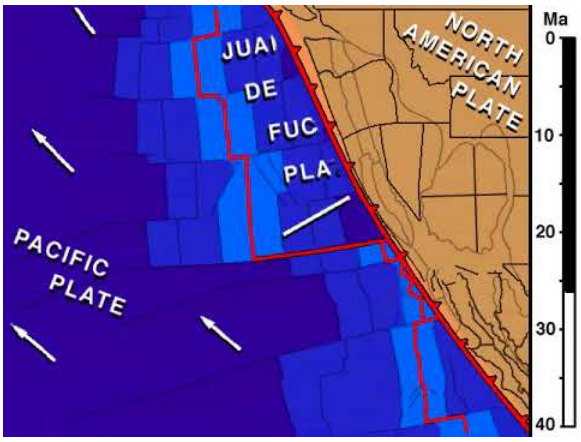


Figure 8.6.1: Paraceratherium, seen in this reconstruction, was a massive (15-20 ton, 15 foot tall) ancestor of rhinos.

Cenozoic Tectonics and Paleogeography

Animation of the last 38 million years of movement in western North America. Note that after the ridge is subducted, the convergent boundary turns to a transform boundary (with a divergent boundary inland).





In the Cenozoic, the plates of the Earth moved into more familiar places, with the biggest change being the closing of the Tethys Sea with collisions such as the Alps, Zagros, and Himalaya, a collision that started about 57 million years ago and continues today. Maybe the most significant tectonic feature that occurred in the Cenozoic of North America was the conversion of the west coast of California from a convergent boundary subduction zone to a transform boundary. Subduction off the coast of the western United States, which had occurred throughout the Mesozoic, had continued in the Cenozoic. After the Sevier Orogeny in the late Mesozoic, a subsequent orogeny called the **Laramide orogeny** occurred in the early Cenozoic. The Laramide was thick-skinned, different than the Sevier Orogeny. It involved deeper crustal rocks and produced bulges that would become mountain ranges like the Rocky Mountains, Black Hills, Wind River Range, Uinta Mountains, and the San Rafael Swell. Instead of descending directly into the mantle, the subducting plate shallowed out and moved eastward beneath the continental plate, affecting the overlying continent hundreds of miles east of the continental margin and building high mountains. This occurred because the subducting plate was so young and near the spreading center that the density of the plate was therefore low and subduction was hindered [156].

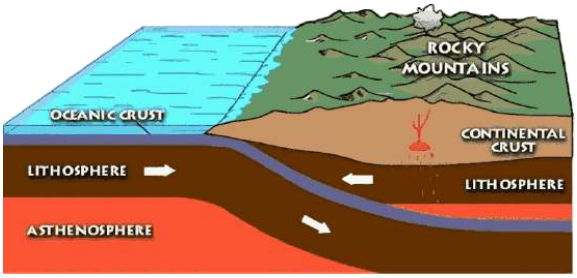


Figure 8.6.2: Shallow subduction during the Laramide Orogeny.

As the mid-ocean ridge itself started to subduct, the relative motion had changed. Subduction caused a relative convergence between the subducting Farallon Plate and the North American Plate. On the other side of the mid-ocean ridge from the Farallon Plate was the Pacific Plate, which was moving away from the North American Plate. Thus, as the subduction zone consumed the mid-ocean ridge, the relative movement became transform instead of convergent, which went on to become the **San Andreas Fault System**. As the San Andreas grew, it caused east-west directed extensional forces to spread over the western United States, creating the **Basin and Range province**. The transform fault switched position over the last 18 million years, twisting the mountains around Los Angeles, and new faults in the southeastern California deserts may become a future San Andreas-style fault [160]. During this switch from subduction to transform, the nearly horizontal Farallon slab began to sink into the mantle. This caused magmatism as the subducting slab sank, allowing asthenosphere material to rise around it. This event is called the **Oligocene ignimbrite flare-up**, which was one of the most significant periods of volcanism ever, including the largest single confirmed eruption, the 5000 cubic kilometer **Fish Canyon Tuff** [162].

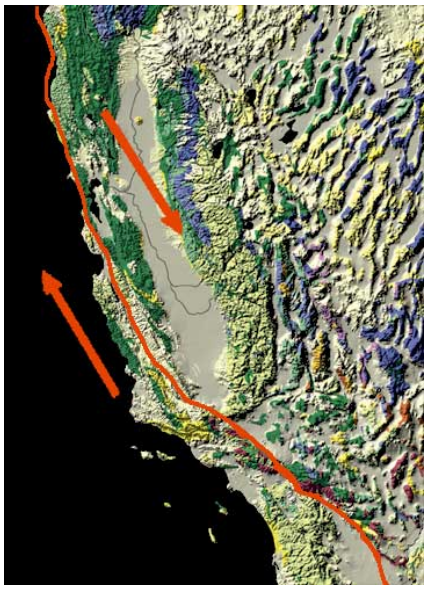


Figure 8.6.3: Map of the San Andreas fault, showing relative motion.

Cenozoic Evolution

There are five groups of early mammals in the fossil record, based primarily on fossil teeth, the hardest bone in vertebrate skeletons. For the purpose of this text, the most important group is the **Eupantotheres**, which diverges into the two main groups of mammals, the **marsupials** (*Sinodelphys*) and **placentals** or **eutherians** (*Eomaia*) in the Cretaceous and then diversified in the Cenozoic. The marsupials dominated on the isolated island continents of South America and Australia, and many went extinct in South America with the introduction of placental mammals. Some well-known mammal groups have been highly studied with interesting evolutionary stories in the Cenozoic. For example, horses started small with four toes, and ended up larger and having just one toe [163]. **Cetaceans** (marine mammals like whales and dolphins) started on land from small bear-like (mesonychids) creatures in the early Cenozoic and gradually took to water [164]. However, no study of evolution has been more studied than human evolution. **Hominids**, the name for human-like primates, started in eastern Africa several million years ago.

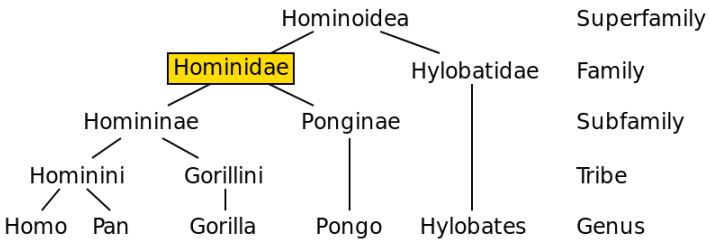


Figure 8.6.4: Family tree of Hominids (Hominidae).

The first critical event in this story is an environmental change from jungle to more of a savanna, probably caused by changes in Indian Ocean circulation. While **bipedalism** is known to have evolved before this shift, it is generally believed that our bipedal ancestors (like *Australopithecus*) had an advantage by covering ground more easily in a more open environment compared to their non-bipedal evolutionary cousins. There is also a growing body of evidence, including the famous “Lucy” fossil of an Australopithecine, that our early ancestors lived in trees. **Arboreal** animals usually demand a high intelligence to navigate through a three-dimensional world. It is from this lineage that humans evolved, using endurance running as a means to acquire more resources and possibly even hunt. This can explain many uniquely human features, from our long legs, strong achilles, lack of lower gut protection, and our wide range of running efficiencies.



Figure 8.6.5: Lucy skeleton from the Cleveland Natural History Museum, showing real fossil (brown) and reconstructed skeleton (white).

Now that the hands are freed up, the next big step is a large brain. There have been arguments from a switch to more meat-eating, cooking with fire [170], tool usage, and even the construct of society itself to explain this increase in brain size. Regardless of how, it was this increased cognitive power that allowed humans to reign as their ancestors moved out of Africa and explored the world, ultimately entering the Americas through land bridges like the Bering Land Bridge. The details of this worldwide migration and the different branches of the hominid evolutionary tree are very complex, and best reserved for its own course.

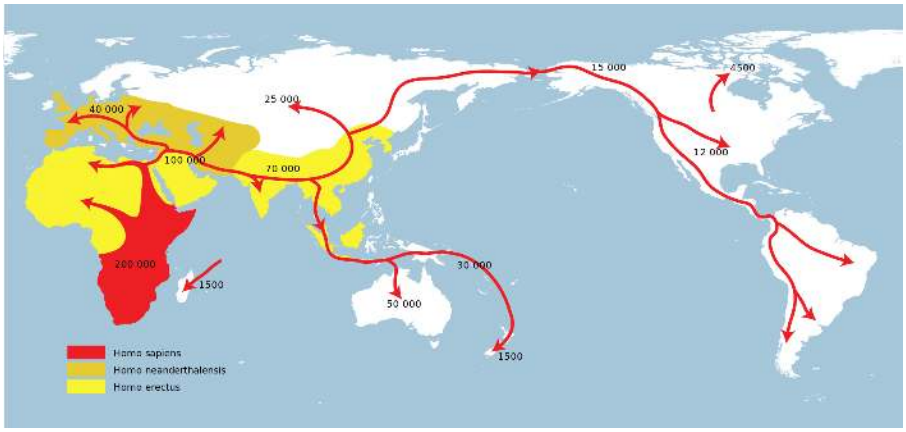


Figure 8.6.6: The hypothesized movement of the homo genus. Years are marked as to the best guess of the timing of movement.

Anthropocene and Extinction

Humans have had an influence on the Earth, its ecosystems and climate. Yet, human activity can not explain all of the changes that have occurred in the recent past. The start of the **Quaternary Period**, the last and current period of the Cenozoic, is marked by the start of our current ice age 2.58 million years ago. During this time period, ice sheets advanced and retreated, most likely due to Milankovitch cycles. Also at this time, various cold-adapted megafauna emerged (like giant sloths, saber-tooth cats, and woolly mammoths), and most of them went extinct as the Earth warmed from the most recent glacial maximum. A long-standing debate is over the cause of these and other extinctions. Is climate warming to blame, or were they caused by humans [175]? Certainly, we know of recent human extinctions of animals like the dodo or passenger pigeon. Can we connect modern extinctions to extinctions in the recent past? If so, there are several ideas as to how this happened. Possibly the most widely accepted and oldest is the **hunting/overkill hypothesis** [176]. The idea behind this hypothesis is that humans hunted large herbivores for food, then carnivores could not find food and both went extinct. Human arrival times in locations have been shown to be tied to increased extinction rates in many cases.

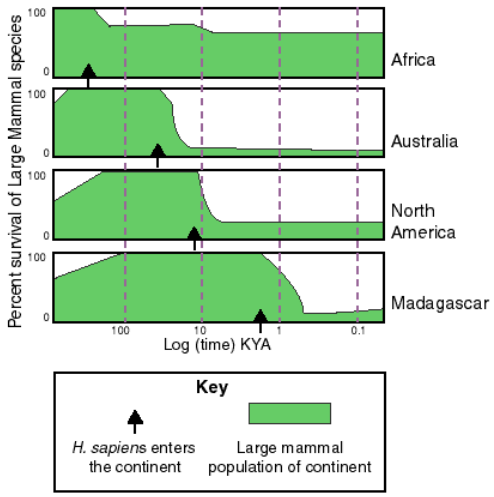


Figure 8.6.7: Graph showing the abundance of large mammals and the introduction of humans.

Modern human impact on the environment and the Earth as a whole is unquestioned. In fact, many scientists are starting to suggest that the rise of human civilization ended and/or replaced the **Holocene** epoch and defines a new geologic time interval: the **Anthropocene** [177]. Evidence for this change includes extinctions, increased tritium (hydrogen with two neutrons) due to nuclear testing, rising pollutants like carbon dioxide, more than 200 never-before-seen mineral species that have occurred only in this epoch, materials such as plastic and metals which will be long-lasting "fossils" in the geologic record, and large amounts of earthen material moved. The biggest scientific debate with this topic is the starting point. Some say that humans' invention of agriculture would be recognized in geologic strata and that should be the starting point, around 12,000 years ago. Others link the start of the Industrial Revolution and the subsequent addition of vast amounts of carbon dioxide in the atmosphere [180] is the starting point. Either way, the idea is that alien geologists visiting Earth in the distant future would easily recognize the impact of humans on the Earth as the beginning of a new geologic period.



Figure 8.6.8: Bingham Canyon Mine, Utah. This open-pit mine is the largest man-made removal of rock in the world.

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8.S: Summary

The changes that have occurred since the inception of Earth are vast and significant. From the oxygenation of the atmosphere, the progression of life forms, the assembly and deconstruction of several supercontinents, to the extinction of more life forms than exist today, having a general understanding of these changes can put present change into a more rounded perspective.

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CHAPTER OVERVIEW

9: Crustal Deformation and Earthquakes

Learning Objectives

- Differentiate between stress and strain.
- Identify the three major types of stress.
- Differentiate between brittle, ductile, and elastic deformation.
- Describe the geological map symbol used for strike and dip of strata.
- Name and describe different fold types.
- Differentiate the three major fault types and describe their associated movements.
- Explain how elastic rebound relates to earthquakes.
- Describe different seismic wave types and how they are measured.
- Explain how humans can induce seismicity.
- Describe how seismographs work to record earthquake waves.
- From seismograph records, locate the epicenter of an earthquake.
- Explain the difference between earthquake magnitude and intensity.
- List earthquake factors that determine ground shaking and destruction.
- Identify secondary earthquake hazards.
- Describe notable historical earthquakes.

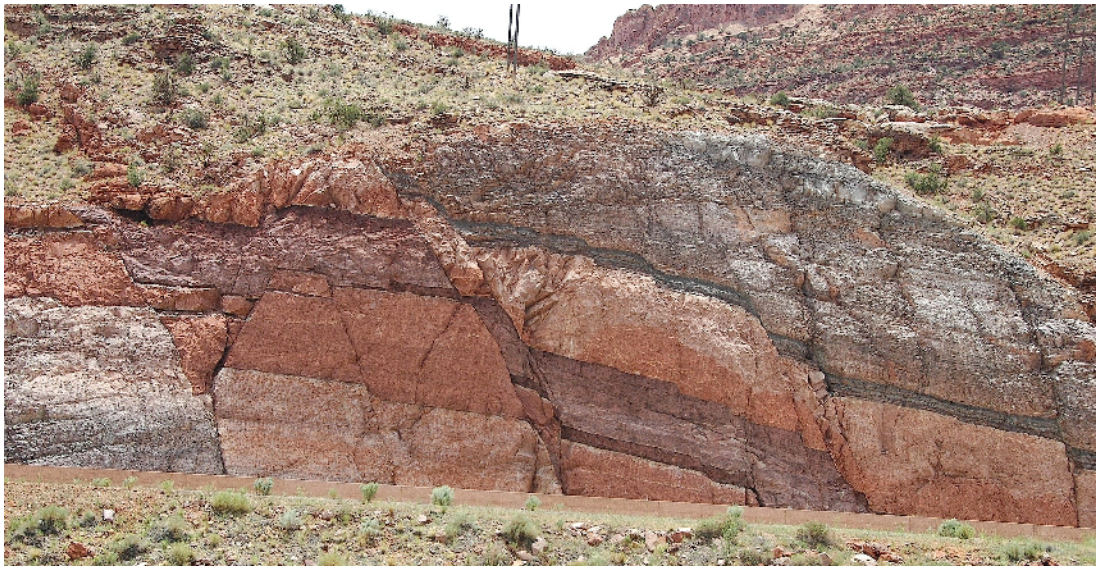


Figure 9.1: Example of normal faulting in an outcrop of the Pennsylvanian Honaker Trail Formation near Moab, Utah.

Crustal deformation occurs when applied forces exceed the internal strength of rocks, physically changing their shapes. These forces are called stress, and the physical changes they create are called strain. Forces involved in tectonic processes as well as gravity and igneous pluton emplacement produce strains in rocks that include folds, fractures, and faults. When rock experiences large amounts of shear stress and breaks with rapid, brittle deformation, energy is released in the form of seismic waves, commonly known as an earthquake.

9.1: Stress and Strain

9.2: Deformation

9.3: Geological Maps

9.4: Folds

9.5: Faults
9.6: Earthquake Essentials
9.7: Measuring Earthquakes
9.8: Earthquake Risk
9.9: Case Studies
9.S: Summary

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9.1: Stress and Strain

Stress is the force exerted per unit area and **strain** is the physical change that results in response to that force. When the applied stress is greater than the internal strength of rock, strain results in the form of deformation of the rock caused by the stress. Strain in rocks can be represented as a change in rock volume and/or rock shape, as well as fracturing the rock. There are three types of stress: tensional, compressional, and shear [1]. **Tensional stress** involves forces pulling in opposite directions, which results in strain that stretches and thins rock. **Compressional stress** involves forces pushing together, and the compressional strain shows up as rock folding and thickening. **Shear stress** involves transverse forces; the strain shows up as opposing blocks or regions of the material moving past each other.

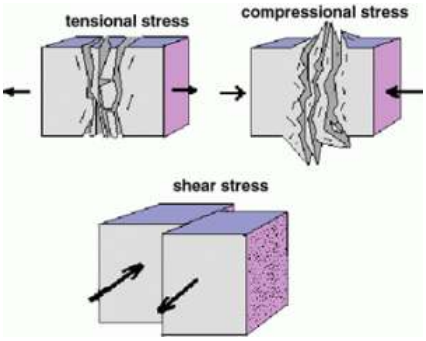


Figure 9.1.1: Types of stress. Clockwise from top left: tensional stress, compressional stress, and shear stress, and some examples of resulting strain.

Table 9.1.1: Types of stress and resulting strain.

Type of Stress	Associated Plate Boundary type	Resulting Strain	Associated fault and offset types
Tensional	divergent	Stretching and thinning	Normal
Compressional	convergent	Shortening and thickening	Reverse
Shear	transform	Tearing	Strike-slip

References

1. Earle S (2015) Physical geology OER textbook. BC Campus OpenEd

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9.2: Deformation

When rocks are stressed, the resulting strain can be elastic, ductile, or brittle. This change is generally called **deformation**. **Elastic deformation** is a strain that is reversible after the stress is released. For example, when you stretch a rubber band, it elastically returns to its original shape after you release it. **Ductile deformation** occurs when enough stress is applied to a material that the changes in its shape are permanent, and the material is no longer able to revert to its original shape. For example, if you bend a metal bar too far, it can be permanently bent out of shape. The point at which elastic deformation is surpassed and strain becomes permanent is called the **yield point**. In the figure, the yield point is where the line transitions from elastic deformation to ductile deformation (the end of the dashed line). **Brittle deformation** is another critical point of no return when rock integrity fails and the rock fractures under increasing stress.

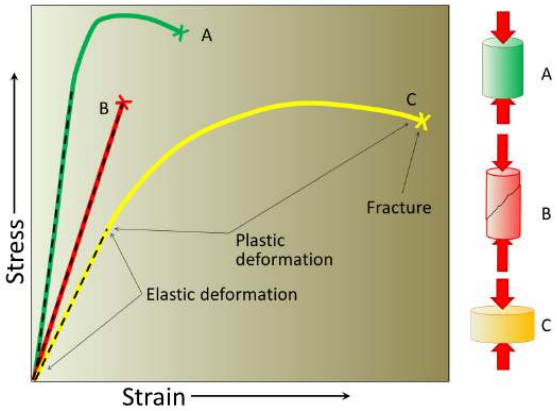


Figure 9.2.1: Different materials deform differently when stress is applied. Material A has relatively little deformation when undergoing large amounts of stress, before undergoing plastic deformation, and finally brittle failure. Material B only elastically deforms before brittle failure. Material C undergoes significant plastic deformation before finally brittle failure.

The type of deformation a rock undergoes depends on pore pressure, strain rate, rock strength, temperature, stress intensity, time, and confining pressure. **Pore pressure** is exerted on the rock by fluids in the open spaces or pores embedded within rock or sediment. **Strain rate** measures how quickly material is deformed. For example, applying stress slowly makes it is easier to bend a piece of wood without breaking it. **Rock strength** measures how easily a rock deforms under stress. Shale has low strength and granite has high strength. Removing heat, or decreasing the temperature, makes materials more rigid and susceptible to brittle deformation. On the other hand, heating materials make them more ductile and less brittle. Heated glass can be bent and stretched.

Table 9.2.1: Relationship between factors operating on rock and the resulting strains

Factor	Strain Response
Increase Temperature	More Ductile
Increase Strain Rate	More Brittle
Increase Rock Strength	More Brittle

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9.3: Geological Maps

Geologic maps are two-dimensional (2D) representations of geologic formations and structures at the Earth's surface, including faults, folds, inclined strata, and rock types. Geologists use geologic maps to represent where geologic formations, faults, folds, and inclined rock units are. **Geologic formations** are recognizable, mappable rock units. Each formation on the map is indicated by a color and a label. For examples of geologic maps, see the California Geological Survey (CGS) geologic map viewer.

Formation labels include symbols that follow a specific protocol. The first one or more letters are uppercase and represent the geologic time period of the formation. More than one uppercase letter indicates the formation is associated with multiple time periods. The following lowercase letters represent the formation name, abbreviated rock description, or both.

Cross Sections

Cross sections are subsurface interpretations made from surface and subsurface measurements. Maps display geology in the horizontal plane while cross sections show subsurface geology in the vertical plane. For more information on cross sections, check out the AAPG wiki.

Strike and Dip

Geologists use a special symbol called strike and dip to represent inclined beds. Strike and dip map symbols look like the capital letter T, with a short trunk and extra-wide top line. The short trunk represents the dip and the top line represents the strike. **Dip** is the angle that a bed plunges into the Earth from the horizontal. A number next to the symbol represents dip angle. One way to visualize the strike is to think about a line made by standing water on the inclined layer. That line is horizontal and lies on a compass direction that has some angle with respect to true north or south (see figure). The **strike** angle is that angle measured by a compass. For example, N30°E (read north 30 degrees east) means the horizontal line points northeast at an angle of 30° from true north. The strike and dip symbol is drawn on the map at the strike angle with respect to true north on the map. The dip of the inclined layer represents the angle down to the layer from horizontal, in the figure 45° SE (read dipping 45 degrees to the SE). The direction of dip would be the direction a ball would roll if set on the layer and released. A horizontal rock bed has a dip of 0° and a vertical bed has a dip of 90°. Strike and dip considered together are called **rock attitude**.

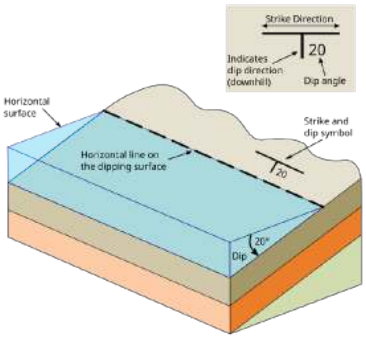


Figure 9.3.1: "Strike" and "dip" are words used to describe the orientation of rock layers with respect to North/South and Horizontal. (By Karla Panchuk; CC BY 4.0 via Physical Geology.)

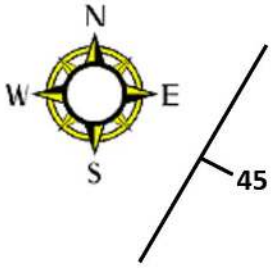


Figure 9.3.2: Attitude symbol on a geologic map (with compass directions for reference) showing strike of N30ºE and dip of 45° to the SE.

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9.4: Folds

Geologic **folds** are layers of rock that are curved or bent by ductile deformation. Folds are most commonly formed by compressional forces at depth, where hotter temperatures and higher confining pressures allow ductile deformation to occur.

Folds are described by the orientation of their axes, axial planes, and limbs. The plane that splits the fold into two halves is known as the **axial plane**. The **fold axis** is the line along which the bending occurs and is where the axial plane intersects the folded strata. The **hinge line** follows the line of greatest bend in a fold. The two sides of the fold are the fold **limbs**.

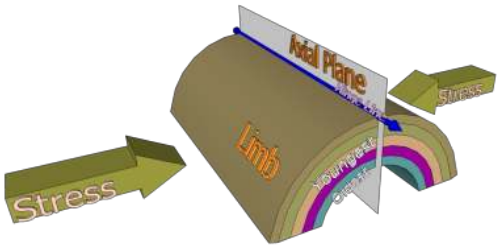


Figure 9.4.1: Model of anticline. Oldest beds are in the center and youngest on the outside. The axial plane intersects the center angle of bend. The hinge line follows the line of greatest bend, where the axial plane intersects the outside of the fold.

Symmetrical folds have a vertical axial plane and limbs have equal but opposite dips. **Asymmetrical folds** have dipping, non-vertical axial planes, where the limbs dip at different angles. **Overturned folds** have steeply dipping axial planes and the limbs dip in the same direction but usually at different dip angles. **Recumbent folds** have horizontal or nearly horizontal axial planes. When the axis of the fold plunges into the ground, the fold is called a **plunging fold**. Folds are classified into five categories: anticline, syncline, monocline, dome, and basin.

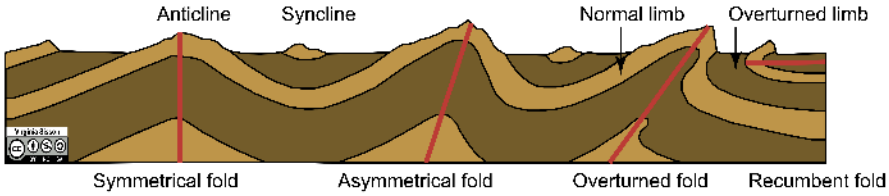


Figure 9.4.2: Types of folds. (By Virginia Sisson; CC BY-NC-SA 4.0 via Exercises for Physical Geology.)

Anticline

Anticlines are arch-like, or A-shaped folds that are convex-upward in shape. They have downward curving limbs and beds that dip down and away from the central fold axis. In anticlines, the oldest rock strata are in the center of the fold, along the axis, and the younger beds are on the outside. Since geologic maps show the intersection of surface topography with underlying geologic structures, an anticline on a geologic map can be identified by both the attitude of the strata forming the fold and the older age of the rocks inside the fold. An **antiform** has the same shape as an anticline, but the relative ages of the beds in the fold cannot be determined. Oil geologists are interested in anticlines because they can form oil traps, where oil migrates up along the limbs of the fold and accumulates in the high point along the fold axis.



Figure 9.4.3: Anticline near Bcharre, Lebanon. (By Not home at English Wikipedia; public domain via Wikimedia Commons.)

Syncline

Synclines are trough-like, or U shaped, folds that are concave-upward in shape. They have beds that dip down and in toward the central fold axis. In synclines, older rock is on the outside of the fold and the youngest rock is inside of the fold axis. A **synform** has the shape of a syncline but like an antiform, does not have distinguishable age zones.



Synclinal fold – Macigno Formation by alanpitts on Sketchfab

Monocline

Monoclines are step-like folds, in which flat rocks are upwarped or downwarped, then continue flat. Monoclines are relatively common on the Colorado Plateau where they form "**reefs**," which are ridges that act as topographic barriers and should not be confused with ocean reefs. Capitol Reef is an example of a monocline in Utah. Monoclines can be caused by bending of shallower sedimentary strata as faults grow below them. These faults are commonly called "blind faults" because they end before reaching the surface and can be either normal or reverse faults.



Figure 9.4.4: Monocline at Colorado National Monument. (By Anky-man at the English Wikipedia; CC BY-SA 3.0 via Wikimedia Commons.)

Dome

A **dome** is a symmetrical to semi-symmetrical upwarping of rock beds. Domes have a shape like an inverted bowl, similar to an architectural dome on a building. Examples of domes in Utah include the San Rafael Swell, Harrisburg Junction Dome, and Henry Mountains [2; 3]. Domes are formed from compressional forces, underlying igneous intrusions [2], by salt diapirs, or even impacts, like upheaval dome in Canyonlands National Park.



Figure 9.4.5: This prominent circular feature with a diameter of almost 50 km (30 mi) in the Sahara desert of Mauritania has attracted attention since the earliest space missions because it forms a conspicuous bulls-eye in the otherwise rather featureless expanse of the desert. Initially interpreted as a meteorite impact structure because of its high degree of circularity, it is now thought to be merely a dome that has been laid bare by erosion. (By NASA; public domain via Wikimedia Commons.)

Basin

A **basin** is the inverse of a dome, a bowl-shaped depression in a rock bed. The Los Angeles Basin in California is an example of a basin. Some structural basins are also sedimentary basins that collect large quantities of sediment over time. Sedimentary basins can form as a result of folding but are much more commonly produced in mountain building, forming between mountain blocks or via faulting. Regardless of the cause, as the basin sinks or subsides, it can accumulate more sediment because the weight of the sediment causes more subsidence in a positive-feedback loop. There are active sedimentary basins all over the world [4]. An example of a rapidly subsiding basin in Utah is the Oquirrh Basin, dated to the Pennsylvanian-Permian age, which has accumulated over 9,144 m (30,000 ft) of fossiliferous sandstones, shales, and limestones. These strata can be seen in the Wasatch Mountains along the east side of Utah Valley, especially on Mt. Timpanogos and in Provo Canyon.



Figure 9.4.6: The Denver Basin is an active sedimentary basin at the eastern extent of the Rocky Mountains. As sediment accumulates, the basin subsides, creating basin-shaped beds that are all dipping towards the center of the basin.

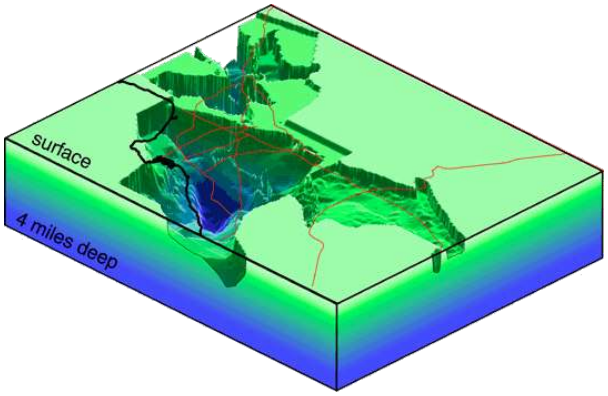


Figure 9.4.7: The depth of sedimentary basins in the Los Angeles region. Major freeways are drawn in red. (By Ned Field; public domain via SCEC.)

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2. Gilbert GK, Dutton CE (1877) Report on the geology of the Henry Mountains. U.S. Government Printing Office, Washington

3. Ressetar R (2013) The San Rafael Swell and Henry Mountains Basin: geologic centerpiece of Utah. Utah Geological Association, Utah Geological Association

4. Coleman JL, Cahan SM (2012) Preliminary catalog of the sedimentary basins of the United States. U.S. Geological Survey

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9.5: Faults

Faults are the places in the crust where brittle deformation occurs as two blocks of rocks move relative to one another. Normal and reverse faults display vertical, also known as dip-slip, motion. **Dip-slip motion** consists of relative up-and-down movement along a dipping fault between two blocks, the hanging wall, and footwall. In a dip-slip system, the **footwall** is below the fault plane and the **hanging wall** is above the fault plane. A good way to remember this is to imagine a mine tunnel running along a fault; the hanging wall would be where a miner would hang a lantern and the footwall would be at the miner's feet.

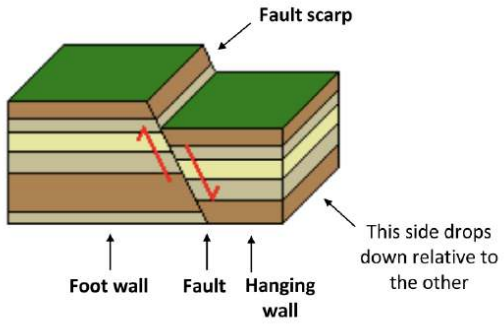


Figure 9.5.1: Common terms used for normal faults. Normal faults form when the hanging wall moves down relative to the footwall. (By Kindred Grey; CC BY-SA 3.0 via Introduction to Earth Science.)

Faulting as a term refers to the rupture of rocks. Such ruptures occur at plate boundaries but can also occur in plate interiors as well. Faults slip along the **fault plane**. The **fault scarp** is the offset of the surface produced where the fault breaks through the surface. **Slickensides** are polished, often grooved surfaces along the fault plane created by friction during the movement.

A **joint** or **fracture** is a plane of brittle deformation in the rock created by the movement that is not offset or sheared. Joints can result from many processes, such as cooling, depressurizing, or folding. Joint systems may be regional affecting many square miles.

Normal Faults

Normal faults move by a vertical motion where the hanging wall moves downward relative to the footwall along the dip of the fault. Normal faults are created by tensional forces in the crust. Normal faults and tensional forces commonly occur at divergent plate boundaries, where the crust is being stretched by tensional stresses. Examples of normal faults in Utah are the Wasatch Fault, the Hurricane Fault, and other faults bounding the valleys in the Basin and Range province.



Figure 9.5.2: Example of a normal fault in an outcrop of the Pennsylvanian Honaker Trail Formation near Moab, Utah.

Grabens, horsts, and half-grabens are blocks of crust or rock bounded by normal faults. **Grabens** drop down relative to adjacent blocks and create valleys. **Horsts** rise up relative to adjacent down-dropped blocks and become areas of higher topography. Where occurring together, horsts and grabens create a symmetrical pattern of valleys surrounded by normal faults on both sides and mountains. **Half-grabens** are a one-sided version of a horst and graben, where blocks are tilted by a normal fault on one side,

creating an asymmetrical valley-mountain arrangement. The mountain-valleys of the Basin and Range Province of Western Utah and Nevada consist of a series of full and half-grabens from the Salt Lake Valley to the Sierra Nevada Mountains.

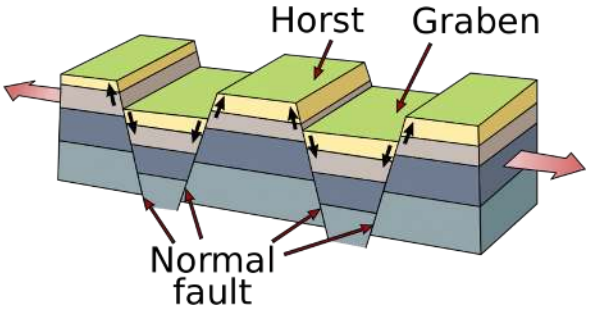


Figure 9.5.3: Faulting that occurs in the crust under tensional stress forming horsts and grabens.

Normal faults do not continue to clear into the mantle. In the Basin and Range Province, the dip of a normal fault tends to decrease with depth, i.e., the fault angle becomes shallower and more horizontal as it goes deeper. Such decreasing dips happen when large amounts of extension occur along very low-angle normal faults, known as **detachment faults**. The normal faults of the Basin and Range, produced by tension in the crust, appear to become detachment faults at greater depths.

Reverse Faults

In **reverse faults**, compressional forces cause the hanging wall to move up relative to the footwall. A **thrust fault** is a reverse fault where the fault plane has a low dip angle of less than 45°. Thrust faults carry older rocks on top of younger rocks and can even cause the repetition of rock units in the stratigraphic record.

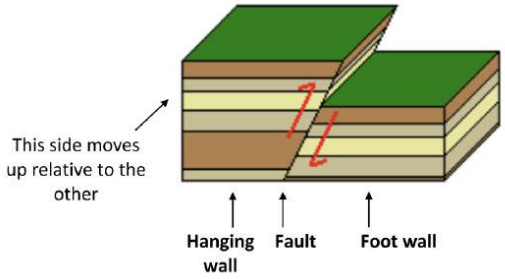


Figure 9.5.4: Simplified block diagram of a reverse fault. (By Kindred Grey; CC BY-SA 3.0 via Introduction to Earth Science.)

Convergent plate boundaries with subduction zones create a special type of “reverse” fault called a **megathrust fault** where denser oceanic crust drives down beneath less dense overlying crust. Megathrust faults cause the largest magnitude earthquakes yet measured and commonly cause massive destruction and tsunamis.

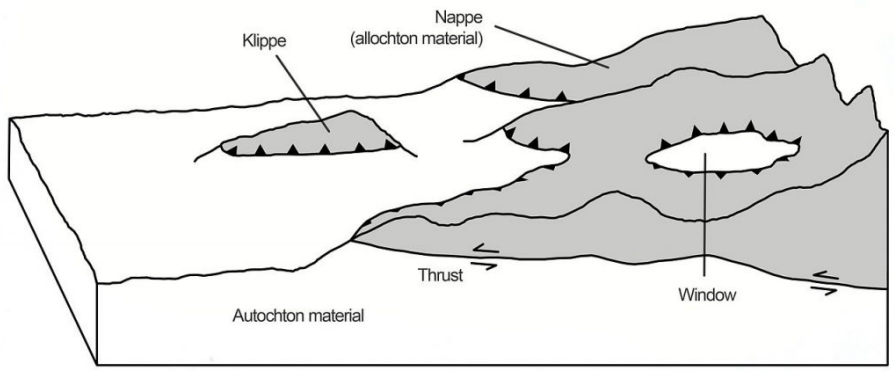


Figure 9.5.5: Terminology of thrust faults (low-angle reverse faults). A klippe is the remnant of the hangingwall (aka nappe), where the surrounding material has been eroded away. A window is where part of the hanging wall has been eroded away to expose the footwall (autochthon). Note the symbol shows flags on the overlying thrust plate.



Figure 9.5.6: Ketobe Knob in the San Rafael Swell of Utah displays an example of a thrust fault. (CC BY-NC-SA 4.0 International; Ron Schott via Flickr)

Strike-Slip Faults

Strike-slip faults have side-to-side motion. Strike-slip faults are most commonly associated with transform plate boundaries and are prevalent in transform fracture zones along mid-ocean ridges. In pure strike-slip motion, fault blocks on either side of the fault do not move up or down relative to each other, rather move laterally, side to side. The direction of the strike-slip movement is determined by an observer standing on a block on one side of the fault. If the block on the opposing side of the fault moves left relative to the observer's block, this is called **sinistral** (or left-lateral) motion. If the opposing block moves right, it is **dextral** (or right-lateral) motion. An example of a dextral, right-lateral strike-slip fault is the San Andreas Fault, which denotes a transform boundary between the North American and Pacific plates. An example of a sinistral, left-lateral strike-slip fault is the Dead Sea fault in Jordan and Israel.

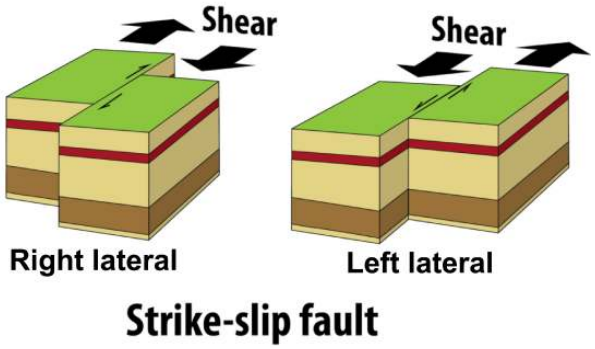


Figure 9.5.7: Right-lateral and left-lateral strike-slip faults resulting from shear stress. (By Karla Panchuk; CC BY-SA 4.0 via Physical Geology)

Bends along strike-slip faults create areas of compression or tension between the sliding blocks. Tensional stresses create **transtensional** features with normal faults and basins, such as the Salton Sea in California. Compressional stresses create **transpressional** features with reverse faults and cause small-scale mountain building, such as the San Gabriel Mountains in California. The faults that splay off transpression or transtension features are known as **flower structures**.

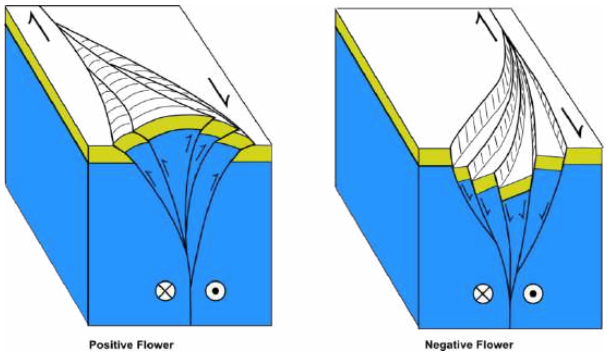
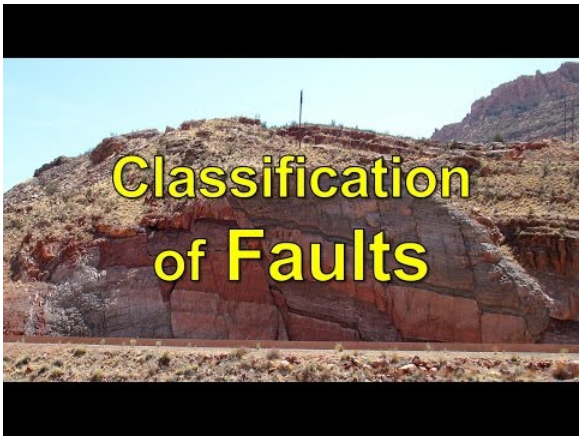


Figure 9.5.8: Flower structures created by strike-slip faults. Depending on the relative movement in relation to the bend in the fault, flower structures can create basins or mountains.

Video showing how faults are classified.





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9.6: Earthquake Essentials

Earthquakes are felt at the surface of the Earth when energy is released by blocks of rock sliding past each other, i.e. faulting has occurred. **Seismic energy** thus released travels through the Earth in the form of seismic waves. Most earthquakes occur along active plate boundaries. Intraplate earthquakes (not along plate boundaries) occur and are still poorly understood. The USGS Earthquakes Hazards Program has a real-time map showing the most recent earthquakes.

How Earthquakes Happen

The release of seismic energy is explained by the **elastic rebound theory**. When the rock is strained to the point that it undergoes brittle deformation, the place where the initial offsetting rupture occurs between the fault blocks is called the **focus**. This offset propagates along the fault, which is known as the fault plane.

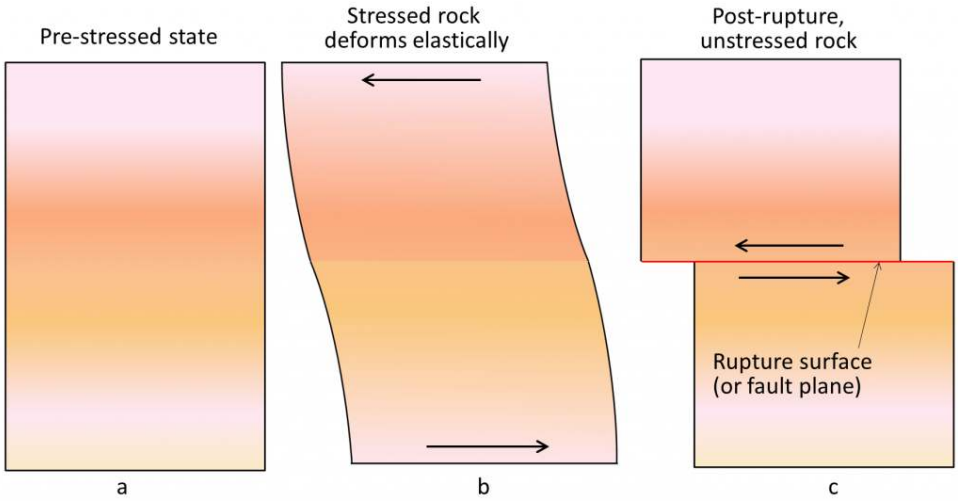


Figure 9.6.1: Process of elastic rebound: a) Undeformed state, b) accumulation of elastic strain, and c) brittle failure and release of elastic strain.

The fault blocks of persistent faults, like the Wasatch Fault (Utah), that show recurring movements are locked together by friction. Over hundreds to thousands of years, stress builds up along the fault until it overcomes frictional resistance, rupturing the rock and initiating fault movement. The deformed, unbroken rocks snap back toward their original shape in a process called elastic rebound**.** Think of bending a stick until it breaks; stored energy is released, and the broken pieces return to near their original orientation.

Bending, the ductile deformation of the rocks near a fault reflects a build-up of stress. In earthquake-prone areas like California, strain gauges are used to measure this bending and help **seismologists**, scientists who study earthquakes, understand more about forecasting them. In locations where the fault is not locked, seismic stress causes continuous, gradual displacement between the fault blocks called **fault creep**. Fault creep occurs along some parts of the San Andreas Fault (California).

After an initial earthquake, continuous application of stress in the crust causes elastic energy to begin to build again during a period of inactivity along the fault. The accumulating elastic strain may be periodically released to produce small earthquakes on or near the main fault called **foreshocks**. Foreshocks can occur hours or days before a large earthquake, or may not occur at all. The main release of energy during the major earthquake is known as the **mainshock**. **Aftershocks** may follow the mainshock to adjust new strain produced during the fault movement and they generally decrease over time.

Focus and Epicenter

The earthquake focus, also called the **hypocenter**, is the initial point of rupture, and displacement of the rock moves from the hypocenter along the fault surface. The earthquake focus or hypocenter is the point along the fault plane from which initial seismic waves spread outward and is always at some depth below the ground surface. From the focus, rock displacement propagates up, down, and laterally along the fault plane. This displacement produces shock waves called seismic waves. The larger the displacement between the opposing fault blocks and the further the displacement propagates along the fault surface, the more seismic energy is released and the greater the amount and time of shaking is produced. The **epicenter** is the location on the Earth's

surface vertically above the focus. This is the location that most news reports give because it is the center of the area where people are affected.

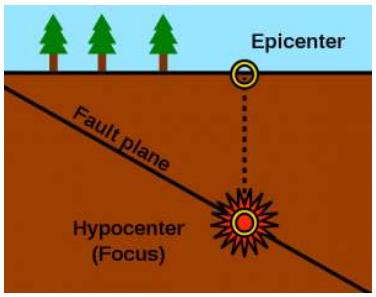


Figure 9.6.2: The hypocenter is the point along the fault plane in the subsurface from which seismic energy emanates. The epicenter is the point on the land surface vertically above the hypocenter.

Seismic Waves

To understand earthquakes and how earthquake energy moves through the Earth, consider the basic properties of waves. **Waves** describe how energy moves through a medium, such as rock or unconsolidated sediments in the case of earthquakes. **Wave amplitude** indicates the magnitude or height of earthquake motion. **Wavelength** is the distance between two successive peaks of a wave. **Wave frequency** is the number of repetitions of the motion over a period of time, cycles per time unit. **Period**, which is the amount of time for a wave to travel one wavelength, is the inverse of frequency. When multiple waves combine, they can interfere with each other (see figure). When waves combine in sync, they produce **constructive interference**, where the influence of one wave adds to and magnifies the other. If waves are out of sync, they produce **destructive interference**, which diminishes the amplitudes of both waves. If two combined waves have the same amplitude and frequency but are one-half wavelength out of sync, the resulting destructive interference can eliminate each wave. These processes of wave amplitude, frequency, period, and constructive and destructive interference determine the magnitude and intensity of earthquakes.

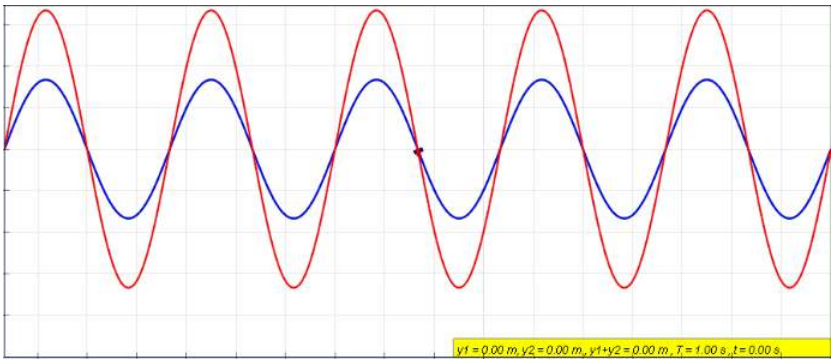


Figure 9.6.3: Example of constructive and destructive interference; note red line representing the results of interference.

Seismic waves are the physical expression of energy released by the elastic rebound of rock within displaced fault blocks and are felt as an earthquake. Seismic waves occur as body waves and surface waves. **Body waves** pass underground through the Earth's interior body and are the first seismic waves to propagate out from the focus. Body waves include primary (P) waves and secondary (S) waves. **P waves** are the fastest body waves and move through the rock via compression, very much like sound waves move through air. Rock particles move forward and back during the passage of the P waves, enabling them to travel through solids, liquids, plasma, and gases. **S waves** travel slower, following P waves, and propagate as shear waves that move rock particles from side to side. Because they are restricted to lateral movement, S waves can only travel through solids but not liquids, plasma, or gases.

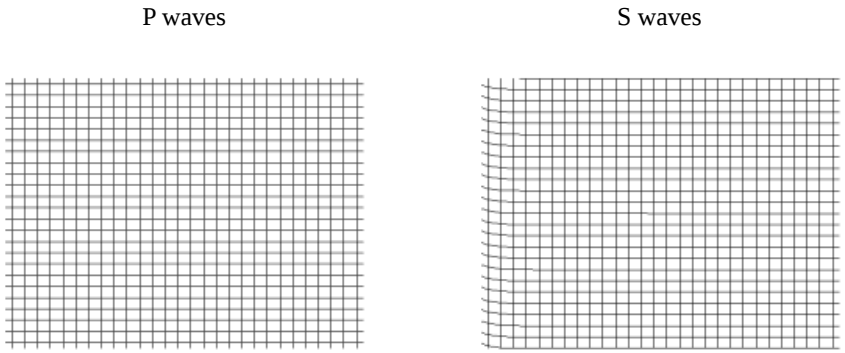


Figure 9.6.4: Left: P waves are compressional. Right: S waves are shear.

During an earthquake, body waves pass through the Earth and into the mantle as a sub-spherical wavefront. Considering a point on a wavefront, the path followed by a specific point on the spreading wavefront is called a **seismic ray**. A seismic ray reaches a specific **seismograph** located at one of thousands of seismic monitoring stations scattered over the Earth. Density increases with depth in the Earth, and since **seismic velocity** increases with density, a process called **refraction** causes earthquake rays to curve away from the vertical and bend back toward the surface, passing through different bodies of rock along the way.

Surface waves are produced when body waves from the focus strike the Earth's surface. Surface waves travel along the Earth's surface, radiating outward from the epicenter. Surface waves take the form of rolling waves called Raleigh Waves and side to side waves called Love Waves (watch videos for wave propagation animations). Surface waves are produced primarily as the more energetic S waves strike the surface from below with some surface wave energy contributed by P waves (videos courtesy blog.Wolfram.com). Surface waves travel more slowly than body waves and because of their complex horizontal and vertical movement, surface waves are responsible for most of the damage caused by an earthquake. Love waves produce predominantly horizontal ground shaking and, ironically from their name, are the most destructive. Rayleigh waves produce an elliptical motion with longitudinal dilation and compression, like ocean waves. However, Raleigh waves cause rock particles to move in a direction opposite to that of water particles in ocean waves.

The Earth has been described as ringing like a bell after an earthquake with earthquake energy reverberating inside it. Like other waves, seismic waves refract (bend) and bounce (reflect) when passing through rocks of differing densities. S waves, which cannot move through liquids, are blocked by the Earth's liquid outer core, creating an **S wave shadow zone** on the side of the planet opposite to the earthquake focus. P waves, on the other hand, pass through the core but are refracted into the core by the difference of density at the core-mantle boundary. This has the effect of creating a cone-shaped **P wave shadow zone** on parts of the other side of the Earth from the focus.

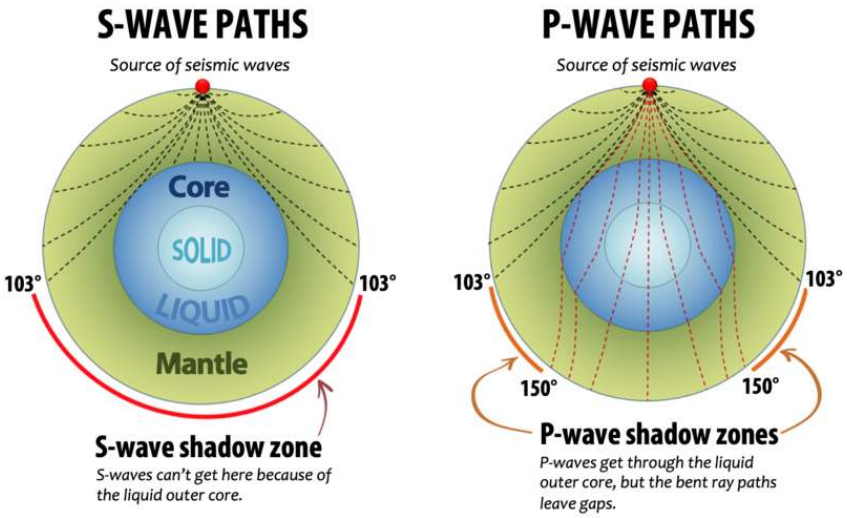
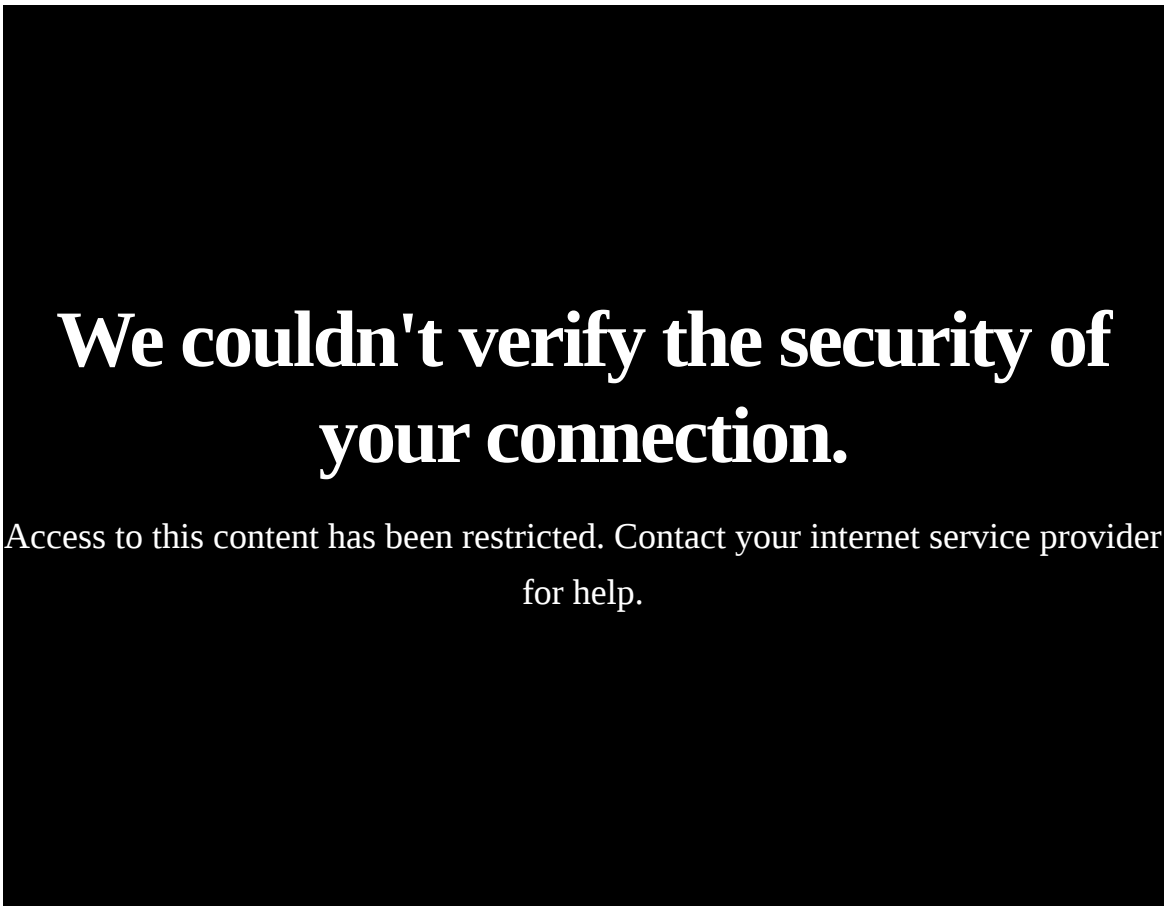


Figure 9.6.5: Patterns of seismic waves travel through Earth's mantle and core. S waves can't travel through the liquid outer core, so they leave a shadow zone opposite the focus. P waves do travel through the core but refraction beds them and creates a P wave shadow zone. (By Karla Panchuk, Steven Earle; CC BY 4.0 via Physical Geology.)

2011 Tohoku Earthquake, Mag. 9.0. Body and Surface Waves from seismicsoundlab on Vimeo.



Induced Seismicity

Earthquakes known as **induced seismicity** occur near natural gas extraction sites because of human activity. Injection of waste fluids in the ground, commonly a byproduct of an extraction process for natural gas known as **fracking**, can increase the outward pressure that liquid in the pores of a rock exerts, known as **pore pressure** [5; 6]. The increase in pore pressure decreases the frictional forces that keep rocks from sliding past each other, essentially lubricating fault planes. This effect is causing earthquakes

to occur near injection sites, which is induced seismicity due to human activity [5]. The significant increase in drilling activity in the central United States has spurred the requirement for the disposal of significant amounts of waste drilling fluid, resulting in a measurable change in the cumulative number of earthquakes experienced in the region.

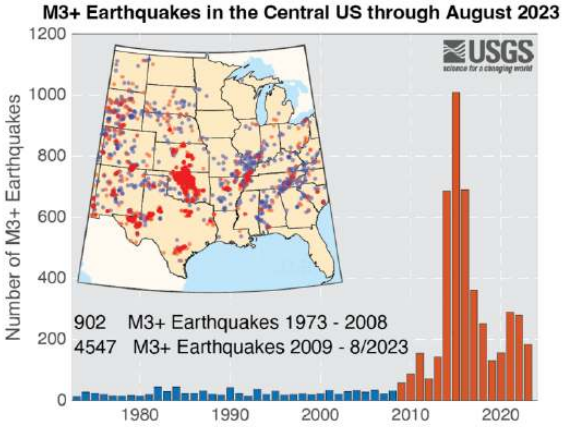


Figure 9.6.6: Frequency of earthquakes in the central United States. Note the sharp increase in the number of earthquakes from 2009 to 2018.

References

5. Ellsworth WL (2013) Injection-Induced Earthquakes. Science 341:1225942. https://doi.org/10.1126/science.1225942

6. Rubinstein JL, Mahani AB (2015) Myths and facts on wastewater injection, hydraulic fracturing, enhanced oil recovery, and induced seismicity. Seismological research letters 86:1060–1067

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9.7: Measuring Earthquakes

Seismographs

People feel approximately 1 million earthquakes a year, usually when they are close to the source and the earthquake registers at least moment magnitude 2.5. Major earthquakes of moment magnitude 7.0 and higher are extremely rare. The U. S. Geological Survey (USGS) Earthquakes Hazards Program real-time map shows the location and magnitude of recent earthquakes around the world.

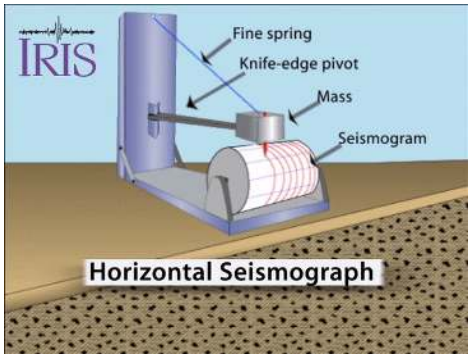


Figure 9.7.1: Animation of a horizontal seismograph.

To accurately study seismic waves, geologists use seismographs that can measure even the slightest ground vibrations. Early 20th-century seismographs used a weighted pen (pendulum) suspended by a long spring above a recording device fixed solidly to the ground. The recording device is a rotating drum mounted with a continuous strip of paper. During an earthquake, the suspended pen stays motionless and records ground movement on the paper strip. The resulting graph is a **seismogram**. Digital versions use magnets, wire coils, electrical sensors, and digital signals instead of mechanical pens, springs, drums, and paper. A **seismograph array** is multiple seismographs configured to measure vibrations in three directions: north-south (x-axis), east-west (y-axis), and up-down (z-axis).

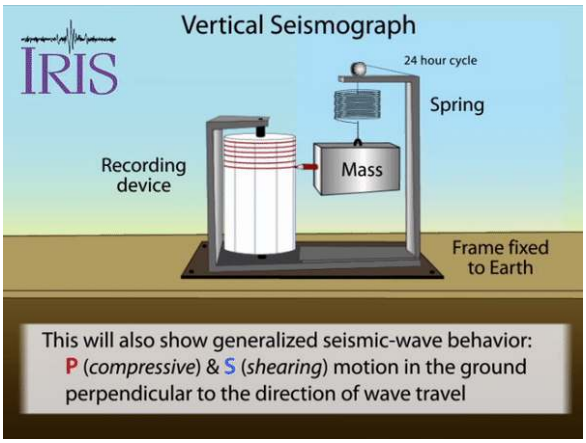


Figure 9.7.2: Animation of how a basic vertical seismograph records seismic waves.

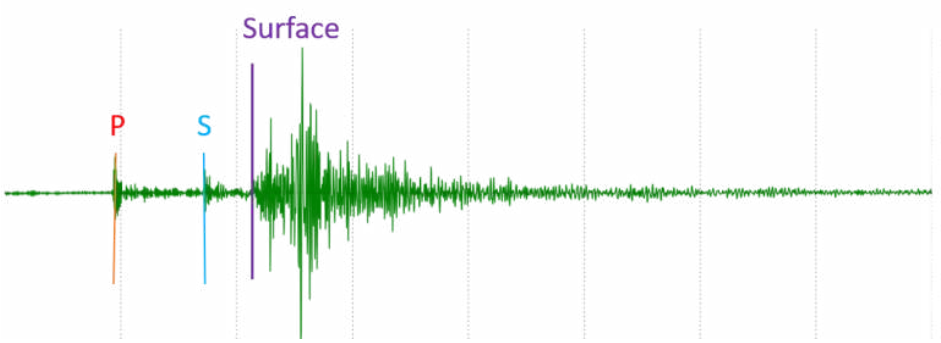


Figure 9.7.3: A seismogram showing the arrivals of the P, S, and surface waves

To pinpoint the location of an earthquake epicenter, seismologists use the differences in arrival times of the P, S, and surface waves. After an earthquake, P waves will appear first on a seismogram, followed by S waves, and finally surface waves, which have the largest amplitude. It is important to note that surface waves lose energy quickly, so they are not measurable at great distances from the epicenter. These time differences determine the distance to but not the direction of the epicenter. By using wave arrival times recorded on seismographs at multiple stations, seismologists can apply **triangulation** to pinpoint the location of the epicenter of an earthquake. At least three seismograph stations are needed for triangulation. The distance from each station to the epicenter is plotted as the radius of a circle. The epicenter is located where all of the circles intersect. This method also works in 3D, using multi-axis seismographs and sphere radii to calculate the underground depth of the focus.

This video shows the method of triangulation to locate the epicenter of an earthquake.





Seismograph Network

The International Registry of Seismograph Stations lists more than 20,000 seismographs on the planet. By comparing data from multiple seismographs, scientists can map the properties of the inside of the Earth, detect detonations of large explosive devices, and predict tsunamis. The **Global Seismographic Network**, a worldwide set of linked seismographs that electronically distribute real-time data, includes more than 150 stations that meet specific design and precision standards. The USArray is a network of hundreds of permanent and transportable seismographs in the United States that are used to map the subsurface activity of earthquakes (see video).

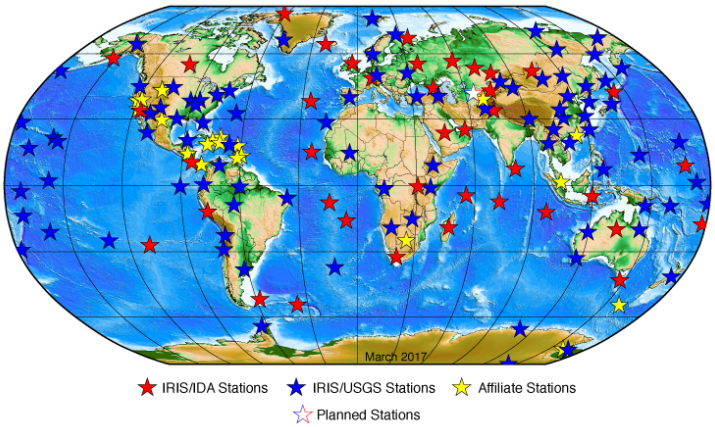
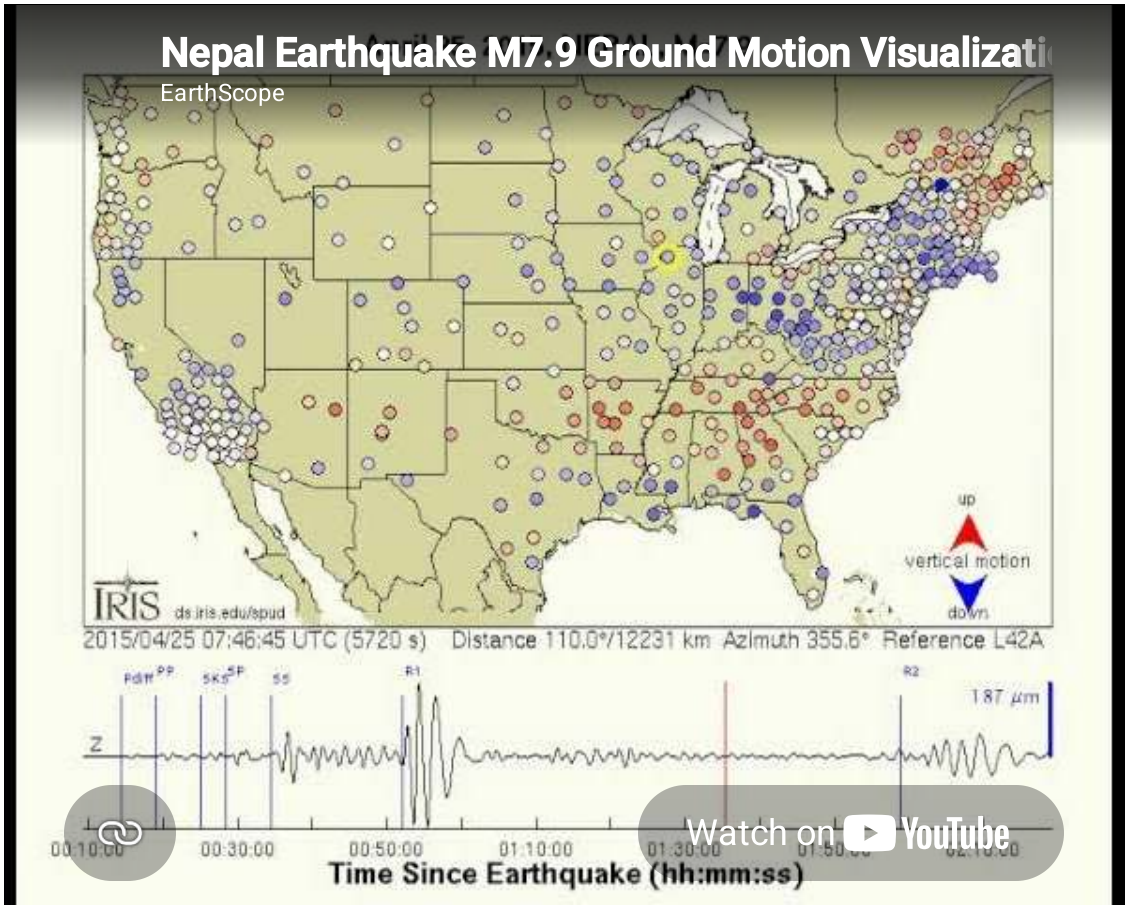
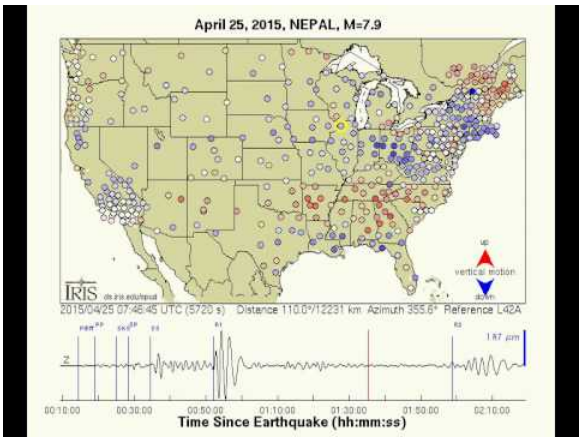


Figure 9.7.4: Global network of seismic stations. Note that this map does not show all of the world's seismic stations, just some of the networks of stations scientists use to measure seismic activity.

Along with monitoring for earthquakes and related hazards, the Global Seismographic Network helps detect nuclear weapons testing, which is monitored by the Comprehensive Nuclear Test Ban Treaty Organization. Most recently, seismographs have been used to determine nuclear weapons testing by North Korea.

Nepal Earthquake M7.9 Ground Motion Visualization





Seismic Tomography

Very much like a CT (Computed Tomography) scan uses X-rays at different angles to image the inside of a body, **seismic tomography** uses seismic rays from thousands of earthquakes that occur each year, passing at all angles through masses of rock, to generate images of internal Earth structures.

Using the assumption that the Earth consists of homogenous layers, geologists developed a model of expected properties of Earth materials at every depth within the Earth called the PREM (**Preliminary Reference Earth Model**). These properties include seismic wave **transmission velocity**, which is dependent on rock density and elasticity. In the mantle, temperature differences affect rock density. Cooler rocks have a higher density and therefore transmit seismic waves faster. Warmer rocks have a lower density and transmit earthquake waves slower. When the arrival times of seismic rays at individual seismic stations are compared to arrival times predicted by PREM, differences are called **seismic anomalies** and can be measured for bodies of rock within the

Earth from seismic rays passing through them at stations of the seismic network. Because seismic rays travel at all angles from lots of earthquakes and arrive at lots of stations of the seismic network, like CT scans of the body, variations in the properties of the rock bodies allow 3D images to be constructed of the rock bodies through which the rays passed. Seismologists are thus able to construct 3D images of the interior of the Earth.

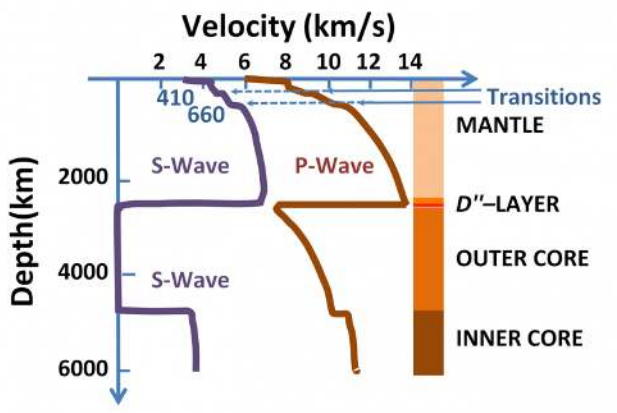


Figure 9.7.5: Speed of seismic waves with depth in the Earth as predicted by the PREM. Two thousand kilometers is 1240 miles.

For example, seismologists have mapped the Farallon Plate, a tectonic plate that subducted beneath North America during the last several million years, and the Yellowstone magma chamber, which is a product of the Yellowstone hot spot under the North American continent. Peculiarities of the Farallon Plate subduction are thought to be responsible for many features of western North America including the Rocky Mountains.

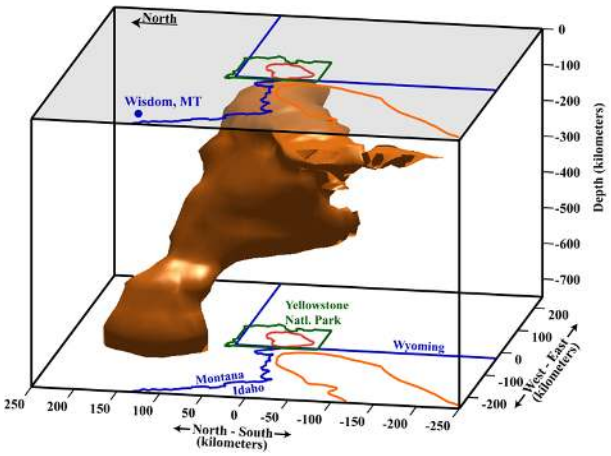


Figure 9.7.6: Seismic tomography showing the magma chamber beneath Yellowstone National Park.

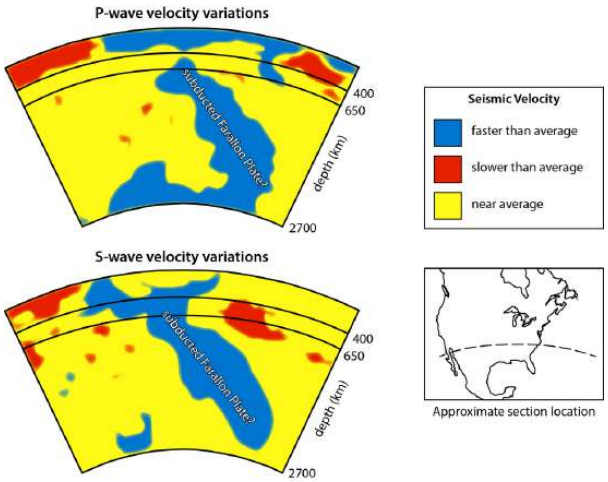


Figure 9.7.7: Simplified and interpreted P and S wave velocity variations in the mantle across southern North America showing the subducted Farallon Plate. (By Oilfieldvegetarian; CC BY-SA 4.0 via Wikimedia Commons.)

Earthquake Magnitude and Intensity

Richter Scale

Magnitude is the measure of the energy released by an earthquake. The **Richter scale** (M_L), the first and most well-known magnitude scale, was developed by Charles F. Richter (1900-1985) at the California Institute of Technology. This was the magnitude scale used historically by early seismologists. Richter magnitude (M_L) is determined from the maximum amplitude of the pen tracing on the seismogram recording. Adjustments for epicenter distance from the seismograph are made using the arrival-time differences of S and P waves [7].

The Richter Scale is logarithmic, based on powers of 10. This means an increase of one Richter unit represents a 10-fold increase in seismic-wave amplitude or in other words, a magnitude 6 earthquake shakes the ground 10 times more than a magnitude 5. However, the *actual energy released* for each magnitude unit is 32 times greater, which means a magnitude 6 earthquake releases 32 times more energy than a magnitude 5.

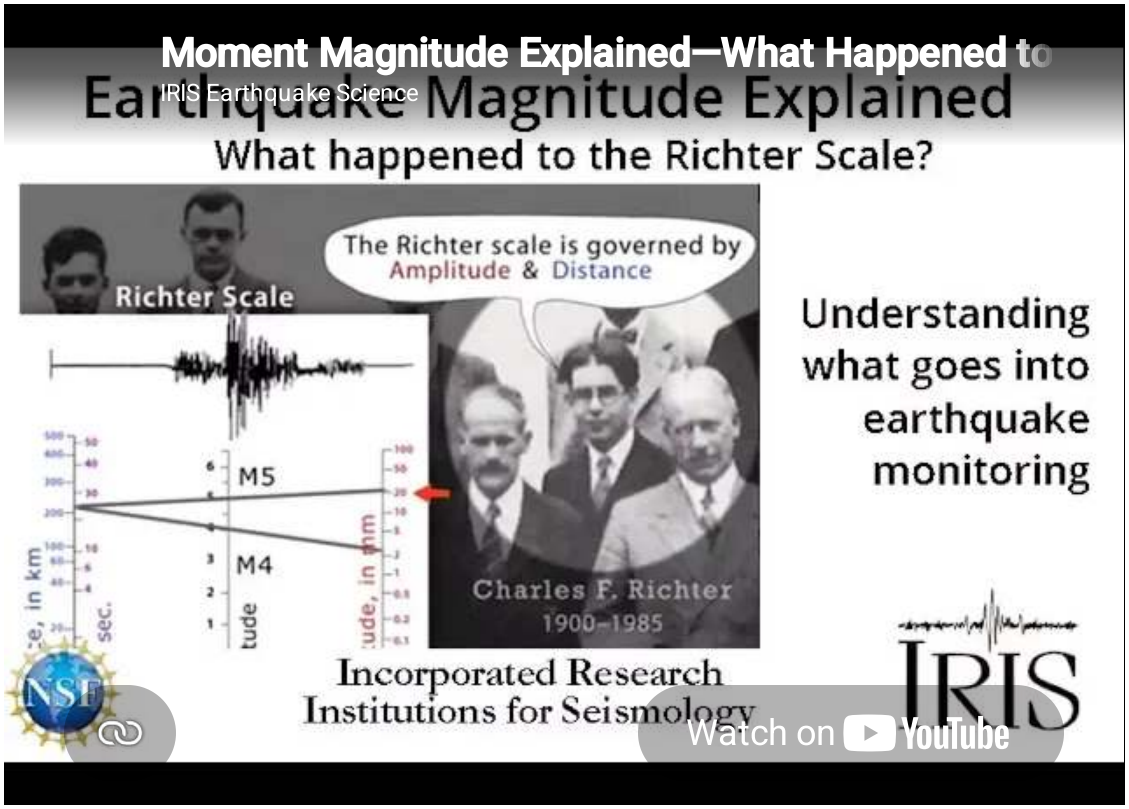
The Richter Scale was developed for earthquakes in Southern California, using local seismographs. It has limited applications for larger distances and very large earthquakes. Therefore, most agencies no longer use the Richter Scale. The moment magnitude (M_W), which is measured using seismic arrays and generates values comparable to the Richter Scale, is more accurate for measuring earthquakes across the Earth, including large earthquakes, although they require more time to calculate. News media often report Richter magnitudes right after an earthquake occurs even though scientific calculations now use moment magnitudes.

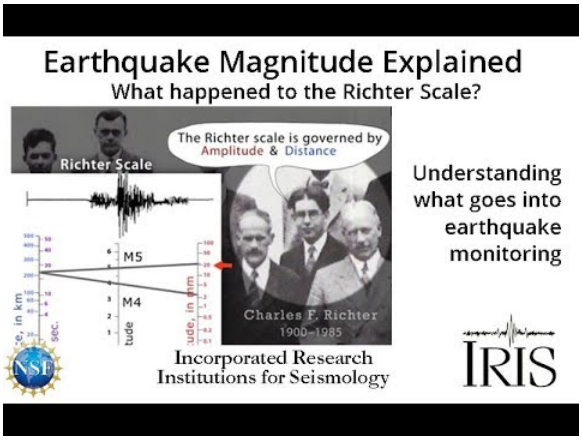
Moment Magnitude Scale

The **moment magnitude scale** depicts the absolute size of earthquakes, comparing information from multiple locations and using a measurement of actual energy released calculated from the cross-sectional area of rupture, amount of slippage, and the rigidity of the rocks. Because each earthquake occurs in a unique geologic setting and the rupture area is often hard to measure, estimates of moment magnitude can take days or even months to calculate.

Like the Richter Scale, the moment magnitude scale is logarithmic. The magnitude values of the two scales are approximately equal, except for very large earthquakes. Both scales are used for reporting earthquake magnitude. The Richter Scale provides a quick magnitude estimate immediately following the quake and thus, is usually reported in news accounts. Moment magnitude calculations take much longer but are more accurate and thus, more useful for scientific analysis.

Watch this video that explains the moment magnitude scale.





Modified Mercalli Intensity Scale

The **Modified Mercalli Intensity Scale** (MMI) is a qualitative rating of ground-shaking intensity based on observable structural damage and people's perceptions. This scale uses a *I* (Roman numeral one) rating for the lowest intensity and *X* (ten) for the highest (see table) and can vary depending on epicenter location and population density, such as urban versus rural settings. Historically, scientists used the MMI Scale to categorize earthquakes before they developed quantitative measurements of magnitude. Intensity maps show locations of the most severe damage, based on residential questionnaires, local news articles, and on-site assessment reports.

Table 9.7.1: Abridged Mercalli Scale from USGS General Interest Publication 1989-288-913.

Intensity	Shaking	Description/Damage
I	Not felt	Not felt except by a very few under especially favorable conditions.
II	Weak	Felt only by a few persons at rest, especially on upper floors of buildings.
III	Weak	Felt quite noticeably by persons indoors, especially on upper floors of buildings. Many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibrations similar to the passing of a truck. Duration estimated.
IV	Light	Felt indoors by many, outdoors by few during the day. At night, some awakened. Dishes, windows, doors disturbed; walls make cracking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably.
V	Moderate	Felt by nearly everyone; many awakened. Some dishes, windows broken. Unstable objects overturned. Pendulum clocks may stop.
VI	Strong	Felt by all, many frightened. Some heavy furniture moved; a few instances of fallen plaster. Damage slight.
VII	Very strong	Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable damage in poorly built or badly designed structures; some chimneys broken.
VIII	Severe	Damage slight in specially designed structures; considerable damage in ordinary substantial buildings with partial collapse. Damage great in poorly built structures. Fall of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned.
IX	Violent	Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb. Damage great in substantial buildings, with partial collapse. Buildings shifted off foundations.
X	Extreme	Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations. Rails bent.

ShakeMaps

Shake maps, written ShakeMaps by the USGS, use high-quality, computer-interpolated data from seismograph networks to show areas of intense shaking. Shake maps are useful in the crucial minutes after an earthquake, as they show emergency personnel where the greatest damage likely occurred and help them locate possibly damaged gas lines and other utility facilities.

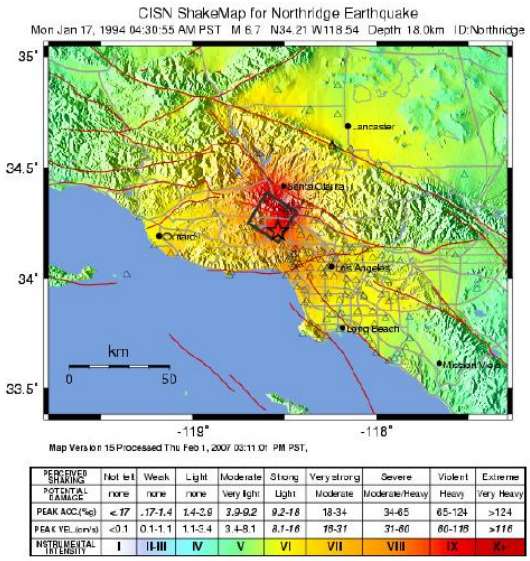


Figure 9.7.8: Example of a shake map. (By CISN; CC BY-SA 4.0 via Wikimedia Commons.)

Reference

7. Richter CF (1935) An instrumental earthquake magnitude scale. Bulletin of the Seismological Society of America 25:1–32

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9.8: Earthquake Risk

Factors that Determine Shaking

Earthquake magnitude is an absolute value that measures pure energy release. **Intensity**, however, i.e. how much the ground shakes, is determined by several factors.

Earthquake magnitude—In general, the larger the magnitude, the stronger the shaking and the longer the shaking will last.

This table is taken from the USGS and shows scales of magnitude and Mercalli Intensity and descriptions of shaking and resulting damage.

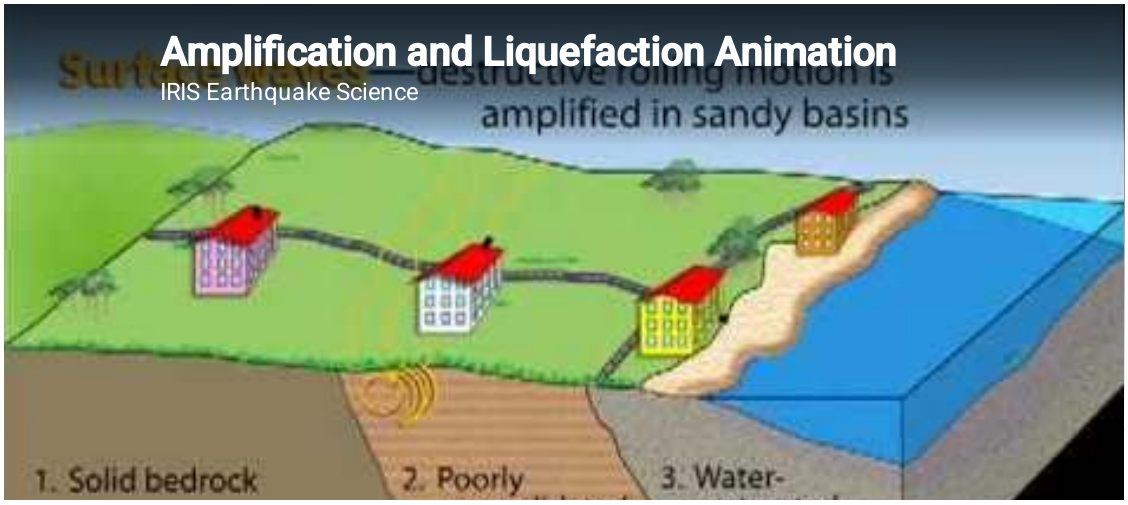
Table 9.8.1: Mercalli Intensity as it relates to magnitude.

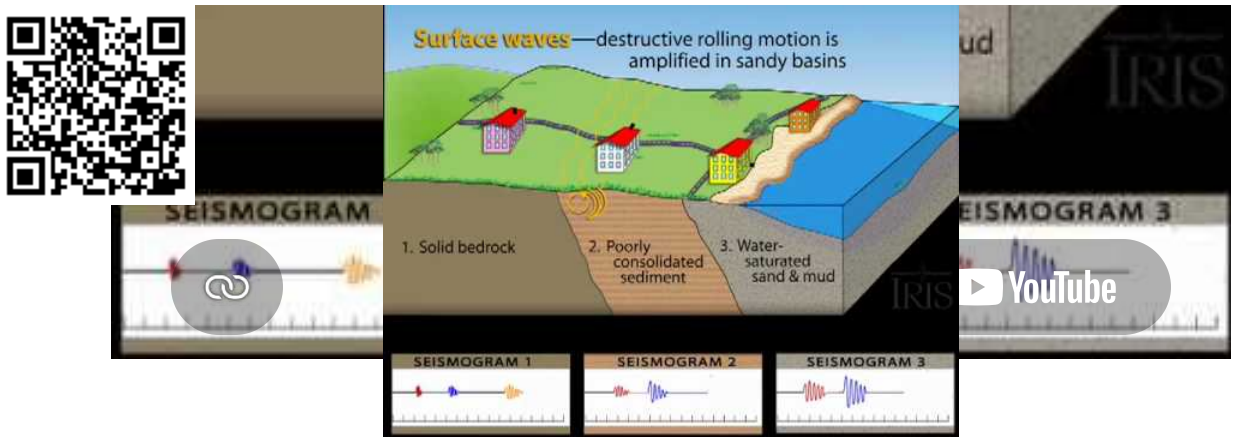
Magnitude	Modified Mercalli Intensity	Shaking/Damage Description
1.0 – 3.0	**I**	Only felt by a very few.
3.0 – 3.9	**II – III**	Noticeable indoors, especially on upper floors.
4.0 – 4.9	**IV – V**	Most to all feel it. Dishes, doors, cars shake and possibly break.
5.0 – 5.9	**VI – VII**	Everyone feels it. Some items knocked over or broken. Building damage possible.
6.0 – 6.9	**VII – VIII**	Frightening amounts of shaking. Significant damage especially with poorly constructed buildings.
≥ 7.0	**≥ VIII**	Significant destruction of buildings. Potential for objects to be thrown in the air from shaking.

Location and direction—Shaking is more severe closer to the epicenter. The severity of shaking is influenced by the location of the observer relative to the epicenter, the direction of rupture propagation, and the path of greatest rupture.

Local geologic conditions—Seismic waves are affected by the nature of the ground materials through which they pass. Different materials respond differently to an earthquake. Think of shaking a block of Jello versus meatloaf, one will jiggle much more when hit by waves of the same amplitude. The ground's response to shaking depends on the degree of substrate consolidation. Solid sedimentary, igneous, or metamorphic bedrock shakes less than unconsolidated sediments.

This video shows how different substrates behave in response to different seismic waves and their potential for destruction.





Seismic waves move fastest through consolidated bedrock, slower through unconsolidated sediments, and slowest through unconsolidated sediments with a high water content. Seismic energy is transmitted by wave velocity and amplitude. When seismic waves slow down, energy is transferred to the amplitude, increasing the motion of surface waves, which in turn amplifies ground shaking.

Focus depth—Deeper earthquakes cause less surface shaking because much of their energy, transmitted as body waves, is lost before reaching the surface. Recall that surface waves are generated by P and S waves impacting the Earth's surface.

Factors that Determine Destruction

Just as certain conditions will impact the intensity of ground shaking, several factors affect how much destruction is caused.



Figure 9.8.1: Example of devastation on unreinforced masonry by seismic motion.

Building materials—The flexibility of a building material determines its resistance to earthquake damage. Unreinforced masonry (URM) is the material most devastated by ground shaking. Wood framing fastened with nails bends and flexes during seismic wave passage and is more likely to survive intact. Steel also has the ability to deform elastically before brittle failure.

Intensity and duration—Greater shaking and duration of shaking causes more destruction than lower and shorter shaking.

Resonance—**Resonance** occurs when seismic wave frequency matches a building's natural shaking frequency and increases the shaking. This happened in the 1985 Mexico City earthquake, where buildings of heights between 6 and 15 stories were especially vulnerable to earthquake damage. Skyscrapers designed with earthquake resilience have dampers and base isolation features to reduce resonance.

Resonance is influenced by the properties of the building materials. Changes in the structural integrity of a building can alter resonance [8]. Conversely, changes in measured resonance can indicate a potentially altered structural integrity.

These two videos discuss why buildings fall during earthquakes and a modern procedure to reduce potential earthquake destruction for larger buildings.


Why do buildings fall in earthquakes? - Vicki V. Ma
TED-Ed
WHY DO BUILDINGS FALL IN EARTHQUAKES?
TEDEd
Watch on YouTube


WHY DO BUILDINGS FALL IN EARTHQUAKES?
TEDEd


Utah State Capital Building - Base Isolators
ReadySetPrepShow
Watch on YouTube



Earthquake Recurrence

A long hiatus in activity along a fault segment with a history of recurring earthquakes is known as a **seismic gap**. The lack of activity may indicate the fault segment is locked, which may produce a buildup of strain and a higher probability of an earthquake recurring. Geologists dig earthquake trenches across faults to estimate the frequency of past earthquake occurrences. Trenches are effective for faults with relatively long **recurrence intervals**, roughly 100s to 10,000s of years between significant earthquakes. Trenches are less useful in areas with more frequent earthquakes because they usually have more recorded data.



Figure 9.8.2: Fault trench near Draper Utah. Trenches allow geologists to see a cross-section of a fault and to use dating techniques to determine how frequently earthquakes occur.

Earthquake Distribution

This video shows the distribution of earthquakes with magnitude 6 or more on Earth during the years 1900 through 2019. Like volcanoes, earthquakes tend to aggregate around active boundaries of tectonic plates. The exception is intraplate earthquakes, which are comparatively rare.





Subduction Zones—Subduction zones, found at convergent plate boundaries, are where the largest and deepest earthquakes, called **megathrust earthquakes**, occur. Examples of subduction-zone earthquake areas include the Sumatran Islands, Aleutian Islands, west coast of South America, and Cascadia Subduction Zone off the coast of Washington and Oregon.

Collision Zones—Collisions between converging continental plates create broad earthquake zones that may generate deep, large earthquakes from the remnants of past subduction events or other deep-crustal processes. The Himalayan Mountains (northern border of the Indian subcontinent) and Alps (southern Europe and Asia) are active regions of collision-zone earthquakes.

Transform Boundaries—Strike-slip faults created at transform boundaries produce moderate-to-large earthquakes, usually having a maximum moment magnitude of about 8. The San Andreas fault (California) is an example of a transform-boundary earthquake zone. Haiti's Enriquillo-Plantain Garden fault system, which caused the 2010 earthquake near Port-au-Prince, and Septentrional Fault, which destroyed Cap-Haïtien in 1842 and shook Cuba in 2020, are also transform faults. Other examples are the Alpine Fault (New Zealand) and Anatolian Faults (Turkey).

Divergent Boundaries—Continental rifts and mid-ocean ridges found at divergent boundaries generally produce moderate earthquakes. Examples of active earthquake zones include the East African Rift System (southwestern Asia through eastern Africa), Iceland, and Basin and Range province (Nevada, Utah, California, Arizona, and northwestern Mexico).

Intraplate Earthquakes—Intraplate earthquakes are not found near tectonic plate boundaries, but generally occur in areas of weakened crust or concentrated tectonic stress. The New Madrid seismic zone, which covers Missouri, Illinois, Tennessee, Arkansas, and Indiana, is thought to represent the failed Reelfoot rift [9]. The failed rift zone weakened the crust, making it more

responsive to tectonic plate movement and interaction. Geologists theorize the infrequently occurring earthquakes are produced by low strain rates.

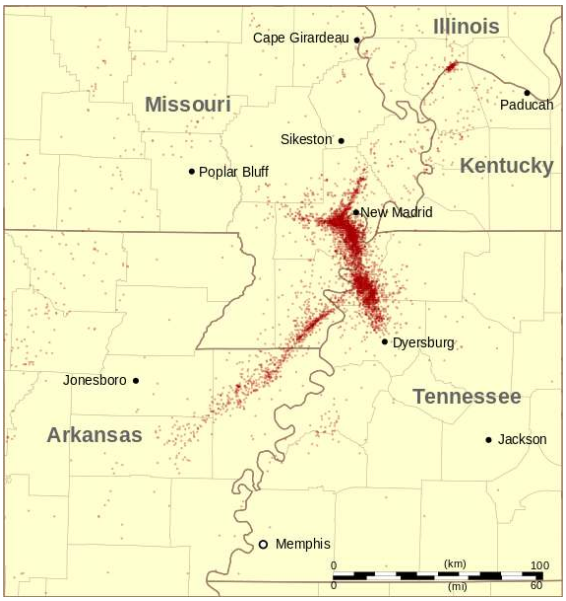


Figure 9.8.3: High density of earthquakes in the New Madrid seismic zone.

Secondary Hazards Caused by Earthquakes

Most earthquake damage is caused by ground shaking and fault-block displacement. In addition, there are secondary hazards that endanger structures and people, in some cases after the shaking stops.

Liquefaction—**Liquefaction** occurs when water-saturated, unconsolidated sediments, usually silt or sand, become fluid-like from shaking. The shaking breaks the cohesion between grains of sediment, creating a slurry of particles suspended in water. Buildings settle or tilt in the liquefied sediment, which looks very much like quicksand in the movies. Liquefaction also creates sand volcanoes, cone-shaped features created when liquefied sand is squirted through an overlying and usually finer-grained layer.



Figure 9.8.4: Buildings toppled from liquefaction during a 7.5 magnitude earthquake in Japan.

This video demonstrates how liquefaction takes place.





Tsunamis—Among the most devastating natural disasters are **tsunamis**, earthquake-induced ocean waves. When the seafloor is offset by fault movement or an underwater landslide, the ground displacement lifts a volume of ocean water and generates the tsunami wave. Tsunami waves are fast-moving with low amplitude in deep ocean water but grow significantly in amplitude in the shallower waters approaching the shore. When a tsunami is about to strike land, the drawback of the trough preceding the wave crest causes the water to recede dramatically from shore. Tragically, curious people wander out and follow the disappearing water, only to be overcome by an oncoming wall of water that can be upwards of a 30 m (100 ft) high. Early warning systems help mitigate the loss of life caused by tsunamis.

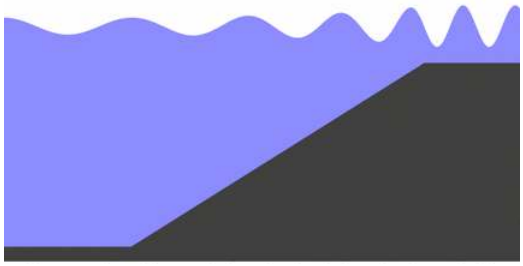


Figure 9.8.5: As the ocean depth becomes shallower, the wave slows down and pile up on top of itself, making large, high-amplitude waves.

Landslides—Shaking can trigger landslides. In 1992 a moment magnitude 5.9 earthquake in St. George, Utah, caused a landslide that destroyed several structures in the Balanced Rock Hills subdivision in Springville, Utah [10].



Figure 9.8.6: House in Springdale, Utah destroyed by an earthquake-triggered landslide.

Seiches—**Seiches** are waves generated in lakes by earthquakes. The shaking may cause water to slosh back-and-forth or sometimes change the lake depth. Seiches in Hebgen Lake during a 1959 earthquake caused major destruction to nearby structures and roads.

This video shows a seiche generated in a swimming pool by an earthquake in Nepal in 2015.





Land Elevation Changes—Elastic rebound and displacement along the fault plane can cause significant land elevation changes, such as subsidence or upheaval. The 1964 Alaska earthquake produced significant land elevation changes, with the differences in height between the hanging wall and footwall ranging from one to several meters (3–30 ft). The Wasatch Mountains in Utah represent an accumulation of fault scarps created a few meters at a time, over a few million years.

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9.9: Case Studies

North American Earthquakes

Basin and Range Earthquakes—Earthquakes in the Basin and Range Province, from the Wasatch Fault (Utah) to the Sierra Nevada (California), occur primarily in normal faults created by tensional forces. The Wasatch Fault, which defines the eastern extent of the Basin and Range province, has been studied as an earthquake hazard for more than 100 years.

New Madrid Earthquakes (1811-1812)—Historical accounts of earthquakes in the New Madrid seismic zone date as far back as 1699 and earthquakes continue to be reported in modern times [11]. A sequence of large (M_w >7) occurred from December 1811 to February 1812 in the New Madrid area of Missouri [12]. The earthquakes damaged houses in St. Louis, affected the stream course of the Mississippi River, and leveled the town of New Madrid. These earthquakes were the result of intraplate seismic activity [9].

Charleston (1868)—The 1868 earthquake in Charleston South Carolina was a moment magnitude 7.0, with a Mercalli intensity of X, caused significant ground motion, and killed at least 60 people. This intraplate earthquake was likely associated with ancient faults created during the breakup of Pangea. The earthquake caused significant liquefaction [13]. Scientists estimate the recurrence of destructive earthquakes in this area with an interval of approximately 1500 to 1800 years.

Great San Francisco Earthquake and Fire (1906)—On April 18, 1906, a large earthquake, with an estimated moment magnitude of 7.8 and MMI of X, occurred along the San Andreas fault near San Francisco California. There were multiple aftershocks followed by devastating fires, resulting in about 80% of the city being destroyed. Geologists G.K. Gilbert and Richard L. Humphrey, working independently, arrived the day following the earthquake and took measurements and photographs [14].



Figure 9.9.1: Remains of San Francisco after the 1906 earthquake and fire.

Alaska (1964)—The 1964 Alaska earthquake, moment magnitude 9.2, was one of the most powerful earthquakes ever recorded. The earthquake originated in a megathrust fault along the Aleutian subduction zone. The earthquake caused large areas of land subsidence and uplift, as well as significant mass wasting.

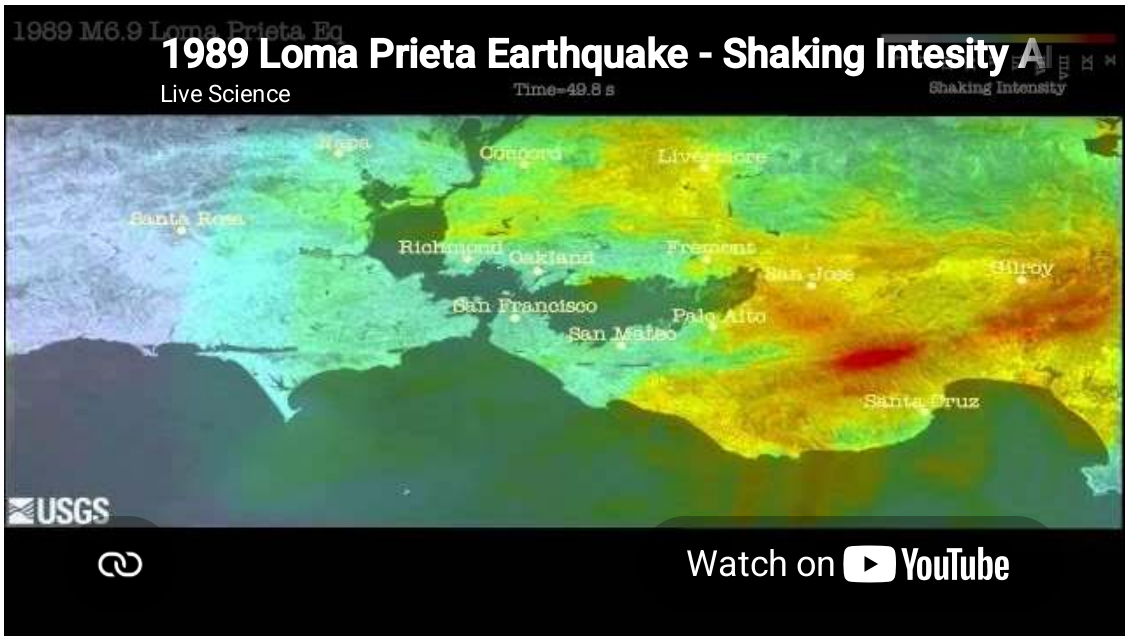
Video from the USGS about the 1964 Alaska earthquake.

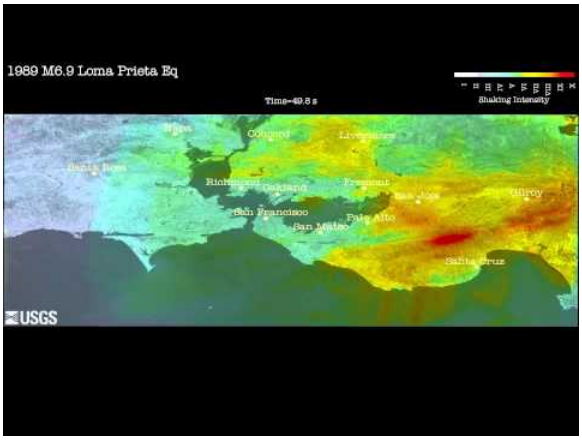




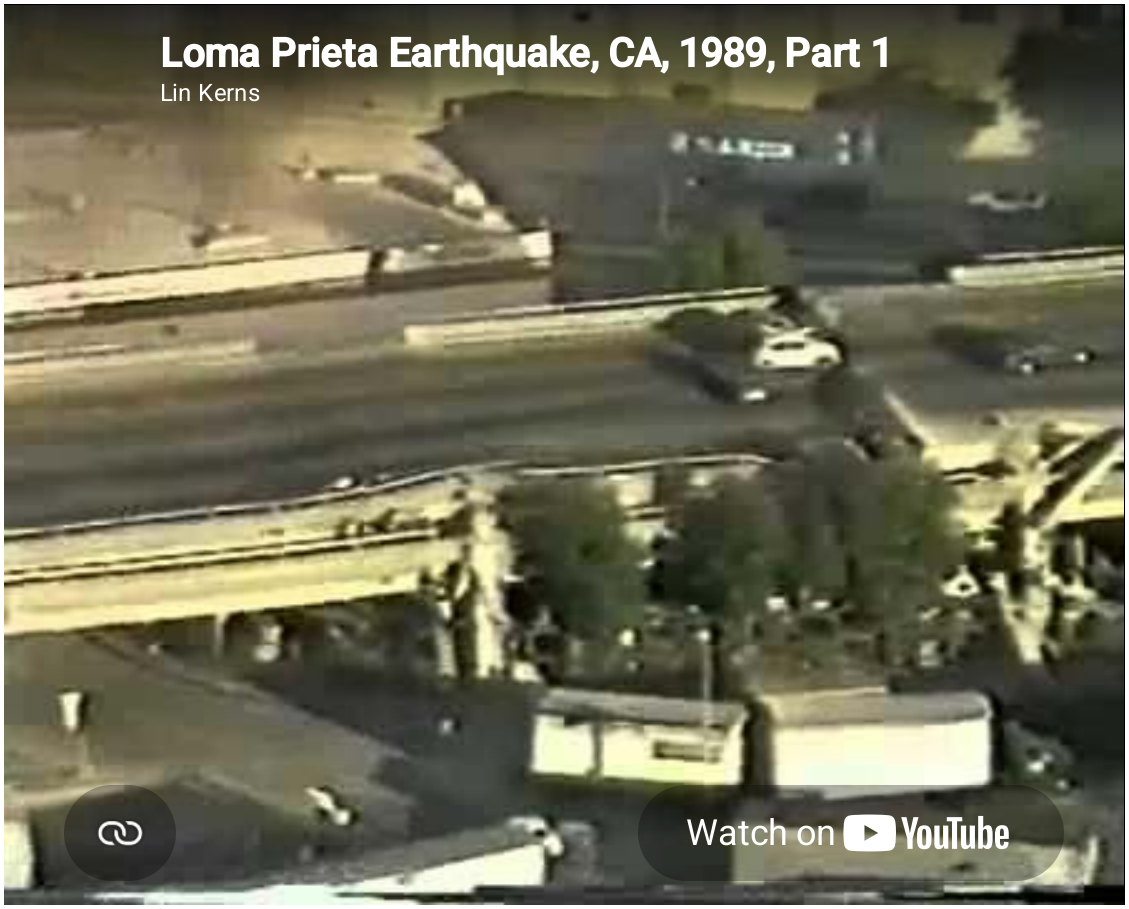
Loma Prieta (1989)—The Loma Prieta, California, earthquake was created by movement along the San Andreas Fault. The moment magnitude 6.9 earthquake was followed by a magnitude of 5.2 aftershock. It caused 63 deaths, buckled portions of the several freeways, and collapsed part of the San Francisco-Oakland Bay Bridge.

This video shows how shaking propagated across the Bay Area during the 1989 Loma Prieta earthquake.





This video shows the destruction caused by the 1989 Loma Prieta earthquake.





Global Earthquakes

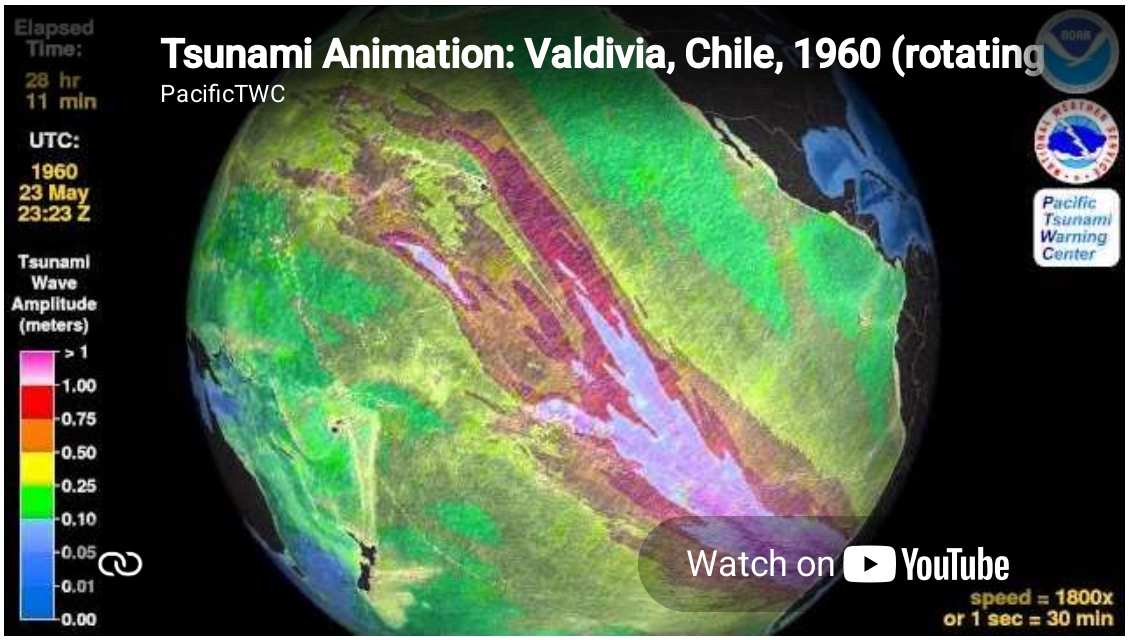
Many of history's largest earthquakes occurred in megathrust zones, such as the estimated magnitude 9 earthquake that occurred on the Cascadia Subduction Zone (Washington and Oregon coasts) in 1700.

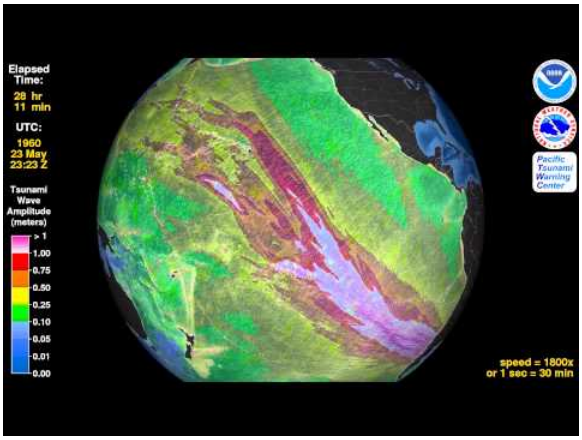
Shaanxi, China (1556)—On January 23, 1556 an earthquake of an approximate moment magnitude 8 hit central China, killing approximately 830,000 people in what is considered the most deadly earthquake in history. The high death toll was attributed to the collapse of cave dwellings (*yaodong*) built in loess deposits, which are large banks of windblown, compacted sediment. Earthquakes in this are region are believed to have a recurrence interval of 1000 years. [15].

Lisbon, Portugal (1755)—On November 1, 1755 an earthquake with an estimated moment magnitude range of 8–9 struck Lisbon, Portugal [13], killing between 10,000 to 17,400 people [16]. The earthquake was followed by a tsunami, which brought the total death toll to between 30,000 to 70,000 people.

Valdivia, Chile (1960)—The May 22, 1960 earthquake was the most powerful earthquake ever measured, with a moment magnitude of 9.4–9.6 and lasting an estimated 10 minutes. It triggered tsunamis that destroyed houses across the Pacific Ocean in Japan and Hawaii and caused vents to erupt on the Puyehue-Cordón Caulle (Chile).

This video describes the tsunami produced by the 1960 Chili earthquake.





Tangshan, China (1976)—Just before 4 a.m. (Beijing time) on July 28, 1976 a moment magnitude 7.8 earthquake struck Tangshan (Hebei Province), China, and killed more than 240,000 people. The high death toll is attributed to people still being asleep or at home and most buildings being made of unreiforced masonry.

Sumatra, Indonesia (2004)—On December 26, 2004, slippage of the Sunda megathrust fault generated a moment magnitude 9.0–9.3 earthquake off the coast of Sumatra, Indonesia [17]. This megathrust fault is created by the Australia plate subducting below the Sunda plate in the Indian Ocean [18]. The resultant tsunamis created massive waves as tall as 24 m (79 ft) when they reached the shore and killed more than an estimated 200,000 people along the Indian Ocean coastline.

Haiti (2010)—The moment magnitude 7 earthquake that occurred on January 12, 2010, was followed by many aftershocks of magnitude 4.5 or higher. More than 200,000 people are estimated to have died as a result of the earthquake. The widespread infrastructure damage and crowded conditions contributed to a cholera outbreak, which is estimated to have caused thousands more deaths.

Tōhoku, Japan (2011)—Because most Japanese buildings are designed to tolerate earthquakes, the moment magnitude 9.0 earthquake on March 11, 2011, was not as destructive as the tsunami it created. The tsunami caused more than 15,000 deaths and tens of billions of dollars in damage, including the destructive meltdown of the Fukushima nuclear power plant.

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9.S: Summary

Summary

Geologic stress, applied force, comes in three types: tension, shear, and compression. Strain is produced by stress and produces three types of deformation: elastic, ductile, and brittle. Geological maps are two-dimensional representations of surface formations which are the surface expression of three-dimensional geologic structures in the subsurface. The map symbol called strike and dip or rock attitude indicates the orientation of rock strata with reference to north-south and horizontal. Folded rock layers are categorized by the orientation of their limbs, fold axes and axial planes. Faults result when stress forces exceed rock integrity and friction, leading to brittle deformation and breakage. The three major fault types are described by the movement of their fault blocks: normal, strike-slip, and reverse.

Earthquakes, or seismic activity, are caused by sudden brittle deformation accompanied by elastic rebound. The release of energy from an earthquake focus is generated as seismic waves. P and S waves travel through the Earth's interior. When they strike the outer crust, they create surface waves. Human activities, such as mining and nuclear detonations, can also cause seismic activity. Seismographs measure the energy released by an earthquake using a logarithmic scale of magnitude units; the Moment Magnitude Scale has replaced the original Richter Scale. Earthquake intensity is the perceived effects of ground shaking and physical damage. The location of earthquake foci is determined from triangulation readings from multiple seismographs.

Earthquake rays passing through rocks of the Earth's interior and measured at the seismographs of the worldwide Seismic Network allow 3-D imaging of buried rock masses as seismic tomographs.

Earthquakes are associated with plate tectonics. They usually occur around the active plate boundaries, including zones of subduction, collision, and transform and divergent boundaries. Areas of intraplate earthquakes also occur. The damage caused by earthquakes depends on a number of factors, including magnitude, location and direction, local conditions, building materials, intensity and duration, and resonance. In addition to damage directly caused by ground shaking, secondary earthquake hazards include liquefaction, tsunamis, landslides, seiches, and elevation changes.

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CHAPTER OVERVIEW

10: Mass Wasting

Learning Objectives

At the end of this chapter, students should be able to:

- Explain what mass wasting is and why it occurs on a slope.
- Explain the basic triggers of mass-wasting events and how they occur.
- Identify types of mass wasting.
- Identify risk factors for mass-wasting events.
- Evaluate landslides and their contributing factors.



Figure 10.1: The 1983 Thistle landslide (foreground) dammed the Spanish Fork River creating a lake that covered the town of Thistle, Utah. The slide covered Hwy 6 and the main railroad between Salt Lake and Denver. (By R.L. Schuster; public domain via Wikimedia Commons.)

This chapter discusses the fundamental processes driving mass-wasting, types of mass wasting, examples, and lessons learned from famous mass-wasting events, how mass wasting can be predicted, and how people can be protected from this potential hazard. **Mass wasting** is the downhill movement of rock and soil material due to gravity. The term landslide is often used as a synonym for mass wasting, but mass wasting is a much broader term referring to all movement downslope. Geologically, a **landslide** is a general term for mass wasting that involves fast-moving geologic material. Loose material along with overlying soils are what typically moves during a mass-wasting event. Moving blocks of bedrock are called rock topples, rock slides, or rock falls, depending on the dominant motion of the blocks. Movements of dominantly liquid material are called flows. Movement by mass wasting can be slow or rapid. Rapid movement can be dangerous, such as during debris flows. Areas with steep topography and rapid rainfall, such as the California coast, Rocky Mountain Region, and Pacific Northwest, are particularly susceptible to hazardous mass-wasting events.

10.1: Slope Strength

10.2: Mass-Wasting Triggers and Mitigation

10.3: Landslide Classification and Identification

10.4: Examples of Landslides

10.S: Summary

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10.1: Slope Strength

Mass wasting occurs when a slope fails. A **slope failure** occurs when it is too steep and unstable for existing materials and conditions. **Slope stability** is ultimately determined by two principal factors: the **slope angle** and the strength of the underlying material. The force of gravity, which plays a part in mass wasting, is constant on the Earth's surface for the most part, although small variations exist depending on the elevation and density of the underlying rock. In the figure below, a block of rock situated on a slope is pulled down toward the Earth's center by the force of gravity (fg). The **gravitational force** acting on a slope can be divided into two components: the **shear** or **driving force** (fs) pushing the block down the slope, and the **normal** or **resisting force** (fn) pushing into the slope, which produces **friction**. The relationship between shear force and normal force is called **shear strength**. When the normal force, i.e., friction, is greater than the shear force, then the block does *not* move downslope. However, if the slope angle becomes steeper or if the earth material is weakened, shear force exceeds normal force, compromising shear strength, and downslope movement occurs.

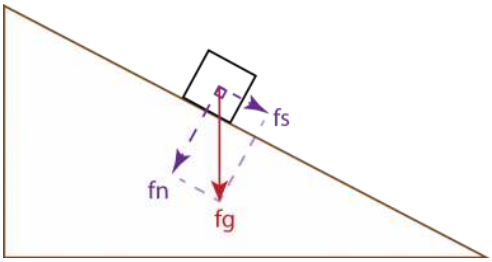


Figure 10.1.1: Forces on a block on an inclined plane (fg = force of gravity; fn = normal force; fs = shear force).

In the figure, the force vectors change as the slope angle increases. The gravitational force doesn't change, but the shear force increases while the normal force decreases. The steepest angle at which rock and soil material is stable and will *not* move downslope is called the **angle of repose.** The angle of repose is measured relative from the horizontal. When a slope is at the angle of repose, the shear force is in equilibrium with the normal force. If the slope becomes just slightly steeper, the shear force exceeds the normal force, and the material starts to move downhill.

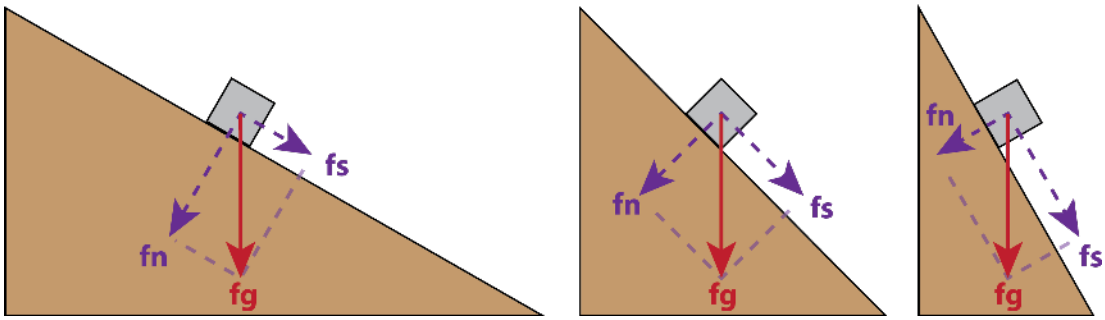


Figure 10.1.2: As slope increases, the force of gravity (fg) stays the same and the normal force decreases while the shear force proportionately increases.

The angle of repose varies for all material and slopes depending on many factors such as grain size, grain composition, and water content. The figure shows the angle of repose for sand that is poured into a pile on a flat surface. The sand grains cascade down the sides of the pile until coming to rest at the angle of repose. At that angle, the base and height of the pile continue to increase, but the angle of the sides remains the same.

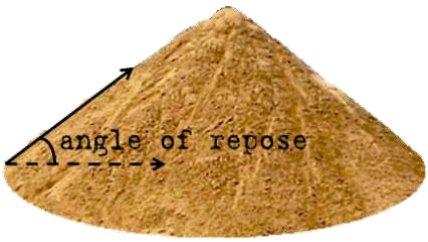


Figure 10.1.3: Angle of repose in a pile of sand.

Water is a common factor that can significantly change the shear strength of a particular slope. Water is located in **pore spaces**, which are empty air spaces between the grains in sediments or rocks. For example, assume a dry sand pile has an angle of repose of 30°. If water is added to the sand, the angle of repose will increase, possibly to 60° or even 90°, such as a sandcastle being built at a beach. But if too much water is added to the pore spaces of the sandcastle, the water decreases the shear strength, lowers the angle of repose, and the sandcastle collapses.

Another factor influencing shear strength is planes of weakness in sedimentary rocks. Bedding planes can act as significant planes of weakness when they are parallel to the slope but less so if they are perpendicular to the slope [1]. At locations A and B, the bedding is nearly perpendicular to the slope and relatively stable. At location D, the bedding is nearly parallel to the slope and quite unstable. At location C, the bedding is nearly horizontal, and the stability is intermediate between the other two extremes [1]. Additionally, if clay minerals form along bedding planes, they can absorb water and become slick. When a bedding plane of shale (clay and silt) becomes saturated, it can lower the shear strength of the rock mass and cause a landslide, such as at the 1925 Gros Ventre, Wyoming rock slide. See the case studies section for details on this and other landslides.

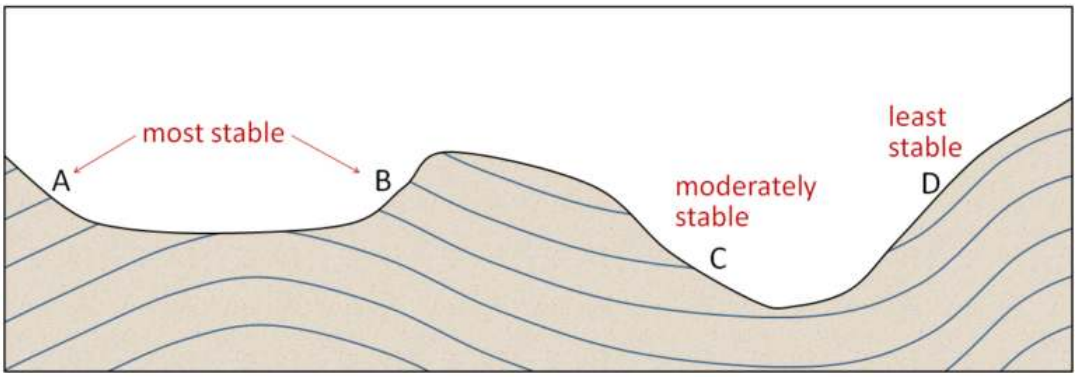


Figure 10.1.4: Locations A and B have bedding nearly perpendicular to the slope, making for a relatively stable slope. Location D has bedding nearly parallel to the slope, increasing the risk of slope failure. Location C has bedding nearly horizontal and the stability is relatively intermediate.

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10.2: Mass-Wasting Triggers and Mitigation

Mass wasting events often have a **trigger**: something changes that cause a landslide to occur at a specific time. It could be rapid snowmelt, intense rainfall, earthquake shaking, volcanic eruption, storm waves, rapid-stream erosion, or human activities, such as grading a new road. Increased water content within the slope is the most common mass wasting trigger. Water content can increase due to rapidly melting snow or ice or an intense rain event. Intense rain events can occur more often during El Niño years. The west coast of North America receives more precipitation than normal, and landslides become more common. Changes in surface-water conditions resulting from earthquakes, previous slope failures that dam up streams or human structures that interfere with runoff, such as buildings, roads, or parking lots [2] can provide additional water to a slope. In the case of the 1959 Hebgen Lake rock slide, Madison Canyon, Montana, the shear strength of the slope may have been weakened by earthquake shaking. Most landslide mitigation efforts divert and drain water away from slide areas. Tarps and plastic sheeting are often used to drain water off of slide bodies and prevent infiltration into the slide. Drains are used to dewater landslides and shallow wells are used to monitor the water content of some active landslides.

An **oversteepened slope** may also trigger landslides. Slopes can be made excessively steep by natural processes of erosion or when humans modify the landscape for building construction. An example of how a slope may be oversteepened during development occurs where the bottom of the slope is cut into, perhaps to build a road or level a building lot, and the top of the slope is modified by depositing excavated material from below. If done carefully, this practice can be very useful in land development, but in some cases, this can result in devastating consequences. For example, this might have been a contributing factor in the 2014 North Salt Lake City, Utah landslide. A former gravel pit was regraded to provide a road and several building lots. These activities may have oversteepened the slope, which resulted in a slow-moving landslide that destroyed one home at the bottom of the slope. Natural processes such as excessive stream erosion from a flood or coastal erosion during a storm can also oversteepen slopes. For example, natural undercutting of the riverbank was proposed as part of the trigger for the famous 1925 Gros Ventre, Wyoming rock slide.

Slope **reinforcement** can help prevent and mitigate landslides. For rockfall-prone areas, sometimes it is economical to use long steel bolts. Bolts, drilled a few meters into a rock face, can secure loose pieces of material that could pose a hazard. **Shotcrete**, a reinforced spray-on form of concrete, can strengthen a slope face when applied properly. **Buttressing** a slide by adding weight at the toe of the slide and removing weight from the head of the slide, can stabilize a landslide. **Terracing**, which creates a stairstep topography, can be applied to help with slope stabilization, but it must be applied at the proper scale to be effective.

A different approach in reducing landslide hazard is to shield, catch, and divert the runout material. Sometimes the most economical way to deal with a landslide hazard is to divert and slow the falling material. Special stretchable fencing can be applied in areas where rockfall is common to protect pedestrians and vehicles. Runout channels, diversion structures, and check dams can be used to slow debris flows and divert them around structures. Some highways have special tunnels that divert landslides over the highway. In all of these cases, the shielding has to be engineered to a scale that is greater than the slide, or catastrophic loss in property and life could result.

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10.3: Landslide Classification and Identification

Mass wasting events are classified by type of movement and type of material, and there are several ways to classify these events. The figure and table show the terms used. In addition, mass wasting types often share common morphological features observed on the surface, such as the head scarp—commonly seen as crescent shapes on a cliff face; hummocky or uneven surfaces; accumulations of talus—loose rocky material falling from above; and toe of the slope, which covers existing surface material.

Types of Mass Wasting

The most common mass wasting types are falls, rotational and translational slides, flows, and creep [2]. **Falls** are abrupt rock movements that detach from steep slopes or cliffs. Rocks separate along existing natural breaks such as fractures or bedding planes. Movement occurs as free-falling, bouncing, and rolling. Falls are strongly influenced by gravity, mechanical weathering, and water. **Rotational slides** commonly show slow movement along a curved rupture surface. **Translational slides** often are rapid movements along a plane of distinct weakness between the overlying slide material and the more stable underlying material. Slides can be further subdivided into **rock slides**, **debris slides**, or **earth slides** depending on the type of the material involved (see table).

Table 10.3.1: Table of Mass Wasting Types. Mass wasting movement type and primary earth material. Modified from [2; 3].

Type of movement	Bedrock	Soil: mostly coarse-grained	Soil: mostly fine-grained
Falls	Rock fall	—	—
Rock avalanche	Rock avalanche	—	—
Rotational slide (slump)	—	Rotational debris slide (slump)	Rotational earth slide (slump)
Translational slide	Translational rock slide	Translational debris slide	Translational earth slide
Flows	—	debris flow	earthflow
Soil creep	—	creep	creep

Flows are rapidly moving mass wasting events in which the loose material is typically mixed with abundant water, creating long runouts at the slope base. Flows are commonly separated into **debris flow** (coarse material) and **earthflow** (fine material) depending on the type of material involved and the amount of water. Some of the largest and fastest flows on land are called **sturzstroms**, or long-runout landslides. They are still poorly understood but are known to travel for long distances, even in places without significant atmospheres like the Moon.

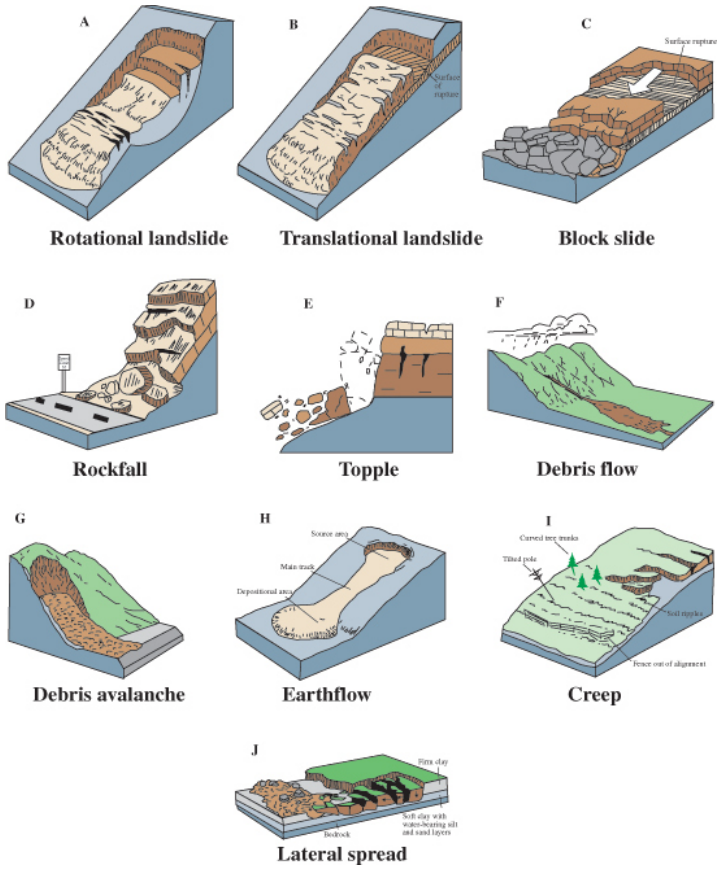


Figure 10.3.1: Examples of some of the types of landslides.

Creep is the imperceptibly slow downward movement of material caused by shear stress sufficient to produce permanent deformation in unconsolidated material [2]. Creep is indicated by curved tree trunks, bent fences or retaining walls, tilted poles or fences, and small soil ripples or ridges. A special type of soil creep is **solifluction**, which is the slow movement of soil lobes on low-angle slopes due to soil repeatedly freezing and thawing in high-latitude, typically sub-Arctic, Arctic, and Antarctic locations.



Landslide Hazards, David Applegate

Parts of a Landslide

Landslides have several identifying features that can be common across the different types of mass wasting. Note that there are many exceptions, and a landslide does not need to have these features. Displacement of material by landslides causes the absence of material uphill and the deposition of new material downhill, and careful observation can identify the evidence of that

displacement. Other signs of landslides include tilted or offset structures or natural features that would normally be vertical or in place.

Many landslides have escarpments or **scarps**. Landslide scarps, like fault scarps, are steep terrain created when the movement of the adjacent land exposes a part of the subsurface. The most prominent scarp is the **main scarp**, which marks the uphill extent of the landslide. As the disturbed material moves out of place, a step slope forms and develops a new hillside escarpment for the undisturbed material. Main scarps are formed by the movement of the displaced material away from the undisturbed ground and are the visible part of the slide rupture surface.

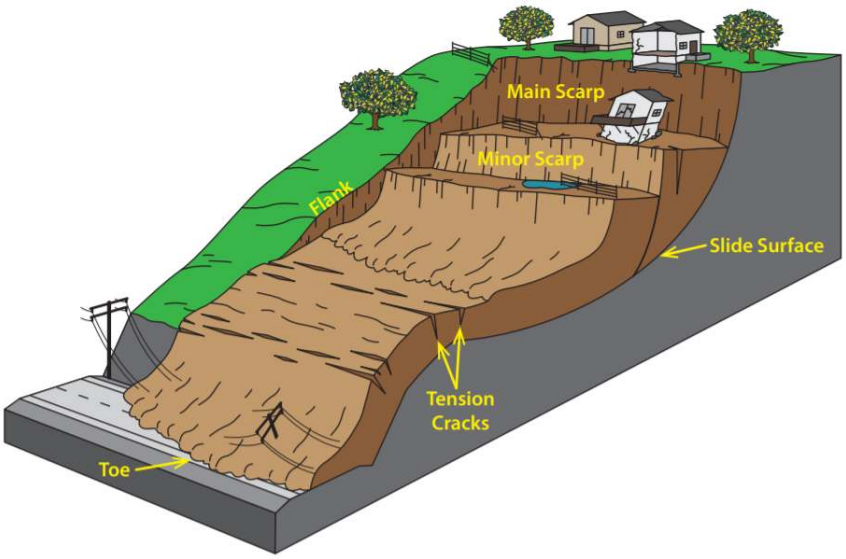


Figure 10.3.2: Diagram of a rotational landslide.

Parts	Description
Crown	The undisturbed material uphill of the scarp. (i.e. the brown house at the top of the hill)
Main Scarp	Steep slope at the upper edge of the landslide (at the head), caused by the movement of displaced material away from the undisturbed ground. The visible part of the slide surface.
Flank	Undisplaced materials adjacent to the sides of the landslide. Flank usually describes the left and right lateral extents of the mass-wasting material.
Rupture Surface/Slide Surface	The lower boundary of the movement below the original ground surface. This is the surface along which material slides. Also known as the surface of rupture.
Main Body	Part of the landslide that overlies the rupture surface.
Tension Cracks	Cracks formed as the result of the middle part of the slide being pulled apart. Usually found in the middle of the slide.
Separation Surface	Part of the original ground surface that is now covered by landslide material (i.e. the location of the landslide where the green and gray surfaces meet)
Foot	The part of the landslide that overlies the original ground surface (i.e. right below the separation surface).
Toe	The downhill end of the slide. The most distant part of the slide from the main scarp.

The **slide rupture surface** is the boundary of the body of the movement of the landslide. The geologic material below the slide surface does not move, and is marked on the sides by the **flanks** of the landslide and at the end by the toe of the landslide.

The **toe** of the landslide marks the end of the moving material. The toe marks the runout, or maximum distance traveled, of the landslide. In rotational landslides, the toe is often a large, disturbed mound of geologic material, forming as the landslide moves past its original rupture surface.

Rotational and translational landslides often have extensional cracks, sag ponds, hummocky terrain, and pressure ridges. **Extensional cracks** form when a landslide's toe moves forward faster than the rest of landslide, resulting in tensional forces. **Sag ponds** are small bodies of water filling depressions formed where landslide movement has impounded drainage. **Hummocky terrain** is undulating and uneven topography that results from the ground being disturbed. **Pressure ridges** develop on the margins of the landslide where the material is forced upward into a ridge structure [4].

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10.4: Examples of Landslides

1925, Gros Ventre, Wyoming: On June 23, 1925, a 38 million cubic meter (50 million yd^3) translational rock slide occurred next to the Gros Ventre River (pronounced “grow vont”) near Jackson Hole, Wyoming. Large boulders dammed the Gros Ventre River and ran up the opposite side of the valley several hundred vertical feet. The dammed river created Slide Lake, and two years later in 1927, lake levels rose high enough to destabilize the dam. The dam failed and caused a catastrophic flood that killed six people in the small downstream community of Kelly, Wyoming [5].



Figure 10.4.1: Scar of the Gros Ventre landslide in the background with landslide deposits in the foreground.

A combination of three factors caused the rock slide: 1) heavy rains and rapidly melting snow saturated the Tensleep Sandstone causing the underlying shale of the Amsden Formation to lose its shear strength, 2) the Gros Ventre River cut through the sandstone creating an oversteepened slope, and 3) soil on top of the mountain became saturated with water due to poor drainage [6]. The cross-section diagram shows how the parallel bedding planes between the Tensleep Sandstone and Amsden Formation offered little friction against the slope surface as the river undercut the sandstone. Lastly, the rockslide may have been triggered by an earthquake.

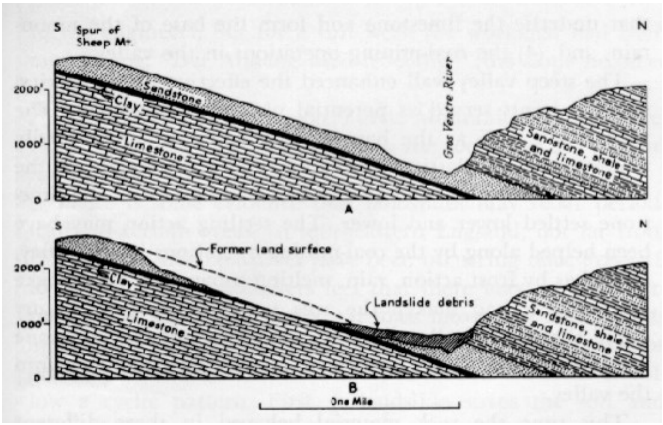


Figure 10.4.2: Cross-section of 1925 Gros Ventre slide showing sedimentary layers parallel with the surface and undercutting (oversteepening) of the slope by the river.

1959, Madison Canyon, Montana: In 1959, the largest earthquake in Rocky Mountain recorded history, magnitude 7.5, struck the Hebgen Lake, Montana area. The earthquake caused a rock avalanche that dammed the Madison River, creating Quake Lake, and ran up the other side of the valley hundreds of vertical feet. Today, there are still house-sized boulders visible on the slope opposite their starting point. The slide moved at a velocity of up to 160.9 kph (100 mph), creating an incredible air blast that swept through the Rock Creek Campground. The slide killed 28 people, most of whom were in the campground and remain buried there [5]. In a manner like the Gros Ventre slide, foliation planes of weakness in metamorphic rock outcrops were parallel with the surface, compromising shear strength.



Figure 10.4.3: 1959 Madison Canyon landslide scar. Photo taken from landslide material.

1980, Mount St. Helens, Washington: On May 18, 1980, a 5.1-magnitude earthquake triggered the largest landslide observed in the historical record. This landslide was followed by the lateral eruption of Mount St. Helens volcano, and the eruption was followed by volcanic debris flows known as lahars. The volume of material moved by the landslide was 2.8 cubic kilometers (0.67 mi^3) [7].

1995 and 2005, La Conchita, California: On March 4, 1995, a fast-moving earthflow damaged nine houses in the southern California coastal community of La Conchita. A week later, debris flow in the same location damaged five more houses. Surface-tension cracks at the top of the slide gave early warning signs in the summer of 1994. During the rainy winter season of 1994/1995, the cracks grew larger. The likely trigger of the 1995 event was unusually heavy rainfall during the winter of 1994/1995 and rising groundwater levels. Ten years later, in 2005, a rapid-debris flow occurred at the end of a 15-day period of near-record rainfall in southern California. Vegetation remained relatively intact as it was rafted on the surface of the rapid flow, indicating that much of the landslide mass simply was being carried on a presumably much more saturated and fluidized layer beneath. The 2005 slide damaged 36 houses and killed 10 people [8].

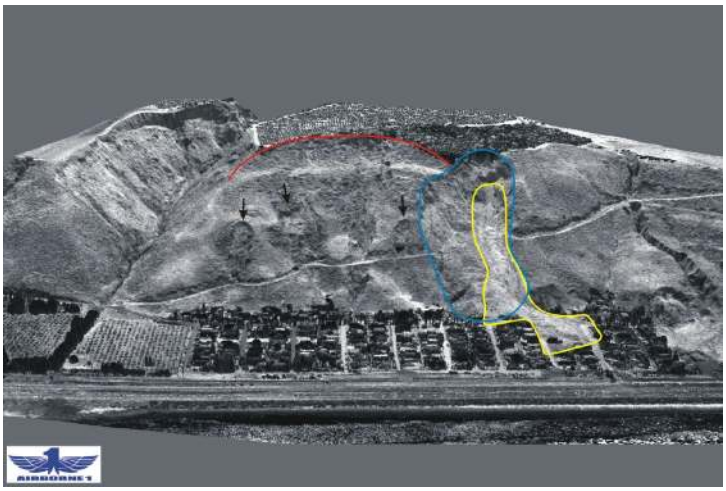


Figure 10.4.4: Oblique LIDAR image of La Conchita after the 2005 landslide. Outline of 1995 (blue) and 2005 (yellow) landslides shown; arrows show examples of other landslides in the area; red line outlines main scarp of an ancient landslide for the entire bluff. Source: Todd Stennett, Airborne 1 Corp., El Segundo. Public domain

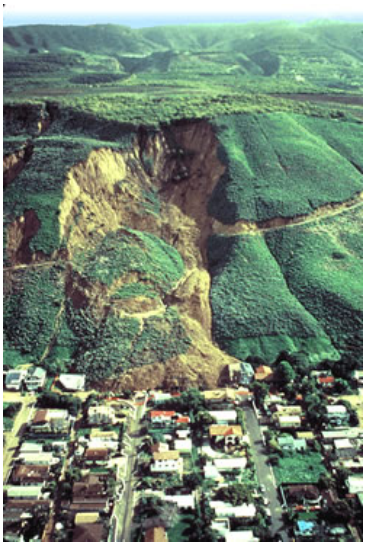


Figure 10.4.5: 1995 La Conchita slide. Source: USGS.

2014, Oso Landslide, Washington: On March 22, 2014, a landslide of approximately 18 million tons (10 million yd^3) traveled at 64 km/hr (40 mph), extended for nearly 1.6 km (1 mi), and dammed the North Fork of the Stillaguamish River. The landslide covered 40 homes and killed 43 people in the Steelhead Haven community near Oso, Washington. It produced a volume of material equivalent to 600 football fields covered in material 3 m (10 ft) deep. The winter of 2013-2014 was unusually wet with almost double the average amount of precipitation. The landslide occurred in an area of the Stillaguamish River Valley historically active with many landslides, but previous events had been small [11].



Figure 10.4.6: 2014 Oso slide in Washington killed 43 people and buried many homes (source: USGS, public domain).

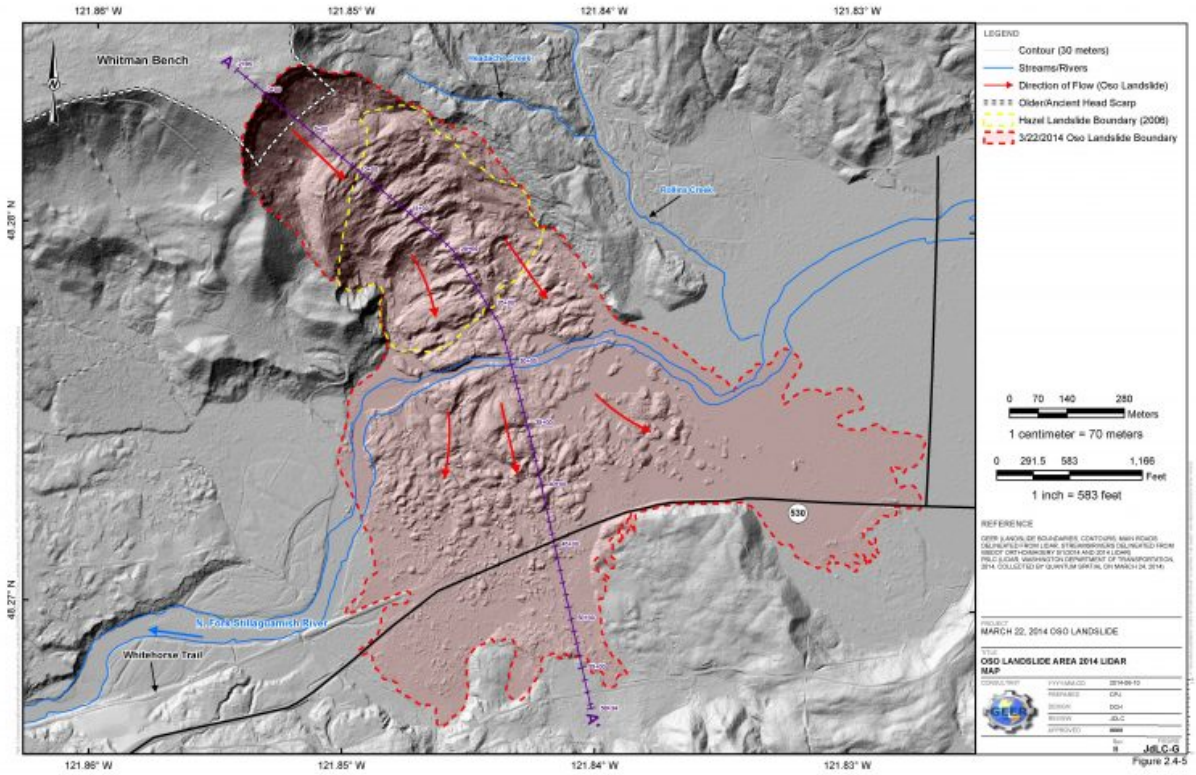


Figure 10.4.7: Annotated LiDAR map of 2014 Oso slide in Washington.

Yosemite National Park Rock Falls: The steep cliffs of Yosemite National Park cause frequent rock falls. Fractures created by tectonic stresses and exfoliation are expanded by frost wedging and can cause house-sized blocks of granite to detach from the cliff faces of Yosemite National Park. The park models potential runout, the distance landslide material travels, to better assess the risk posed to the millions of park visitors.





Rockfalls in Yosemite.

Utah Landslides

Markagunt Gravity Slide: About 21–22 million years ago, one of the biggest land-based landslides yet discovered in the geologic record displaced more than 1,700 km^3 (408 mi^3) of material in one relatively fast event. Evidence for this slide includes breccia conglomerates, glassy pseudotachylyte, slip surfaces (similar to faults), and dikes. The landslide is estimated to encompass an area the size of Rhode Island and to extend from near Cedar City, Utah to Panguitch, Utah. This landslide was likely the result of material released from the side of a growing laccolith (a type of igneous intrusion), after being triggered by an eruption-related earthquake.

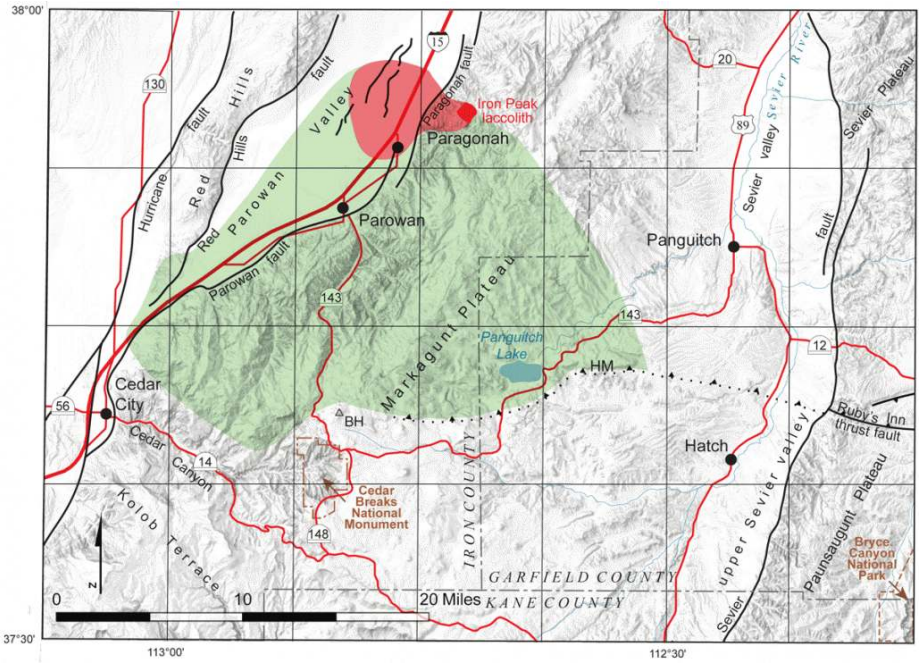


Figure 10.4.8: Approximate extent of Markagunt Gravity slide.

1983, Thistle Slide: Starting in April of 1983 and continuing into May of that year, a slow-moving landslide traveled 305 m (1000 ft) downhill and blocked Spanish Fork Canyon with an earthflow dam 61 m (200 ft) high. This caused disastrous flooding upstream in the Soldier Creek and Thistle Creek valleys, submerging the town of Thistle. As part of the emergency response, a spillway was constructed to prevent the newly formed lake from breaching the dam. Later, a tunnel was constructed to drain the lake, and currently, the river continues to flow through this tunnel. The rail line and US-6 highway had to be relocated at a cost of more than $200 million [13].



Figure 10.4.9: The 1983 Thistle landslide (foreground) dammed the Spanish Fork River creating a lake.

2013, Rockville Rock Fall: Rockville, Utah is a small community near the entrance to Zion National Park. In December of 2013, a 2700 ton (1400 yd^3) block of Shinarump Conglomerate fell from the Rockville Bench cliff, landed on the steep 35° slope below, and shattered into several large pieces that continued downslope at a high speed. These boulders completely destroyed a house located 375 feet below the cliff (see the before and after photographs) and killed two people inside the home. The topographic map shows other rockfalls in the area prior to this catastrophic event [14].



Figure 10.4.10: House before and after destruction from 2013 Rockville rockfall.

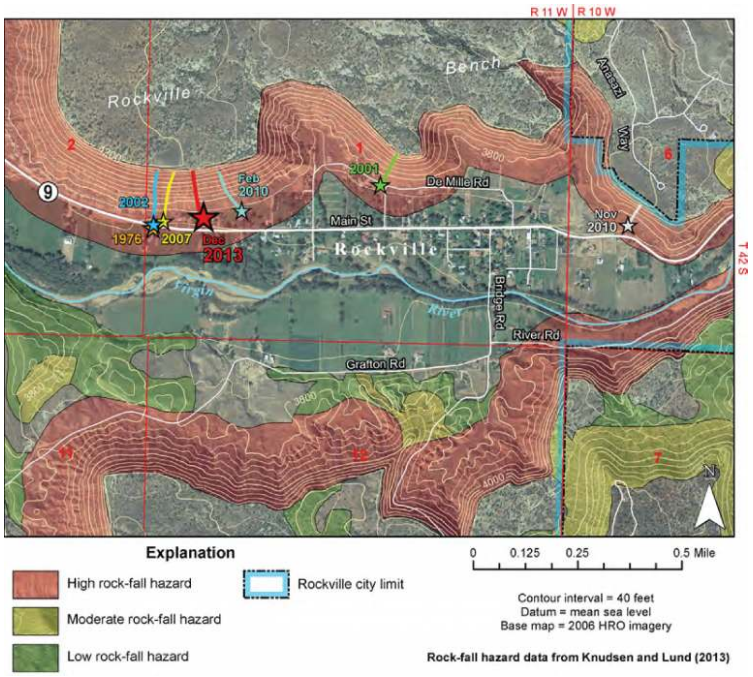


Figure 10.4.11: Tracks of deadly 2013 Rockville rockfall and earlier documented rockfall events.

2014, North Salt Lake Slide: In August 2014 after a particularly wet period, a slow-moving rotational landslide destroyed one home and damaged nearby tennis courts.



Figure 10.4.12: Scarp and displaced material from the North Salt Lake (Parkview) slide of 2014.

Reports from residents suggested that ground cracks had been seen near the top of the slope at least a year prior to the catastrophic movement. The presence of easily-drained sands and gravels overlying more impermeable clays weathered from volcanic ash, along with recent regrading of the slope, may have been contributing causes of this slide. Local heavy rains seem to have provided the trigger. In the two years after the landslide, the slope has been partially regraded to increase its stability. Unfortunately, in January 2017, parts of the slope have shown reactivation movement. Similarly, in 1996 residents in a nearby subdivision started reporting distress to their homes. This distress continued until 2012 when 18 homes became uninhabitable due to extensive damage and were removed. A geologic park was constructed in the now vacant area.





North Salt Lake Landslide

2013, Bingham Canyon Copper Mine Landslide, Utah: At 9:30 pm on April 10, 2013, more than 65 million cubic meters of steep terraced mine wall slid down into the engineered pit of Bingham Canyon mine, making it one of the largest historic landslides not associated with volcanoes. Radar systems maintained by the mine operator warned of movement of the wall, preventing the loss of life and limiting the loss of property.



Figure 10.4.13: Before (left) vs. After (right)

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10.S: Summary

Summary

Mass wasting is a geologic term describing all downhill rock and soil movement due to gravity. Mass wasting occurs when a slope is too steep to remain stable with existing material and conditions. Loose rock and soil, called regolith, are what typically move during a mass-wasting event. Slope stability is determined by two factors: the angle of the slope and the shear strength of the accumulated materials. Mass-wasting events are triggered by changes that oversteepen slope angles and weaken slope stability, such as rapid snow melt, intense rainfall, earthquake shaking, volcanic eruption, storm waves, stream erosion, and human activities. Excessive precipitation is the most common trigger. Mass-wasting events are classified by their type of movement and material, and they share common morphological surface features. The most common types of mass-wasting events are rockfalls, slides, flows, and creep.

Mass-wasting movement ranges from slow to dangerously rapid. Areas with steep topography and rapid rainfall, such as the California coast, Rocky Mountain Region, and Pacific Northwest, are particularly susceptible to hazardous mass-wasting events. By examining examples and lessons learned from famous mass-wasting events, scientists have a better understanding of how mass-wasting occurs. This knowledge has brought them closer to predicting where and how these potentially hazardous events may occur and how people can be protected.

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CHAPTER OVERVIEW

11: Water

Learning Objectives

- Describe the processes of the water cycle.
- Describe drainage basins, watershed protection, and water budget.
- Describe reasons for water laws, who controls them, and how water is shared in the western U.S.
- Describe zone of transport, zone of sediment production, zone of deposition, and equilibrium.
- Describe stream landforms: channel types, alluvial fans, floodplains, natural levees, deltas, entrenched meanders, and terraces.
- Describe the properties required for a good aquifer; define confining layer water table.
- Describe three major groups of water contamination and three types of remediation.
- Describe karst topography, how it is created, and the landforms that characterize it.



Figure 11.1: Entrenched meander of the Colorado River, downstream of Page, Arizona.

All life on Earth requires water. The **hydrosphere** (Earth's water) is an important agent of geologic change. Water shapes our planet through weathering and erosion, deposits minerals that aid in lithification, and alters rocks after they are lithified. Water carried by subducted oceanic plates causes flux melting in the upper mantle material. Water is among the volatiles in magma and emerges at the surface as steam in volcanoes.



Figure 11.2: Example of a Roman aqueduct in Segovia, Spain.

Humans rely on suitable water sources for consumption, power generation, agriculture, and many other puposes. In pre-industrial civilizations, control of water resources was a symbol of power [1; 2]. Two thousand-year-old Roman aqueducts still grace European, Middle Eastern, and North African skylines. Ancient Mayan architecture depicts water imagery such as frogs, water lilies, and waterfowl to illustrate the importance of water in their societies [3]. In the drier lowlands of the Yucatan Peninsula, mask facades of the hooked-nosed rain god, *Chac* (or *Chaac)* are prominent on Mayan buildings such as the Kodz Poop (Temple of the Masks, sometimes spelled Coodz Poop) at the ceremonial site of Kabah. To this day, government control over water continues to be an integral part of most modern societies.



Figure 11.3: Chac mask in Mexico.

11.1: Water Cycle
11.2: Water Basin and Budgets
11.3: Water Use and Distribution
11.4: Water Law
11.5: Surface Water
11.6: Groundwater
11.7: Water Contamination
11.8: Karst
11.S: Summary

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11.1: Water Cycle

The **water cycle** is the continuous circulation of water in the Earth's atmosphere. During circulation, water changes between solid, liquid, and gas (water vapor) and changes location. The processes involved in the water cycle are evaporation, transpiration, condensation, precipitation, and runoff.

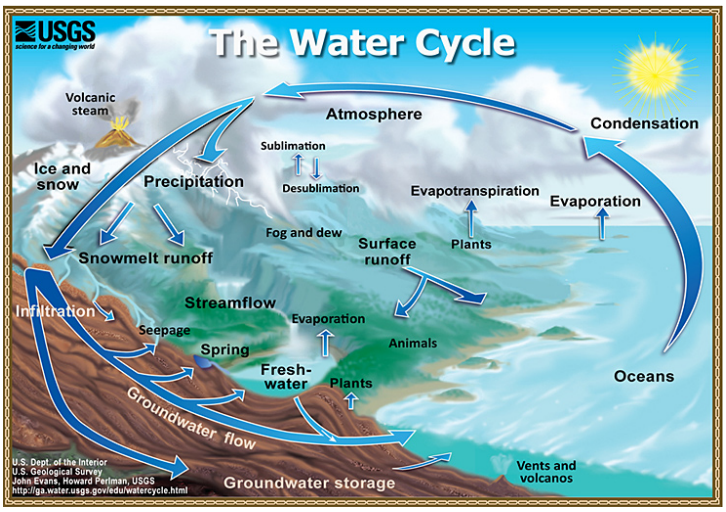


Figure 11.1.1: The water cycle.

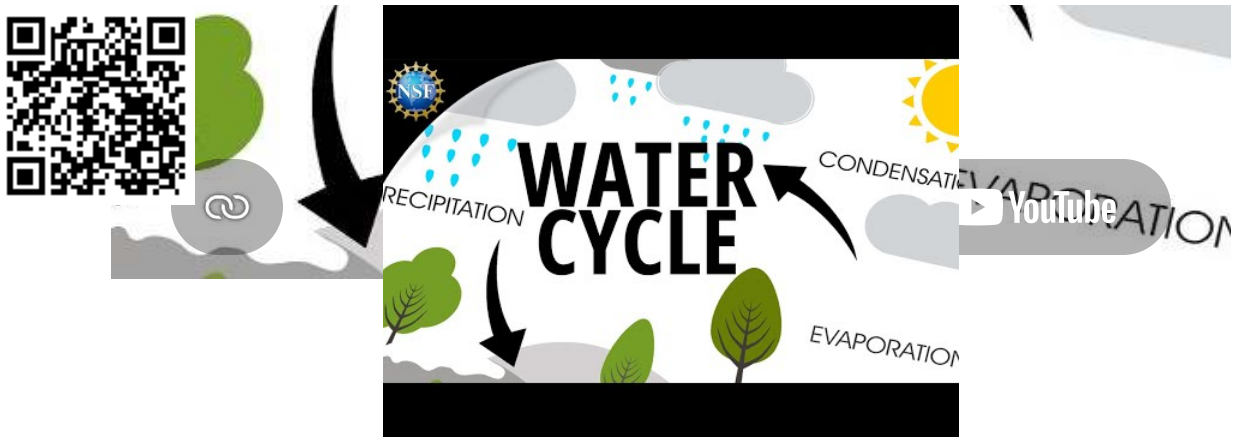
Evaporation is the process by which a liquid is converted to a gas. Water evaporates when solar energy warms the water sufficiently to excite the water molecules to the point of vaporization. Evaporation occurs from oceans, lakes, streams and the land surface. Plants contribute significant amounts of water vapor as a byproduct of photosynthesis in a process called **transpiration** that occurs through the minute pores of plant leaves. The term **evapotranspiration** refers to these two sources of water entering the atmosphere and is commonly used by geologists.

Water vapor is invisible. **Condensation** is the process of water vapor transitioning to a liquid. Winds carry water vapor in the atmosphere long distances. When water vapor cools or when air masses of different temperatures mix, water vapor may condense back into droplets of liquid water. These water droplets usually form around a microscopic piece of dust or salt called **condensation nuclei**. These small droplets of liquid water suspended in the atmosphere become visible as in a cloud. Water droplets inside clouds collide and stick together, growing into larger droplets. Once the water droplets become big enough, they fall to Earth as **precipitation** such as rain, snow, hail, or sleet.

Once precipitation has reached the Earth's surface, it can evaporate or flow as **runoff** into streams, lakes, and eventually back to the oceans. Water in streams and lakes is called **surface water**. In addition, water can also **infiltrate** into the soil and fill the pore spaces in the rock or sediment underground to become **groundwater**, the name given to all subsurface water. Groundwater slowly moves through rock and unconsolidated materials. Some groundwater may reach the surface again, where it discharges as springs and into streams, lakes, and the ocean. Also, surface water in streams and lakes can infiltrate again to recharge groundwater. Therefore, the surface water and groundwater systems are connected.

Watch a video on the water cycle.





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11.2: Water Basin and Budgets

Drainage Basins

The basic unit of division of the landscape is the drainage basin. A **drainage basin**, also known as a **catchment** or **watershed**, is the area of land that captures precipitation and contributes runoff to a stream or stream segment. **Drainage divides** are local topographic high points that separate one drainage basin from another [5]. Water that falls on one side of the divide goes to one stream, and water that falls on the other side of the divide goes to a different stream. Each stream, tributary and streamlet has its own drainage basin. In areas with flatter topography, drainage divides are not as easily identified but they still exist [6].

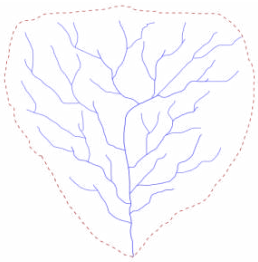


Figure 11.2.1: Map view of a drainage basin with main trunk streams and many tributaries with drainage divide in the dashed red line.

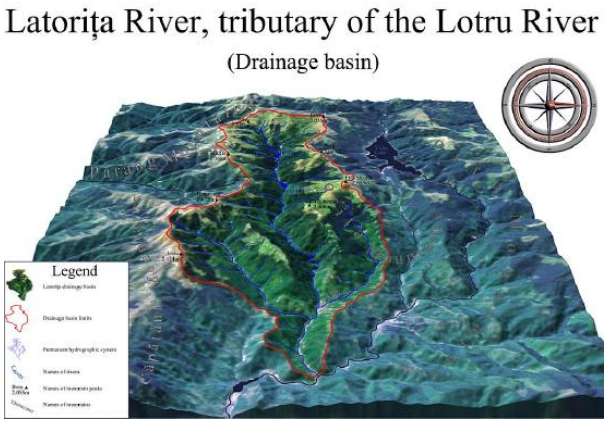


Figure 11.2.2: Oblique view of the drainage basin and divide of the Latorita River, Romania.

The **headwater** is where the stream begins. Streams only flow downhill and smaller tributary streams combine downhill to make the larger **trunk** of the stream. The **mouth** is where the stream finally reaches its end. The mouth of most streams is at the ocean. However, a rare number of streams do not flow to the ocean, but rather end in a **closed basin** (or **endorheic basin**) where the water evaporates from a stream or lake before it can reach the ocean. Most streams in the Great Basin of western North America are in closed basins. For example, in Utah, the Little Cottonwood Creek and the Jordan River flow into the Great Salt Lake where the water evaporates.

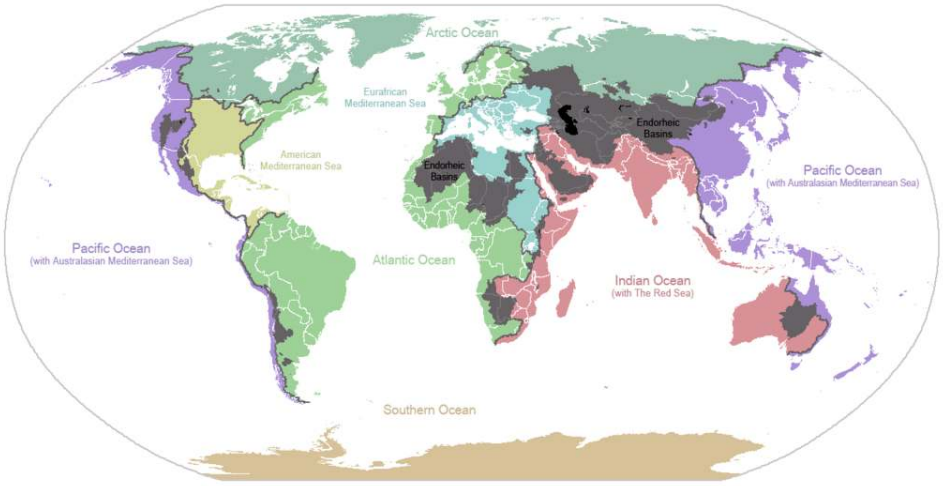


Figure 11.2.3: Major drainage basins color-coded to match the related ocean. Closed basins (or endorheic basins) are shown in gray.

Perennial streams flow all year round. Perennial streams occur in humid or temperate climates where there is sufficient rainfall and low evaporation rates. Water levels rise and fall with the seasons, depending on the discharge. **Ephemeral streams** flow only during rain events or the wet season. In arid climates, like Utah, many streams are ephemeral. These streams occur in dry climates with low amounts of rainfall and high evaporation rates. Their channels are often dry washes or **arroyos** for much of the year and their sudden flow causes flash floods.

Along Utah's Wasatch Front, the urban area extending north to south from Brigham City to Provo, there are several watersheds that are designated as "watershed protection areas" that limit the type of use allowed in those drainages in order to protect culinary water. Dogs and swimming are limited in those watersheds because of the possibility of contamination by harmful bacteria and substances to the drinking supply of Salt Lake City and surrounding municipalities.

Water Budgets

Water in the water cycle is very much like money in a personal budget. Income includes precipitation and stream and groundwater inflow. Expenses include groundwater withdrawal, evaporation, and stream and groundwater outflow. If the expenses outweigh the income, the **water budget** is not balanced. In this case, water is removed from savings, i.e. water storage, if available. Reservoirs, snow, ice, soil moisture, and aquifers all serve as storage in a water budget. In dry regions, water is critical for sustaining human activities. Understanding and managing the water budget is an ongoing political and social challenge.

Hydrologists can make groundwater budgets within any designated boundary, but they are generally made for watershed (basin) boundaries because groundwater and surface water are easier to account for within these boundaries. However, water budgets can be created for state, county, or aquifer extent boundaries as well. The groundwater budget is an essential component of the hydrologic model; hydrologists use measured data with a conceptual workflow of the model to better understand the water system.

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11.3: Water Use and Distribution

In the United States, 355 billion gallons (1344 billion liters) of ground and surface water are used each day, of which 76 billion gallons (288 billion liters) are fresh groundwater. The state of California accounts for 16% of national groundwater [8].

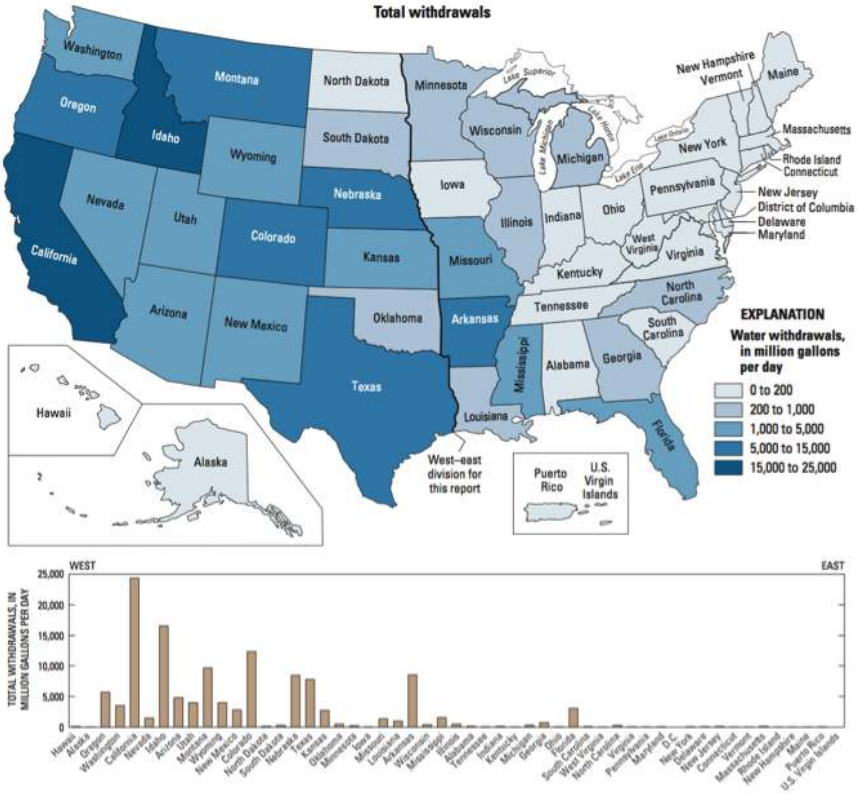


Figure 11.3.1: Agricultural water use in the United States by state.

Utah is the second driest state in the United States behind its neighbor Nevada, having a mean statewide precipitation of 12.2 inches (31 cm) per year. Utah also has the second-highest per capita rate of total domestic water use [8] of 167 gallons (632.16 liters) per day per person. With the combination of relatively high demand and limited quantity, Utah is at risk for water budget deficits.

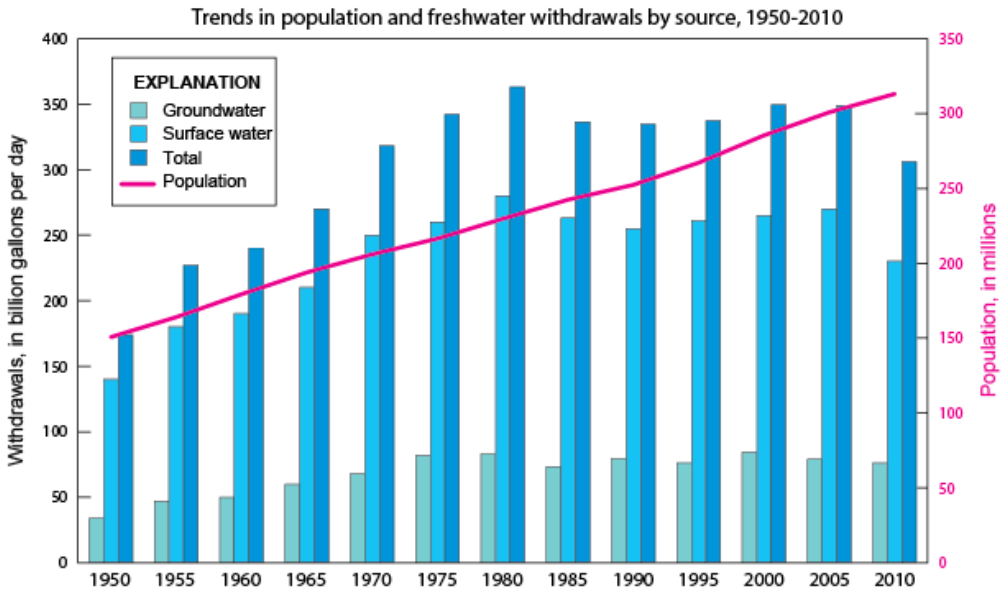


Figure 11.3.2: Trends in water use by source.

Surface Water Distribution

Fresh water is a precious resource and should not be taken for granted, especially in dry climates. Surface water makes up only 1.2% of the fresh water available on the planet, and 69% of that surface water is trapped in ground ice and permafrost. Stream water accounts for only 0.006% of all fresh water and lakes contain only 0.26% of the world's fresh water [9].

Global circulation patterns are the most important factor in distributing surface water through precipitation. Due to the Coriolis effect and the uneven heating of the Earth, air rises near the equator and near 60° north and south latitude. Air sinks at the poles and 30° north and south latitude. Land masses near rising air are more prone to humid and wet climates. Land masses near sinking air, which inhibits precipitation, are prone to dry conditions [10; 11]. Prevailing winds, ocean circulation patterns such as the Gulf Stream's effects on eastern North America, rain shadows (dry leeward sides of mountains), and even the proximity of bodies of water can affect local climate patterns. For example, when cold winds blow across the relatively warm Great Salt Lake, the air warms, which causes it to pick up moisture. This local increase in the moisture content of the air may eventually fall as snow or rain on nearby mountains, a phenomenon known as "**lake-effect precipitation**" [12].

In the United States, the **100th Meridian** roughly marks the boundary between the humid and arid parts of the country. Growing crops west of the 100th Meridian requires irrigation [13]. In the west, surface water is stored in reservoirs and mountain snowpacks [14], then strategically released through a system of canals during times of high use.

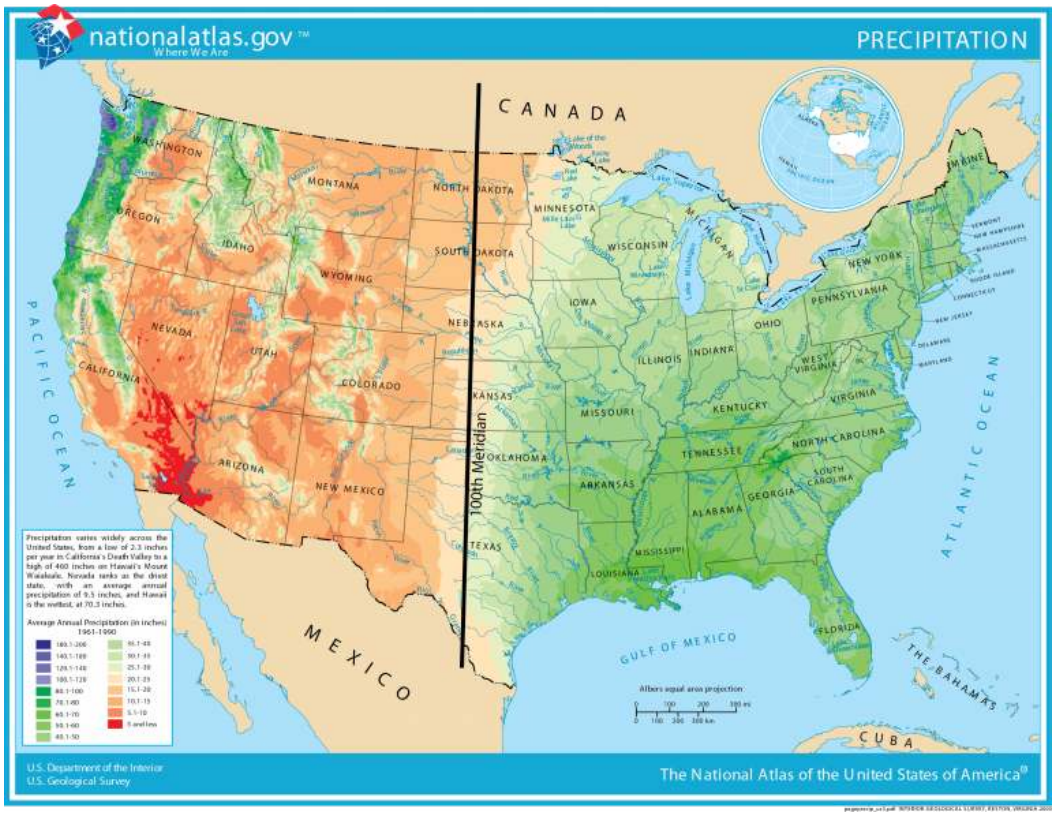


Figure 11.3.3: Distribution of precipitation in the United States. The 100th Meridian is approximately where the average precipitation transitions from relatively wet to dry. (Source: U.S. Geological Survey)

Some of the driest parts of the western United States are in the Basin and Range Province. The Basin and Range has multiple mountain ranges that are oriented north to south. Most of the basin valleys in the Basin and Range are dry, receiving less than 12 inches (30 cm) of precipitation per year. However, some of the mountain ranges can receive more than 60 inches (1.52 m) of water as snow or snow-water-equivalent. The **snow-water equivalent** is the amount of water that would result if the snow were melted, as the snowpack is generally much thicker than the equivalent amount of water that it would produce [12].

Groundwater Distribution

Groundwater makes up 30.1% of the fresh water on the planet, making it the most abundant reservoir of fresh water accessible to most humans. The majority of fresh water, 68.7%, is stored in glaciers and ice caps as ice [9]. As the glaciers and ice caps melt due to global warming, this fresh water is lost as it flows into the oceans.

Table 11.3.1: Groundwater distribution on Earth.

Water source	Water volume (cubic miles)	Freshwater (%)	Total water (%)
Oceans, seas, & bays	321,000,000	—	96.5%
Ice caps, glaciers, & permanent snow	5,773,000	68.7%	1.74%
Groundwater (total)	5,614,000	—	1.69%
Groundwater (fresh)	2,526,000	30.1%	0.76%
Groundwater (saline)	3,088,000	—	0.93%
Soil moisture	3,959	0.05%	0.001%
Ground ice & permafrost	71,970	0.86%	0.022%
Lakes (total)	42,320	—	0.013%
Lakes (fresh)	21,830	0.26%	0.007%
Lakes (saline)	20,490	—	0.006%
Atmosphere	3,095	0.04%	0.001%
Swamp water	2,752	0.03%	0.0008%
Rivers	509	0.006%	0.0002%
Biological water	269	0.003%	0.0001%

Source: Igor Shiklomanov's chapter "World freshwater resources" in Peter H. Gleick (editor), 1993, Water in Crisis: A Guide to the World's Fresh Water Resources (Oxford University Press, New York).

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11.4: Water Law

Federal and state governments have put laws in place to ensure the fair and equitable use of water. In the United States, the states are tasked with creating a fair and legal system for sharing water.

Water Rights

Because of the limited supply of water, especially in the western United States, states disperse a system of legal **water rights** defined as a claim to a portion or all of a water source, such as a spring, stream, well, or lake. Federal law mandates that states control water rights, with the special exception of federally reserved water rights, such as those associated with national parks and Native American tribes, and **navigational servitude**, which maintains navigable water bodies. Each state in the United States has a different way to disperse and manage water rights.

A person or entity (company, organization, etc.) must have a water right to legally extract or use surface or groundwater in their state. Water rights in some western states are dictated by the concept of **prior appropriation**, or "first in time, first in right," where the person with the oldest water right gets priority water use during times when there is not enough water to fulfill every water right.

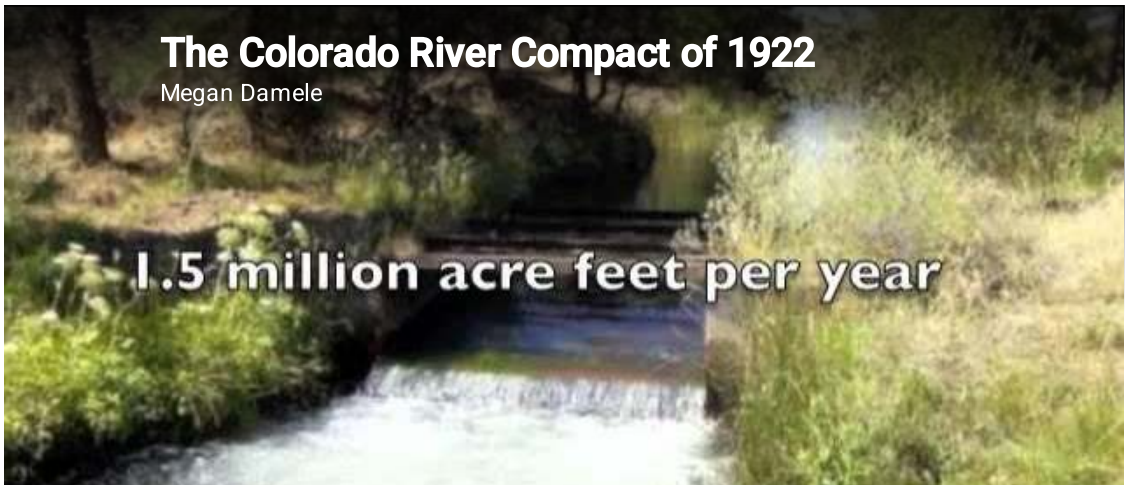
The Law of the River and the Colorado River Compact

The Colorado River and its tributaries pass through a desert region, including seven states (Wyoming, Colorado, Utah, New Mexico, Arizona, Nevada, California), Native American reservations, and Mexico. As the western United States became more populated and while California was becoming a key agricultural producer, the states along the Colorado River realized that the river was important to sustaining life in the West.

The states recognized a water budget was necessary for the Colorado River Basin to guarantee each state's certain perceived water rights. Thus was enacted the **Colorado River Compact** in 1922 to ensure that each state got a fair share of the river water. The Compact granted each state a specific volume of water based on the total measured flow at the time. However, in 1922, the flow of the river was higher than its long-term average flow, consequently, more water was allocated to each state than is typically available in the river.

Over the next several decades, many other agreements and modifications would follow the Colorado River Compact, including agreements that brought about the Hoover Dam (formerly Boulder Dam) and Glen Canyon Dam, and a treaty between the American and Mexican governments. Combined, the agreements became known as "**The Law of the River**" by the United States Bureau of Reclamation. Despite adjustments to the Colorado River Compact, many believe that over-allocation is still prevalent, as the Colorado River no longer reaches the Pacific Ocean, its original terminus (base level). Dams causing diversion and evaporation of Colorado River water have resulted in serious water budget concerns in the Colorado River basin. Predicted drought associated with global warming causes additional concerns about future over-allocation of Colorado River flow.

The Law of the River highlights the complex and prolonged nature of interstate water rights agreements, as well as the importance of water. Watch a video on the Colorado River Compact.





Snake Valley

The Snake Valley straddles the border of Utah and Nevada with more of the irrigable land area lying on the Utah side of the border. In 1989, the Southern Nevada Water Authority (SNWA) submitted applications for water rights to pipe up to 155,000 acre-feet (191,189,707 cubic meters) of water per year (an acre-foot of water is one acre covered with water one foot deep) from Spring, Snake, Delamar, Dry Lake, and Cave valleys to southern Nevada, mostly for Las Vegas [17]. Nevada and Utah attempted a comprehensive agreement, but negotiations have yet to be settled.

Check out these additional resources:

A story map of the groundwater issues of Utah

NPR story on Snake Valley



Howard Berkes, NPR

npr

Share

Struggling over Water

Las Vegas Water Battle: 'Crops vs. Craps'

Listen

© 2007 NPR

Dean Baker Story

Water Quality and Protection

Two major federal laws that protect water quality in the United States are the Clean Water Act and the Safe Drinking Water Act. The **Clean Water Act**, an amendment of the Federal Water Pollution Control Act, protects navigable waters from dumping and point-source pollution. The **Safe Drinking Water Act** ensures that water that is provided by public water suppliers, like cities and towns, is safe to drink [18].

The U.S. **Environmental Protection Agency Superfund program** ensures the cleanup of hazardous contamination, and can be applied to situations of surface water and groundwater contamination. It is part of the Comprehensive Environmental Response, Compensation, and Liability Act of 1980. Under this act, state governments and the U.S. Environmental Protection Agency can use the superfund to pay for remediation of a contaminated site and then file a lawsuit against the polluter to recoup the costs. Or to avoid being sued, the polluter that caused the contamination may take direct action or provide funds to remediate the contamination.

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11.5: Surface Water

Geologically, a **stream** is a body of flowing surface water confined to a channel. Terms such as river, creek and brook are social terms not used in geology. Streams erode and transport sediments, making them the most important agents of the Earth's surface, along with wave action, in eroding and transporting sediments. They create much of the surface topography and are an important water resource.

Several factors cause streams to erode and transport sediment, but the two main factors are stream–channel gradient and velocity. **Stream–channel gradient** is the slope of the stream usually expressed in meters per kilometer or feet per mile. A steeper channel gradient promotes erosion. When tectonic forces elevate a mountain, the stream gradient increases, causing the mountain stream to erode downward and deepen its channel, eventually forming a valley. **Stream–channel velocity** is the speed at which channel water flows. Factors affecting channel velocity include channel gradient which decreases downstream, discharge and channel size which increase as tributaries coalesce, and channel roughness which decreases as sediment lining the channel walls decreases in size thus reducing friction. The combined effect of these factors is that channel velocity actually increases from mountain headwaters to the mouth of the stream.

Discharge

Stream size is measured in terms of **discharge,** the volume of water flowing past a point in the stream over a defined time interval. Volume is commonly measured in cubic units (length x width x depth), shown as feet^3 (ft^3) or meter^3 (m^3). Therefore, the units of discharge are cubic feet per second (ft^3/sec or cfs) or cubic meters per second (m^3/s or cms). Smaller streams have less discharge than larger streams. For example, the Mississippi River is the largest river in North America, with an average flow of about 600,000 cfs (16,990.11 cms) [19]. For comparison, the average discharge for the Jordan River at Utah Lake is about 574 cfs (16.25 cms) [20] and for the Amazon River (the world's largest river), annual discharge is about 6,200,000 cfs (175,565 cms) [21].

Discharge can be expressed by the following equation:

Q = V A

- Q = discharge (ft^3/sec or cms),
- A = cross-sectional area of the stream channel [width times average depth] (in ft^2, in^2, or m^2)
- V = average channel velocity (ft/sec or m/s).

At a given location along the stream, velocity varies with stream width, shape, and depth within the stream channel as well. When the stream channel narrows but discharge remains constant, the same volume of water flows through a narrower space causing the velocity to increase, similar to putting a thumb over the end of a backyard water hose. In addition, during rain storms or heavy snow melt, runoff increases, which increases stream discharge and thus velocity.

Velocity varies within the stream channel as well. Generally, when the channel is straight and uniformly deep, the highest velocity is in the center of the channel along the top of the water where it is the farthest from frictional contact with the stream channel bottom and sides. When the channel curves, the highest velocity will be on the outside of the bend. In hydrology, the **thalweg** of a stream is the line drawn that shows its natural progression and deepest channel, as is shown in the diagram.

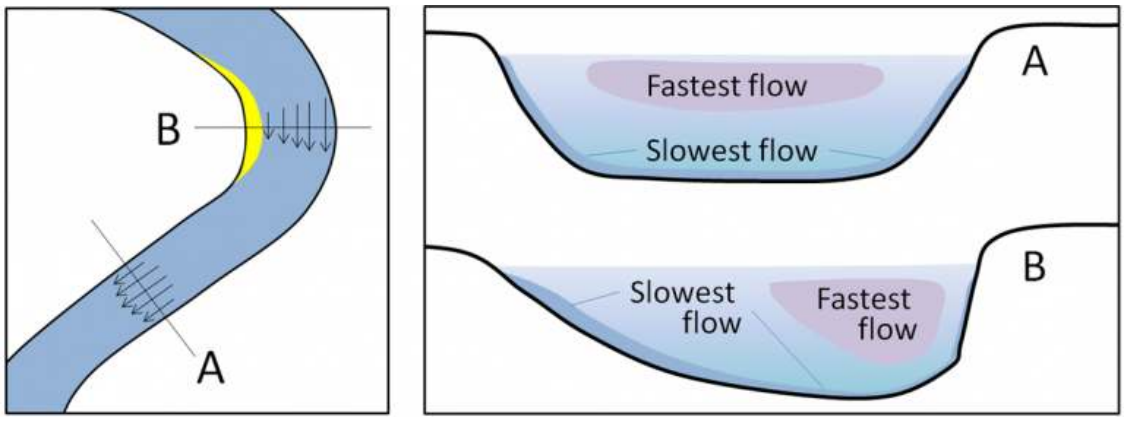


Figure 11.5.1: Left: Thalweg of a stream. In a stream bend (B), the fastest moving particles are on the outside of the bend, near the cutbank. Stream velocity is higher on the outside bend and the surface which is farthest from the friction of the stream bed (bottom right). Longer arrows indicate faster velocity (By Steven Earle 2015; CC BY 4.0 via Physical Geology).

Runoff vs. Infiltration

There are many factors dictating whether water will infiltrate into the ground or run off over the land after precipitation. These include but are not limited to the amount, type, and intensity of precipitation, the type and amount of vegetation cover, the slope of the land, the temperature and aspect of the land, preexisting conditions, and the type of soil in the area of infiltration. High-intensity rain will cause more runoff than the same amount of rain spread out over a longer duration. If the rain falls faster than the properties of the soil allow it to infiltrate, then the water that cannot infiltrate becomes runoff. Dense vegetation can increase infiltration, as the vegetative cover slows the overland flow of water particles, giving them more time to infiltrate. If a parcel of land has more direct solar radiation and/or higher seasonal temperatures, there will likely be less infiltration and runoff, as evapotranspiration rates will be higher. As the slope of the land increases, so does runoff, as the water is more inclined to move downslope than infiltrate into the ground. Extreme examples are a basin and a cliff, where water infiltrates much quicker into a basin than a cliff having the same soil properties. Because saturated soil does not have the capacity to take more water, runoff is generally greater over saturated soil. Clay-rich soil cannot accept infiltration as quickly as gravel-rich soil.

Drainage Patterns

The pattern of tributaries within a region is called a **drainage pattern**. They depend largely on the type of rock beneath, and on structures within that rock (such as folds and faults). The main types of drainage patterns are dendritic, trellis, rectangular, radial, and deranged. **Dendritic patterns** are the most common and develop in areas where the underlying rock or sediments are uniform in character, mostly flat-lying, and can be eroded equally easily in all directions. Examples are alluvial sediments or flat-lying sedimentary rocks. **Trellis patterns** typically develop where sedimentary rocks have been folded or tilted and then eroded to varying degrees depending on their strength. The Appalachian Mountains in the eastern United States have many good examples of trellis drainage. **Rectangular patterns** develop in areas that have very little topography and a system of bedding planes, joints, or faults that form a rectangular network. A **radial pattern** forms when streams flow away from a central high point such as a mountain top or volcano, with the individual streams typically having dendritic drainage patterns. In places with extensive limestone deposits, streams can disappear into the ground via caves and subterranean drainage and this creates a **deranged pattern**.

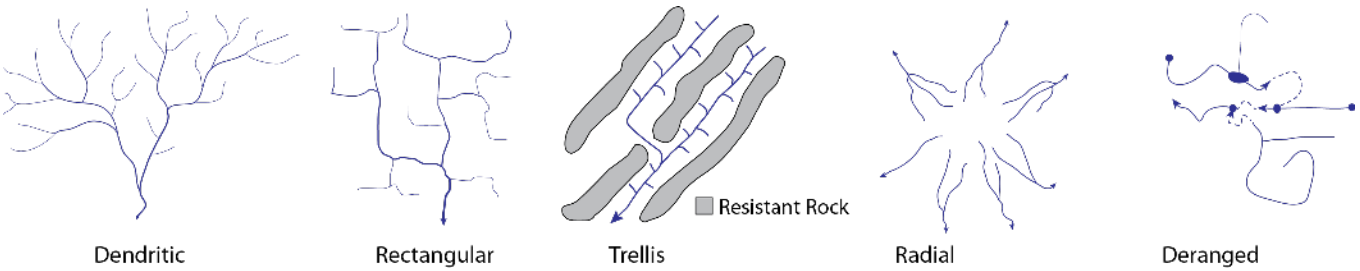


Figure 11.5.2: Various stream drainage patterns.

Fluvial Processes

Fluvial processes are the mechanisms that dictate how a stream functions and include factors controlling fluvial sediment production, transportation, and deposition. Fluvial processes include velocity, slope and gradient, erosion, transportation, deposition, stream equilibrium, and base level.

Streams can be divided into three main sections: the many smaller **tributaries** in the source area, the main trunk stream in the floodplain and the **distributaries** at the mouth of the stream. Major stream systems like the Mississippi are composed of many source areas, many tributaries and trunk streams, all coalescing into the one main stream draining the region. The zones of a stream are defined as zones of sediment production (erosion), transportation, and deposition. The zone of sediment production is located in the headwaters of the stream. In the zone of sediment transportation, there is a general balance between erosion of the finer sediment in its channel and transportation of sediment across the floodplain. Streams eventually flow into the ocean or end in quiet water with a delta which is a zone of sediment deposition located at the mouth of a stream. [6]. The longitudinal profile of a stream is a plot of the elevation of the stream channel at all points along its course and illustrates the location of the three zones [22].

Zone of Sediment Production (Erosion)

The **zone of sediment production** is located in the headwaters of a stream where rills and gullies erode sediment and contribute to larger tributary streams. These tributaries carry sediment and water further downstream to the main trunk of the stream. Tributaries at the headwaters have the steepest gradient, and erosion there produces considerable sediment carried by the stream. Headwater streams tend to be narrow and straight with small or non-existent floodplains adjacent to the channel. Since the zone of sediment production is generally the steepest part of the stream, many headwaters are located in relatively high elevations. For example, the Rocky Mountains of Wyoming and Colorado west of the Continental Divide contain much of the headwaters for the Colorado River, which then flows from Colorado through Utah, Arizona, to Mexico. Headwaters of the Mississippi River system lie east of the Continental Divide in the Rocky Mountains and west of the Appalachian Divide.

Zone of Sediment Transportation

Streams transport sediment great distances from the headwaters to the ocean, the ultimate depositional basins. **Sediment transportation** is directly related to stream gradient and velocity. Faster and steeper streams can transport larger sediment grains. When velocity slows down, larger sediments settle to the channel bottom. When the velocity increases, those larger sediments are entrained and move again.

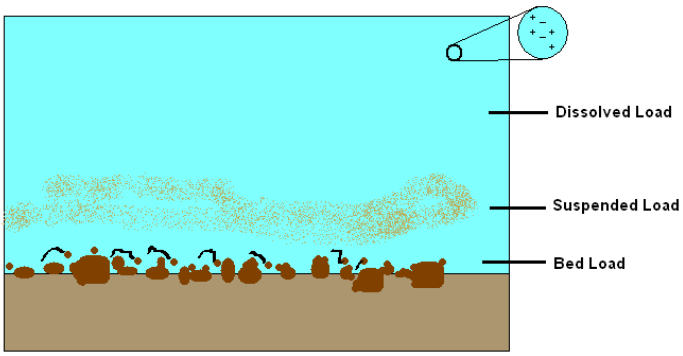


Figure 11.5.3: A stream carries dissolved load, suspended load, and bedload.

Transported sediments are grouped into bedload, suspended load, and dissolved load. Sediments moved along the channel bed are the **bedload** and typically are the largest and densest particles. Bedload is moved by **saltation** (bouncing) and **traction** (being pushed or rolled along by the force of the flow). When stream velocity increases, smaller particles can be picked up by flowing water and carried in suspension as **suspended load**. The particle size that is carried as suspended load and bedload depends on the flow velocity of the stream. **Dissolved load** in a stream is the total of the ions in solution from chemical weathering. The dissolved load includes ions such as bicarbonate (HCO_3^-), calcium (Ca^{2+}), chloride (Cl^-), potassium (K^+), and sodium (Na^+). The solubility of these ions is not affected by flow velocity.

Watch a video on bed load sediment transport. (https://youtu.be/is-qcxrKKBI)

Stream flooding is a natural process that adds sediment to floodplains. A **floodplain** is the generally flat area of land located adjacent to a stream channel that is inundated with floodwater on a regular basis. A stream typically reaches its greatest velocity when it is close to flooding, known as the **bankfull stage**. As soon as the flooding stream overtops its banks and flows onto its floodplain, the velocity decreases. At this point, sediment that was being carried by the swiftly moving water is deposited near the edge of the channel, forming a low ridge or **natural levee.** In addition, sediments are added to the floodplain during this flooding process contributing to fertile soils.

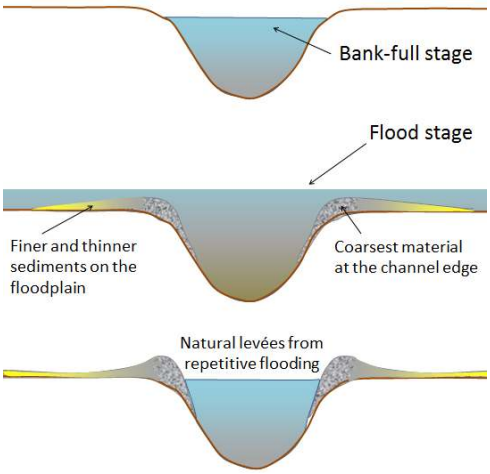


Figure 11.5.4: Profile of stream channel at bankfull stage, flood stage, and deposition of natural levee.

Zone of Deposition

Deposition occurs when bedload and suspended load come to rest on the bottom of the water column in a stream channel, lake, or ocean. The two major factors causing deposition are the decrease in stream gradient and the reduction in velocity. These can be associated with a decrease in discharge or increased in cross-sectional area. Deposition occurs temporarily in the zone of transportation such as along meandering stream point bars, floodplains, and alluvial fans, however, ultimate deposition occurs at the mouth of the stream where it reaches a lake or ocean. These deposits at the mouth of a stream form landforms called **deltas**. Deposition at the mouth of a stream is generally of the finest sediment such as fine sand, silt, and clay, because as the stream exits its channel, the energy of the water is completely dispersed, causing the deposition of all particles in the stream.

Equilibrium and Base Level

All three stream zones are present in the typical **longitudinal profile** of a stream which plots the elevation of the channel at all points along its course. All streams have a longitudinal profile, some of which have been measured, plotted, and published. The longitudinal profile shows the stream gradient from headwater to mouth and represents the balance among erosion, transportation, gradient, velocity, discharge, and channel characteristics at each point along the stream's course. This balance is called **equilibrium**. When mountains are uplifted, streams become steeper which erodes downward, cutting a valley. This uplift is balanced against downward erosion of the stream. Eventually, streams erode enough downward that the gradient is reduced, downward erosion slows, and the stream starts to erode from side to side. This point is generally characterized by a stream with a floodplain [6].

Another factor influencing equilibrium is the **base level**, the elevation of the stream's mouth. The base level represents the lowest level to which a stream can erode. The ultimate base level is, of course, sea level. A lake or reservoir may also represent the base level for a stream entering it. The Great Basin of western Utah, Nevada, and parts of some surrounding states contains no outlets to the sea and provides internal base levels for streams within it. The base level for a stream entering the ocean can change if sea level rises or falls or if a natural or human-made dam is added along its profile. When base level is lowered, a stream will downcut and deepen its channel, perhaps into a canyon. When base level is raised, deposition increases along the stream profile as the stream adjusts to the change and establishes a new state of equilibrium. A stream that has approximately achieved equilibrium is called a **graded stream**.

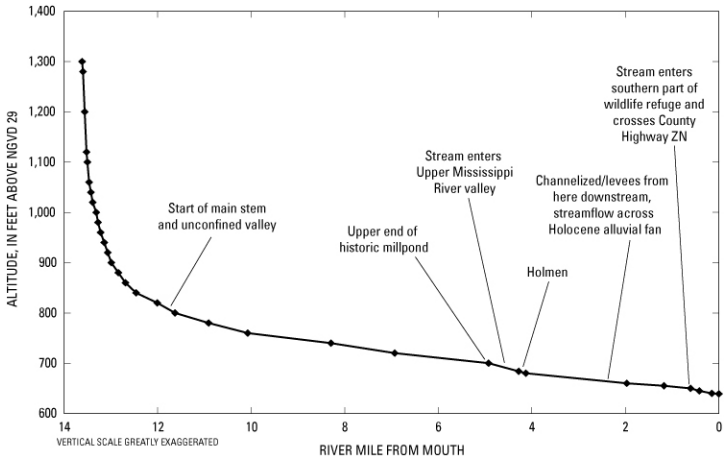


Figure 11.5.5: Example of a longitudinal profile of a stream; Halfway Creek, Indiana.

Fluvial Landforms

Fluvial landforms are the land features formed by a stream on the surface of Earth by either erosion or deposition. The stream-related landforms described here are primarily related to channel types.

Channel Types

Stream channels can be straight, braided, meandering, or entrenched. The gradient, sediment load, discharge, and location of the base level all influence channel type. **Straight channels** are relatively straight, located near the headwaters, have steep gradients, low discharge, and narrow V-shaped valleys. Good examples of these are located in mountainous areas.

Braided streams have multiple channels splitting and recombining around numerous mid-channel bars. These are found in floodplains with low gradients near coarse sediment source areas such as mountains or in front of glaciers.



Figure 11.5.6: Braided stream pattern on the Waimakariri River in New Zealand.

Meandering streams are composed of a single channel that curves back and forth like a snake within its floodplain. Meandering streams tend to have a wide floodplain, high discharge, natural levees, and flood regularly. Meandering streams are usually located on low gradient slopes where the stream emerges from its headwaters into the zone of transportation and extends close to the zone of deposition at the stream's mouth.



Figure 11.5.7: Air photo of the meandering stream, Río Cauto, Cuba.

An **entrenched channel** occurs when a meandering channel rapidly down cuts due to a drop in base level. This causes the original meandering shape to be preserved within a deeply entrenched channel. This channel type is rare worldwide but is common in the Colorado Plateau region which is a broad flat area near four-corners where Utah, Colorado, Arizona, and New Mexico meet. For example, the Green, Colorado, and San Juan Rivers famously form entrenched channels.

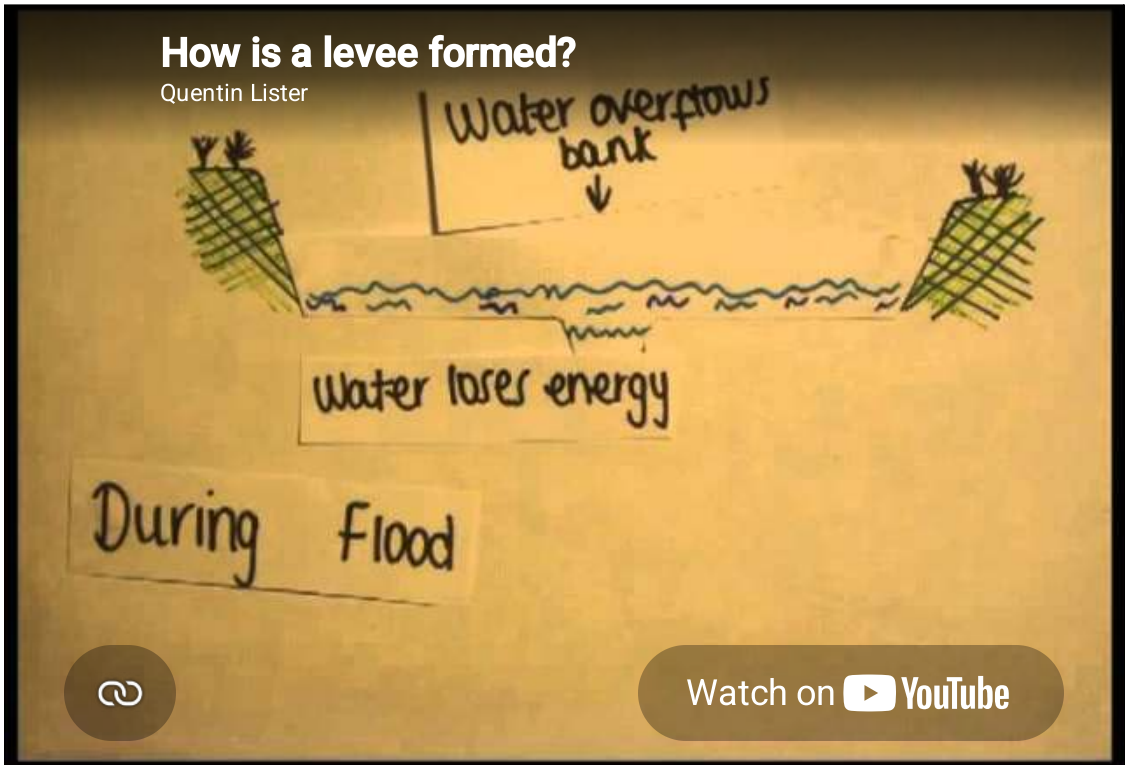


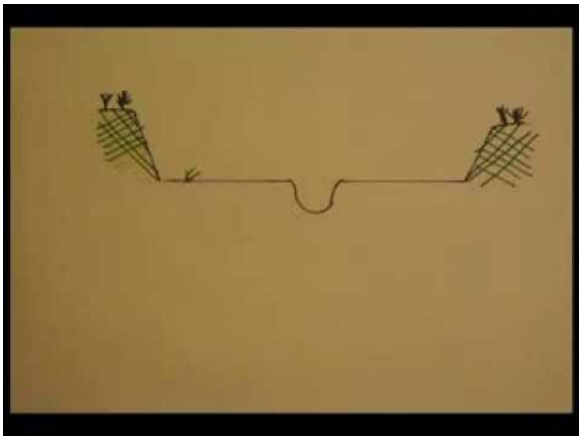
Figure 11.5.8: The San Juan River has entrenched meanders into the flat Colorado Plateau at Goosenecks State Park, southeastern Utah.

Floodplains, Natural Levees and Meandering Levels

Many fluvial landforms occur in a floodplain near a meandering stream. A floodplain is the broad, mostly flat area next to a meandering stream that is regularly flooded. A stream creates its floodplain as the channel meanders back and forth over thousands, even millions of years. Regular flooding contributes to creating the floodplain by eroding **uplands** next to the floodplain. The stream channels are confined by small natural levees that have been built up over many years of regular flooding. Natural levees can isolate and direct flow from tributary channels from immediately reaching the main channel on the floodplain. The smaller isolated streams, called **yazoo streams**, will flow parallel to the main trunk stream until there is an opening in the levee to allow for a belated confluence [23].

Watch a video on how a levee is formed.





The location and width of floodplains naturally vary, however, humans build artificial levees on flood plains to limit flooding. Sediment that breaches the levees during a flood stage is called a **crevasse splay.** The floodwaters deliver silt and clay into the floodplain. These deposits are nutrient-rich and thus often make good farmland. Floodplains are also easy to build on due to their flat nature, however, when floodwaters crest over human-made levees, the levees quickly erode with potentially catastrophic impacts. Because of the good soils, farmers regularly return after floods and rebuild year after year.

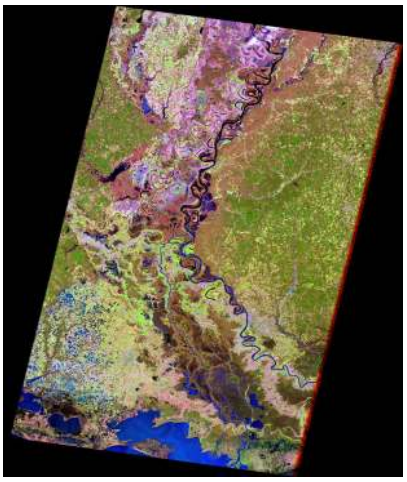


Figure 11.5.9: Landsat image of lower Mississippi River floodplain.

Meandering streams create additional landforms as the channel migrates within the floodplain. Meandering streams erode side-to-side because the highest velocity water having the most capacity to erode is located on the outside of the bend. Erosion of the outside of the bend of a stream channel creates a **cut bank** and the meander extends its loop by this erosion. The thalweg of the stream is the deepest part of the stream channel. In the straight parts of the channel, the thalweg and highest velocity are in the center of the channel. But at the bend of a meandering stream, the thalweg moves to the cut bank. Opposite the cut bank on the inside bend of the channel is the lowest stream velocity and therefore creates an area of deposition called a **point bar**.



Figure 11.5.10: Point bar and cut bank on the Cirque de la Madeleine in France.

Through erosion on the outsides of the meanders and deposition on the insides, the channels of meandering streams move back and forth across their floodplain over time. Sometimes on very broad floodplains with very low gradients, the meander bends can become so extreme that they cut across themselves at a narrow neck called a **cutoff**. The former channel becomes isolated and forms an **oxbow lake**. Eventually, the oxbow lake fills in with sediment and becomes a wetland and eventually a **meander scar**. Stream meanders can migrate and form oxbow lakes in a relatively short amount of time. Where stream channels form geographic and political boundaries, this shifting of channels can cause conflicts.



Figure 11.5.11: Meander nearing cutoff on the Nowitna River in Alaska. Note the oxbow lake in the right of the photo.

The Rincon on the Colorado River at Lake Powell is an entrenched meander cutoff.

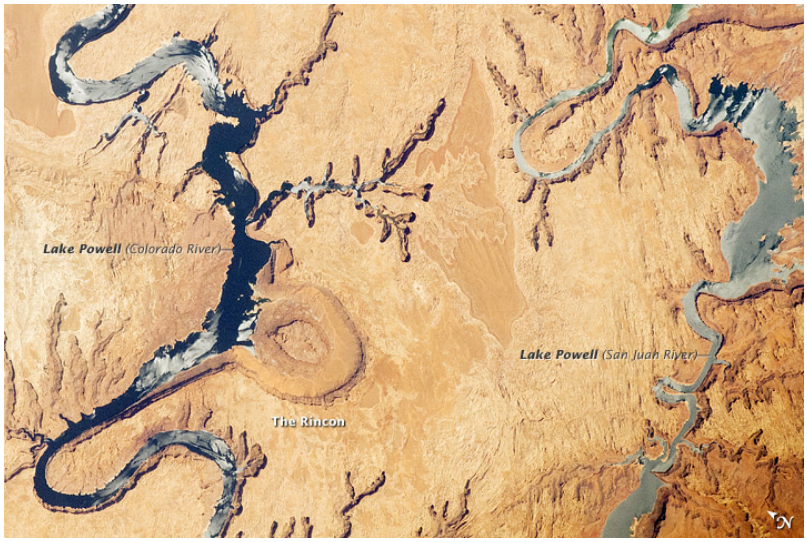
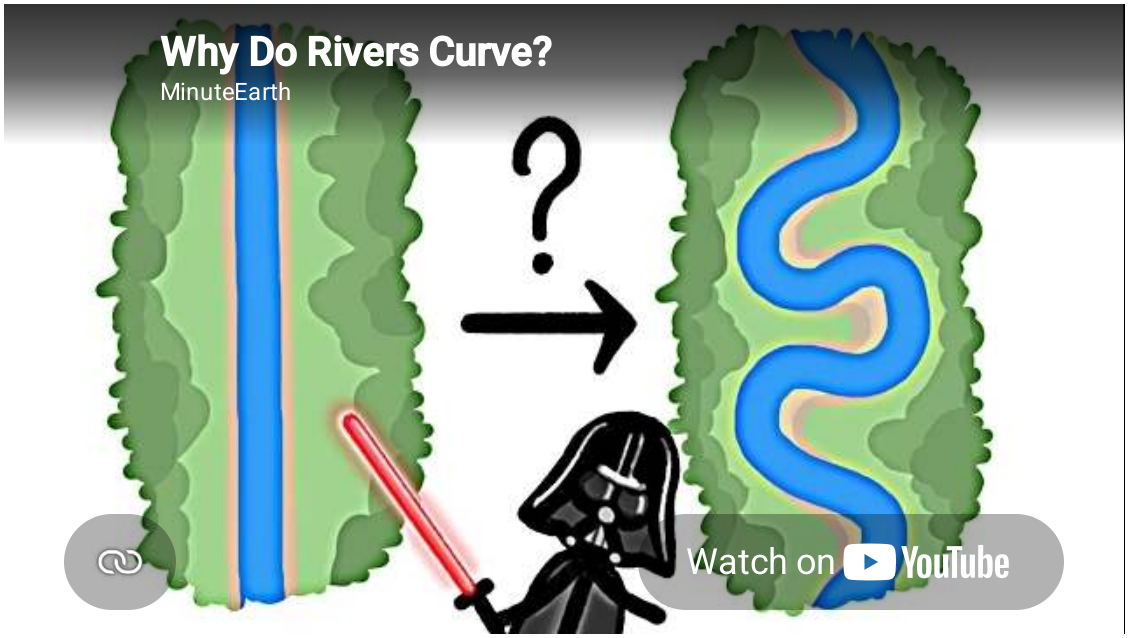
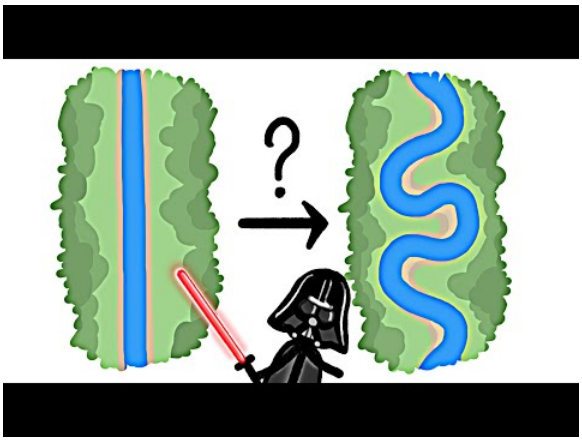


Figure 11.5.12: The Rincon is an abandoned meander loop on the entrenched Colorado River in Lake Powell.

Watch a video on why rivers curve.





Terraces

Stream terraces are remnants of older floodplains located above the existing floodplain and stream. Like entrenched meanders, stream terraces form when uplift occurs or base level drops and streams erode downward, leaving behind their old floodplains. In other cases, stream terraces can form from extreme flood events associated with retreating glaciers. A classic example of multiple stream terraces is along the Snake River in Grand Teton National Park in Wyoming [32, 33].



Figure 11.5.13: Terraces along the Snake River, Wyoming. The two flat areas at different levels above the river in the right half of the photo are terraces.



Figure 11.5.14: Terraces in Glen Roy, Scotland. The terraces are prominently shown as stair steps in the cliff.

Alluvial Fans

Alluvial fans are depositional landforms created where streams emerge from mountain canyons into a valley. The channel that had been confined by the canyon walls is suddenly no longer confined. The stream water slows down and spreads out, dropping its bedload of all sizes, forming a delta in the valley. As distributary channels fill with sediment, the stream is diverted laterally, and the alluvial fan develops into a cone shape with distributaries radiating from the canyon mouth. Alluvial fans are common in the dry climates of the west where ephemeral streams emerge from canyons in the Basin and Range.



Figure 11.5.15: Alluvial fan in Iraq seen by a NASA satellite. A stream emerges from the canyon and creates this cone-shaped deposit.

Deltas

When a stream reaches a low-energy body of water such as a lake or some parts of the ocean, the velocity slows and the bedload and suspended load sediment are deposited, forming a delta. If wave erosion from the water body is greater than deposition from the stream, the delta will not form. The largest and most famous delta in the United States is the Mississippi River delta formed where the Mississippi River flows into the Gulf of Mexico. The Mississippi River drainage basin is the largest in North America, draining 41% of the contiguous U.S. [24]. Because of the large drainage area, the stream carries a large amount of sediment that is supplied to the delta. The Mississippi River is a major shipping route and the river has been artificially straightened so that it meanders less and is now 229 km shorter than it was before humans began engineering it [24]. Because of these restraints, the delta is now solely focused in one area and thus has created a “bird’s foot” pattern. These two NASA images of the delta show how the shoreline has retreated and the land was inundated with water while deposition of sediment was located at end of the delta. These images have changed over a 25 year period from 1976 to 2001. These are stark changes illustrating sea level rise and land subsidence from the compaction of peat due to the lack of sediment resupply [25].



Figure 11.5.16: Location of the Mississippi River drainage basin and Mississippi River delta.

Deltas represent stream deposits protruding into a quiet water body and can be further categorized as wave-dominated or tide-dominated. **Wave-dominated deltas** occur where the tides are small and wave energy dominates. An example is the Nile River delta in the Mediterranean Sea that has the classic shape like the Greek character (Δ) from which the landform is named.



Figure 11.5.17: Nile Delta in green showing its classic “delta” shape.

A **tide-dominated delta** forms when ocean tides are powerful and influence the shape of the delta. For example, Ganges-Brahmaputra Delta in the Bay of Bengal (near India and Bangladesh) is the world’s largest delta called the Sundarban. Tidal forces create linear intrusions of seawater into the delta deposits. This delta also holds the world’s largest mangrove swamp.



Figure 11.5.18: Sundarban Delta in Bangladesh, a tide-dominated delta of the Ganges River.

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11.6: Groundwater

Groundwater is an important source of freshwater. It can be found in all places under the ground but is limited by extractable quantity and quality.



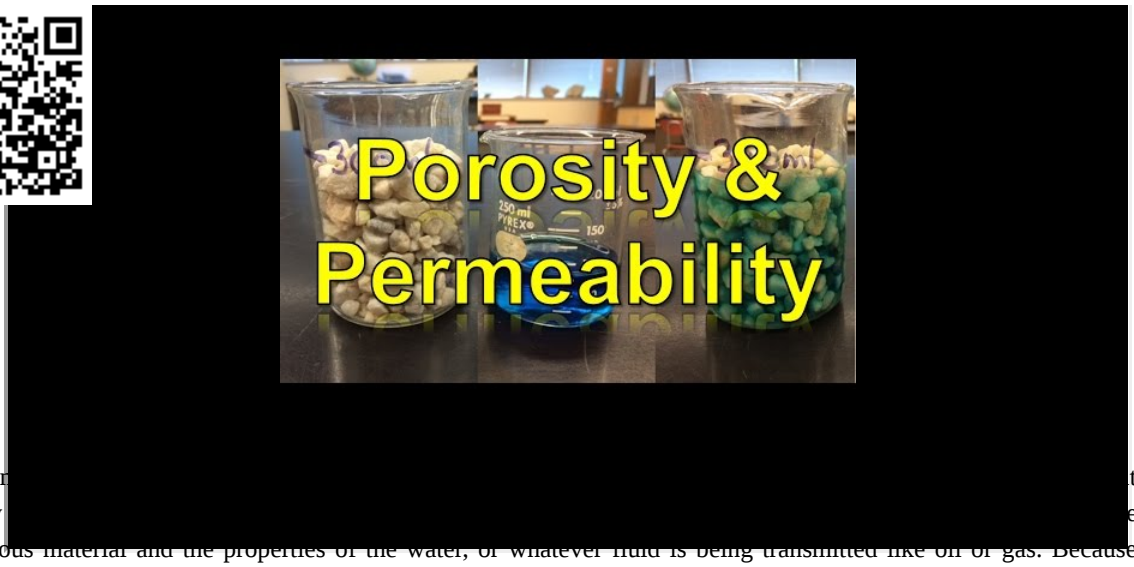


Permeability and Porosity

Most rocks are not entirely solid and contain a certain amount of open space between grains or crystals, known as **pores**. **Porosity** is a measure of the open space in rocks –expressed as the percentage of open space that makes up the total volume of the rock or sediment material. Porosity can occur as **primary porosity**, which represents the original pore spaces in the rock (e.g. space between sand grains, vesicles in volcanic rocks), or **secondary porosity** which occurs after the rock forms (e.g. fractures, dissolved portions of rock). Lithification of unconsolidated sediments will reduce porosity because it compacts grains and adds cement.

Permeability is a measure of the interconnectedness of pores in a rock or sediment. The connections between pores allows for that material to transmit water. A combination of a place to put water (porosity) and the ability to move water (permeability) makes a good **aquifer**—a rock unit or sediment that contains extractable groundwater. Well-sorted sediments have higher porosity because there are not smaller sediment particles filling in the spaces between the larger particles. Shales made of clays generally have very high porosity, but the pores are poorly connected, thereby causing low permeability.





While perm ty is more commonly ermeability of the porous material and the properties of the water, or whatever fluid is being transmitted like oil or gas. Because hydraulic conductivity also measures the properties of the fluid, such as viscosity, it is used by both petroleum geologists and hydrologists to describe both the production capability of oil reservoirs and of aquifers. High hydraulic conductivity indicates that fluid transmits rapidly through an aquifer.

Aquifers and Confining Layers

Aquifers are rock layers with sufficient porosity and permeability to allow water to be both contained and move within them. For rock or sediment to be considered an aquifer, its pores must be at least partially filled with water and it must be permeable enough to transmit water. Drinking water aquifers must also contain **potable**, or drinkable water. Aquifers can vary dramatically in scale, from spanning several formations covering large regions to being a local formation in a limited area. Aquifers adequate for water supply are permeable, porous, and potable.

Groundwater Flow

From the Surface into the Ground

When surface water infiltrates or seeps into the ground, it usually enters the unsaturated zone called the vadose zone or **zone of aeration**. The **vadose zone** is the volume of geologic material between the land surface and the zone of saturation where the pore spaces are not completely filled with water [34]. Plant roots inhabit the upper vadose zone and fluid pressure in the pores is less than atmospheric pressure. Below the vadose zone is the capillary fringe. The **capillary fringe** is the usually thin zone below the vadose zone where the pores are completely filled with water (**saturation**), but the fluid pressure is less than atmospheric pressure. The pores in the capillary fringe are filled because of capillary action, which occurs because of a combination of adhesion and cohesion. Below the capillary fringe is the **saturated zone** (a.k.a. **phreatic zone**), where the pores are completely saturated and the fluid in the pores is at or above atmospheric pressure [34]. The interface between the capillary fringe and the saturated zone marks the location of the water table.

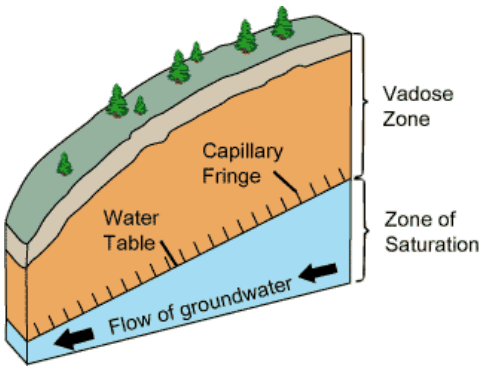


Figure 11.6.1: Zone of saturation.

Water Table

Wells are conduits that extend into the ground with openings to the aquifers, to extract from, measure, and sometimes add water to the aquifer. Wells are generally the way that geologists and hydrologists measure the depth to groundwater from the land surface as well as withdraw water from aquifers.

Water is found throughout the pore spaces in sediments and bedrock. The **water table** is the area below which the pores are fully saturated with water. The simplest case of a water table is when the aquifer is **unconfined**, meaning it does not have a confining layer above it. A **confining layer** is a low-permeability layer above and/or below an aquifer that restricts the water from moving in and out of the aquifer. Confining layers include **aquicludes**, which are so impermeable that no water travels through them, and **aquitards**, which significantly decrease the speed at which water travels through them. Confining layers can pressurize aquifers by trapping water that is recharged at a higher elevation underneath the confining layer, allowing for a potentiometric surface higher than the top of the aquifer, and sometimes higher than the land surface. The **potentiometric surface** represents the height that water would rise in a well penetrating the pressurized aquifer system. Breaches in the pressurized aquifer system, like faults or wells, can cause springs or **flowing wells**, also known as **artesian wells**.

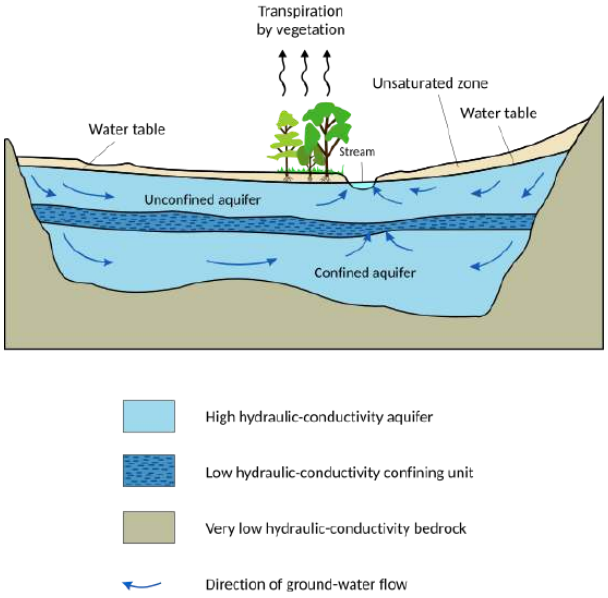


Figure 11.6.2: An aquifer cross-section. This diagram shows two aquifers with one aquitard (confining layer) between them, surrounded by the bedrock aquiclude, which is in contact with a gaining stream (typical in humid regions).

The water table will generally mirror surface topography, though more subdued, because hydrostatic pressure is equal to atmospheric pressure along the surface of the water table. If the water table intersects the ground surface, the result will be water at the surface in the form of a gaining stream, **spring**, lake, or wetland. The water table intersects the channel for **gaining streams** which then gains water from the water table. The channels for **losing streams** lie below the water table, thus losing streams lose water to the water table. Losing streams may be seasonal during a dry season or ephemeral in dry climates where they may normally be dry and carry water only after rainstorms. Ephemeral streams pose a serious danger of flash flooding in dry climates.





Using wells, geologists measure the water table's height and the potentiometric surface. Graphs of the depth to groundwater over time, are known as **hydrographs** and show changes in the water table over time. Well–water level is controlled by many factors and can change very frequently, even every minute, seasonally, and over longer periods of time.

Darcy's Law

In 1856, French engineer Henry Darcy developed a hypothesis to show how discharge through a porous medium is controlled by permeability, pressure, and cross-sectional area. To prove this relationship, Darcy experimented with tubes of packed sediment with water running through them. The results of his experiments empirically established a quantitative measure of hydraulic conductivity and discharge that is known as **Darcy's law**. The relationships described by Darcy's law have close similarities to Fourier's law in the field of heat conduction, Ohm's law in the field of electrical networks, or Fick's law in diffusion theory. Darcy's Law provides a quantitative measure of hydraulic conductivity and discharge.

$$Q = K \cdot A \cdot \frac{\Delta h}{L}$$

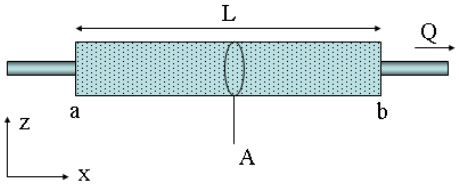


Figure 11.6.3: Pipe showing apparatus that would demonstrate Darcy's law. Δh would be measured across L from a to b.

- Q = flow (volume/time)
- K = hydraulic conductivity (length/time)
- A = cross-sectional area of flow (area)
- Δh = change in pressure head (pressure difference)
- L = distance between pressure (h) measurements (length)
- $\frac{\Delta h}{L}$ is commonly referred to as the hydraulic gradient

Cone of Depression

Pumping water from an aquifer lowers the water table or potentiometric surface around the well. In an unconfined aquifer, the water table is lowered as water is removed from the aquifer near the well. The amount of change from before pumping to pumping level is termed **drawdown**. Drawdown is the greatest nearest the well, resulting in a concentric pattern of drawdown termed the **cone of depression**. In a confined aquifer, pumping water reduces the pressure or potentiometric surface around the well.

When one cone of depression intersects another cone of depression or a barrier feature like an impermeable mountain block, drawdown is intensified. When a cone of depression intersects a recharge zone, the cone of depression is lessened.

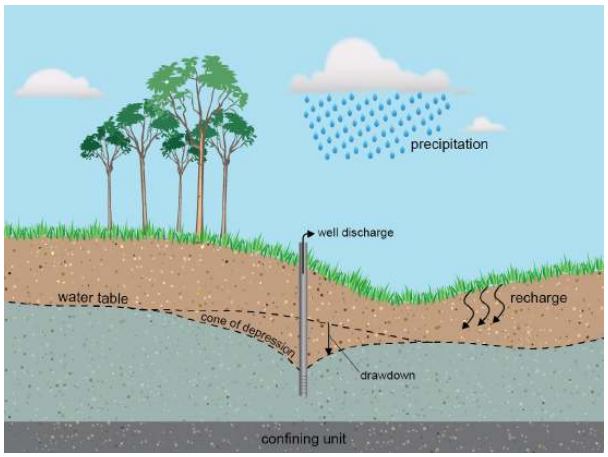


Figure 11.6.4: Cone of depression.

Recharge

The **recharge area** is where surface water enters an aquifer through the process of infiltration. Recharge areas are generally topographically high locations of an aquifer. They are characterized by losing streams and sediment or permeable rock that allows infiltration into the aquifer. Recharge areas mark the beginning of groundwater flow paths. In the Basin and Range region, the recharge zones for the unconsolidated aquifers of the valley areas are along the valley margins, near the foothills of the mountains.

Recharge can be induced through the aquifer management practice of aquifer storage and recovery. **Injection wells** and **infiltration galleries** (basins) allow for humans to increase the rate of recharge into an aquifer system [35]. Injection wells pump water into an aquifer where it can be stored. Injection wells are regulated by state and federal governments to ensure that the injected water is not negatively impacting the quality or supply of the existing groundwater in the aquifer. Some aquifers are capable of storing significant quantities of water, allowing water managers to use the aquifer system like a surface reservoir. Water is stored in the

aquifer during periods of low water demand and high water supply and later extracted during times of high water demand and low water supply.

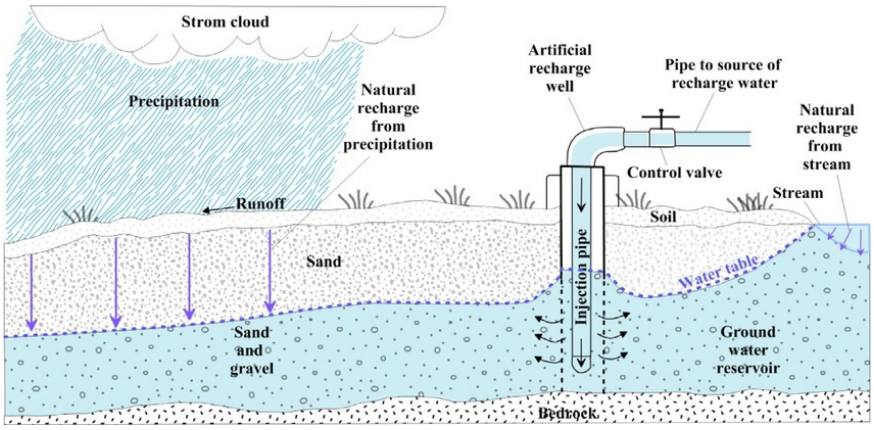


Figure 11.6.5: Different ways an aquifer can be recharged.

Discharge

Discharge areas are where groundwater emerges at the land surface. Groundwater emerges at the land surface when the potentiometric surface or water table intersects the land surface. These areas are characterized by springs, flowing (artesian) wells, gaining streams, and **playas** (dry lake bed). Discharge areas mark the end of groundwater flow paths. In the Basin and Range of the western United States, discharge zones are typically in the middle of the valley basins, where playa lakes, springs, and gaining streams signify groundwater emerging at the land surface.

Groundwater Mining and Subsidence

Groundwater as a Limited Resource

Like other natural resources on our planet, the quantity of fresh and potable water is finite. The only natural source of water on land is from the sky in the form of precipitation. Because of a slow rate of travel, limited recharge areas, and intensifying extraction and demand, in many places groundwater is being extracted faster than it is being replenished. When groundwater is extracted faster than it is recharged, groundwater levels (potentiometric surfaces) decline and areas of discharge can diminish or dry up completely. Regional pumping-induced groundwater decline is known as **groundwater mining** or **groundwater overdraft**. Groundwater mining can lead to dry wells, reduced spring and streamflow, and subsidence. Groundwater mining is happening in places where more water is extracted by pumping than is being replenished by precipitation, and the water table is continually lowered. In these situations, groundwater must be viewed as a ore body and in its depletion, the possibility of producing ghost towns.





Subsidence

In many places, water actually helps hold up the skeleton of the aquifer by the water pressure exerted on the grains in an aquifer. This pressure is called **pore pressure** and comes from the weight of overlying water. If pore pressure decreases because of groundwater mining, the aquifer can compact, causing **subsidence**, or the sinking of the surface of the ground. Areas especially susceptible to this effect are aquifers made of unconsolidated sediments. Unconsolidated sediments with multiple layers of clay and other fine-grained material are at higher risk because clay can compact considerably when drained of water [36, 37].

In many cases, the amount of compaction in one area will be greater than the amount of compaction in an adjacent area. The different amounts of compaction in areas that are next to each other can cause the land to offset and develop cracks and fissures. Subsidence from groundwater mining has been documented in southwestern Utah, notably Cedar Valley, Iron County, Utah. Groundwater levels have declined more than 100 feet in certain parts of Cedar Valley, causing earth fissures and measurable amounts of land subsidence.



Figure 11.6.6: Example of an earth fissure in Cedar City.

The photo shows documentation of subsidence from the pumping of groundwater for irrigation in the Central Valley in California. The dates marked on the pole show land elevation in the past.



Figure 11.6.7: The pole shows subsidence from groundwater pumping for irrigation in the San Joaquin Valley of California.

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11.7: Water Contamination

Types of Contamination

Water can be contaminated by various human activities or by existing natural features, like mineral-rich geologic formations. Agricultural activities, industrial operations, landfills, animal operations, and small and large scale sewage treatment processes, among many other things, all can potentially contribute to contamination. As water runs over the land or infiltrates into the ground, it dissolves material left behind by these potential contaminant sources. There are three major groups of contamination: inorganic chemicals, organic chemicals, and biological agents. Small sediments that cloud water, causing **turbidity**, are also an issue with some wells but it is not considered contamination. The risks and type of remediation for a contaminant depend on the type of chemicals present.

Point source pollution can be attributed to a single, definable source, while **nonpoint source pollution** is from multiple dispersed sources. Point sources include waste disposal sites, storage tanks, sewage treatment plants, and chemical spills. Nonpoint sources are dispersed and indiscreet, where the whole of the contribution of pollutants is harmful, but the individual components do not have harmful concentrations of pollutants. A good example of nonpoint source pollution is residential areas, where lawn fertilizer on one person's yard may not contribute much pollution to the system, but the combined effect of many residents using fertilizer can lead to significant nonpoint source pollution. Other nonpoint sources include nutrients (nitrate and phosphate), herbicides, pesticides contributed by farming, nitrate contributed by animal operations, and nitrate contributed by septic systems.

Organic chemicals are common pollutants. They consist of strands and rings of carbon atoms, usually connected by covalent bonds. Other types of atoms, like chlorine, and molecules like hydroxide (OH^-), are attached to the strands and rings. The number and arrangement of atoms will decide how the chemical behaves in the environment, its danger to humans or ecosystems, and where the chemical ends up in the environment. The different arrangements of carbon allow for tens of thousands of organic chemicals, many of which have never been studied for negative effects on human health or the environment. Common organic pollutants are herbicides and pesticides, pharmaceuticals, fuel, and industrial solvents and cleansers.

Organic chemicals include **surfactants** (cleaning agents) and synthetic hormones associated with pharmaceuticals, which can act as endocrine disruptors. **Endocrine disruptors** mimic hormones and can cause long-term effects in developing sexual reproduction systems in developing animals. Only very small quantities of endocrine disruptors are needed to cause significant changes in animal populations.

An example of organic chemical contamination is the Love Canal, in Niagara Falls, New York. From 1942 to 1952, the Hooker Chemical Company disposed of over 21,000 tons of chemical waste, including chlorinated hydrocarbons, into a canal and covered it with a thin layer of clay. Chlorinated hydrocarbons are a large group of organic chemicals that have chlorine functional groups, most of which are toxic and carcinogenic to humans. The company sold the land to the New York School Board, who developed it into a neighborhood. After residents began to suffer from serious health ailments and pools of oily fluid started rising into residents' basements, the neighborhood had to be evacuated. This site became a U.S. Environmental Protection Agency **Superfund Site**, a site with federal funding and oversight to ensure its cleanup.

Inorganic chemicals are another set of chemical pollutants. They can contain carbon atoms, but not in long strands or links. Inorganic contaminants include chloride, arsenic, and nitrate (NO_3). Nutrients can be from geologic material, like phosphorus-rich rock, but are most often sourced from fertilizer and animal and human waste. Untreated sewage and agricultural runoff contain concentrates of nitrogen and phosphorus which are essential for the growth of microorganisms. Nutrients like nitrate and phosphate in surface water can promote the growth of microbes, like blue-green algae (cyanobacteria), which in turn use oxygen and create toxins (microcystins and anatoxins) in lakes [38]. This process is known as **eutrophication**.

Metals are common inorganic contaminants. Lead, mercury, and arsenic are some of the more problematic inorganic groundwater contaminants. Bangladesh has a well-documented case of arsenic contamination from natural geologic material dissolving into the groundwater. Acid mine drainage can also cause significant inorganic contamination.

Salt, typically sodium chloride, is a common inorganic contaminant. It can be introduced into groundwater from natural sources, such as evaporite deposits like the Arapien Shale of Utah, or from anthropogenic sources like the salts applied to roads in the winter

to keep ice from forming. Salt contamination can also occur from saltwater intruding into the cones of depression around fresh groundwater pumping near ocean coasts, inducing the encroachment of saltwater into the freshwater body.

Another common groundwater contaminant is **biological agents**, which includes harmful bacteria and viruses. A common bacteria contaminant is *Escherichia coli (E. coli)*. Generally, harmful bacteria are not present in groundwater unless the source of groundwater is closely connected with a contaminated surface source, such as a septic system. Karst, landforms created from dissolved limestone, is especially susceptible to this form of contamination because water moves relatively quickly through the dissolved conduits of limestone. Bacteria can also be used for remediation (see below).

View the *USGS Table on Groundwater Contaminants.*

Remediation

Remediation is the act of cleaning contamination. **Biological remediation** usually consists of using specific strains of bacteria to break down a contaminant into safer chemicals. This type of remediation is usually used on organic chemicals but also works on reducing or oxidizing inorganic chemicals like nitrate. **Phytoremediation** is a type of bioremediation that uses plants to absorb the chemicals over time.

Chemical remediation uses the introduction of chemicals to remove the contaminant or make it less harmful. One example is a **reactive barrier**, a permeable wall in the ground or at a discharge point that chemically reacts with contaminants in the water. Reactive barriers made of limestone can increase the pH of acid mine drainage, making the water less acidic and more basic, which removes dissolved contaminants by precipitation into a solid form.

Physical remediation consists of removing the contaminated water and either treating it (aka pump and treat) with filtration or disposing of it. All of these options are technically complex, expensive, and difficult, with physical remediation typically being the most costly.

Reference

38. U.S. Environmental Protection Agency (2014) Cyanobacteria/Cyanotoxins

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11.8: Karst

Karst refers to landscapes and hydrologic features created by the dissolution of limestone. Karst can be found anywhere where there are limestone and other soluble subterranean substances like salt deposits. The dissolution of limestone creates features like sinkholes, caverns, disappearing streams, and **towers**.



Figure 11.8.1: Steep karst towers in China left as remnants as limestone is dissolved away by acidic rain and groundwater. (By chensiyuan; CC BY-SA 4.0 via Wikimedia Commons.)

The dissolution of underlying salt deposits has caused **sinkholes** to form in the Kaibab Limestone on the Colorado Plateau in Arizona.



Figure 11.8.2: Sinkholes of the McCauley Sink in Northern Arizona, produced by the collapse of Kaibab Limestone into caverns caused by the solution of underlying salt deposits.



Figure 11.8.3: Sinkhole from the collapse of the surface into an underground cavern that appeared in the front yard of this home in Florida.

CO_2 in the atmosphere dissolves readily in the water droplets that form clouds from which precipitation comes in the form of rain and snow. Thus precipitation is slightly acidic with carbonic acid. Karst forms when carbonic acid dissolves calcite (calcium carbonate) in limestone.

Water + carbon dioxide gas equals carbonic acid:

$$H_2O + CO_2 \longrightarrow H_2CO_3$$

Solid calcite + carbonic acid equals calcium ion + dissolved bicarbonate ion:

$$CaCO_3 + H_2CO_3 \longrightarrow Ca^{2+} + 2\,HCO_3^-$$

After the slightly acidic water dissolves the calcite, changes in temperature or gas content in the water can cause the water to redeposit the calcite in a different place as **tufa** (travertine), often deposited by a spring or in a cave. **Speleothems** are secondary deposits, typically made of travertine, deposited in a cave. Travertine speleothems form by water dripping through cracks and dissolved openings in caves and evaporating, leaving behind the travertine deposits. Speleothems commonly occur in the form of **stalactites**, when extending from the ceiling, and **stalagmites**, when extending from the floor.

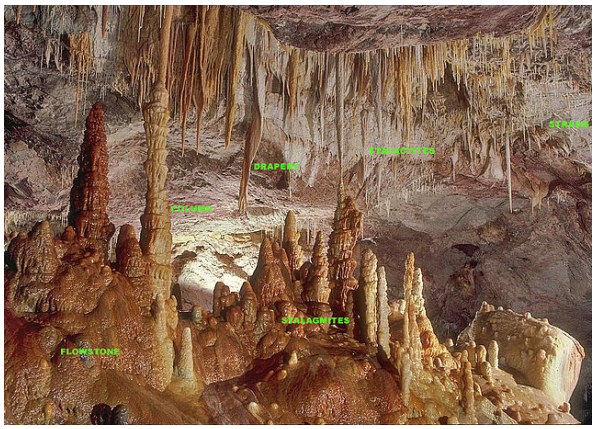


Figure 11.8.4: Varieties of speleotherms.

Surface water enters the karst system through sinkholes, losing streams, and **disappearing streams** which flow underground. Changes in base level can cause rivers running over limestone to dissolve the limestone and sink into the ground. As the water continues to dissolve its way through the limestone, it can leave behind intricate networks of caves and narrow passages. Often dissolution will follow and expand fractures in the limestone. Water exits the karst system as springs and rivers. In mountainous terrain, dissolution can extend all the way through the vertical profile of the mountain, with caverns dropping thousands of feet.



Figure 11.8.5: This stream disappears into a subterranean cavern system to re-emerge a few hundred yards downstream.

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11.S: Summary

Summary

Water is essential for all living things. It continuously cycles through the atmosphere, over land, and through the ground. In much of the United States and other countries, water is managed through a system of regional laws and regulations and distributed on paper in a system collectively known as “water rights”. Surface water follows a watershed, which is separate from other areas by its divides (highest ridges). Groundwater exists in the pores within rocks and sediment. It moves predominantly due to pressure and gravitational gradients through the rock. Human and natural causes can make water unsuitable for consumption. There are different ways to deal with this contamination. Karst is when limestone is dissolved by water, forming caves and sinkholes.

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CHAPTER OVERVIEW

12: Coastlines

Learning Objectives

- Describe how waves occur, move, and carry energy.
- Describe the pattern of the main ocean currents and explain the different factors involved in surface currents and deep ocean currents.
- Explain how ocean tides occur and distinguish among diurnal, semidiurnal, and mixed tide patterns.
- Explain wave behavior approaching the shoreline.
- Describe shoreline features and zones.
- Describe wave refraction and its contribution to longshore currents and longshore drift.
- Explain how longshore currents cause the formation of spits and baymouth bars.
- Distinguish submergent and emergent coasts and describe coastal features associated with each.
- Describe the relationship between the natural river of sand in the littoral zone and human attempts to alter it for human convenience.
- Describe marine depositional environments.



Figure 12.1: The beautiful California coastline near Big Sur. (By dconvertini; CC BY-SA 2.0 via Wikimedia Commons.)

The Earth's surface is 29% land and 71% water. Coastlines are the interfaces between land and water, and as such, the longest visible boundaries on Earth. To understand the processes that occur at these boundaries, it is important to first understand wave energy.

12.1: Waves and Wave Processes
12.2: Currents and Tides
12.3: Shoreline Features
12.4: Marine Depositional Environments
12.S: Summary

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12.1: Waves and Wave Processes

Waves are created when wind blows over the surface of the water. Energy is transferred from wind to the water by friction and carried in the upper part of the water by waves. Waves move across the water surface with individual particles of water moving in circles. As the wave crest passes, the water is moving forward. As the wave trough passes, the water is moving backward. This can be demonstrated by watching the movement of a cork or some floating object as a wave passes.

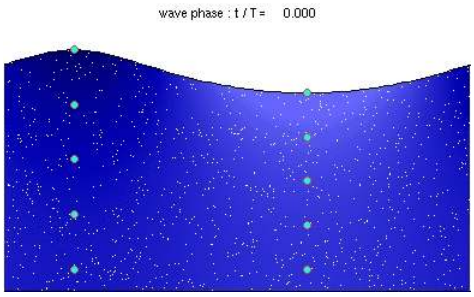


Figure 12.1.1: Particle motion within a wind-blown wave.

Important terms to understand in the operation of waves include: the **wave crest** is the highest point of the wave; the **trough** is the lowest point of the wave. **Wave height** is the vertical distance from the trough to the crest and depends on the amount of energy carried in the wave. Even before reaching shore, wave height increases with increasing wave energy. Wave **amplitude** is half the wave height, or the distance from either the crest or trough to the still water line. **Wavelength** is the horizontal distance between consecutive wave crests. **Wave velocity** is the speed at which a wave crest moves forward, which is related to the energy carried by the wave. **Wave period** is the time interval it takes for consecutive wave crests to pass a given point.

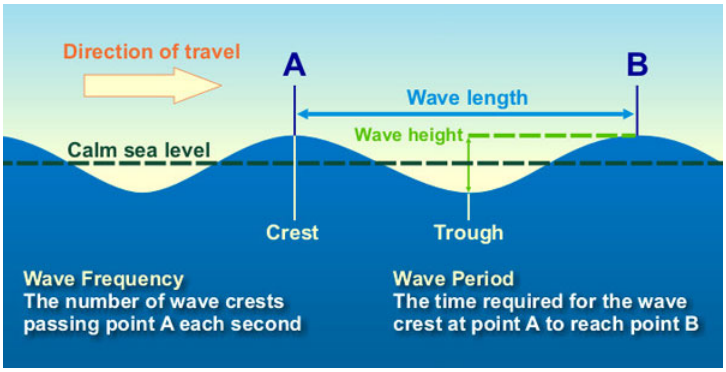


Figure 12.1.2: Aspects of water waves, labeled.

The circular motion of water particles diminishes with depth and is negligible at about one-half wavelength, an important dimension to remember in connection with waves. **Wave base** is the vertical depth at which water ceases to be disturbed by waves. In water shallower than wave base, waves will disturb the bottom and ripple shore sand. Wave base is measured at a depth of about one-half wavelength, where the water particles' circular motion diminishes to zero. If waves approaching a beach have crests at about 6 m (~20 ft) intervals, this wave motion disturbs water to about 3 m (~10 ft) deep. This motion is known as **fair-weather wave base**. In strong storms such as hurricanes, both wavelength and the maximum depth of water disturbance increase dramatically. The effective depth to which waves can erode sediment is thus called a **storm wave base,** which is approximately 91 m (300 feet) [1].

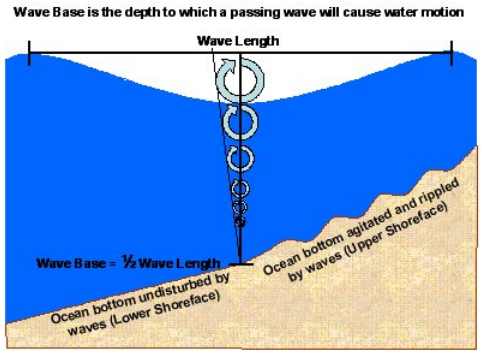


Figure 12.1.3: Diagram describing the wave base.

Waves are generated by wind blowing across the ocean surface. The amount of energy imparted to the water depends on the wind velocity and the distance across which the wind is blowing. This distance is called **fetch**. Waves striking a shore are typically generated hundreds of miles from the coast by storms and may have been traveling across the ocean for days.

Winds blowing in a relatively constant direction generate waves moving in that direction. Such a group of approximately parallel waves traveling together is called a **wave train**. As wave trains spread from different areas of generation, they may move in different directions and carry different amounts of energy. Interaction of these different wave trains produces the choppy sea surface seen in the open ocean. A wave train coming from one fetch can produce various wavelengths. Longer wavelengths travel at a faster velocity than shorter wavelengths, so they arrive first at a distant shore. Thus, there is a wavelength–sorting process that takes place during the wave train's travel. This sorting process is called **wave dispersion**.

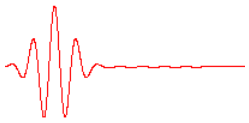


Figure 12.1.4: Model of a wave train moving with dispersion.

Behavior of Waves Approaching Shore

On the open sea, waves generally appear choppy because wave trains from many directions are interacting with each other, a process called **wave interference. Constructive interference** occurs where crests align with other crests. The aligned wave height is the sum of the individual wave heights, a process referred to as **wave amplification**. Constructive interference also produces hollows where troughs align with other troughs. **Destructive interference** occurs where crests align with troughs and cancel each other out. As waves approach the shore and begin to make frictional contact with the seafloor (i.e., water depth is a half wavelength or less) they begin to slow down. However, the energy carried by the wave remains the same so they build up higher. Remember that the water moves in a circular motion as the wave passes, with the water that feeds each circle being drawn from the trough in front of the advancing wave. As the wave encounters shallower water at the shore, there is eventually insufficient water in the trough in front of the wave to supply a complete circle, and the crest pours over creating a **breaker**.

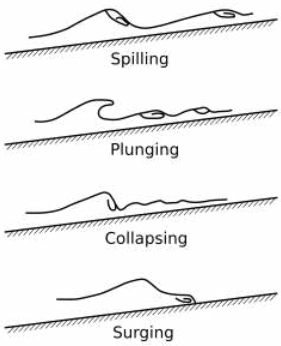


Figure 12.1.5: Types of breakers

A special type of wave is called a **tsunami**, sometimes incorrectly called a “tidal wave.” Tsunamis are generated by energetic events affecting the seafloor, such as earthquakes, submarine landslides, and volcanic eruptions. In the case of earthquakes, tsunamis can be produced when the moving crustal rocks below the sea abruptly elevate a portion of the seafloor. Water is suddenly lifted creating a bulge at the surface and a wave train spreads out in all directions traveling at tremendous speeds [over 322 km/hr (200 mph)] and carrying enormous energy. Tsunamis may pass unnoticed in the open ocean because they move so fast, the wavelength is very long and the wave height is very low. But as the wave train approaches the shore, each wave begins to interact with the shallow seafloor, friction increases and the wave slows down. Still carrying its enormous energy, wave height builds up and the wave strikes the shore as a wall of water that can be over 30 m (~100 ft) high. The massive wave, called a **tsunami runup**, may sweep inland well beyond the beach destroying structures far inland. Tsunamis can deliver a catastrophic blow to people at the beach. As the trough water in front of the tsunami wave is drawn back, the seafloor is exposed. Curious and unsuspecting people on the beach may run out to see exposed offshore sea life only to be overwhelmed when the breaking crest hits.

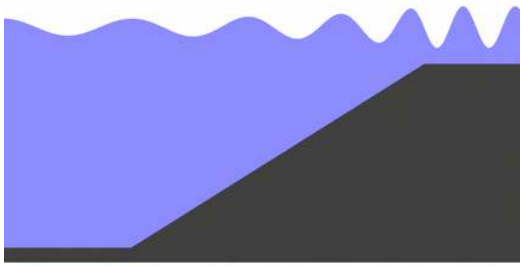


Figure 12.1.6: All waves, like tsunamis, slow down as they reach shallow water. This causes the wave to increase in height.

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12.2: Currents and Tides

Ocean water moves as waves, currents, and tides. Ocean currents are driven by persistent global winds blowing over the water's surface and by water density. Ocean currents are part of Earth's heat engine in which solar energy is absorbed by ocean water and distributed by ocean currents. Water has another unique property, high specific heat, that relates to ocean currents. **Specific heat** is the amount of heat necessary to raise a unit volume of a substance one degree. For water it takes one calorie per cubic centimeter to raise its temperature one degree Celsius. This means the oceans, covering 71% of the Earth's surface, soak up solar heat with little temperature change and distribute that heat around the Earth by ocean currents.

Surface Currents

The Earth's rotation and the Coriolis effect exert significant influence on ocean currents. In the figure, the black arrows show global surface currents. Notice the large circular currents in the Northern and Southern Hemispheres in the Atlantic, Pacific, and Indian Oceans. These are driven by prevailing atmospheric circulation (air movement) and are called **gyres** [9]. Gyres rotate clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere because of the **Coriolis effect**, which, due to the Earth's rotation, causes wind and ocean currents to be deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Currents flowing from the equator toward the poles tend to be narrow as a result of the Earth's rotation and carry warm water poleward along the east coasts of adjacent continents. These are called **western boundary currents**, and they are key contributors to the local climate. The Gulf Stream and the Kuroshio currents in the Northern Hemisphere and the Brazil, Mozambique, and Australian currents in the Southern Hemisphere are such western boundary currents. Currents returning cold water toward the equator tend to be broad and diffuse along the western coasts of adjacent landmasses. These warm western boundary and cold **eastern boundary currents** affect climate of nearby lands making them warmer or colder than other areas at equivalent latitudes. For example, the warm Gulf Stream makes northern Europe much milder than similar latitudes in northeastern Canada and Greenland. Another example is the cool Peru Current flowing north along the west coast of South America. This cold current limits evaporation in the ocean and contributes to the cool and arid condition of the Atacama Desert in Chile [10].

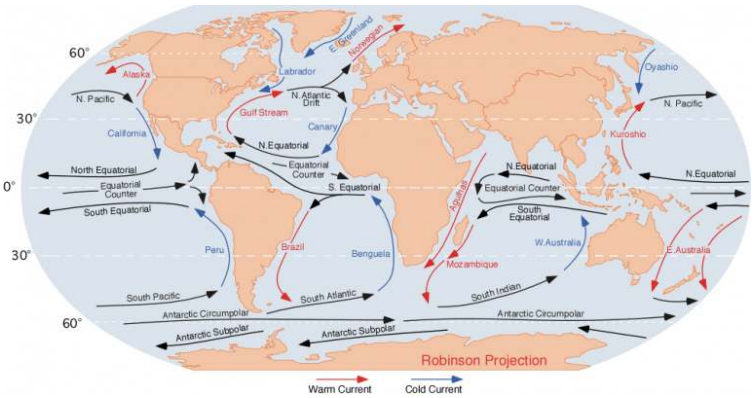


Figure 12.2.1: World ocean currents.

Deep Currents

Whether an ocean current moves horizontally or vertically depends on its density. The density of seawater is determined by temperature and **salinity,** or the amount of salt dissolved in the water. Evaporation and the influx of freshwater from rivers affect salinity and therefore the density of seawater. As the western boundary currents cool at high latitudes and salinity increases due to evaporation and ice formation (recall that ice floats; water is densest just above its freezing point), the cold, denser water sinks to become the ocean's deep waters. Movement of this deep water is called the **thermohaline circulation** (*thermo* refers to temperature, *haline* refers to salinity). This circulation connects the deep waters of all the world's oceans. Movement of the Gulf Stream illustrates the beginning of thermohaline circulation. Heat in the warm poleward moving Gulf Stream promotes evaporation which takes heat from the water and as the heat dissipates, the water cools. The resulting water is much colder, saltier, and denser. As the denser water reaches the North Atlantic and Greenland, it begins to sink and becomes a deep-water current. As shown in the illustration below, this worldwide connection between shallow and deep-ocean circulation overturns and mixes the entire world ocean, bringing nutrients to marine life, and is sometimes referred to as the **global conveyor belt** [11].

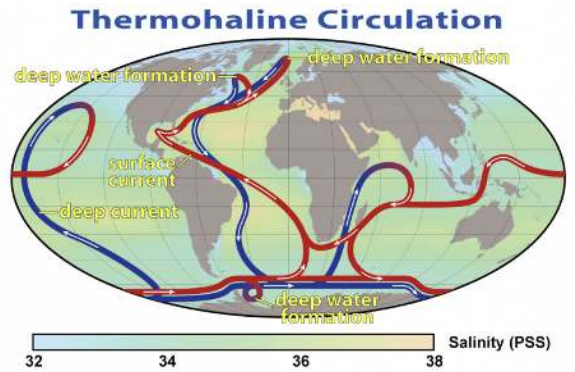


Figure 12.2.2: Global thermohaline circulation. PSS=practical salinity scale units.

Tides

The gravitational effects of the Sun and Moon on the oceans create **tides,** the rising and lowering of sea level during the day [12]. The Earth rotates daily within the gravity fields of the Moon and Sun. Although the Sun is much larger and its gravitational pull is more powerful, its great distance from Earth means that the gravitational influence of the Moon on tides is dominant. The magnitude of the tide at a given location and the difference between high and low tide (the **tidal range**) depends primarily on the configuration of the Moon and Sun with respect to the Earth's orbit and rotation. When the Sun, Moon, and Earth line up with each other at full moon or new moon, the tidal range is at a maximum. This is called **spring tide.** Approximately two weeks later when the Moon and Sun are at right angles with the Earth, the tidal range is lowest. This is called **neap tide.**

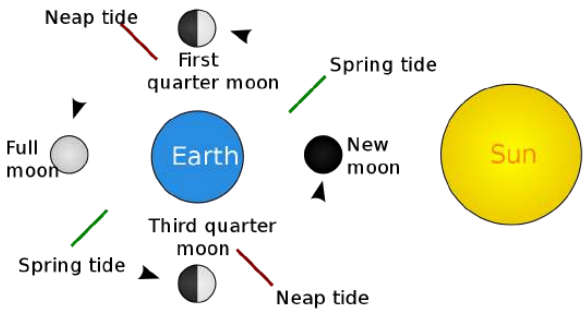


Figure 12.2.3: Diagram showing tides, in relation to the Sun and Moon.

The Earth rotates within a tidal envelope so we experience the rising and ebbing of the tide on a daily basis. Tides are measured at coastal locations and these measurements and tidal predictions based on them are published for those who depend on this information (e.g. this NOAA website) [13]. The rising and falling of the tides (**tidal patterns**) as experienced at a given shore location are of three types: diurnal, semidiurnal, and mixed**.**

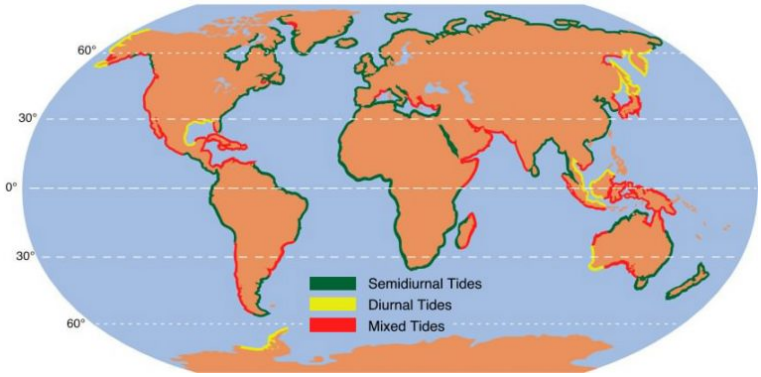


Figure 12.2.4: Global tide types.

Diurnal tides go through a complete cycle once in each tidal day. Keeping in mind that the Moon orbits around the Earth, a **tidal day** is the amount of time for the Moon to align with a point on the Earth as the Earth rotates, which is slightly longer than 24

hours. **Semidiurnal tides** go through two complete cycles in each tidal day—approximately 24 hours and 50 minutes, with the tidal range typically varying in each cycle. **Mixed tides** are a combination of diurnal and semidiurnal patterns and show two tidal cycles per tidal day, but the relative amplitudes of each cycle and their highs and lows vary during the tidal month. For example, there is a high high-tide and a high low-tide. The next day, there is a low high-tide and a low low-tide.

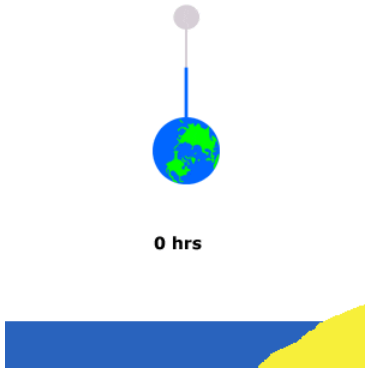


Figure 12.2.5: A tidal day lasts slightly longer than 24 hours. (By NOAA; public domain.)

The pattern at a given shore location and the times of arrival of tidal phases are complicated and determined by the **bathymetry** (depth) of the ocean basins and continental obstacles in the way of the tidal envelope within which the Earth rotates. Local tidal experts make 48-hour tidal forecasts using tidal charts based on daily observations, as can be seen in the chart of different tide types. Typical tidal ranges are on the order of 1 meter (3 feet). Extreme tidal ranges occur where the tidal wave enters a narrow restrictive zone that funnels the tidal energy. An example is the English Channel between Great Britain and the European continent where tidal ranges of 7 to 9.5 meters (25 to 32 feet) have been observed. The Earth's highest tidal ranges occur at the Bay of Fundy, the funnel-like bay between Nova Scotia and New Brunswick, Canada, where the average range is nearly 12 meters (40 feet) and the extreme range of around 18 meters (60 feet). At these locations of extreme tidal range, a person who ventures out onto the seafloor exposed during **ebb tide** may not be able to outrun the advancing water during **flood tide**. Additional information on tides can be found at this NOAA website.

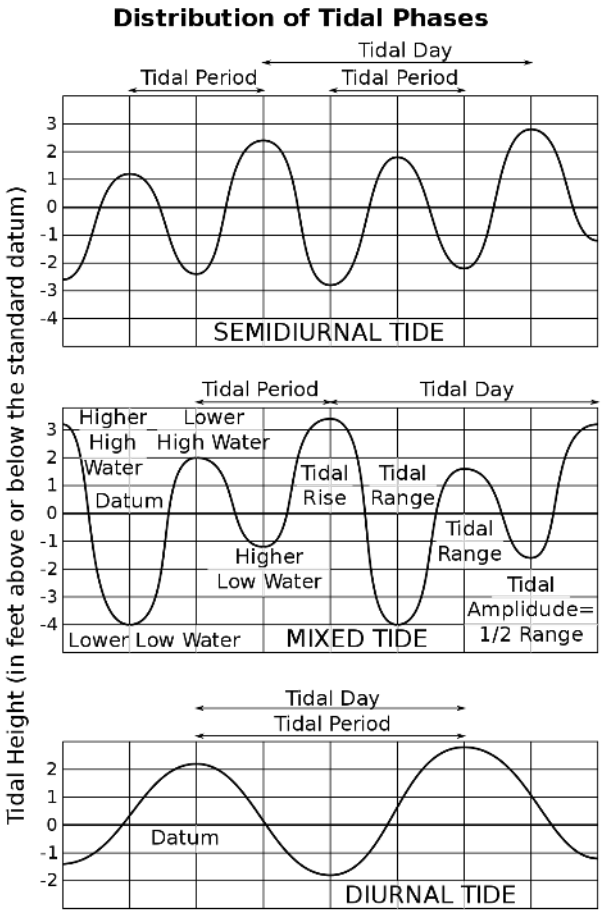


Figure 12.2.6: Different tide types.

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12.3: Shoreline Features

Many different erosional and depositional features exist in the high energy of the coast. The **coast** or **coastline** includes all parts of the land-sea boundary area that are directly affected by the sea. This includes land far above high tide and well below normal wave base. But the **shore** or **shoreline** itself is the direct interface between water and land that migrates with the tides and with deposition and erosion of sediment. Processes at the shoreline are called **littoral** processes.

Shoreline Zones

Shorelines are divided into five primary zones—offshore, nearshore, surf, foreshore, and backshore. The **offshore zone** is below water, but it is still geologically active due to flows of turbidity currents that cascade over the continental slope and accumulate in the continental rise. The **nearshore zone** is the area of the shore affected by the waves where water depth is one-half wavelength or less. The width of this zone depends on the maximum wavelength of the approaching wave train and the slope of the seafloor. The nearshore zone includes the **shoreface**, which is where sand is disturbed and deposited. The shoreface is broken into two segments: upper and lower shoreface. **Upper shoreface** is affected by everyday wave action and consists of finely-laminated and cross-bedded sand. The **lower shoreface** is the only area moved by storm waves and consists of hummocky cross-stratified sand. The **surf zone** is where the waves break.

The **foreshore zone** overlaps the surf zone and is periodically wet and dry due to waves and tides. The foreshore zone is where planar-laminated, well-sorted sand accumulates. The **beach face** is the part of the foreshore zone where the breaking waves swash up and the backwash flows back down. Low ridges above the beach face in the foreshore zone are called **berms**. During the summer in North America, when most people visit the beach, the zone where people spread their towels and beach umbrellas is the **summer berm**. Wave energy is typically lower in the summer, which allows sand to pile onto the beach. Behind the summer berm is a low ridge of sand called the **winter berm**. In winter, higher storm energy moves the summer berm sand off the beach and piles it in the nearshore zone. The next year, that sand is replaced on the beach and moved back onto the summer berm. The **backshore zone** is the area always above sea level in normal conditions. In the backshore zone, onshore winds may blow sand behind the beach and the berms, creating dunes.

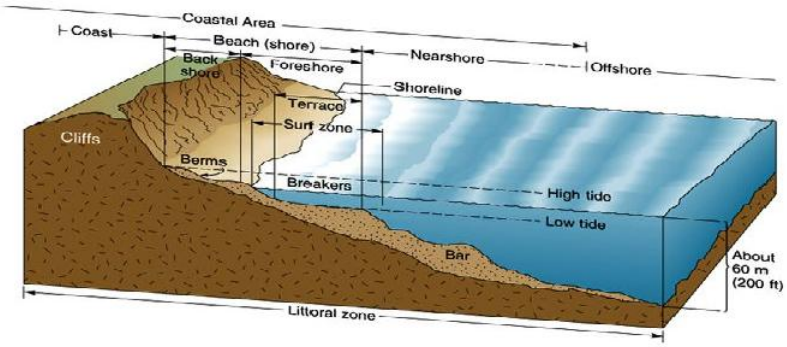


Figure 12.3.1: Diagram of zones of the shoreline.

Refraction, Longshore Currents, and Longshore Drift

As waves enter shallower water less than one-half wavelength depth, they slow down. Waves usually approach the shoreline at an angle, with the end of the waves nearest the beach slowing down first. This causes the wave crests to bend, called **wave refraction**. From the beach face, this causes it to look like waves are approaching the beach straight on, parallel to the beach. However, as refracted waves actually approach the shoreline at a slight angle, they create a slight difference between the swash as it moves up the beach face at a slight angle and the backwash as it flows straight back down under gravity. This slight angle between swash and backwash along the beach creates a current called the **longshore current**. Waves stir up sand in the surf zone and move it along the shore. This movement of sand is called **longshore drift**. Longshore drift along both the west and east coasts of North America moves sand north to south on average.

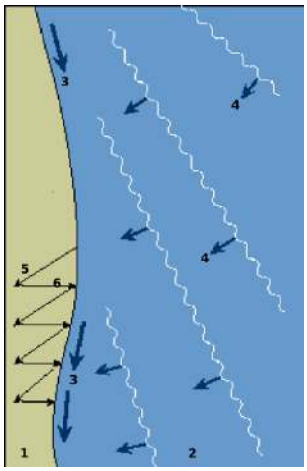


Figure 12.3.2: Longshore Drift. 1=beach, 2=sea, 3=longshore current direction, 4=incoming waves, 5=swash, 6=backwash

Longshore currents can carry longshore drift down a coast until it reaches a bay or inlet where it will deposit sand in the quieter water. Here, a **spit** can form. As the spit grows, it may extend across the mouth of the bay forming a barrier called a **baymouth bar**. Where the bay or inlet serves as boat anchorage, spits and baymouth bars are a severe inconvenience. Often, inconvenienced communities create methods to keep their bays and harbors open.



Figure 12.3.3: Farewell Spit, New Zealand.

One way to keep a harbor open is to build a jetty, a long concrete or stone barrier constructed to deflect the sand away from a harbor mouth or other ocean waterway. If the jetty does not deflect the sand far enough out, sand may continue to flow along the shore, forming a spit around the end of the jetty. A more expensive but effective method to keep a bay mouth open is to dredge the sand from the growing spit, put it on barges, and deliver it back to the drift downstream of the harbor mouth. An even more expensive but more effective option is to install large pumps and pipes to draw in the sand upstream of the harbor, pump it through pipes, and discharge it back into the drift downstream of the harbor mouth. Because natural processes work continuously, human efforts to mitigate inconvenient spits and baymouth bars require ongoing modifications. For example, the community of Santa Barbara, California, tried several methods to keep their harbor open before settling on pumps and piping. [2].



Figure 12.3.4: Jetties near Marina Del Rey, California. Notice the top jetty is loaded with sand, while the lower jetty is lacking sand. This is due to the longshore drift going from north to south. (By Google Maps.)

Rip currents are another coastal phenomenon related to longshore currents. Rip currents occur in the nearshore seafloor when wave trains come *straight* onto the shoreline. In areas where wave trains push water directly toward the beach face or where the shape of the nearshore seafloor refracts waves toward a specific point on the beach, the water piles up on shore. But this water must find an outlet back to the sea. The outlet is relatively narrow, and rip currents carry the water directly away from the beach. Swimmers caught in rip currents are carried out to sea. Swimming back to shore directly against the strong current is fruitless. A solution for good swimmers is to ride out the current to where it dissipates, swim around it, and return to the beach. Another solution for average swimmers is to swim parallel to the beach until out of the current, then return to the beach. Where rip currents are known to exist, warning signs are often posted. The best solution is to understand the nature of rip currents, have a plan before entering the water, or watch the signs and avoid them all together.

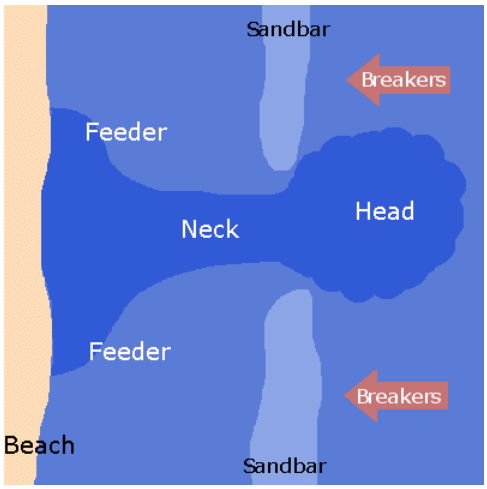


Figure 12.3.5: Animation of rip currents.

Like rip currents, **undertow** is a current that moves away from the shore. However, unlike rip currents, undertow occurs underneath the approaching waves and is strongest in the surf zone where waves are high and water is shallow. Undertow is another return flow for water transported onshore by waves.

Emergent and Submergent Coasts

Emergent coasts occur where sea levels fall relative to land level. Submergent coasts occur where sea levels rise relative to land level. Tectonic shifts and sea level changes cause the long-term rise and fall of sea level relative to land. Some features associated with emergent coasts include high cliffs, headlands, exposed bedrock, steep slopes, rocky shores, arches, stacks, tombolos, wave-cut platforms, and wave notches.

In emergent coasts, wave energy, wind, and gravity erode the coastline. The erosional features are elevated relative to the wave zone. **Sea cliffs** are persistent features as waves cut away at their base and higher rocks calve off by mass wasting. Refracted waves that attack bedrock at the base of **headlands** may erode or carve out a **sea arch**, which can extend below sea level in a sea cave. When a sea arch collapses, it leaves one or more rock columns called stacks.



Figure 12.3.6: Island Arch, a sea arch in Victoria, Australia.

A **stack** or near shore island creates a quiet water zone behind it. Sand moving in the longshore drift accumulates in this quiet zone forming a **tombolo**: a sand strip that connects the island or stack to the shoreline.



Figure 12.3.7: This tombolo, called “Angel Road,” connects the stack of Shodo Island, Japan.

Where sand supply is low, wave energy may erode a **wave-cut platform** across the surf zone, exposed as bare rock with tidal pools at low tide. This bench-like terrace extends to the cliff’s base. When wave energy cuts into the base of a sea cliff, it creates a **wave notch**.



Figure 12.3.8: Wave-cut platform off Dorset Coast, England. (By Jim Probert; CC BY-SA 2.0 via Wikimedia Commons.)

Submergent coasts occur where sea levels rise relative to land. This may be due to tectonic subsidence—when the Earth’s crust sinks—or when sea levels rise due to glacier melt. Features associated with submergent coasts include flooded river mouths, fjords, barrier islands, lagoons, estuaries, bays, tidal flats, and tidal currents. In submergent coastlines, river mouths are flooded by the rising water, for example Chesapeake Bay. **Fjords** are glacial valleys flooded by post-ice age sea level rise. **Barrier islands** are elongated bodies of sand that formed from old beach sands that used to parallel the shoreline. Often, lagoons lie behind barrier islands. [3]. Barrier island formation is controversial: some scientists believe that they formed when ice sheets melted after the last ice age, raising sea levels. Another hypothesis is that barrier islands formed from spits and bars accumulating far offshore.



Figure 12.3.9: Satellite image of Cape Hatteras National Seashore on the outer banks of North Carolina. Note the barrier islands off the coastline. (By NASA; public domain.)

Tidal flats or **mudflats** form where tides alternately flood and expose low areas along the coast. Combinations of symmetrical ripple marks, asymmetrical ripple marks from tidal currents, and mud cracks from drying form on these flats. Typically tidal flats

are broken into three different sections, which may be abundant or absent in each individual tidal flat. **Barren zones** are areas with strong, flowing water and coarser sediment, with ripple marks and cross-bedding common. **Marshes** are vegetated and typically have sand and mud. **Salt pans** are the finest-grained parts of the tidal flats, with silty sediment, mud cracks, and are less often submerged [4].

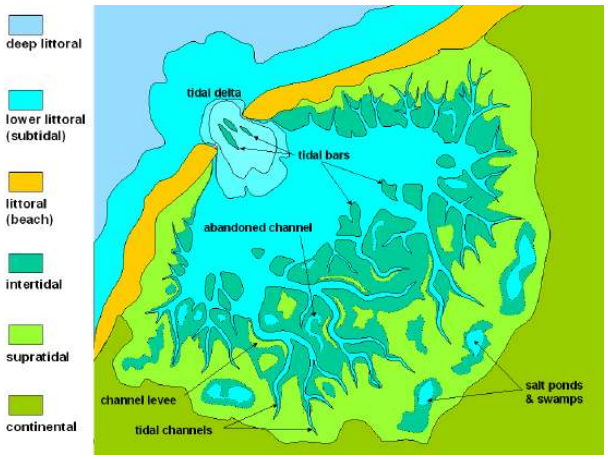


Figure 12.3.10: General diagram of a tidal flat and associated features.

Lagoons are locations where spits, barrier islands, or other features have partially cut off a body of water from the ocean. **Estuaries** are a vegetated type of lagoon where freshwater is flowing into the area, making the water **brackish** (between salt and freshwater). However, terms like a lagoon, estuary, and even **bay** are often loosely used in place of one another [5]. Lagoons and estuaries are certainly transitional between terrestrial and marine environments, where littoral, lacustrine (lakes or lagoons), and fluvial processes can overlap.



Figure 12.3.11: Kara-Bogaz Gol lagoon, Turkmenistan.

Human Impact on Coastal Beaches

Humans impact coastal beaches when they build homes, condominiums, hotels, businesses, and harbors—and then again when they try to manage the natural processes of erosion. Waves, currents, longshore drift, and dams at river mouths deplete sand from expensive beachfront property and expose once calm harbors to high-wave energy. To protect their investment, keep sand on their beach, and maintain calm harbors, cities and landowners find ways to mitigate the damage by building jetties, groins, dams, and breakwaters.

Jetties are large manmade piles of boulders or concrete barriers built at river mouths and harbors. A jetty is designed to divert the current or tide, to keep a channel to the ocean open, and to protect a harbor or beach from wave action. **Groins** (also spelled groyne) are similar but smaller than jetties. Groins are fences of wire, wood or concrete built across the beach perpendicular to the shoreline and downstream of a property. Unlike jetties, groins are used to preserve sand on a beach rather than to divert it. Sand erodes on the downstream side of the groin and collects against the upstream side. Every groin on one property thus creates a need for another one on the property downstream. A series of groins along a beach develops a scalloped appearance along the shoreline.

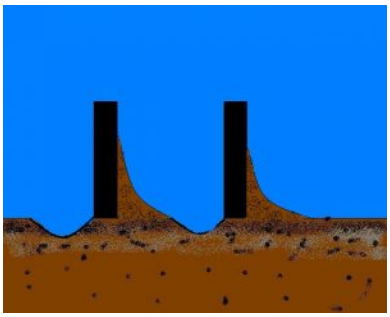


Figure 12.3.12: Groins gathering sediment from longshore drift. The current direction is from right to left.

Inland streams and rivers flow to the ocean carrying sand to the longshore current which distributes it to beaches. When dams are built, they trap sand and keep sediment from reaching beaches. To replenish beaches, sand may be hauled in from other areas by trucks or barges and dumped on the depleted beach. Unfortunately, this can disrupt the ecosystem that exists along the shoreline by exposing native creatures to foreign ecosystems and microorganisms and by introducing foreign objects to humans. For example, visitors to one replenished east coast beach found munitions and metal shards in the sand, which had been dredged from abandoned military test ranges [6].

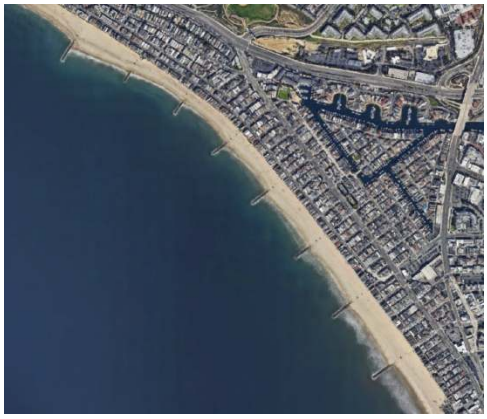


Figure 12.3.13: Groin system along Newport Beach, California. (By Google Maps.)

Another approach to reduce erosion or provide protected areas for boat anchoring is the construction of a **breakwater**, an offshore structure against which the waves break, leaving calmer waters behind it. Unfortunately, this means that waves can no longer reach the beach to keep the longshore drift of sand moving. The drift is interrupted, the sand is deposited in the quieter water, and the shoreline builds out forming a tombolo behind the breakwater, eventually covering the structure with sand [7]. The image shows this result at the breakwater constructed by the city of Venice, California in an attempt to create a quiet water harbor. The tombolo behind the breakwater is now acting as a large groin in the beach drift.



Figure 12.3.14: A tombolo has formed behind the breakwater at Venice, California. (By Google Maps.)

Submarine Canyons

Submarine canyons are narrow, deep underwater canyons located on continental shelves. Submarine canyons typically form at the mouths of large landward river systems. They form when rivers cut down into the continental shelf during low sea level and when

material continually slumps or flows down from the mouth of a river or a delta. Underwater currents rich in sediment and more dense than sea water, can flow down the canyons, even erode and deepen them, then drain onto the ocean floor. Underwater landslides, called turbidity flows, occur when steep delta faces and underwater sediment flows are released down the continental slope. Turbidity flows in submarine canyons can continue to erode the canyon, and eventually, fan-shaped deposits develop at the mouth of the canyon on the continental rise. [8].

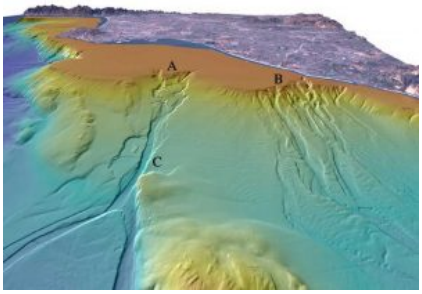


Figure 12.3.1: Submarine canyons off of Los Angeles. A=San Gabriel Canyon, B=Newport Canyon. At point C, the canyon is 815 m wide and 25 m deep.

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12.4: Marine Depositional Environments

Depositional Environments

Broadly, depositional environments along the coast are marine or a transitional zone between terrestrial and marine. **Marine** refers to environments associated with saltwater seas and oceans. Transitional depositional environments include environments such as deltas, where freshwater rivers empty into saltwater seas or oceans. These environments form the landforms around the coast.

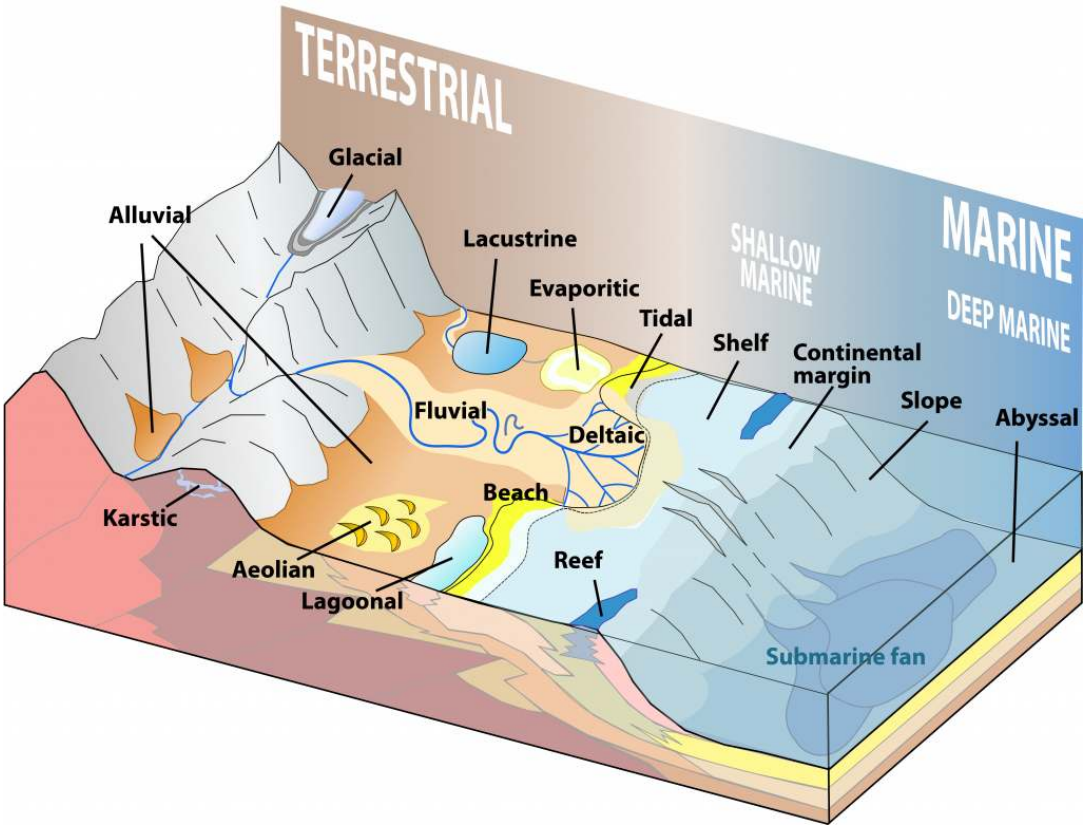


Figure 12.4.1: Some of the important depositional environments for sediments and sedimentary rocks. Source: Karla Panchuk (2021) CC BY-SA 4.0. Modified after Mike Norton (2018) CC BY-SA 3.0. View source.

Table 12.4.1 provides a summary of the processes and sediment types that pertain to the various marine and transitional depositional environments illustrated in the figure above.

Table 12.4.1: Marine and Transitional Depositional Environments.

Environment	Key Transport Processes	Depositional Settings	Typical Sediments
Deltaic	Moving water	Deltas	Sand, silt, clay, organic matter
Beach	Waves, long-shore currents	Beaches, spits, sand bars	Gravel, sand
Tidal	Tidal currents	Tidal flats	Fine-grained sand, silt, clay
Reef	Waves, tidal currents	Reefs and adjacent basins	Carbonates
Shallow marine	Waves, tidal currents	Shelves, slopes, lagoons	Carbonates in tropical climates; sand/silt/clay elsewhere
Lagoonal	Little transportation	Lagoon bottom	Carbonates in tropical climates; silt, clay elsewhere
Submarine fan	Underwater gravity flows	Continental slopes, abyssal plains	Gravel, sand, silt, clay
Deep water	Ocean currents	Deep-ocean abyssal plains	Clay, carbonate mud, silica mud

Source: Karla Panchuk (2018), CC BY 4.0. Modified after Steven Earle (2015), CC BY 4.0. View source.

Marine

Marine depositional environments are completely and constantly submerged in seawater. Their depositional characteristics are largely dependent on the depth of water with two notable exceptions: submarine fans and turbidites. The table shows the sediment, rocks, fossils and sedimentary structures typically found in marine depositional environments.

Table 12.4.2: Rock record and marine depositional environments.

Location	Sediment	Common Rock Types	Typical Fossils	Sedimentary Structures
Abyssal	very fine muds and oozes, diatomaceous earth	chert	diatoms	few
Submarine fan	graded Bouma sequences, alternating sand/mud	clastic rocks	rare	channels, fan shape
Continental slope	mud, possible sand, contourites	shale, siltstone, limestone	rare	swaths
Lower shoreface	laminated sand	sandstone	bioturbation	hummocky cross beds
Upper shoreface	planar sand	sandstone	bioturbation	plane beds, cross beds
Littoral (beach)	very well sorted sand	sandstone	bioturbation	few
Tidal flat	mud and sand with channels	shale, mudstone, siltstone	bioturbation	mudcracks, symmetric ripples
Reef	lime mud with coral	limestone	many, commonly coral	few
Lagoon	laminated mud	shale	many, bioturbation	laminations
Delta	channelized sand with mud, ±swamp	clastic rocks	many to few	cross beds

Abyssal

Abyssal sedimentary rocks form on the **abyssal plain**. The plain encompasses a relatively flat ocean floor with some minor topographical features, called **abyssal hills**. These small seafloor mounts range from 100 m to 20 km in diameter and are possibly created by extension [38]. Most abyssal plains do not experience significant fluid movement, so sedimentary rocks formed there are very fine-grained [39].

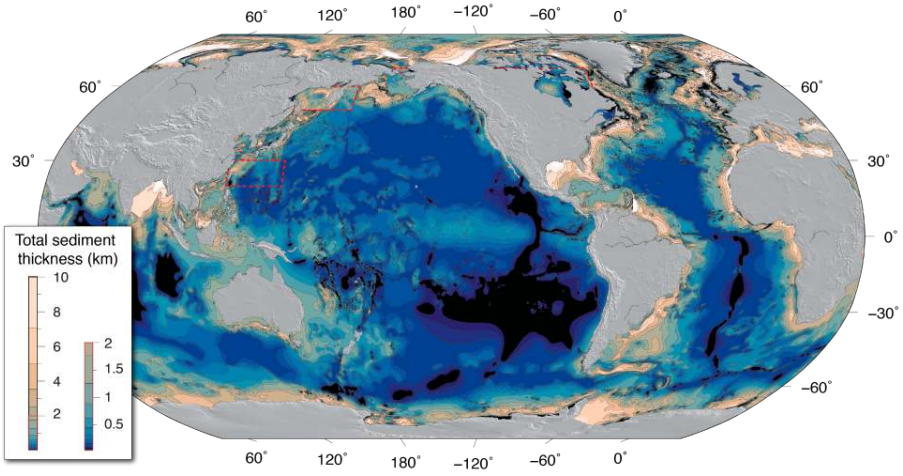


Figure 12.4.2: Marine sediment thickness. Note the lack of sediment away from the continents.

There are three categories of abyssal sediment. **Calcareous oozes** consist of calcite-rich plankton shells that have fallen to the ocean floor. An example of this type of sediment is **chalk**. **Siliceous oozes** are also made of plankton debris, but these organisms build their shells using silica or hydrated silica. In some cases such as with **diatomaceous earth**, sediment is deposited below the **calcite compensation depth**, a depth where calcite solubility increases. Any calcite-based shells are dissolved, leaving only silica-based shells. **Chert** is another common rock formed from these types of sediment. These two types of abyssal sediment are also classified as biochemical in origin.



Figure 12.4.3: Diatomaceous earth

The third sediment type is **pelagic clay**. Very fine-grained clay particles, typically brown or red, descend through the water column very slowly. Pelagic clay deposition occurs in areas of remote open ocean, where there is little plankton accumulation.

Two notable exceptions to the fine-grained nature of abyssal sediment are submarine fan and turbidite deposits [40]. **Submarine fans** occur offshore at the base of large river systems. They are initiated during times of low sea level, as strong river currents carve submarine canyons into the continental shelf. When sea levels rise, sediments accumulate on the shelf typically forming large, fan-shaped floodplains called deltas. Periodically, the sediment is disturbed creating dense slurries that flush down the underwater canyons in large gravity-induced events called **turbidites**. The submarine fan is formed by a network of turbidites that deposit their sediment loads as the slope decreases, much like what happens above-water at alluvial fans and deltas. This sudden flushing transports coarser sediment to the ocean floor where they are otherwise uncommon. Turbidites are also the typical origin of graded Bouma sequences.

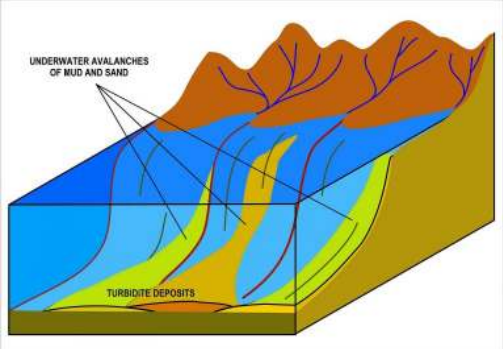


Figure 12.4.4: Turbidites inter-deposited within submarine fans.

Continental Slope

Continental slope deposits are not common in the rock record. The most notable type of continental slope deposits are contourites [41]. **Contourites** form on the slope between the continental shelf and the deep ocean floor. Deep-water ocean currents deposit sediment into smooth drifts of various architectures, sometimes interwoven with turbidites.

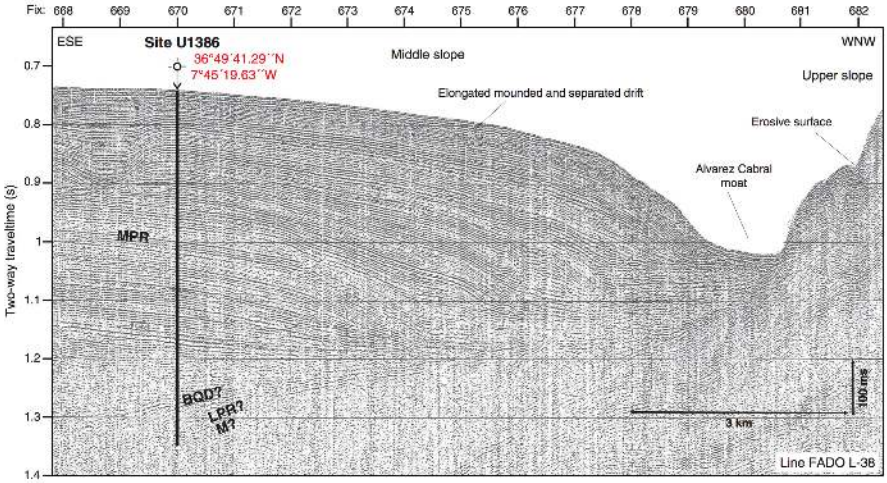


Figure 12.4.5: Contourite drift deposit imaged with seismic waves.

Lower Shoreface

The lower shoreface lies below the normal depth of wave agitation, so the sediment is not subject to daily winnowing and deposition. These sediment layers are typically finely laminated and may contain hummocky cross-stratification. Lower shoreface beds are affected by larger waves, such those as generated by hurricanes and other large storms [42].

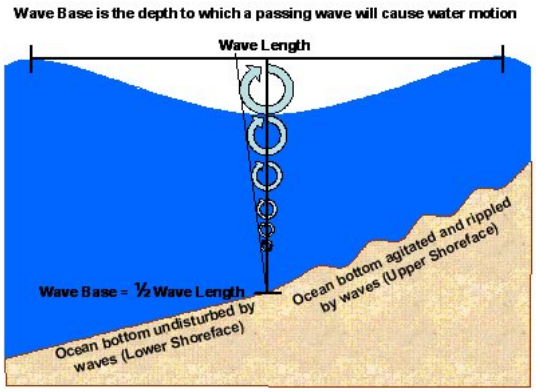


Figure 12.4.6: Diagram describing wavebase in the lower shoreface.

Upper Shoreface

The upper shoreface contains sediments within the zone of normal wave action but still submerged below the beach environment. These sediments usually consist of very well-sorted, fine sand. The main sedimentary structure is planar bedding consistent with the lower part of the upper flow regime, but it can also contain cross-bedding created by longshore currents [43].

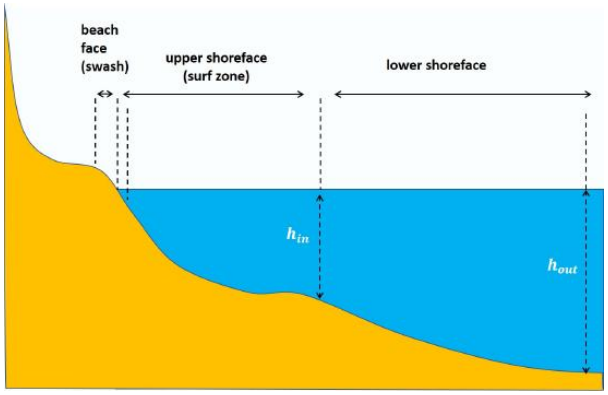


Figure 12.4.7: Diagram of a shoreface profile. (By Dronkers J; via Coastal Wiki.)

Transitional Coastline Environments

Transitional environments, more often called shoreline or **coastline environments**, are zones of complex interactions caused by ocean water hitting land. The sediment preservation potential is very high in these environments because deposition often occurs on the **continental shelf** and underwater. Shoreline environments are an important source of hydrocarbon deposits (petroleum, natural gas).

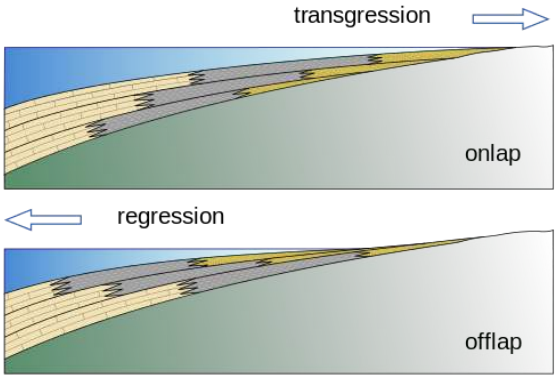


Figure 12.4.8: The rising sea levels of transgressions create onlapping sediments, regressions create offlapping.

The study of shoreline depositional environments is called **sequence stratigraphy**. Sequence stratigraphy examines depositional changes and 3D architectures associated with rising and falling sea levels, which is the main force at work in shoreline deposits. These sea-level fluctuations come from the daily tides, as well as climate changes and plate tectonics. A steady rise in sea level relative to the shoreline is called **transgression**. **Regression** is the opposite, a relative drop in sea level. Some common components of shoreline environments are littoral zones, tidal flats, reefs, lagoons, and deltas.

Littoral

The **littoral zone**, better known as the **beach**, consists of highly weathered, homogeneous, well-sorted sand grains made mostly of quartz. There are black sand and other types of sand beaches, but they tend to be unique exceptions rather than the rule. Because beach sands, past or present, are so highly evolved, the amount of grain weathering can be discerned using the minerals zircon, tourmaline, and rutile. This tool is called the ZTR (zircon, tourmaline, rutile) index [44]. The **ZTR index** is higher in more weathered beaches because these relatively rare and weather-resistant minerals become concentrated in older beaches. In some beaches, the ZTR index is so high the sand can be harvested as an economically viable source of these minerals. The beach environment has no sedimentary structures, due to the constant bombardment of wave energy delivered by surf action. Beach sediment is moved around via multiple processes. Some beaches with high sediment supplies develop dunes nearby.



Figure 12.4.9: Lithified heavy mineral sand (dark layers) from a beach deposit in India.

Tidal Flats

Tidal flats, or mudflats, are sedimentary environments that are regularly flooded and drained by ocean tides. Tidal flats have large areas of fine-grained sediment but may also contain coarser sands. Tidal flat deposits typically contain gradational sediments and may include multi-directional ripple marks. Mudcracks are also commonly seen due to the sediment being regularly exposed to air during low tides; the combination of mud cracks and ripple marks is distinctive to tidal flats [45].

Tidal water carries in sediment, sometimes focusing the flow through a narrow opening called a **tidal inlet**. **Tidal channels**, creek channels influenced by tides, can also focus on tidally-induced flow. Areas of higher flow like inlets and tidal channels feature coarser grain sizes and larger ripples, which in some cases can develop into dunes.



Figure 12.4.10: Aerial view of tidal channels draining tidal flat, Everglades National Park, Florida. Gulf of Mexico in background. (By Marli Miller; via Geologic Time Pics.)

Reefs

Reefs, which most people would immediately associate with tropical coral reefs found in the oceans, are not only made by living things. Natural buildups of sand or rock can also create reefs, similar to barrier islands. Geologically speaking, a **reef** is any topographically-elevated feature on the continental shelf, located oceanward of and separate from the beach. The term reef can also be applied to terrestrial features. Capitol Reef National Park in Utah contains a topographic barrier, a reef, called the Waterpocket Fold.



Figure 12.4.11: Waterpocket fold, Capitol Reef National Park, Utah.

Most reefs, now and in the geologic past, originate from the biological processes of living organisms [46]. The growth habits of coral reefs provide geologists important information about the past. The hard structures in coral reefs are built by soft-bodied marine organisms, which continually add new material and enlarge the reef over time. Under certain conditions, when the land beneath a reef is subsiding, the coral reef may grow around and through existing sediment, holding the sediment in place, and thus preserving the record of environmental and geological conditions around it.



Figure 12.4.12: A modern coral reef.

Sediment found in coral reefs is typically fine-grained, mostly carbonate, and tends to deposit between the intact coral skeletons. Water with high levels of silt or clay particles can inhibit the reef growth because coral organisms require sunlight to thrive; they host symbiotic algae called zooxanthellae that provide the coral with nourishment via photosynthesis. Inorganic reef structures have much more variable compositions. Reefs have a big impact on sediment deposition in lagoon environments since they are natural storm breaks, wave and storm buffers, which allow fine grains to settle and accumulate.



Figure 12.4.13: The light blue reef is fringing the island of Vanatinai. As the island erodes away, only the reef will remain, forming a reef-bound seamount.

Reefs are found around shorelines and islands; coral reefs are particularly common in tropical locations. Reefs are also found around features known as **seamounts**, which is the base of an ocean island left standing underwater after the upper part is eroded away by waves. Examples include the Emperor Seamounts, formed millions of years ago over the Hawaiian hotspot. Reefs live and grow along the upper edge of these flat-topped seamounts. If the reef builds up above sea level and completely encircles the top of

the seamount, it is called a **coral-ringed atoll**. If the reef is submerged, due to erosion, subsidence, or sea-level rise, the seamount-reef structure is called a **guyot**.

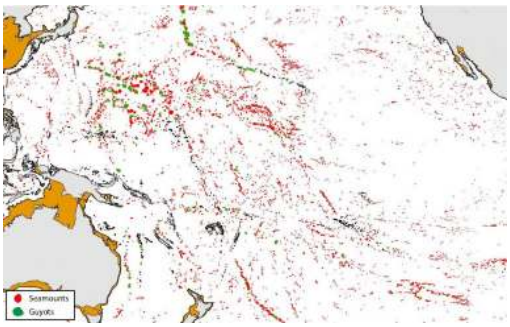


Figure 12.4.14: Seamounts and guyots in the North Pacific.

Lagoon

Lagoons are small bodies of seawater located inland from the shore or isolated by another geographic feature, such as a reef or barrier island. Because they are protected from the action of tides, currents, and waves, lagoon environments typically have very fine-grained sediments [47]. Lagoons, as well as estuaries, are ecosystems with high biological productivity. Rocks from these environments often include bioturbation marks or coal deposits. Around lagoons where evaporation exceeds water inflow, **salt flats**, also known as **sabkhas**, and sand dune fields may develop at or above the high tide line.



Figure 12.4.15: Oludeniz beach with blue lagoon in Turkey. (By Kotangens; education license via Adobe Stock.)

Deltas

Deltas form where rivers enter lakes or oceans and are organized based on the dominant process that controls their shape: wave-dominated deltas, river-dominated deltas, and tide-dominated deltas. The name delta comes from the Greek letter Δ (delta, uppercase) [48], which resembles the triangular shape of the Nile River delta. The velocity of water flow is dependent on riverbed slope or gradient, which becomes shallower as the river descends from the mountains. At the point where a river enters an ocean or lake, its slope angle drops to zero degrees (0°). The flow velocity quickly drops as well, and sediment is deposited, from coarse clasts to fine sand, and mud to form the delta. As one part of the delta becomes overwhelmed by sediment, the slow-moving flow gets diverted back and forth, over and over, and forms a spread out network of smaller distributary channels.

Wave-dominated deltas generally have smooth coastlines and beach-ridges on the land that represent previous shorelines. The Nile River delta is a wave-dominated type.



Figure 12.4.16: The Nile River delta, in Egypt.

The Mississippi River delta is a **river-dominated delta**, shaped by levees along the river and its distributaries that confine the flow forming a shape called a **bird-foot delta**. Other times the tides or the waves can be a bigger factor and can reshape the delta in various ways.



Figure 12.4.17: Birdfoot river-dominated delta of the Mississippi River.

A **tide-dominated delta** is dominated by tidal currents. During flood stages when rivers have lots of water available, it develops distributaries that are separated by sand bars and sand ridges. The tidal delta of the Ganges River is the largest delta in the world.

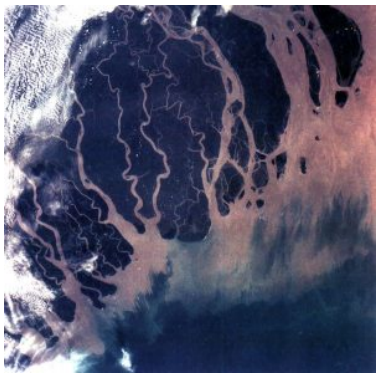


Figure 12.4.18: Tidal delta of the Ganges River.

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12.S: Summary

Summary

Shoreline processes are complex, but important for understanding coastal processes. Waves, currents, and tides are the main agents that shape shorelines. Most coastal landforms can be attributed to moving sand via longshore drift, and long-term rising or falling sea levels.

Oceans absorb solar energy, which is distributed by currents throughout the world. Circular surface currents, called gyres, rotate clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere. Thermohaline deep circulation connects the world's deep ocean waters: when shallow poleward moving warm water evaporates, the colder, saltier, and denser water sinks and becomes deep-water currents. The connection between shallow and deep-ocean circulation is called the global conveyor belt.

Tides are the rising and lowering of sea level during the day and are caused by the gravitational effects of the Sun and Moon on the oceans. There are three types of tidal patterns: diurnal, semidiurnal, and mixed. Typical tidal ranges are approximately 1 m (3 ft). Extreme tidal ranges are around 18 m (60 ft).

The shoreline is the interface between water and land and is divided into five zones. Processes at the shoreline are called littoral processes. Waves approach the beach at an angle, which cause the waves to bend towards the beach. This bending action is called wave refraction and is responsible for creating the longshore current and longshore drift—the process that moves sand along the coasts. When the longshore current deposits sand along the coast into quieter waters, the sand can accumulate, creating a spit or barrier called a baymouth bar, which often blocks bays and harbors. Inconvenienced humans create methods to keep their harbors open and preserve sand on their beaches by creating jetties and groins, which negatively affect natural beach processes. Emergent coasts are created by sea levels falling, while submergent coasts are caused by sea levels rising.

Marine depositional environments include abyssal plains, submarine fans, the continental slope, the continental rise and the lower and upper shoreface. Transitional depositional environments include beaches, deltas, tidal flats, lagoons and reefs. Sediments range from fine clay and mud to sand to coarse gravel, depending on water depth and movement of water. Fossils and shoreline depositional environments can tell the geologic history of the region, including rise and fall of sea level. Hydrocarbon deposits used by humans can be found in shoreline environments.

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CHAPTER OVERVIEW

13: Deserts

Learning Objectives

- Explain the defining characteristic of a desert and distinguish between the three broad categories of deserts.
- Explain how geographic features, latitude, atmospheric circulation, and Coriolis effect influence where deserts are located.
- List the primary desert weathering and erosion processes.
- Identify desert landforms.
- Explain how desert landforms are formed by erosion and deposition.
- Describe the main types of sand dunes and the conditions that form them.
- Identify the main features of the Basin and Range desert (United States).



Figure 13.1: A playa filled with evaporite minerals (such as gypsum) erodes and forms ripple-covered dunes in White Sands National Monument, New Mexico.

Approximately 30% of Earth's terrestrial surface is desert. **Deserts** are defined as locations of low precipitation. While temperature extremes are often associated with deserts, they do not define them. Deserts exhibit extreme temperatures because of the lack of moisture in the atmosphere, including low humidity and scarce cloud cover. Without cloud cover, the Earth's surface absorbs more of the Sun's energy during the day and emits more heat at night.

13.1: Locations of Deserts
13.2: The Origin of Deserts
13.3: Desert Weathering and Erosion
13.4: Desert Landforms
13.5: The Great Basin and the Basin and Range
13.S: Summary

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13.1: Locations of Deserts

Deserts are not randomly located on the Earth's surface. Many deserts are located at latitudes between 15° and 30° in both hemispheres and at both the North and South Poles, created by prevailing wind circulation in the atmosphere. Sinking, dry air currents occurring at 30° north and south of the equator produce trade winds that create deserts like the African Sahara and Australian Outback. [1].

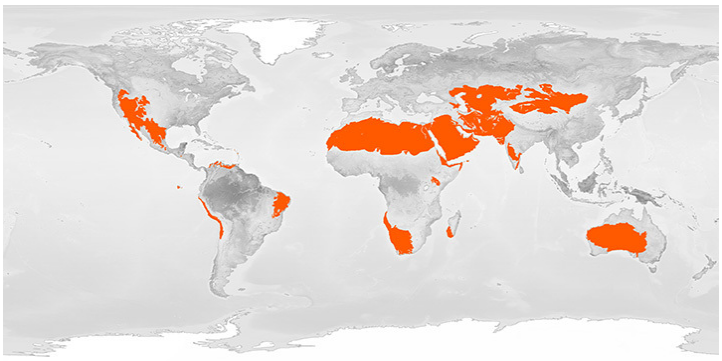


Figure 13.1.1: Map of the world's major deserts. (By NASA; public domain.)

Another type of desert is found in the rain shadow created from prevailing winds blowing over mountain ranges. As the wind drives air up and over mountains, atmospheric moisture is released as snow or rain. Atmospheric pressure is lower at higher elevations, causing the moisture-laden air to cool. Cool air holds less moisture than hot air, and precipitation occurs as the wind rises up the mountain. After releasing its moisture on the windward side of the mountains, the dry air descends on the leeward or downwind side of the mountains to create an arid region with little precipitation called a **rain shadow**. Examples of rain-shadow deserts include the Western Interior Desert of North America and Atacama Desert of Chile, which is the Earth's driest, warm desert.

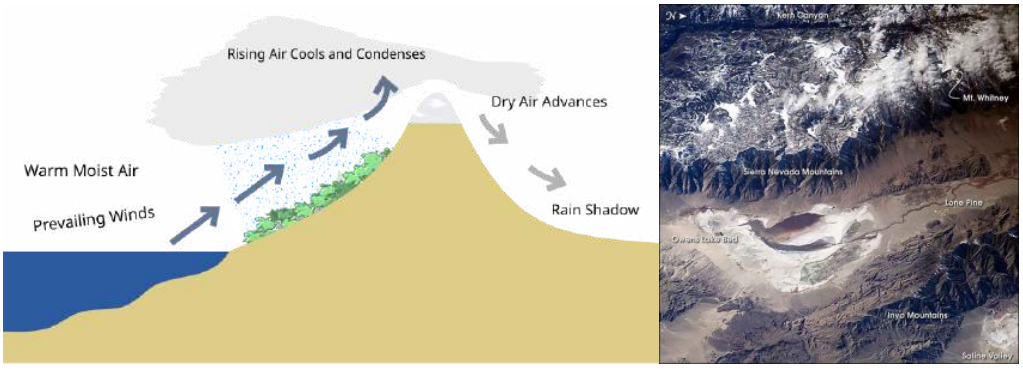


Figure 13.1.2: Left: Mountainous areas in front of the prevailing winds create a rain shadow. (By dodomegg; CC BY 4.0 via Wikimedia Commons.) Right: In this image from the ISS, the Sierra Nevada Mountains are perpendicular to prevailing westerly winds, creating a rain shadow to the east (down in the image). Note the dramatic decrease in snow on the Inyo Mountains.

Finally, **polar deserts**, such as the vast areas of the Antarctic and Arctic are created from sinking cold air that is too cold to hold much moisture. Although they are covered with ice and snow, these deserts have very low average annual precipitation. As a result, Antarctica is Earth's driest continent.

Watch this video on rain shadow deserts:





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13.2: The Origin of Deserts

Atmospheric Circulation

Geographic location, atmospheric circulation, and the Earth's rotation are the primary causal factors of deserts. Solar energy converted to heat is the engine that drives the circulation of air in the atmosphere and water in the oceans. The strength of the circulation is determined by how much energy is absorbed by the Earth's surface, which in turn is dependent on the average position of the Sun relative to the Earth. In other words, the Earth is heated unevenly depending on latitude and angle of incidence. **Latitude** is a line circling the Earth parallel to the equator and is measured in degrees. The equator is 0° and the North and South Poles are 90° N and 90° S respectively. **Angle of incidence** is the angle from perpendicular to the ground surface made by a ray of sunlight shining on the Earth's surface. Tropical zones are located near the equator, where the latitude and angle of incidence are close to 0°, and receive high amounts of solar energy. The poles, which have latitudes and angles of incidence approaching 90°, receive little or almost no solar energy.

The figure shows the generalized air circulation within the atmosphere. Three cells of circulating air span the space between the equator and poles in both hemispheres, the **Hadley cell**, the **Ferrel** or **Mid-latitude cell**, and the **Polar cell**. In the Hadley cell located over the tropics and closest to the equatorial belt, the Sun heats the air and causes it to rise. The rising air cools and releases its contained moisture as tropical rain. The rising dried air spreads away from the equator and toward the North and South Poles, where it collides with dry air in the Ferrel cell. The combined dry air sinks back to the Earth at 30° latitude. This sinking drier air creates belts of predominantly high pressure at approximately 30° north and south of the equator, called the **horse latitudes**. Arid zones between 15° and 30° north and south of the equator thus exist within which desert conditions predominate [2]. The descending air flowing north and south in the Hadley and Ferrel cells also creates prevailing winds called **trade winds** near the equator, and **westerlies** in the temperate zone. Note the arrows indicating general directions of winds in these zones.

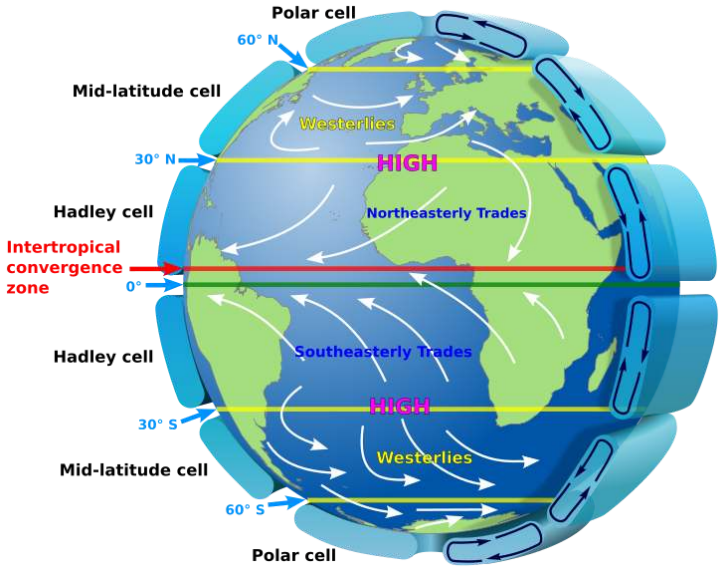


Figure 13.2.1: Generalized atmospheric circulation. (By Kaidor; CC BY-SA 3.0 via Wikimedia Commons.)

Other deserts, like the **Great Basin Desert** that covers parts of Utah and Nevada, owe at least part of their origin to other atmospheric phenomena [3]. The Great Basin Desert, while somewhat affected by sinking air effects from global circulation, is a rain-shadow desert. As westerly moist air from the Pacific rises over the Sierra Nevada and other mountains, it cools and loses moisture as condensation and precipitation on the upwind or rainy side of the mountains.



Figure 13.2.2: USGS Map of the Great Basin Desert.

One of the driest places on Earth is the **Atacama Desert** of northern Chile [4]. The Atacama Desert occupies a strip of land along Chile's coast just north of latitude 30° S, at the southern edge of the trade-wind belt. The desert lies west of the Andes Mountains, in the rain shadow created by prevailing trade winds blowing west. As this warm, moist air crossing the Amazon basin meets the eastern edge of the mountains, it rises, cools, and precipitates much of its water out as rain. Once over the mountains, the cool, dry air descends onto the Atacama Desert. Onshore winds from the Pacific are cooled by the Peru (Humboldt) ocean current. This super-cooled air holds almost no moisture and, with these three factors, some locations in the Atacama Desert have received no measured precipitation for several years. [5]. This desert is the driest, non-polar location on Earth.



Figure 13.2.3: Map of the Atacama Desert (yellow) and surrounding related climate areas (orange).

Notice that the polar regions are also areas of predominantly high pressure created by descending cold dry air, the Polar cell [6]. As with the other cells, cold air, which holds much less moisture than warm air, descends to create polar deserts. This is why historically, land near the North and South Poles has always been so dry.

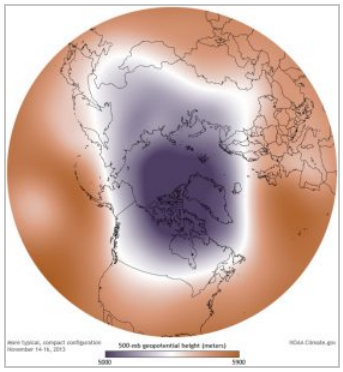


Figure 13.2.4: The polar vortex of mid-November, 2013. This cold, descending air (shown in purple) is characteristic of polar circulation.

Coriolis Effect

In a non-rotating Earth, air would rise at the equator and sink at the poles, creating one circulation cell. However, as noted above, Earth has three atmospheric circulation cells. Why? As objects move on a rotating sphere, an effect called the **Coriolis effect**

occurs which causes a deflection in the motion. In the Northern Hemisphere, this deflection is to the right; in the Southern Hemisphere, it is to the left.

The Earth rotates toward the east where the Sun rises. Think of spinning a weight on a string around your head. The speed of the weight depends on the length of the string. The speed of an object on the rotating Earth depends on its horizontal distance from the Earth's axis of rotation. Higher latitudes are a smaller distance from the Earth's rotational axis, and therefore do not travel as fast eastward as lower latitudes that are closer to the equator. When a fluid like air or water moves from a lower latitude to a higher latitude, the fluid maintains its momentum from moving at a higher speed, so it will travel relatively faster eastward than the Earth beneath at the higher latitudes. This factor causes deflection of movements that occur in north-south directions.

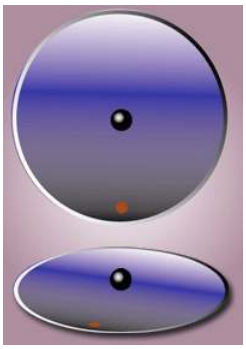


Figure 13.2.5: Example of the motion of a ball thrown on a rotating body. In the inertial frame of reference of the top picture (bird's eye view), the ball moves in a straight line. The observer, represented as a red dot, standing in the rotating frame of reference sees the ball following a curved path. This perceived curvature is due to the Coriolis effect and centrifugal forces.

Another factor in the Coriolis effect also causes deflection of east-west movement due to the angle between the centripetal effect of Earth's spin and gravity pulling toward the Earth's center. This produces a net deflection toward the equator. The total Coriolis deflection on a mass moving in any direction on the rotating Earth results from a combination of these two factors.

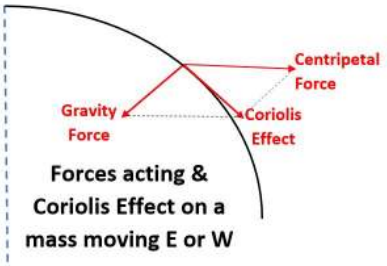


Figure 13.2.6: Forces acting on a mass moving East-West in the Northern Hemisphere on the rotating Earth that produce the Coriolis effect.

The Coriolis effect has two consequences on masses of air (and water) moving on the Earth. In the Northern Hemisphere Hadley cell, the lower altitude air currents are flowing south toward the equator. These are deflected to the right (or west) by the Coriolis effect. This deflected air generates the prevailing trade winds that European sailors used to cross the Atlantic Ocean and reach South America and the Caribbean Islands in their tall ships. This air movement is mirrored in the Hadley cell in the Southern Hemisphere; the lower altitude air current flowing equatorward is deflected to the left, creating trade winds that blow to the northwest.

In the northern Mid-latitude or Ferrel cell, surface air currents flow from the horse latitudes (30°) toward the North Pole, and the Coriolis effect deflects them toward the east, or to the right, producing the zone of westerly winds. In the Southern Hemisphere Mid-latitude or Ferrel cell, the poleward flowing surface air is deflected to the left and flows southeast creating the Southern Hemisphere westerlies.

Another Coriolis-generated deflection produces the Polar cells. At 60° north and south latitude, relatively warmer rising air flows poleward, cooling and converging at the poles, where it sinks in the polar high. This sinking dry air creates the polar deserts, the driest deserts on Earth. Persistence of ice and snow is a result of cold temperatures at these dry locations.



Figure 13.2.7: Inertia of air masses caused by the Coriolis effect in the absence of other forces.

The Coriolis effect operates on all motions on the Earth. Artillerymen must take the Coriolis effect into account on ballistic trajectories when making long-distance targeting calculations. Geologists note how its effect on air and oceanic currents creates deserts in designated zones around the Earth. The Coriolis effect causes the ocean gyres to turn clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere. These currents bring cold water along the west coasts of both North and South America contributing to the drier climates of the Atacama and Central and Southern California. It also affects weather by creating high-altitude, polar jet streams that sometimes push lobes of cold arctic air into the temperate zone, down to as far as latitude 30° from the usual 60°. It also causes low pressure systems and intense tropical storms to rotate counter-clockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. The Coriolis effect acting on both the atmosphere and ocean is a major contributor to climate and weather on the Earth.

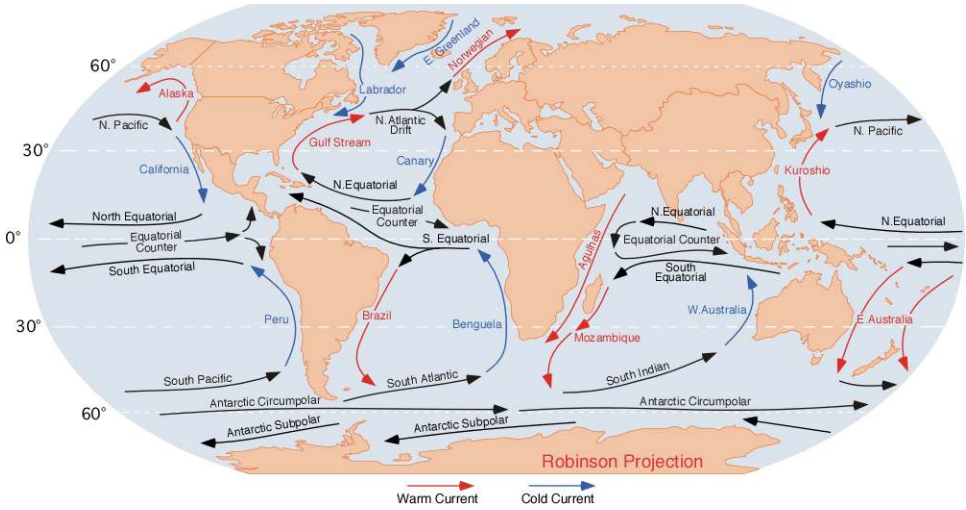


Figure 13.2.8: Gyres of the Earth's oceans.

Watch a video on the Coriolis effect:

https://geo.libretexts.org/@go/page/32401





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13.3: Desert Weathering and Erosion

Weathering takes place in desert climates by the same means as other climates, only at a slower rate. While higher temperatures typically spur faster chemical weathering, water is the main agent of weathering, and lack of water slows both mechanical and chemical weathering. Low precipitation levels also mean less runoff as well as ice wedging. When precipitation does occur in the desert, it is often heavy and may result in flash floods in which a lot of material may be dislodged and moved quickly.



Figure 13.3.1: Weathering and erosion of Canyonlands National Park have created a unique landscape, including arches, cliffs, and spires.

One unique weathering product of deserts is **desert varnish**. Also known as desert **patina** or rock rust, it is thin dark brown layers of clay minerals and iron and manganese oxides that form on very stable surfaces within arid environments. The exact way this material forms is still unknown, though cosmogenic and biologic mechanisms have been proposed.

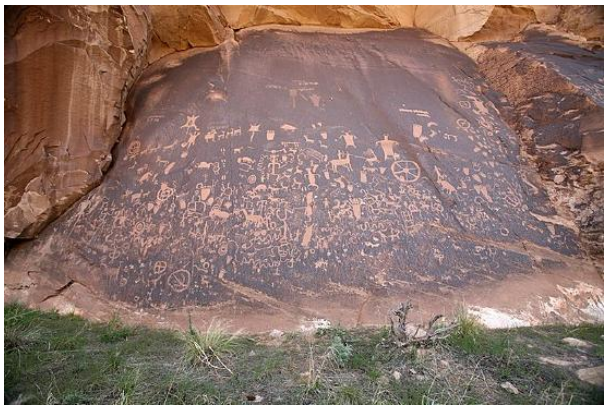


Figure 13.3.2: Newspaper rock, near Canyonlands National Park, has many petroglyphs carved into the desert varnish.

While water is still the dominant agent of erosion in most desert environments, wind is a notable agent of weathering and erosion in many deserts. This includes suspended sediment traveling in **haboobs**, or dust storms, that frequent deserts. Deposits of windblown dust are called **loess**. Loess deposits cover wide areas of the midwestern United States, much of it from rock flour carried by the ice sheets that melted after the last ice age [7]. Loess was also blown from desert regions in the West. Like water, wind is a fluid and can transport sediment. Possessing lower energy than water, wind transport nevertheless moves sand, silt, and dust [8]. As noted in the chapter on Water, the load carried by a fluid is distributed among bedload and suspended load. Similar to water, the loads carried by wind depends on wind velocity.



Figure 13.3.3: A dust storm (haboob) appears on the front of a storm in Texas. (By Jakeorin; CC BY-SA 4.0 via Wikimedia Commons.)

Sand-sized material moves by a process called **saltation** in which sand grains are lifted into the moving air and carried a short distance where they drop and splash into the surface dislodging other sand grains which are then carried a short distance and splash dislodging still others [8].

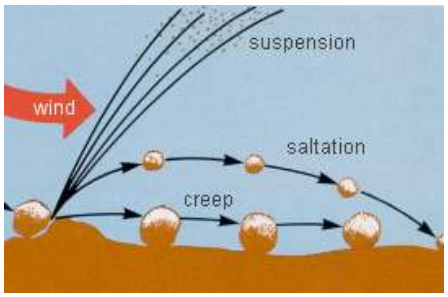


Figure 13.3.4: Diagram showing the mechanics of saltation.

Since saltating sand grains are constantly impacting other sand grains, windblown sand grains are commonly pretty well rounded with frosted surfaces. Saltation is a cascading effect of sand movement creating a zone of windblown sand up to a meter or so above the ground. This zone of saltating sand is a powerful erosive agent in which bedrock features are effectively sandblasted. The fine-grained suspended load is effectively sorted from the sand near the surface, carrying silt and dust in haboobs. Wind is thus an effective sorting agent separating sand and dust-sized (≤70 μm) particles [9]. When wind velocity is high enough to slide or roll materials along the surface, the process is called **creep**.

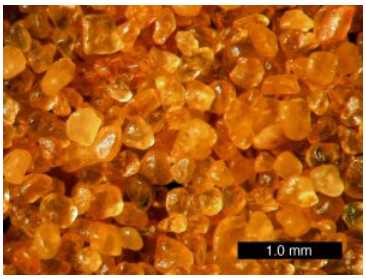


Figure 13.3.5: Enlarged image of frosted and rounded windblown sand grains.

One extreme version of sediment movement was shrouded in mystery for years: **sliding stones**. Also called sailing stones and sliding rocks, these are large moving boulders along flat surfaces in deserts, leaving trails. This includes the famous example of the Racetrack Playa in Death Valley National Park, California. For years, scientists and enthusiasts attempted to explain their movement, with little definitive results [10; 11]. In recent years, several experimental and observational studies have confirmed that thin layers of ice allow the stones to move with high winds providing propulsive energy [12; 13]. These studies include measurements of actual movement, as well as re-creations of the conditions, with resulting movement in the lab.



Figure 13.3.6: A sliding stone at Racetrack Playa in Death Valley National Park, California.

The zone of saltating sand is an effective agent of erosion through sand abrasion. A bedrock outcrop which has such a sandblasted shape is called a **yardang** [14]. Rocks and boulders lying on the surface may be blasted and polished by saltating sand. When predominant wind directions shift, multiple sandblasted and polished faces may appear. Such polished desert rocks are called **ventifacts** [15].



Figure 13.3.7: Left: A yardang in Bolivia. Right: Ventifact from Mojave Desert near Barstow, California.

In places with sand and silt accumulations, clumps of vegetation often anchor sediment that has accumulated on the desert surface. Yet, winds may be sufficient to remove materials not anchored by vegetation. This causes a bowl-shaped depression in the sand called a **blowout** [16].



Figure 13.3.8: Blowout in Texas.

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13.4: Desert Landforms

In the American Southwest, as streams emerge into the valleys from the adjacent mountains, they create desert landforms called **alluvial fans**. When the stream emerges from the narrow canyon, the flow is no longer constrained by the canyon walls [18] and spreads out. At the lower slope angle, the water slows down [17] and drops its coarser load. As the channel fills with this conglomeratic material, the stream is deflected around it. This process develops a system of radial distributary channels in a process similar to how a delta is made by a river entering a body of water [18]. The fan-shaped feature called an alluvial fan.



Figure 13.4.1: Aerial image of an alluvial fan at Badwater in Death Valley National Park, California. (By Marli Miller; via Geologic Time Pics.)

Alluvial fans continue to grow and may eventually coalesce with neighboring fans to form an apron of alluvium along the mountain front called a **bajada** [19].



Figure 13.4.2: Bajadas form in Death Valley Wash in California. (By Dicklyon; CC BY-SA 4.0 via Wikimedia Commons.)

As the mountains erode away and the sediment accumulates first in alluvial fans, then bajadas, the mountains eventually are buried in their own erosional debris. Such buried mountain remnants are called **inselbergs** [20], “island mountains,” as first described by the German geologist Wilhelm Bornhardt (1864-1946).



Figure 13.4.3: Inselbergs in the Mojave Desert, California.

Where the desert valley is an enclosed basin, i.e. streams entering it do not drain out but the water is removed by evaporation, a dry lake bed is formed called a **playa.**



Figure 13.4.4: Lake Tecopa, a playa in the Mojave Desert of southern California. (By Marli Miller; via Geologic Time Pics.)

Playas are among the flattest of all landforms. Such a dry lake bed may cover a large area and be filled after a heavy thunderstorm to only a few inches deep. Playa lakes and desert streams that flow only after rainstorms are called **intermittent** or **ephemeral**. Because of intense thunderstorms, the volume of water transported by ephemeral dtrainage in arid environments can be substantial during a short period of time. Desert soil structures lack organic matter that promotes infiltration by absorbing water. Instead of percolating into the soil, the runoff compacts the ground surface, making the soil **hydrophobic** or water-repellant. Because of this hardpan surface, ephemeral streams may gather water across large areas, suddenly filling with water from storms many miles away.



Figure 13.4.5: Dry wash (or ephemeral stream). (By Finetooth; CC BY-SA 3.0 via Wikimedia Commons.)

High-volume ephemeral flows, called **flash floods**, may move as sheet flows or **sheetwash**, as well as be channeled through normally dry riverbeds called **arroyos** or through canyons. Flash floods are a major factor in desert deposition. Dry channels can fill quickly with ephemeral drainage, creating a mass of water and debris that charges down the arroyo, and even overflowing the banks. Flash floods pose a serious hazard for desert travelers because the storm activity feeding the runoff may be miles away. People hiking or camping in arroyos that have been bone dry for months or years have been swept away by sudden flash floods.



Figure 13.4.6: Flash flood from Hurricane Hilary in 2023 over a road in Darwin Wash in Death Valley National Park in California. (By California Highway Patrol; public domain via NPS.)

Sand

The popular concept of a typical desert is a broad expanse of sand. Geologically, deserts defined by a lack of water and arid regions resembling a sea of sand belong to the category of desert call an erg. An **erg** consists of fine-grained, loose sand, often blown by **aeolian** forces (wind) into dunes [21]. Probably the best-known erg is the Rub' al Khali, which means Empty Quarter, of the Arabian Peninsula. Ergs are also found in Colorado (Great Sand Dunes National Park), Utah (Little Sahara Recreation Area), New Mexico (White Sands National Monument), and California (parts of Death Valley National Park). Ergs are not restricted to deserts, but may form anywhere there is a substantial supply of sand, including as far north as 60°N in Saskatchewan, Canada in the Athabasca Sand Dunes Provincial Park. Coastal ergs exist along lakes and oceans as well, and can be found in places like Oregon, Michigan, and Indiana.



Figure 13.4.7: The Sahara Desert, a sand sea or erg. (By Wonker; CC BY 2.0 via Wikimedia Commons.)

An internal cross section of a sand dune shows a feature called **cross-bedding**. As wind blows up the windward side of the dune, it carries sand to the **dune crest**, depositing layers of sand parallel to the windward (or "**stoss**") side. The sand builds up the crest of the dune and pours over the top until the leeward (downwind or **slip**) face of the dune reaches the angle of repose, the maximum angle which will support the sand pile. Dunes are unstable features and move as the sand erodes from the stoss side and continues to drop down the leeward side covering previous stoss and slip-face layers and creating the cross-beds. Mostly, these are reworked over and over again, but occasionally, the features are preserved in a depression, then lithified. Shifting wind directions and abundant sand sources create chaotic patterns of cross-beds like those seen in Zion National Park of Utah.

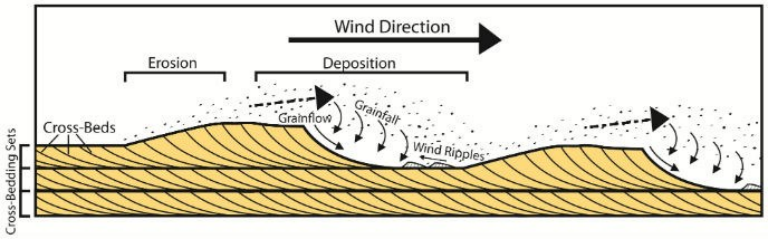


Figure 13.4.8: The formation of cross-bedding in sand dunes. (By David Tarailo; public domain via NPS.)

In the Mesozoic Era, Utah was covered by a series of ergs, with the thickest in Southern Utah, which lithified into sandstone. Perhaps the best known of these sandstone formations is the Navajo Sandstone of Jurassic age. This formation consists of the dramatic cliffs and spires in Zion National Park and covers a large part of the Colorado Plateau. In Arches National Park, a later series of sand dunes covered the Navajo Sandstone and lithified to become the Entrada Formation also during the Jurassic. Erosion of overlying layers exposed fins of the underlying Entrada Sandstone and carved out weaker parts of the fins forming the arches.



Figure 13.4.9: Cross-beds in the Navajo Sandstone at Zion National Park. (By Dr. Igor Smolyar; public domain via Wikimedia Commons.)

As the cement that holds the grains together in these modern sand cliffs disintegrates and the freed grains gather at the base of the cliffs and move down the washes, sand grains may be recycled and redeposited. These Mesozoic sand ergs may represent ancient quartz sands recycled many times from igneous origins in the early Precambrian, just passing now through another cycle of erosion and deposition. One example of this is Coral Pink Sand Dunes State Park in Southwestern Utah, which contains sand that is being eroded from the Navajo Sandstone forming new dunes.

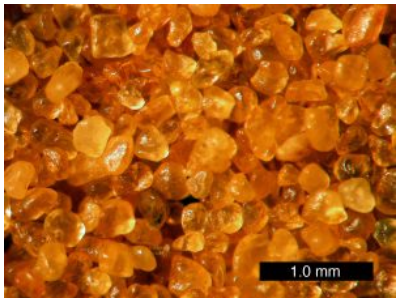


Figure 13.4.9: Enlarged image of frosted and rounded windblown sand grains from Coral Pink Sand Dunes.

Dune Types

Dunes are complex features formed by a combination of wind direction and sand supply, in some cases interacting with vegetation. There are several types of dunes representing the variable of wind direction, sand supply, and vegetative anchoring. **Barchan dunes** or **crescent dunes** form where sand supply is limited and there is a fairly constant wind direction. Barchans move downwind and develop a crescent shape with wings on either side of a dune crest. Barchans are known to actually move over homes, even towns.

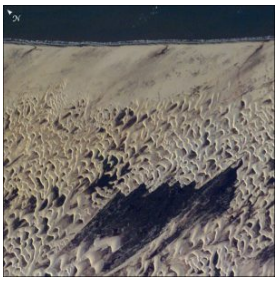


Figure 13.4.10: NASA image of a barchan dune field in coastal Brazil.

Longitudinal dunes or **linear dunes** form where sand supply is greater and the wind blows around a dominant direction, in a back-and-forth manner. They may form ridges tens of meters high lined up with the predominant wind directions.



Figure 13.4.11: Satellite image of longitudinal dunes in Egypt.

Parabolic dunes form where vegetation anchors parts of the sand and unanchored parts blowout. Parabolic dune shape is similar to barchan dunes but usually reversed, and it is determined more by the anchoring vegetation than a strict parabolic form.



Figure 13.4.12: Parabolic dunes, Cape Cod.

Star dunes form where the wind direction is variable in all directions. Sand supply can range from limited to quite abundant. It is the variation in wind direction that forms the star.

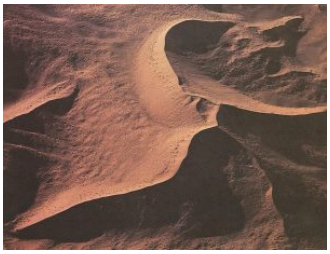


Figure 13.4.13: Star dune in the Sahara Desert.

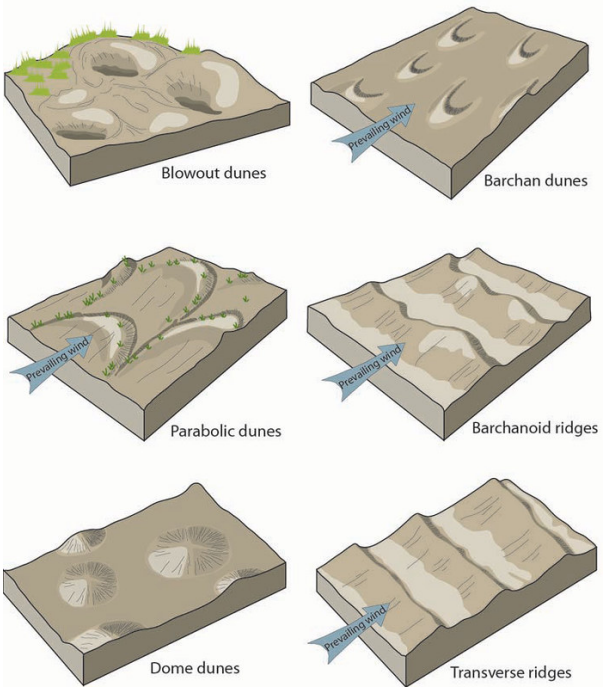


Figure 13.4.14: Types of dunes with prevailing wind direction. (By NPS; public domain via Flickr.)

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13.5: The Great Basin and the Basin and Range

The Great Basin is the largest area of interior drainage in North America, meaning there is no outlet to the ocean and all precipitation remains in the basin or is evaporated. It covers western Utah, most of Nevada, and extends into eastern California, southern Oregon, and southern Idaho. Streams in the Great Basin gather runoff and groundwater discharge and deliver it to lakes and playas within the basin. A subregion within the Great Basin is the **Basin and Range** which extends from the Wasatch Front in Utah, west across Nevada to the Sierra Nevada Mountains of California. The basins and ranges referred to in the name are horsts and grabens, formed by normal fault blocks from the crustal extension, as discussed in the chapters on Plate Tectonics and Crustal Deformation and Earthquakes. The lithosphere of the entire area has stretched by a factor of about 2, meaning from end to end, the distance has doubled over the past 30 million years or so. Valleys without outlets form individual basins, each of which is filled with alluvial sediments leading into playa depositional environments. During the recent Ice Age, the climate was more humid and while glaciers were forming in some of the mountains, pluvial lakes formed covering large areas. During the Ice Age, valleys in much of western Utah and eastern Nevada were covered by Lake Bonneville. As the climate became arid after the Ice Age, Lake Bonneville dried leaving the Great Salt Lake in Utah as a remnant.



Figure 13.5.1: Map of the Great Basin.

The desert of the Basin and Range extends from about 35°N to near 40°N and results from a rain shadow effect created by westerly winds from the Pacific rising and cooling over the Sierras, becoming depleted of moisture by precipitation on the western side. The result is relatively dry air descending across Nevada and western Utah. A journey from the Wasatch Front southwest to the Pacific Ocean will show stages of desert landscape evolution from the fault block mountains of Utah with sharp peaks and alluvial fans at the mouths of canyons, through landscapes in Southern Nevada with bajadas along the mountain fronts, to the landscapes in the Mojave Desert of California with subdued inselbergs sticking up through a sea of coalesced bajadas. These landscapes illustrate the evolutionary stages of desert landscape development.



Figure 13.5.2: Aerial view of several parallel mountain ranges in and intervening basins in the Basin and Range, between the Last Chance Range and Eureka Valley (in Death Valley National Park in California) and the Sierra Nevada Mountains. The Eureka Sand Dunes can be seen in the lower left part of the photo. (By Marli Miller; via Geologic Time Pics.)

Desertification

When previously arable land suitable for agriculture transforms into desert, this process is called **desertification**. Plants and humus-rich soil promote groundwater infiltration and water retention. When an area becomes more arid due to changing environmental conditions, the plants and soil become less effective in retaining water, creating a positive feedback loop of desertification. This self-reinforcing loop spirals into increasingly arid conditions and further enlarges the desert regions.

Desertification may be caused by human activities, such as unsustainable crop cultivation practices, overgrazing by livestock, overuse of groundwater, and global climate change. Human-caused desertification is a serious worldwide problem [22]. The world map figure shows what areas are most vulnerable to desertification. Note the red and orange areas in the western and midwestern regions of the United States, which also cover large areas of arable land used for raising food crops and animals. The creation of the Dust Bowl in the 1930s is a classic example of a high-vulnerability region impacted by human–caused desertification. As demonstrated in the Dust Bowl, conflicts may arise between agricultural practices and conservation measures. Mitigating desertification while allowing farmers to make a survivable living requires public and individual education to create community support and understanding of sustainable agriculture alternatives.

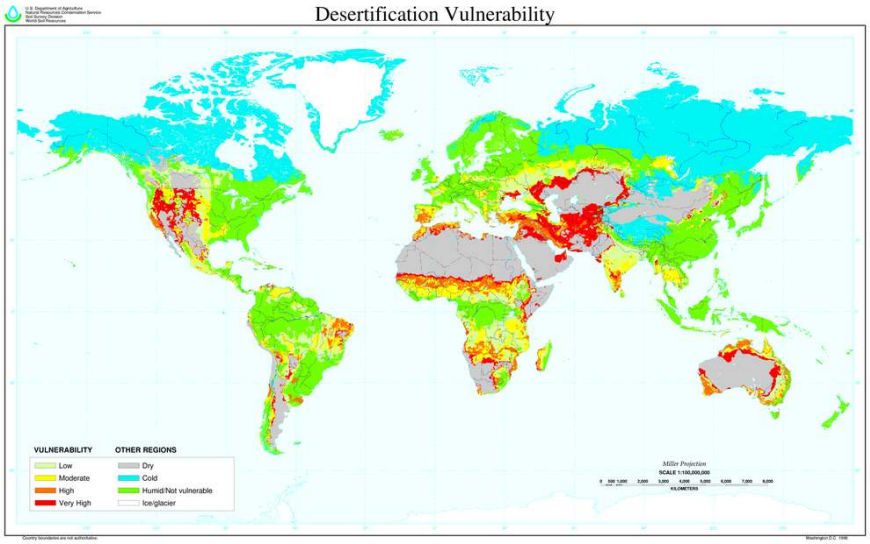


Figure 13.5.3: World map showing desertification vulnerability.

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13.S: Summary

Summary

Approximately 30% of Earth's surface is arid lands, the location of which is determined by latitude, atmospheric circulation, and terrain. The arid belts between 15° and 30° north and south latitudes are produced by descending air masses associated with major cells in the atmosphere and include the major deserts like the Sahara in Africa and the Middle East. Rain shadow deserts lie behind mountain ranges or long land expanses in zones of prevailing winds like the deserts of western North America, the Atacama of South America, and the Gobi of Asia. Dry descending air also creates the polar deserts at the poles.

Major atmospheric circulation involves the Hadley cells, midlatitude or Ferrel cells, and the polar cells in each hemisphere. Warmed and rising air in the Hadley cells rains back on the tropics and moves toward the poles as drier air. That air meets the drier equatorward moving air of the Ferrel cells. This dry air descends in the arid zones, called the horse latitudes, to produce the arid belts in each hemisphere. Rotation of the Earth creates the Coriolis effect that deflects these moving air masses to produce zones of prevailing winds, the trade winds in the subtropics and the westerlies in the midlatitudes. A combination of latitude, rain shadow, and cold adjacent ocean currents causes the Atacama Desert of northern Chile, the driest desert on Earth.

Weathering in deserts takes place just as in other climes only slower because of less water. Desert varnish is a weathering product unique to desert environments. As in more humid climes, water is the main agent of erosion although wind is a notable agent. Large dust storms called haboobs transport large amounts of sediment that may accumulate in sand seas called ergs or finer-grained loess deposits. Sand transport occurs mainly by saltation in which grain to grain impact causes frosting of grain surfaces. Sandblasting by persistent winds produces stones with polished surfaces called ventifacts and sculpted bedrock features called yardangs.

Landforms produced in desert environments include alluvial fans, bajadas, inselbergs, and playas. Windblown sand can accumulate as dunes. The forms of dunes, like barchans, parabolic, longitudinal, and star dunes, relate to the abundance of sand supply and wind direction as well as presence of vegetation. The internal structure of dunes shows cross bedding. Fossil dunes in an ancient desert leave cross bedding in places like Zion and Arches National Parks in Utah showing shifting wind directions in these ancient environments. Ephemeral streams in desert regions may carry water only after storms and pose risk of flash floods.

The Great Basin of North America is an enclosed basin with no drainage to the ocean. The only exit for precipitation there is by evaporation. Travels in the Great Basin show stages of development of desert landscapes from playas and alluvial fans, to bajadas, to inselbergs which are eroded mountains buried in their own erosional debris.

Poor land management can result in dying vegetation and loss of soil moisture producing an accelerating process of desertification in which once productive land is degraded into unproductive desert. This is a serious worldwide problem.

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CHAPTER OVERVIEW

14: Glaciers

Learning Objectives

- Differentiate the different types of glaciers and contrast them with sea ice.
- Describe how glaciers form, move, erode, and create landforms.
- Describe glacial budget: zones of accumulation, equilibrium, and melting.
- Identify glacial erosional and depositional landforms and interpret their origin.
- Describe the history and causes of past glaciations and their relationship to climate and sea-level changes and isostatic rebound.



Figure 14.1: Glacier in the Bernese Alps. (By Dirk Beyer; CC BY-SA 3.0 via Wikimedia Commons.)

The Earth's **cryosphere**, or ice, has a unique set of erosional and depositional features compared to its **hydrosphere**, or liquid water. This ice exists primarily in two forms, glaciers and icebergs. **Glaciers** are large accumulations of ice that exist year-round on the land surface. In contrast, masses of ice floating on the ocean are **icebergs**, although they may have had their origin in glaciers.

Glaciers cover about 10% of the Earth's surface and are powerful erosional agents that sculpt the planet's surface. These enormous masses of ice usually form in mountainous areas that experience cold temperatures and high precipitation. Glaciers also occur in low-lying areas such as Greenland and Antarctica that remain extremely cold year-round.

14.1: Glacier Formation
14.2: Glacier Movement
14.3: Glacial Budget
14.4: Glacial Landforms
14.5: Ice Age Glaciations
14.S: Summary

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14.1: Glacier Formation

Glaciers form when repeated annual snowfall accumulates deep layers of snow that are not completely melted in the summer. Thus there is an accumulation of snow that builds up into deep layers. **Perennial snow** is a snow accumulation that lasts all year. A thin accumulation of perennial snow is a **snow field**. Over repeated seasons of perennial snow, the snow settles, compacts, and bonds with underlying layers. The amount of void space between the snow grains diminishes. As the old snow gets buried by more new snow, the older snow layers compact into **firn**, or **névé**, a granular mass of ice crystals. As the firn continues to be buried, it compresses and recrystallizes, the void spaces become smaller, and the ice becomes less porous, eventually turning into **glacial ice**. Solid glacial ice still retains a fair amount of void space that traps air. These small air pockets provide records of the past atmospheric composition.

There are three general types of glaciers: alpine or valley glaciers, ice sheets and ice caps. Most **alpine glaciers** are located in the world's major mountain ranges such as the Andes, Rockies, Alps, and Himalayas, usually occupying long, narrow valleys. Alpine glaciers may also form at lower elevations in areas that receive high annual precipitation such as the Olympic Peninsula in Washington state.



Figure 14.1.1: Greenland ice sheet.

Ice sheets (also called **continental glaciers**) form across millions of square kilometers of land and are thousands of meters thick. These are thick accumulations of ice that occupy a large geographical area. Earth's largest ice sheets are located on Greenland and Antarctica. The Greenland Ice Sheet is the largest ice mass in the Northern Hemisphere with an extensive surface area over 2 million square kilometers (1,242,700 square miles) and average thickness up to 1500 meters (5000 feet or almost a mile) and has a volume estimated at nearly 3 million cubic kilometers (~102 billion cubic feet) [1].



Figure 14.1.2: Thickness of Greenland ice sheet in meters (Source: Eric Gaba).

The Antarctic Ice Sheet is much larger and covers almost the entire continent. The thickest parts of the ice sheet are over 4000 meters thick (>13,000 feet or 2.5 miles) and its weight depresses the Antarctic bedrock to below sea level in many places beneath the ice [2]. The Antarctic Ice Sheet contains the most ice as illustrated by the figure below comparing cross-sectional views of both ice sheets.

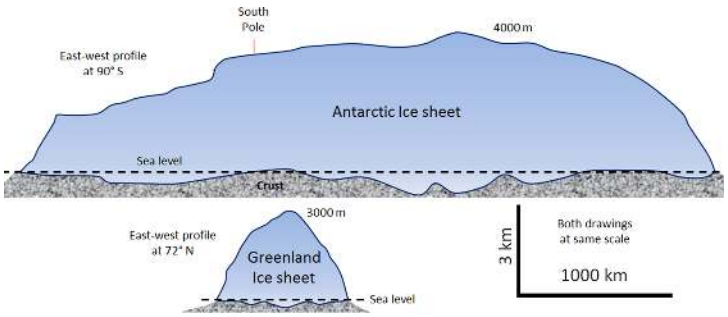


Figure 14.1.3: Cross-sectional view of both Greenland and Antarctic ice sheets drawn to scale for size comparison. (Source: Steve Earle)

Ice cap glaciers are smaller versions of ice sheets that cover less than 50,000 square kilometers, usually occupy higher elevations and may cover tops of mountains. There are several ice caps on Iceland. A small ice cap called Snow Dome is near Mt. Olympus on the Olympic Peninsula in the state of Washington.



Figure 14.1.4: Left: Snow Dome ice cap near Mt. Olympus, Washington. (By NPS; public domain via Wikimedia Commons.) Right: Vatnajokull ice cap in Iceland. (By NASA; public domain via Wikimedia Commons.)

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14.2: Glacier Movement

As the ice accumulates, it begins to flow downward under its own weight. In an early study of glacier movement conducted in 1948 on the Jungfraufirn Glacier in the Alps, glaciologists installed hollow vertical rods in the ice and measured the changes in movement over two years. The study showed that the ice at the surface was fairly rigid and the ice within the glacier was actually flowing downhill. Ice actually melts under pressure (one of the unique properties of water) so ice at the base of a typical glacier is actually melting. About half of the overall glacial movement was from sliding on a film of meltwater along the bedrock surface and half from internal flow [3]. These studies show that the ice near the surface (about the upper 165 feet [50 meters] depending on location, temperature, and flow rate) is rigid and brittle [4]. This upper zone is the **brittle zone**, the portion of the ice in which ice breaks when it moves. Large cracks along the top of a glacier are called **crevasses**. These crevasses can be covered and hidden by a snow bridge and thus are a hazard for glacier travelers.



Figure 14.2.1: Left: Glacial crevasses. Right: Crevasse on the Easton Glacier in the North Cascades.

Below the brittle zone, the pressure typically exceeds 100 kilopascals (kPa), which is approximately 100,000 times atmospheric pressure. Under this applied force, the ice no longer breaks, but rather it bends or flows in a zone called the **plastic zone**. This plastic zone represents the great majority of glacier ice. The plastic zone contains a fair amount of sediment of various grades from boulders to silt and clay. As the bottom of the glacier slides and grinds across the bedrock surface, these sediments act as grinding agents and create a zone of significant erosion.

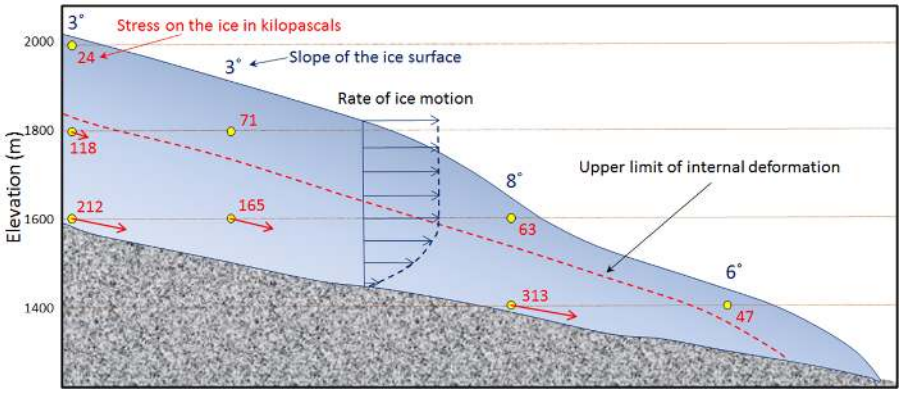


Figure 14.2.2: Cross-section of an alpine glacier showing stress (red numbers) increases with depth under the ice. The ice will deform and flow where the stress is greater than 100 kilopascals, and the relative extent of that deformation is depicted by the red arrows. Down slope movement is shown with blue arrows. The upper ice above the red dashed line does not flow but is pushed along with the lower ice. The rate of ice movement is slow near the bottom, and fastest in the middle, with the top ice being carried along on the ice below. (Source: Steve Earle)

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14.3: Glacial Budget

A **glacial budget** is like a bank account, with the ice being the existing balance. If there is more income (snow accumulating in winter) than expense (snow and ice melting in summer), then the glacial budget shows growth. A positive or negative balance of ice in the overall glacial budget determines whether a glacier advances or retreats, respectively. The area where the ice balance is growing is called the **zone of accumulation**. The area where ice balance is shrinking is called the **zone of ablation**.

The diagram shows these two zones and the equilibrium line. In the zone of accumulation, the snow accumulation rate exceeds the snow melting rate and the ice surface is always covered with snow. The **equilibrium line**, also called the **snow line** or **firn line**, marks the boundary between the zones of accumulation and ablation. Below the equilibrium line in the zone of ablation, the melting rate exceeds snow accumulation, leaving the bare ice surface exposed. The position of the firn line changes during the season and from year to year as a reflection of a positive or negative ice balance in the glacial budget. Of the two variables affecting a glacier‘s budget, winter accumulation and summer melt, summer melt matters most to a glacier’s budget. Cool summers promote glacial advance and warm summers promote glacial retreat [5].

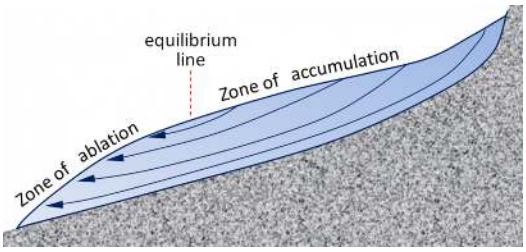


Figure 14.3.1: Cross-sectional view of an alpine glacier showing internal flow lines, zone of accumulation, equilibrium line, and zone of ablation.

If warmer summers promote glacial retreat, then global climate warming over decades and centuries will accelerate glacial melting and retreating significantly. Global warming due to human burning of fossil fuels [6] is causing the ice sheets to melt (or lose mass) at an increasing rate over years and decades rather than over centuries [7]. Current glacial melting is contributing to rising sea-levels faster than expected based on the previous history.

As the Antarctica and Greenland ice sheets melt during global warming, they become thinner or deflate [8; 9]. The edges of the ice sheets break off and fall into the ocean, a process called **calving**, becoming floating **icebergs**. A **fjord** is a steep-walled valley flooded with sea water. The narrow shape of a fjord has been carved out by a glacier during a cooler climate period. During a warming trend, glacial meltwater may raise the sea level in fjords and flood formerly dry valleys.

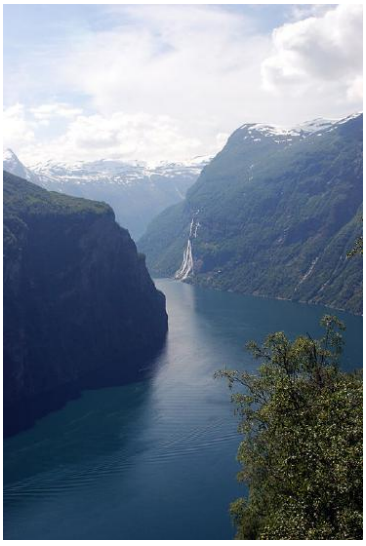


Figure 14.3.2: Fjord in Norway.

Glacial retreat and deflation are well-illustrated in the 2009 TED Talk by James Balog.



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14.4: Glacial Landforms

Both alpine and continental glaciers create two categories of landforms: erosional and depositional. Erosional landforms are formed by the removal of material. Depositional landforms are formed by the addition of material. Because glaciers were first studied by 18th and 19th century geologists in Europe, the terminology applied to glaciers and glacial features contains many terms derived from European languages.

Erosional Glacial Landforms

Erosional landforms are created when moving masses of glacial ice slide and grind over bedrock. Glacial ice contains large amounts of poorly sorted sand, gravel, and boulders that have been plucked and pried from the bedrock. As the glaciers slide across the bedrock, they grind these sediments into a fine powder called **rock flour**. Rock flour acts as fine grit that polishes the surface of the bedrock to a smooth finish called **glacial polish**. Larger rock fragments scrape over the surface creating elongated grooves called **glacial striations**.



Figure 14.4.1: Left: A polished and striated rock in Yosemite. Right: Glacial striations in Mt. Rainier National Park.

Alpine glaciers produce a variety of unique erosional landforms, such as U-shaped valleys, arêtes, cirques, tarns, horns, cols, hanging valleys, and truncated spurs. In contrast, stream-carved canyons have a V-shaped profile when viewed in cross-section. Glacial erosion transforms a former V-shaped stream valley into a **U-shaped valley**. Glaciers are typically wider than streams of similar length, and since glaciers tend to erode both at their bases and their sides, they erode V-shaped valleys into relatively flat-bottomed broad valleys with steep sides and a distinctive “U” shape [5]. As seen in the images, Little Cottonwood Canyon near Salt Lake City, Utah was occupied by an Ice Age glacier that extended down to the mouth of the canyon and into Lake Bonneville [10]. Today, that U-shaped valley hosts many erosional landforms, including polished and striated rock surfaces. In contrast, Big Cottonwood Canyon to the north of Little Cottonwood Canyon has retained the V-shape in its lower portion, indicating that its glacier did not extend clear to its mouth, but was confined to its upper portion.

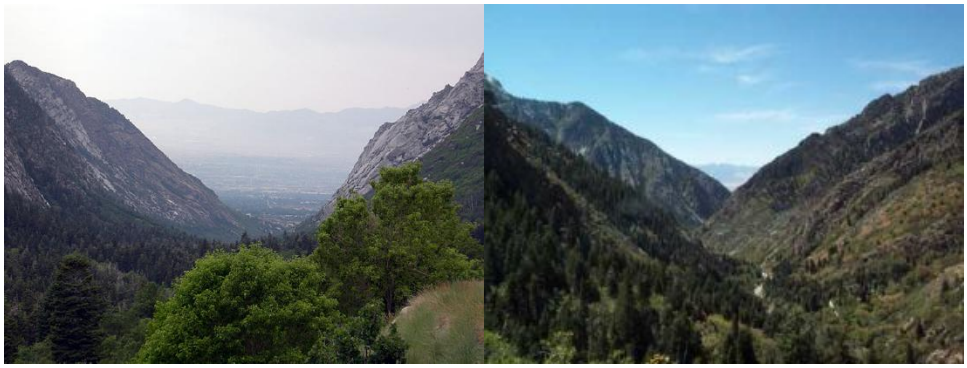


Figure 14.4.2: Left: The U-shape of the Little Cottonwood Canyon, Utah, as it enters into the Salt Lake Valley. Right: V-shaped Big Cottonwood Canyon, carved by a stream.

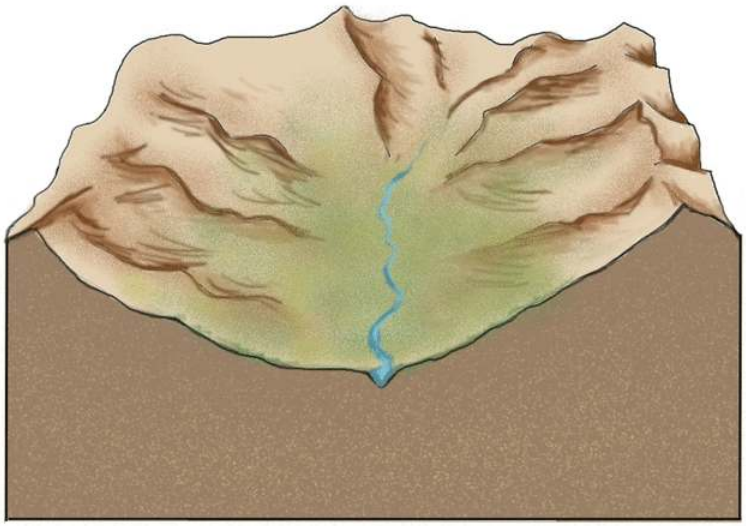


Figure 14.4.3: Formation of a glacial valley. Glaciers change the shape of the valley from a “V” shape to a “U”.

When glaciers carve two U-shaped valleys adjacent to each other, the ridge between them tends to be sharpened into a sawtooth feature called an **arête**. At the head of a glacially carved valley is a bowl-shaped feature called a **cirque**. The cirque represents where the head of the glacier eroded the mountain by plucking rock away from it and the weight of the thick ice eroded out a bowl. After the glacier is gone, the bowl at the bottom of the cirque often fills with precipitation and is occupied by a lake, called a **tarn**. When three or more mountain glaciers erode headward at their cirques, they produce **horns**, steep-sided, spire-shaped mountains. Low points along arêtes or between horns are mountain passes termed **cols**. Where a smaller tributary glacier flows into a larger trunk glacier, the smaller glacier cuts down less. Once the ice has gone, the tributary valley is left as a **hanging valley**, sometimes with a waterfall plunging into the main valley. As the trunk glacier straightens and widens a V-shaped valley and erodes the ends of side ridges, a steep triangle-shaped cliff is formed called a **truncated spur**.



Figure 14.4.4: Left: Cirque with Upper Thornton Lake in the North Cascades National Park, Washington. Photo by Walter Siegmund. Middle: An example of a horn, Kinnerly Peak, Glacier National Park, Montana. Photo by USGS. Right: Bridalveil Fall in Yosemite National Park, California, is a good example of a hanging valley.

Depositional Glacial Landforms

Depositional landforms and materials are produced from deposits left behind by a retreating glacier. All glacial deposits are called **drift**. These include till, tillites, diamictites, terminal moraines, recessional moraines, lateral moraines, medial moraines, ground moraines, silt, outwash plains, glacial erratics, kettles, kettle lakes, crevasses, eskers, kames, and drumlins.

Glacial ice carries a lot of sediment, which when deposited by a melting glacier is called **till**. Till is poorly sorted with grain sizes ranging from clay and silt to pebbles and boulders. These clasts may be striated. Many depositional landforms are composed of till. The term **tillite** refers to lithified rock having glacial origins. **Diamictite** refers to a lithified rock that contains a wide range of clast sizes; this includes glacial till but is a more objective and descriptive term for any rock with a wide range of clast sizes.



Figure 14.4.5: Proterozoic diamictite from southern California. (By Marli Miller; via Geologic Time Pics.)

Moraines are mounded deposits consisting of glacial till carried in the glacial ice and rock fragments dislodged by mass wasting from the U-shaped valley walls. The glacier acts like a conveyor belt, carrying and depositing sediment at the end of and along the sides of the ice flow. Because the ice in the glacier is always flowing downslope, all glaciers have moraines build up at their terminus, even those not advancing.

Moraines are classified by their location with respect to the glacier. A **terminal moraine** is a ridge of till located at the end or **terminus** of the glacier [11]. **Recessional moraines** are left as glaciers retreat and there are pauses in the retreat. **Lateral moraines** accumulate along the sides of the glacier from material mass wasted from the valley walls. When two tributary glaciers merge, the two lateral moraines combine to form a **medial moraine** running down the center of the combined glacier. **Ground moraine** is a veneer of till left on the land as the glacier melts.



Figure 14.4.6: Left: Lateral moraines of Little Cottonwood and Bell canyons, Utah. Right: Medial moraines in the middle of the tributary glaciers and lateral moraines on the sides of the glaciers.

In addition to moraines, glaciers leave behind other depositional landforms. Silt, sand, and gravel produced by the intense grinding process are carried by streams of water and deposited in front of the glacier in an area called the **outwash plain**. Retreating glaciers may leave behind large boulders that don't match the local bedrock. These are called **glacial erratics**. When continental glaciers retreat, they can leave behind large blocks of ice within the till. These ice blocks melt and create a depression in the till called a **kettle**. If the depression later fills with water, it is called a **kettle lake**.

If meltwater flowing over the ice surface descends into crevasses in the ice, it may find a channel and continue to flow in sinuous channels within or at the base of the glacier. Within or under continental glaciers, these streams carry sediments. When the ice recedes, the accumulated sediment is deposited as a long sinuous ridge known as an **esker**. Meltwater descending down through the ice or over the margins of the ice may deposit mounds of till in hills called **kames**.

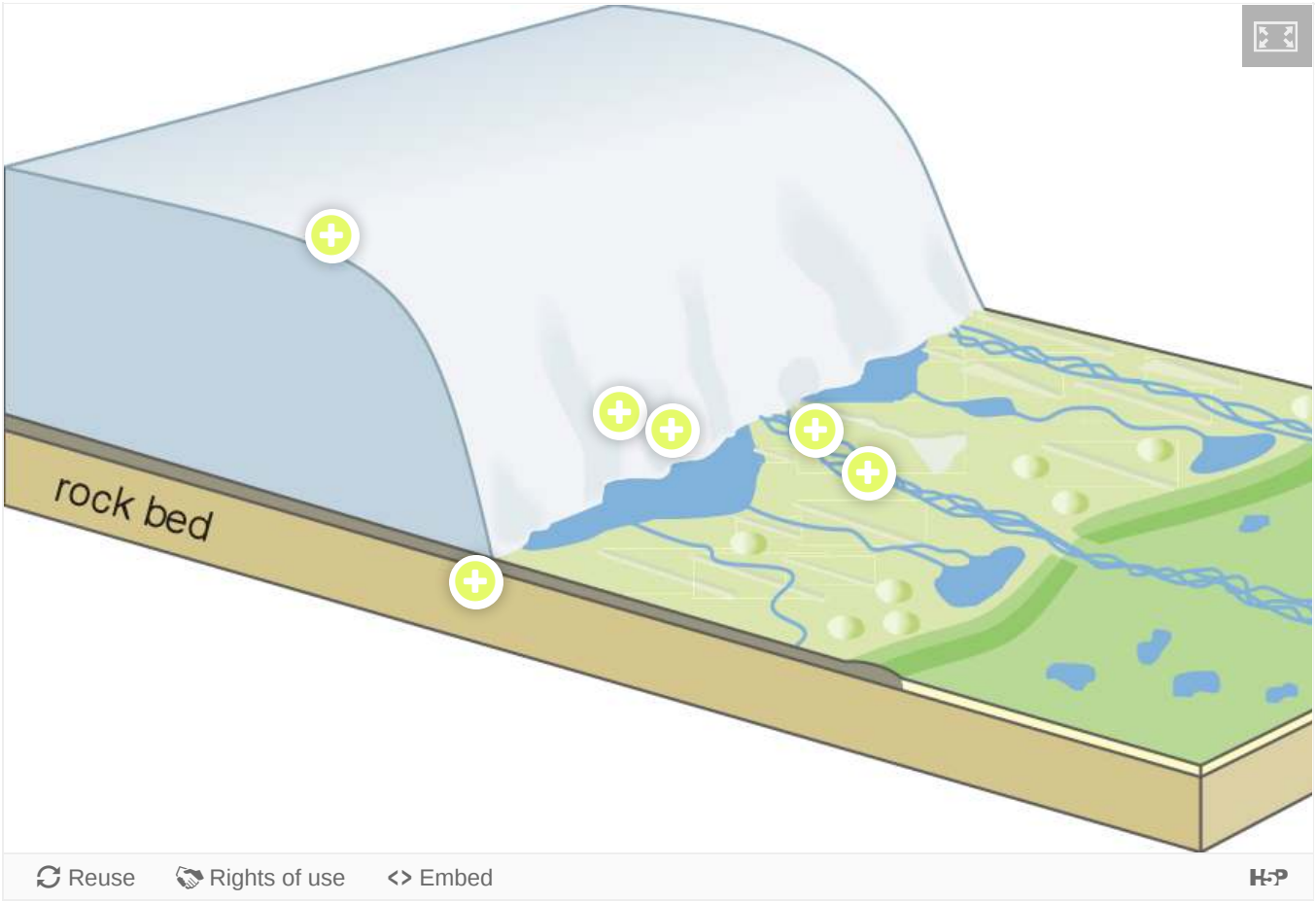
Drumlins are common in continental glacial areas of Germany, New York, and Wisconsin, where they typically are found in fields with great numbers. A **drumlin** is an elongated asymmetrical teardrop-shaped hill reflecting ice movement with its steepest side pointing upstream to the flow of ice and its streamlined or low-angled side pointing downstream in the direction of ice movement.



Figure 14.4.7: A small group of Ice Age drumlins in Germany.

Glacial scientists debate the origins of drumlins. A leading idea is that drumlins are created from accumulated till being compressed and sculpted under a glacier that retreated then advanced again over its own ground moraine. Another idea is that meltwater catastrophically flooded under the glacier and carved the till into these streamlined mounds. Still another proposes that the weight of the overlying ice statically deformed the underlying till [12; 13].

Check your understanding of glacial landforms:



Glacial Lakes

Glacial lakes are commonly found in alpine environments. A lake confined within a glacial cirque is called a tarn. A tarn forms when the depression in the cirque fills with precipitation after the ice is gone. Examples of tarns include Silver Lake near Brighton Ski resort in Big Cottonwood Canyon, Utah and Avalanche Lake in Glacier National Park, Montana.



Figure 14.4.8: Tarn in a cirque.

When recessional moraines create a series of isolated basins within a glaciated valley, the resulting chain of lakes is called **paternoster lakes**.



Figure 14.4.9: Paternoster lakes.

Lakes filled by glacial meltwater often look milky due to finely ground material called rock flour suspended in the water. Lakes that occupy long glacially carved depressions are known as **finger lakes** [5].



Figure 14.4.10: Satellite view of Finger Lakes region of New York.

Proglacial lakes form along the edges of all the largest continental ice sheets, such as Antarctica and Greenland. The crust is depressed isostatically by the overlying ice sheet and these basins fill with glacial meltwater. Many such lakes, some of them huge, existed at various times along the southern edge of the Laurentide Ice Sheet during the last ice age. Lake Agassiz, Manitoba, Canada, is a classic example of a proglacial lake. Lake Winnipeg serves as the remnant of a much larger proglacial lake.

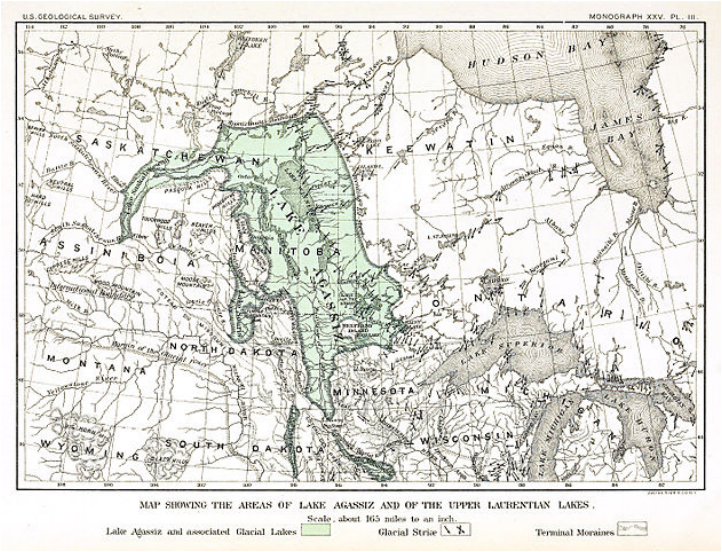


Figure 14.4.11: Extent of Lake Agassiz.

Other proglacial lakes were formed when glaciers dammed rivers and flooded the river valley. A classic example is Lake Missoula, which formed when a lobe of the Laurentide Ice Sheet blocked the Clark Fork River about 18,000 years ago. Over about 2000 years the ice dam holding back Lake Missoula failed several times. During each breach, the lake emptied across parts of eastern Washington, Oregon, and Idaho into the Columbia River Valley and eventually the Pacific Ocean. After each breach, the dam reformed and the lake refilled. Each breach produced a catastrophic flood over a few days. Scientists estimate that this cycle of ice dam, proglacial lake, and torrential massive flooding happened at least 25 times over a span of 20 centuries. The rate of each outflow is believed to have equaled the combined discharge of all of Earth's current rivers combined. The landscape produced by these massive floods is preserved in the Channeled Scablands of Idaho, Washington, and Oregon [5].



Figure 14.4.12: View of Channeled Scablands in eastern Washington showing huge potholes and massive erosion.

Pluvial lakes form in humid environments that experience low temperatures and high precipitation. During the last glaciation, most of the western United States' climate was cooler and more humid than today. Under these low-evaporation conditions, many large lakes, called **pluvial lakes**, formed in the basins of the Basin and Range Province. Two of the largest were Lake Bonneville and Lake Lahontan. Lake Lahontan was in northwestern Nevada. The figure illustrates the tremendous size of Lake Bonneville, which occupied much of western Utah and into eastern Nevada. The lake level fluctuated greatly over the centuries leaving several pronounced old shorelines marked by wave-cut terraces. These old shorelines can be seen on mountain slopes throughout the western portion of Utah, including the Salt Lake Valley, indicating that the now heavily urbanized valley was once filled with hundreds of feet of water. Lake Bonneville's level peaked around 18,000 years ago [14] when a breach occurred at Red Rock Pass in Idaho and water spilled into the Snake River. The flooding rapidly lowered the lake level and scoured the Idaho landscape across the Pocatello Valley, the Snake River Plain, and Twin Falls. The floodwaters ultimately flowed into the Columbia River across part of the scablands area at an incredible discharge rate of about 4,750 cu km/sec (1,140 cu mi/sec) [15]. For comparison, this discharge rate would drain the volume of Lake Michigan completely dry within a few days.

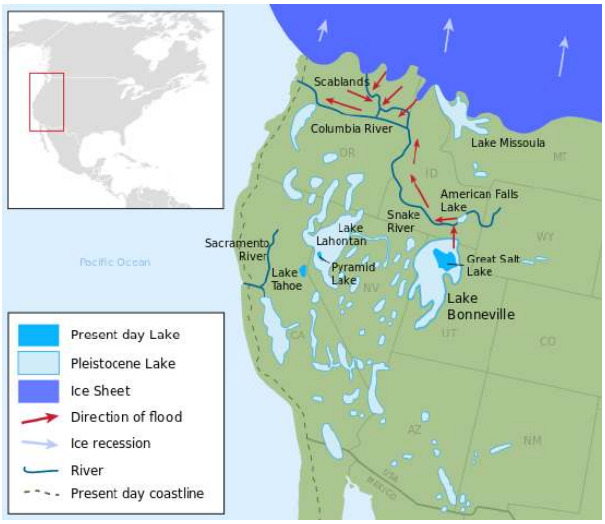


Figure 14.4.13: Pluvial lakes in the western United States

The five Great Lakes in North America's upper Midwest are proglacial lakes that originated during the last ice age. The lake basins were originally carved by the encroaching continental ice sheet. The basins were later exposed as the ice retreated about 14,000 years ago and filled by precipitation.



Figure 14.4.14: The Great Lakes.

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14.5: Ice Age Glaciations

A **glaciation** (or **ice age**) occurs when the Earth's climate becomes cold enough that continental ice sheets expand, covering large areas of land. Four major, well-documented glaciations have occurred in Earth's history: one during the Archean-early Proterozoic Eon, ~2.5 billion years ago; another in late Proterozoic Eon, ~700 million years ago; another in the Pennsylvanian, 323 to 300 million years ago, and most recently during the Pliocene-Pleistocene epochs starting 2.5 million years ago. Some scientists also recognize a minor glaciation around 440 million years ago in Africa.

The best-studied glaciation is, of course, the most recent. This **infographic** illustrates the glacial and climate changes over the last 20,000 years, ending with those caused by human actions since the Industrial Revolution. The Pliocene-Pleistocene glaciation was a series of several glacial cycles, possibly 18 in total. Antarctic ice-core records exhibit especially strong evidence for eight glacial advances occurring within the last 420,000 years [16]. The last of these is known in popular media as "The Ice Age," but geologists refer to it as the **Last Glacial Maximum**. The glacial advance reached its maximum between 26,500 and 19,000 years ago [10; 17].

Former ice sheets, present during the Last Glacial Maximum event (or the last ice age) in North America, are called the **Laurentide Ice Sheet**.

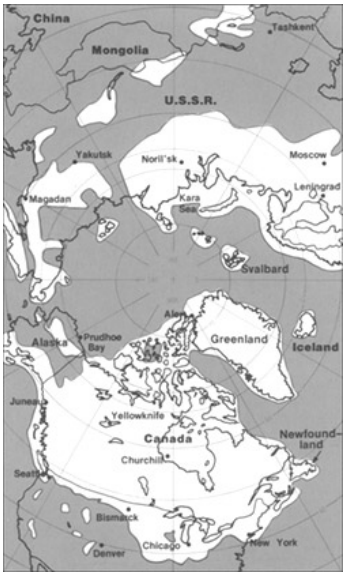


Figure 14.5.1: Maximum extent of Laurentide Ice Sheet.

Causes of Glaciations

Glaciations occur due to both long-term and short-term factors. In the geologic sense, long-term means a scale of tens to hundreds of millions of years and short-term means a scale of hundreds to thousands of years.

Long-term causes include plate tectonics breaking up the supercontinents, moving land masses to high latitudes near the north or south poles, and changing ocean circulation. For example, the closing of the Panama Strait and isolation of the Pacific and Atlantic Oceans may have triggered a change in precipitation cycles, which combined with a cooling climate to help expand the ice sheets.

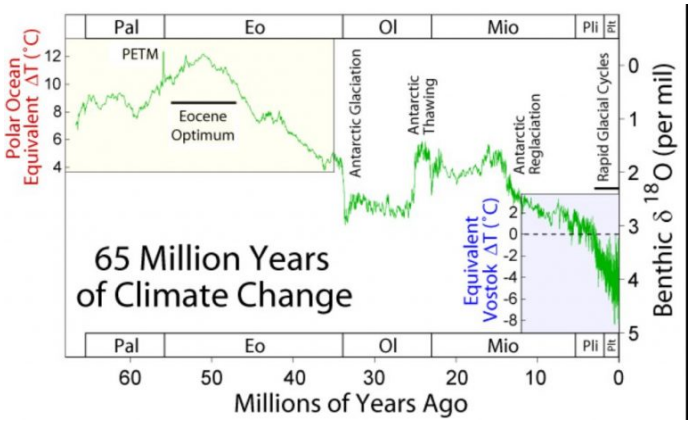
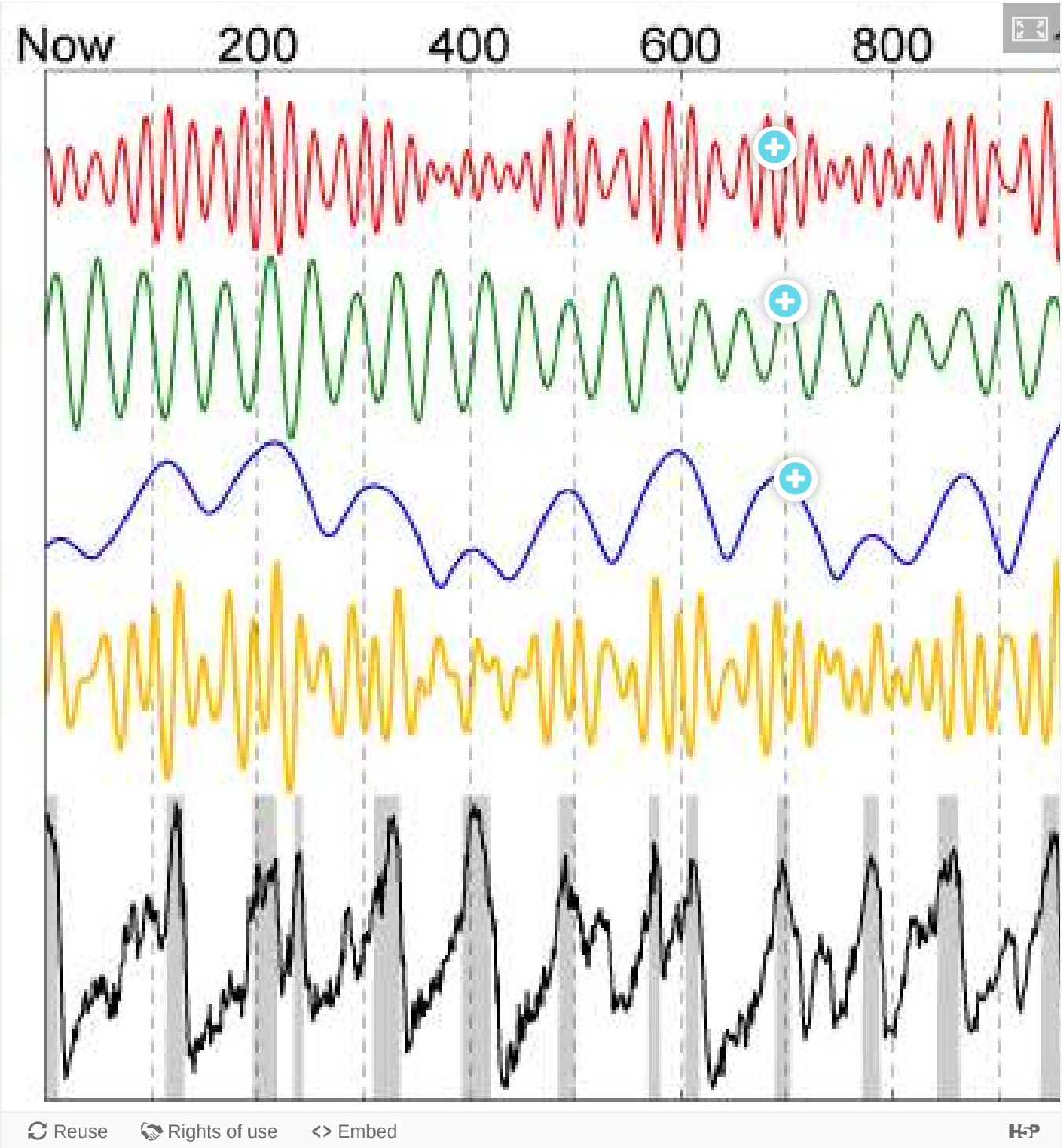


Figure 14.5.2: Oxygen isotope measurements as a proxy for temperature over the past 65 million years. Lowest levels indicate glaciations.

Short-term causes of glacial fluctuations are attributed to the cycles in the Earth's rotational axis and to variations in the Earth's orbit around the Sun which affect the distance between Earth and the Sun. Called **Milankovitch Cycles**, these cycles affect the amount of incoming solar radiation, causing short-term cycles of warming and cooling. During the Cenozoic Era, carbon dioxide levels steadily decreased from a maximum in the Paleocene, causing the climate to gradually cool [18]. By the Pliocene, ice sheets began to form. The effects of the Milankovitch Cycles created short-term cycles of warming and cooling within the larger glaciation event.

Milankovitch Cycles are three orbital changes named after the Serbian astronomer Milutan Milankovitch. The three orbital changes are called precession, obliquity, and eccentricity. **Precession** is the wobbling of Earth's axis with a period of about 21,000 years; **obliquity** is changes in the angle of Earth's axis with a period of about 41,000 years; and **eccentricity** is variations in the Earth's orbit around the Sun leading to changes in distance from the Sun with a period of 93,000 years. [19]. These orbital changes created a 41,000-year-long glacial–interglacial Milankovitch Cycle from 2.5 to 1.0 million years ago, followed by another longer cycle of about 100,000 years from 1.0 million years ago to today (see Milankovitch Cycles).

Enhance your understanding by clicking the plus signs:



This video summarizes ice ages: their characteristics and causes.

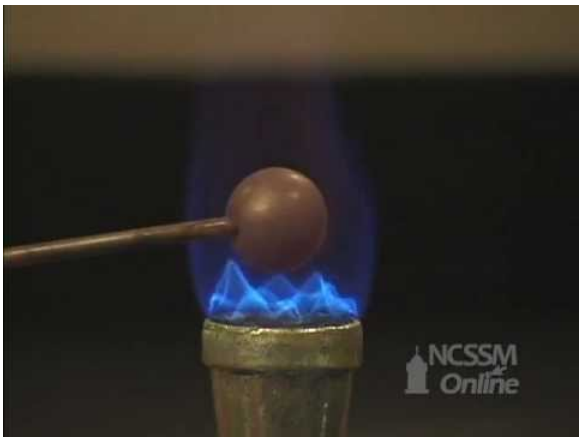




Sea-Level Change and Isostatic Rebound

When glaciers melt and retreat, two things happen: water runs off into the ocean causing sea levels to rise worldwide, and the land, released from its heavy covering of ice, rises due to isostatic rebound. For example, since the Last Glacial Maximum about 19,000 years ago [17] sea-level has risen about 400 feet (125 meters) [20]. A global change in sea level is called **eustatic** sea-level change. During a warming trend, sea-level rises due to more water being added to the ocean and also thermal expansion of sea water. About half of the Earth's eustatic sea-level rise during the last century has been the result of glaciers melting and about half due to thermal expansion [21; 22]. According to basic physics, **thermal expansion** occurs when a solid, liquid, or gas expands in volume when the temperature increases. Watch the 30-second video demonstrating thermal expansion with the classic brass ball and ring experiment.





Relative sea-level change includes vertical movement of both eustatic sea level and continents on tectonic plates. In other words, sea-level change is measured relative to land elevation. For example, if the land rises a lot and sea level rises only a little, then the relative sea level would appear to drop.

Continents sitting on the lithosphere can move vertically upward as a result of two main processes, tectonic uplift and isostatic rebound. Tectonic uplift occurs when tectonic plates collide. **Isostatic rebound** describes the upward movement of lithospheric crust sitting on top of the asthenospheric layer below it. Continental crust bearing the weight of continental ice sinks into the asthenosphere, displacing it. After the ice sheet melts away, the asthenosphere flows back in and continental crust floats back upward. Erosion can also create isostatic rebound by removing large masses like mountains and transporting the sediment away (think of the Mesozoic removal of the Alleghanian Mountains and the uplift of the Appalachian plateau), albeit this process occurs more slowly than relatively rapid glacier melting.

The isostatic rebound map below shows rates of vertical crustal movement worldwide. The highest rebound rate is indicated by the blue-to-purple zones (top end of the scale). The orange-to-red zones (bottom end of the scale) surrounding the high-rebound zones indicate isostatic lowering as adjustments in displaced subcrustal material have taken place.

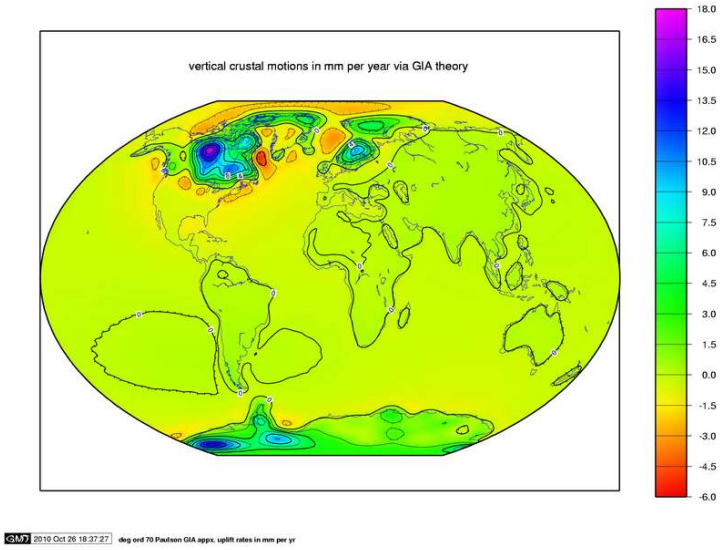


Figure 14.5.3: Rates of isostatic rebound worldwide, greatest in regions of recent glaciation.

Most glacial isostatic rebound is occurring where continental ice sheets rapidly melted about 19,000 years ago, such as in Canada and Scandinavia. Its effects can be seen wherever Ice Age ice or water bodies are or were present on continental surfaces and in terraces on river floodplains that cross these areas. Isostatic rebound occurred in Utah when the water from Lake Bonneville drained away [23]. North America's Great Lakes also exhibit emergent coastline features caused by isostatic rebound since the continental ice sheet retreated.

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14.S: Summary

Summary

Glaciers form when average annual snowfall exceeds melting and snow compresses into glacial ice. There are three types of glaciers, alpine or valley glaciers that occupy valleys, ice sheets that cover continental areas, and ice caps that cover smaller areas usually at higher elevations. As the ice accumulates, it begins to flow downslope and outward under its own weight. Glacial ice is divided into two zones, the upper rigid or brittle zone where the ice cracks into crevasses and the lower plastic zone where under the pressure of overlying ice, the ice bends and flows exhibiting ductile behavior. Rock material that falls onto the ice by mass wasting or is plucked and carried by the ice is called moraine and acts as grinding agents against the bedrock creating significant erosion.

Glaciers have a budget of income and expense. The zone of income for the glacier is called the zone of accumulation, where snow is converted into firn then ice by compression and recrystallization, and the zone of expense called the zone of ablation, where ice melts or sublimes away. The line separating these two zones latest in the year is the equilibrium or firn line and can be seen on the glacier separating bare ice from snow-covered ice. If the glacial budget is balanced, even though the ice continues to flow downslope, the end or terminus of the glacier remains in a stable position. If income is greater than expense, the position of the terminus moves downslope. If expense is greater than income, a circumstance now affecting glaciers and ice sheets worldwide due to global warming, the terminus recedes. If this situation continues, the glaciers will disappear. An average of cooler summers affects the stability or growth of glaciers more than higher snowfall. As the Greenland and Antarctic ice sheets flow seaward, the edges calve off forming icebergs.

Glaciers create two kinds of landforms, erosional and depositional. Alpine glaciers carve U-shaped valleys, and moraine carried in the ice polishes and grooves or striates the bedrock. Other landscape features produced by erosion include horns, arête ridges, cirques, hanging valleys, cols, and truncated spurs. Cirques may contain eroded basins that are occupied by post glacial lakes called tarns. Depositional features result from deposits left by retreating ice called drift. These include till, and moraine deposits (terminal, recessional, lateral, medial, and ground), eskers, kames, kettles and kettle lakes, erratics, and drumlins. A series of recessional moraines in glaciated valleys may create basins that are later filled with water to become paternoster lakes. Glacial meltwater carries fine-grained sediment onto the outwash plain. Lakes containing glacial meltwater are milky in color from suspended finely ground rock flour. Ice Age climate was more humid and precipitation that did not become glacial ice filled regional depressions to become pluvial lakes. Examples of pluvial lakes include Lake Missoula, which dammed behind an ice sheet lobe, and Lake Bonneville in Utah whose shoreline remnants can be seen on mountainsides. Repeated breaching of the ice lobe allowed Lake Missoula to rapidly drain causing massive floods that scoured the Channeled Scablands of Idaho, Washington, and Oregon.

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CHAPTER OVERVIEW

15: Energy and Mineral Resources

Learning Objectives

- Describe how a renewable resource is different from a nonrenewable resource.
- Compare the pros and cons of fossil fuel extraction and use, including unconventional resources.
- Describe the process of metallic mineral formation and extraction.
- Understand the use of nonmetallic mineral resources.

This text has discussed pioneers in the scientific study of geology like James Hutton and Charles Lyell, but the first "geologists" were the hominids who picked up stones, beginning the stone age. Maybe stones were first used as curiosity pieces, maybe as weapons, but ultimately, they were used as tools. This was the **Paleolithic Period**, the beginning of the study of geology and it goes back 2.6 million years ago to east Africa [1].



Figure 15.1: A Mode 1 Oldowan tool used for chopping. (By Locutus Borg; public domain via Wikimedia Commons.)

In modern times, an important use of geologic knowledge is locating economically valuable materials for use in society. All items we use can come from only three sources: they can be farmed, hunted or fished, or they can be mined. At the turn of the twentieth century, speculation was rampant that food supplies would not keep pace with world demand, and artificial fertilizers would need to be developed [2]. The ingredients for fertilizers are mined: nitrogen from the atmosphere using the Haber process [3], potassium from the hydrosphere (lakes or oceans) by evaporation, and phosphorus from the lithosphere (minerals like apatite from phosphorite rock, found in Florida, North Carolina, Idaho, Utah, and around the world). Thus, without mining and processing of natural materials, modern civilization would not exist. Geologists are essential in the process of mining.

15.1: Mining

15.2: Fossil Fuels

15.3: Mineral Resources

15.S: Summary

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15.1: Mining

Mining is defined as extracting valuable materials from the Earth for society's use. Usually, these include solid materials such as gold, iron, coal, diamond, sand, and gravel, but can also include fluid resources such as oil and natural gas. Modern mining has a long relationship with modern society. The oldest mine dates back 40,000 years to the Lion Cavern in Swaziland where there is evidence of concentrated digging into the Earth for hematite, an important iron ore used as red dye [4]. Resources extracted by mining are generally considered to be nonrenewable.

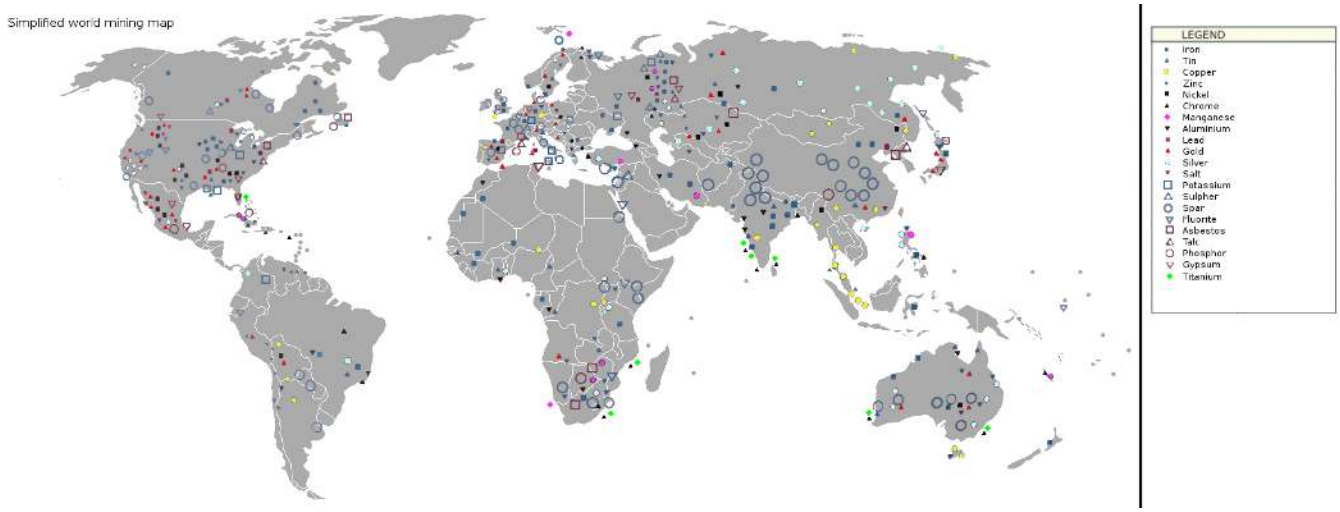


Figure 15.1.1: Map of world mining areas.

Renewable versus Nonrenewable Resources

Resources generally come in two major categories: renewable and nonrenewable. Renewable resources can be reused over and over or their availability replicated over a short human life span; nonrenewable resources cannot.

Renewable resources are materials present in our environment that can be exploited and replenished. Some common renewable energy sources are linked with **green energy sources** because they are associated with relatively small or easily remediated environmental impact. For example, **solar energy** comes from fusion within the Sun, which radiates electromagnetic energy. This energy reaches the Earth constantly and consistently and should continue to do so for about 5 billion more years [5]. **Wind energy**, also related to solar energy, is maybe the oldest form of renewable energy, and is used to sail ships and power windmills. Both solar and wind-generated energy are variable on the Earth's surface. These limitations are offset because we can use energy-storing devices, such as batteries or electricity exchanges between producing sites. The Earth's heat, known as **geothermal energy**, can be viable anywhere that geologists drill deeply enough. In practice, geothermal energy is more useful where heat flow is great, such as volcanic zones or regions with a thinner crust [6]. **Hydroelectric** dams provide energy by allowing water to fall through the dam under gravity, which activates turbines that produce energy. Ocean tides (**tidal energy**) can also be a reliable energy source. All of these renewable resources provide energy that powers society. Other renewable resources are plant and animal matter, which are used for food, clothing, and other necessities, but are being researched as possible energy sources.



Figure 15.1.2: Hoover Dam provides hydroelectric energy and stores water for southern Nevada.

Nonrenewable resources cannot be replenished at a sustainable rate. They are finite within a human lifetime. Many nonrenewable resources come from planetary, tectonic, or long-term biologic processes, and include materials such as gold, lead, copper, diamonds, marble, sand, natural gas, oil, and coal. Most nonrenewable resources include specific concentrated elements listed on the periodic table; some are compounds of those elements. For example, if society needs iron (Fe) sources, then an exploration geologist will search for iron-rich deposits that can be economically extracted. Nonrenewable resources may be abandoned when other materials become cheaper or serve a better purpose. For example, coal is abundantly available in England and other nations, but because oil and natural gas are available at a lower cost and lower environmental impact, coal use has decreased. Economic competition among nonrenewable resources is shifting use away from coal in many developed countries.

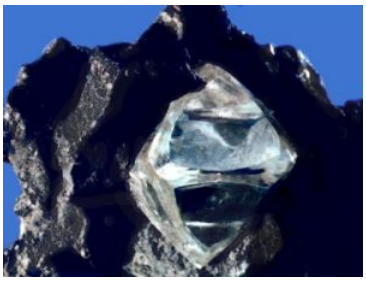


Figure 15.1.3: Natural, octahedral shape of diamond.

Ore

Earth's materials include the periodic table elements. However, it is rare that these elements are concentrated to the point where it is profitable to extract and process the material into usable products. Any place where a valuable material is concentrated is a geologic and geochemical anomaly. A body of material from which one or more valuable substances can be mined at a profit, is called an **ore deposit**. Typically, the term ore is used for only metal-bearing minerals, but it can be applied to valuable nonrenewable resource concentrations such as fossil fuels, building stones, and other nonmetal deposits, even groundwater. If a metal-bearing resource is not profitable to mine, it is referred to as a **mineral deposit**. The term **natural resource** is more common than the term ore for non-metal-bearing materials.

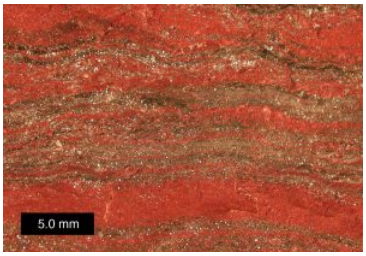


Figure 15.1.4: Banded-iron formations are an important ore of iron (Fe).

It is implicit that the technology to mine is available, economic conditions are suitable, and political, social and environmental considerations are satisfied in order to classify a natural resource deposit as ore. Depending on the substance, it can be concentrated in a narrow vein or distributed over a large area as a low-concentration ore. Some materials are mined directly from bodies of water (e.g. sylvite for potassium; water through desalination) and the atmosphere (e.g. nitrogen for fertilizers). These differences lead to various methods of mining, and differences in terminology depending on the certainty. **Ore mineral resource** is used for an indication of ore that is potentially extractable, and the term **ore mineral reserve** is used for a well-defined (proven), profitable amount of extractable ore.

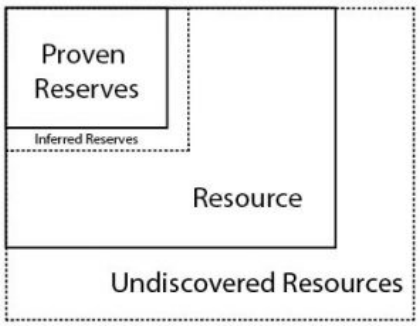


Figure 15.1.5: Diagram illustrating the relative abundance of proven reserves, inferred reserves, resources, and undiscovered resources. (Source: Chris Johnson)

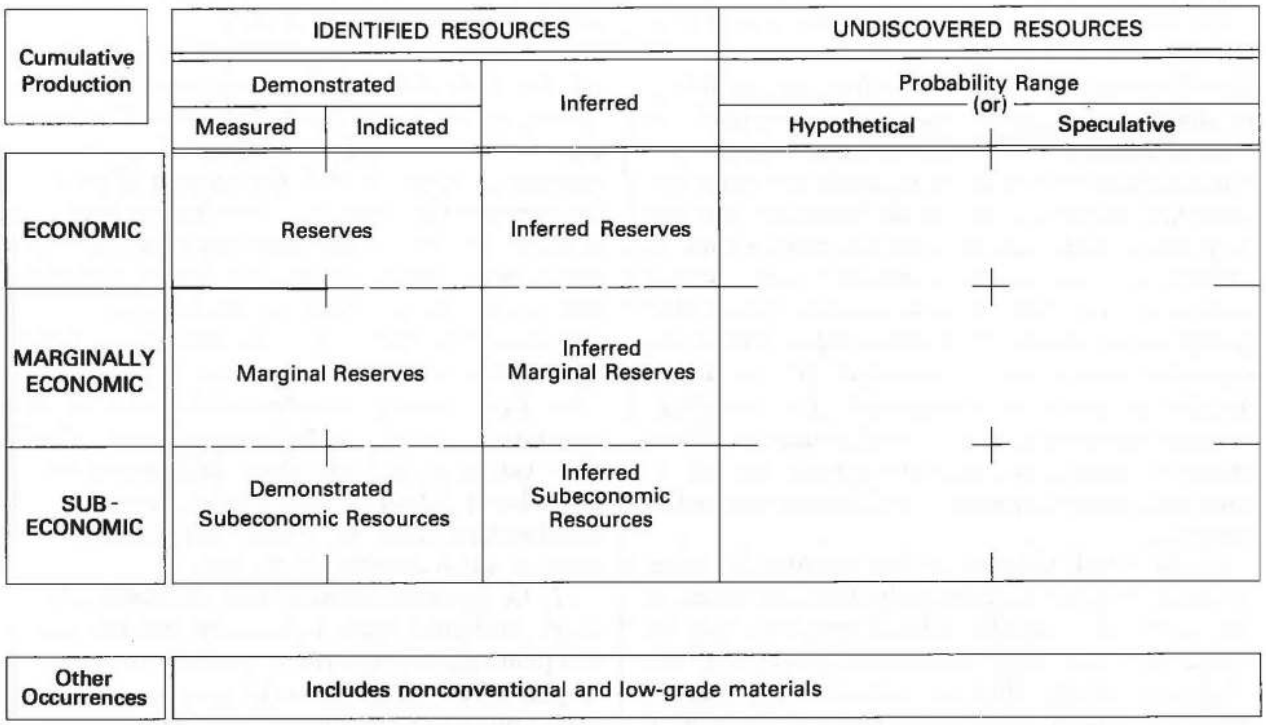


Figure 15.1.6: McKelvey diagram showing different definitions for different degrees of concentration and understanding of mineral deposits.

Mining Techniques

The mining style is determined by technology, social license, and economics. It is in the best interest of the company extracting the resources to do so in a cost-effective way. Fluid resources, such as oil and gas, are extracted by drilling wells and pumping. Over the years, drilling has evolved into a complex discipline in which directional drilling can produce multiple bifurcations and curves originating from a single drill collar at the surface. Using geophysical tools like seismic imaging, resources can be pinpointed and extracted efficiently.



Figure 15.1.7: Bingham Canyon Mine, Utah. This open-pit mine is the largest man-made removal of rock in the world.

Solid resources are extracted by two principal methods, of which there are many variants. **Surface mining** is the practice of removing material from the outermost part of the Earth. **Open-pit mining** is used to target shallow, broadly disseminated resources. Open-pit mining requires careful study of the ore body through surface mapping and drilling exploratory cores. The pit is progressively deepened through additional mining cuts to extract the ore. Typically, the pit's walls are as steep as can safely be managed. Once the pit is deepened, widening the top is very expensive. A steep wall is thus an engineering balance between efficient and profitable mining (from the company's point of view) and mass wasting (angle of repose from a safety point of view) so that there is less waste to remove. The waste is called **non-valuable rock** or **overburden** and moving it is costly. Occasionally landslides do occur, such as the very large landslide in the Kennecott Bingham Canyon Mine, Utah, in 2013. These events are costly and dangerous. The job of engineering geologists is to carefully monitor the mine; when company management heeds their warnings, there is ample time and action to avoid or prepare for any slide.

Strip mining and **mountaintop mining** are surface mining techniques that are used to mine resources that cover large areas, especially layered resources like coal. In this method, an entire mountaintop or rock layer is removed to access to the ore below. The environmental impacts of surface mining are usually much greater due to the larger surface footprint that is disturbed [7].



Figure 15.1.8: A surface coal mine in Wyoming.

Underground mining is often used for higher-grade, more localized, or very concentrated resources. For example, geologists mine some underground ore minerals by introducing chemical agents, which dissolve the target mineral. Then, they bring the solution to the surface where precipitation extracts the material. But more often, a mining shaft tunnel or a large network of these shafts and tunnels is dug to access the material. The decision to mine underground or from Earth's surface is dictated by the ore deposit's concentration, depth, geometry, land-use policies, economics, surrounding rock strength, and physical access to the ore. For example, to use surface mining techniques for deeper deposits might require removing too much material, or the necessary method may be too dangerous or impractical, or removing the entire overburden may be too expensive, or the mining footprint would be too large. These factors may prevent materials from being mined from the surface, and cause a project to be mined underground. The mining method and its feasibility depends on the price of the commodity and the cost of the technology needed to remove it and deliver it to market. Thus, mines and the towns that support them come and go as the commodity price varies. And conversely, technological advances and market demands may reopen mines and revive ghost towns.



Figure 15.1.9: Underground mining in Estonia of oil shale.

Concentrating and Refining

All ore minerals occur mixed with less desirable components called **gangue**. The process of physically separating gangue minerals from ore-bearing minerals is called **concentrating**. Separating the desired element from a host mineral by chemical means, including heating, is called **smelting**. Finally, taking a metal such as copper and removing other trace metals such as gold or silver is done through the process of **refining**. Typically, refining is done one of three ways:

1. Materials can either be mechanically separated and processed based on the unique physical properties of the ore mineral, like recovering placer gold based on its high density.
2. Materials can be heated to chemically separate desired components, such as refining crude oil into gasoline.
3. Materials can be smelted, in which controlled chemical reactions unbind metals from the minerals they are contained in, such as when copper is taken out of chalcopyrite ($CuFeS_2$).

Mining, concentrating, smelting and refining processes require enormous energy. Continual advances in metallurgy and mining practices strive to develop ever more energy-efficient and environmentally benign processes and practices.

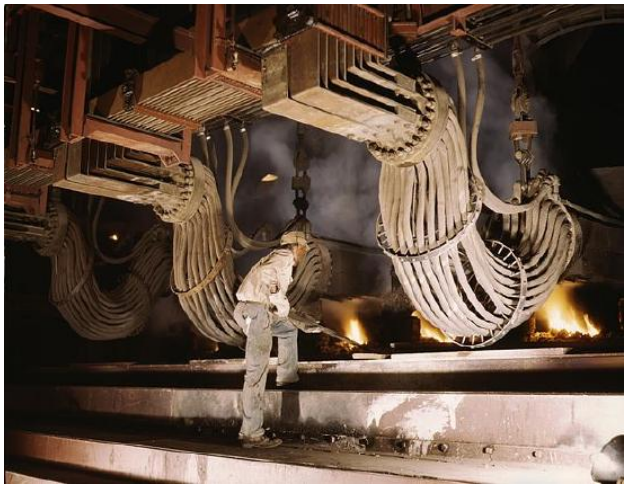


Figure 15.1.10: A phosphate smelting operation in Alabama, 1942.

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15.2: Fossil Fuels

Fossils fuels are extractable sources of stored energy created by ancient ecosystems. The natural resources that typically fall under this category are coal, oil, petroleum, and natural gas. These resources were originally formed via photosynthesis by living organisms such as plants, phytoplankton, algae, and cyanobacteria. This energy is actually fossil solar energy, since the Sun's ancient energy was converted by ancient organisms into tissues that preserved the chemical energy within the fossil fuel. Of course, as the energy is used, just like respiration from photosynthesis that occurs today, carbon enters the atmosphere as CO_2, causing climate consequences. Today, fossil fuels account for a large portion of the energy used in the world.



Figure 15.2.1: Coal power plant in Helper, Utah.

Converting solar energy by living organisms into hydrocarbon fossil fuels is a complex process. As organisms die, they decompose slowly, usually due to being buried rapidly, and the chemical energy stored within the organisms' tissues is buried within surrounding geologic materials. All fossil fuels contain carbon that was produced in an ancient environment. In environments rich with organic matter such as swamps, coral reefs, and planktonic blooms, there is a higher potential for fossil fuels to accumulate. Indeed, there is some evidence that over geologic time, organic hydrocarbon fossil fuel material was highly produced globally. [8]. Lack of oxygen and moderate temperatures in the environment seem to help preserve these organic substances [9; 10]. Also, the heat and pressure applied to organic material after it is buried contribute to transforming it into higher quality materials, such as brown coal to anthracite and oil to gas. Heat and pressure can also cause mobile materials to migrate to conditions suitable for extraction [11].



Figure 15.2.2: Modern coral reefs and other highly-productive shallow marine environments are thought to be the sources of most petroleum resources.

Oil and Gas

Petroleum is principally derived from organic-rich shallow marine sedimentary deposits where the remains of micro-organisms like plankton accumulated in fine-grained sediments. [12]. Petroleum's liquid component is commonly called **oil** and its gas component called **natural gas**, which is mostly made up of methane (CH_4). As rocks such as shale, mudstone, or limestone lithify, increasing pressure and temperature cause the oil and gas to be squeezed out and migrate from the **source rock** to a different rock unit higher in the rock column. Similar to the discussion of good aquifers, if the rock is sandstone, limestone, or other porous and permeable rock, and involved in a suitable stratigraphic or structural trapping process, then that rock can act as an oil and gas **reservoir**.

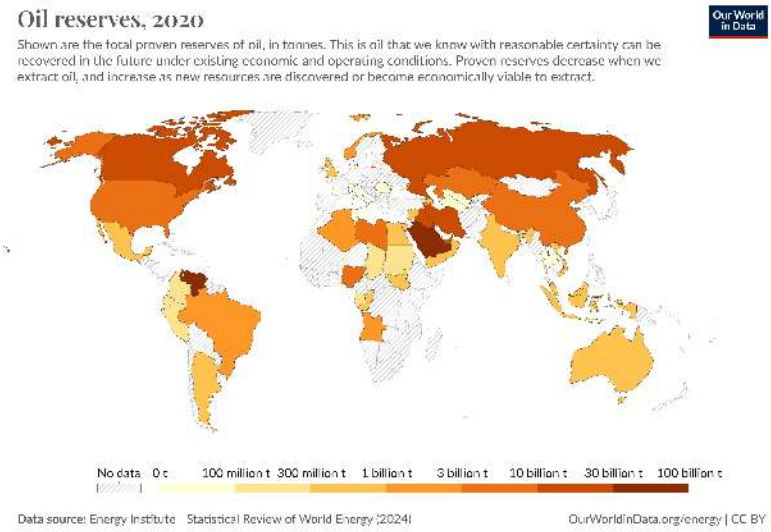


Figure 15.2.3: World Oil Reserves in 2020. Scale in billions of tons. (By Our World in Data; CC BY via ourworldindata.org)

A **trap** is a combination of a subsurface geologic structure, a porous and permable rock, and an impervious layer that helps block the movement of oil and gas and concentrates it for later human extraction [13; 14]. The development of a trap could be a result of many different geologic situations. Common examples include an anticline or domal structure, an impermeable salt dome, or a fault-bounded stratigraphic block, which is porous rock next to non-porous rock. The different traps have one thing in common: they pool the fluid fossil fuels into a configuration in which extraction is more likely to be profitable. Oil or gas in strata outside of a trap renders it less viable to extract.

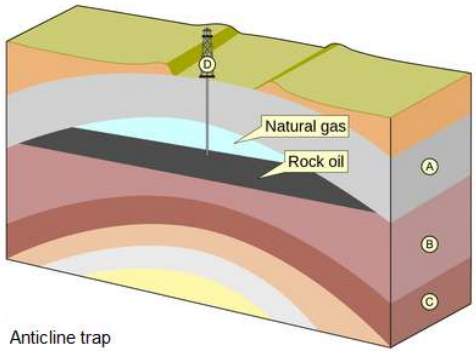


Figure 15.2.4: A structural or anticline trap. Gas and oil can get trapped under an impermeable rock layer. (By MagentaGreen; CC BY-SA 4.0 via Wikimedia Commons.)

Sequence stratigraphy is a branch of geology that studies sedimentary facies both horizontally and vertically and is devoted to understanding how changing sea level creates organic-rich shallow marine muds, carbonates, and sands in areas that are close to each other [15]. For example, shoreline environments may have beaches, lagoons, reefs, nearshore and offshore deposits, all next to each other. Beach sand, lagoonal and nearshore muds, and coral reef layers accumulate into sediments that include sandstones—good reservoir rocks— next to mudstones, next to limestones, both of which are potential source rocks. As sea level either rises or falls, the location of the shoreline changes and the locations of sands, muds, and reefs with it. This places oil and gas producing rocks such as mudstones and limestones next to oil and gas reservoirs such as sandstones and some limestones. Understanding the lithology and the facies/stratigraphic relationships interplay is very important in finding new petroleum resources. Using sequence stratigraphy as a model allows geologists to predict favorable locations of source rocks and reservoirs.

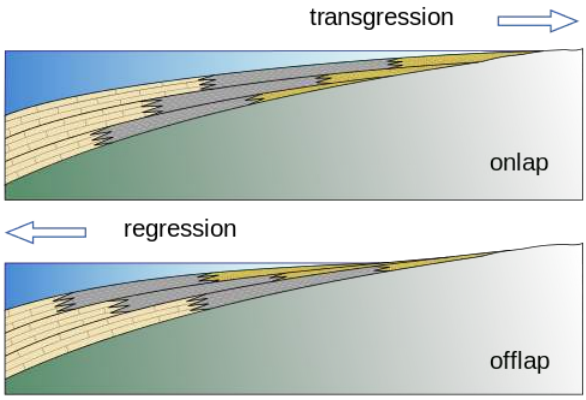


Figure 15.2.5: The rising sea levels of transgressions create onlapping sediments, while regressions create offlapping.

Tar Sands

Conventional oil and gas, which is pumped from a reservoir, are not the only way to obtain hydrocarbons. There are a few fuel sources known as **unconventional petroleum sources**. However, they are becoming more important as conventional sources become scarce. **Tar sands**, or **oil sands**, are sandstones that contain petroleum products that are highly viscous, like tar, and thus, cannot be drilled and pumped out of the ground readily like conventional oil. The unconventional fossil fuel is **bitumen**, which can be pumped as a fluid only at very low rates of recovery and only when heated or mixed with solvents. So, using injections of steam and solvents, or directly mining tar sands for later processing are ways to extract the tar from the sands. Alberta, Canada is known to have the largest reserves of tar sands in the world [16].



Figure 15.2.6: Tar sandstone from the Miocene Monterey Formation of California.

An energy resource becomes **uneconomic** once the total extraction and processing costs exceed the extracted material's sales revenue. Environmental costs may also contribute to a resource becoming uneconomic.

Oil Shale

Oil shale, or **tight oil,** is a fine-grained sedimentary rock that has a significant quantity of petroleum or natural gas locked tightly in the sediment. Shale is a common source of fossil fuels with high porosity but it has very low permeability. To extract the oil directly from the shale, the material has to be mined and heated, which, like with tar sands, is expensive and typically has a negative impact on the environment [17].

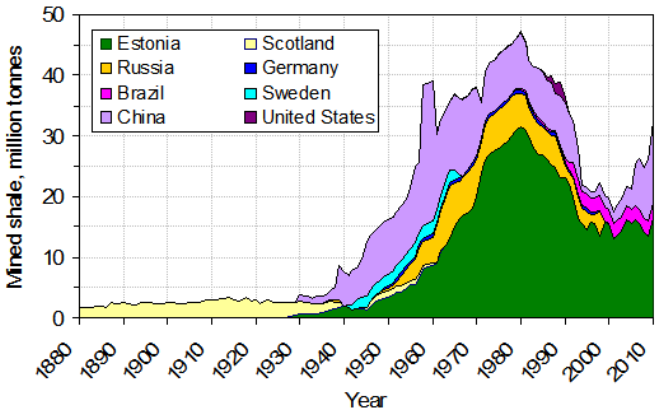


Figure 15.2.7: Global production of oil shale, 1880-2010.

Fracking

Another process which is used to extract the oil and gas from shale and other unconventional tight resources is called **hydraulic fracturing**, better known as **fracking** [18]. In this method, high-pressure injections of water, sand grains, and added chemicals are pumped underground. Under high pressure, this creates and holds open fractures in the rocks, which aids in the release of the hard-to-access fluids, mostly natural gas. Fracking is more useful in tighter sediments, especially shale, which has a high porosity to store the hydrocarbons but low permeability to transmit the hydrocarbons. Fracking has become controversial because its methods contaminate groundwater [19] and induce seismic activity [20]. This has created much controversy between public concerns, political concerns, and energy value.

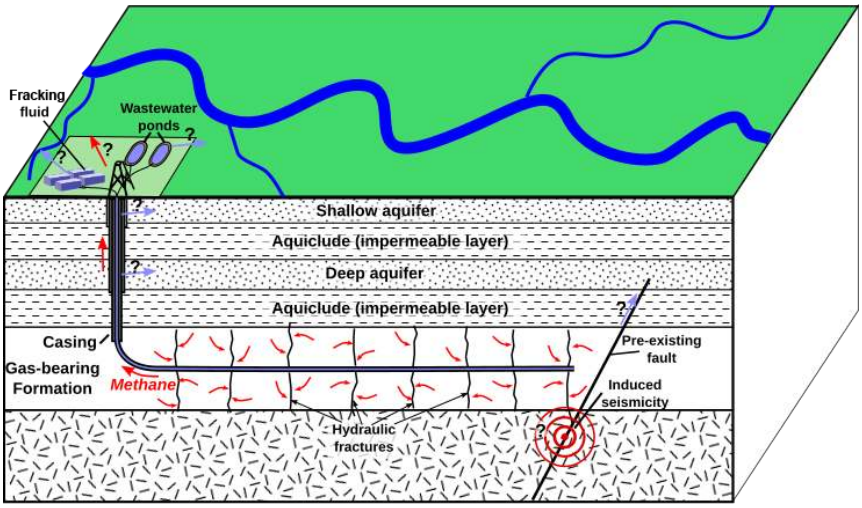


Figure 15.2.8: Schematic diagram of fracking.

Coal

Coal is the product of fossilized swamps [21], though some older coal deposits that predate terrestrial plants are presumed to come from algal buildups [22]. Coal is chiefly carbon, hydrogen, nitrogen, sulfur, and oxygen, with minor amounts of other elements [23]. As plant material is incorporated into sediments, heat and pressure cause several changes that concentrate the fixed carbon, which is the combustible portion of the coal. So, the more heat and pressure that coal undergoes, the greater its carbon concentration and fuel value, and the more desirable is the coal.

The general sequence of a swamp progressing through the various stages of coal formation and becoming more concentrated in carbon is:

$$\text{Swamp} \rightarrow \text{Peat} \rightarrow \text{Lignite} \rightarrow \text{Sub-bituminous Coal} \rightarrow \text{Bituminous Coal} \rightarrow \text{Anthracite} \rightarrow \text{Graphite}$$

As swamp materials collect on the swamp floor and are buried under accumulating materials, they first turn to **peat**. Peat itself is an economic fuel in some locations like the British Isles and Scandinavia. As lithification occurs, peat turns to **lignite**. With increasing heat and pressure, lignite turns to **sub-bituminous coal**, **bituminous coal**, and then, in a process like metamorphism, **anthracite**. Anthracite is the highest metamorphic grade and most desirable coal since it provides the highest energy output. With even more heat and pressure driving out all the volatiles and leaving pure carbon, anthracite can turn to **graphite**.

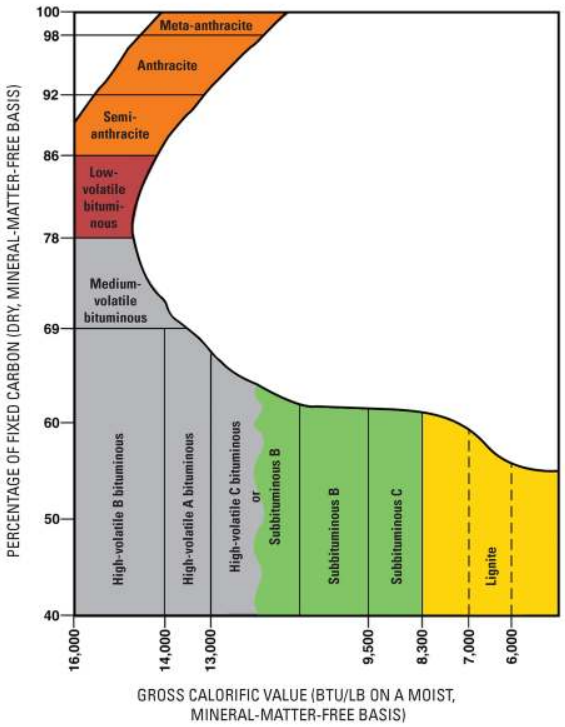


Figure 15.2.9: USGS diagram of different coal rankings.

Humans have used coal for at least 6000 years [23], mainly as a fuel source. Coal resources in Wales are often cited as a primary reason for the rise of Britain, and later, the United States during the Industrial Revolution [24; 25; 26]. According to the US Energy Information Administration, the production of coal in the US has decreased due to cheaper prices of competing energy sources and due to society recognizing its negative environmental impacts, including increased very fine-grained particulate matter as an air pollutant, greenhouse gases [27], acid rain [28], and heavy metal pollution [29]. Seen from this point of view, the coal industry as a source of fossil energy is unlikely to revive.





Figure 15.2.10: Left: Peat (also known as turf) consists of partially decayed organic matter. The Irish have long mined peat to be burned as fuel though this practice is now discouraged for environmental reasons. (By David Stanley; CC BY 2.0 via Wikimedia Commons.) Right: Anthracite coal, the highest grade of coal.

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15.3: Mineral Resources

Mineral resources, while principally nonrenewable, are generally placed in two main categories: **metallic**, which contain metals, and **nonmetallic**, which contain other useful materials. Most mining has been traditionally focused on extracting metallic minerals. Human society has advanced significantly because we've developed the knowledge and technologies to yield metal from the Earth. This knowledge has allowed humans to build the machines, buildings, and monetary systems that dominate our world today. Locating and recovering these metals has been a key facet of the study of geology since its inception. Every element across the periodic table has specific applications in human civilization. Metallic mineral mining is the source of many of these elements.



Figure 15.3.1: Gold-bearing quartz vein from California.

Types of Metallic Mineral Deposits

The various ways in which minerals and their associated elements concentrate to form ore deposits are too complex and numerous to fully review in this text. However, entire careers are built around them. Some of the more common types of these deposits are described, along with their associated elemental concentrations and world-class occurrences.

Magmatic Processes

When a magmatic body crystallizes and differentiates, it can cause certain minerals and elements to concentrate. **Layered intrusions**, typically ultramafic to mafic, can host deposits that contain copper, nickel, platinum, palladium, rhodium, and chromium. The Stillwater Complex in Montana is an example of economic quantities of a layered mafic intrusion [30]. Associated deposit types can contain chromium or titanium-vanadium. The largest magmatic deposits in the world are the chromite deposits in the Bushveld Igneous Complex in South Africa [31]. These rocks have an areal extent larger than the state of Utah. The chromite occurs in layers, which resemble sedimentary layers, except these layers occurred within a crystallizing magma chamber.



Figure 15.3.2: Layered intrusion of dark chromium-bearing minerals, Bushveld Complex, South Africa.

Water and other volatiles that are not incorporated into mineral crystals when a magma crystallizes can become concentrated around the margins of these crystallizing magmas. Ions in these hot fluids are very mobile and can form exceptionally large crystals. Once crystallized, these large crystal masses are called **pegmatites**. They form from magma fluids that are expelled from solidifying magma when nearly the entire magma body has crystallized. In addition to minerals that are predominant in the main igneous mass, such as quartz, feldspar, and mica, pegmatite bodies may also contain very large crystals of unusual minerals that contain rare elements like beryllium, lithium, tantalum, niobium, and tin, as well as native elements like gold [32]. Such pegmatites are ores of these metals.

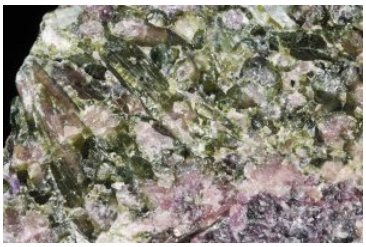


Figure 15.3.3: This pegmatite from Brazil contains lithium-rich green elbaite (a tourmaline) and purple lepidolite (a mica).

An unusual magmatic formation is a **kimberlite pipe**, which is a volcanic conduit that transports ultramafic magma from depths in the mantle to the surface. Diamonds, which are formed at great temperatures and pressures of depth, are transported by a kimberlite pipe to locations where they can be mined. The process that created these kimberlite ultramafic rocks is no longer common on Earth. Most of the known deposits are from the Archean Eon [33].

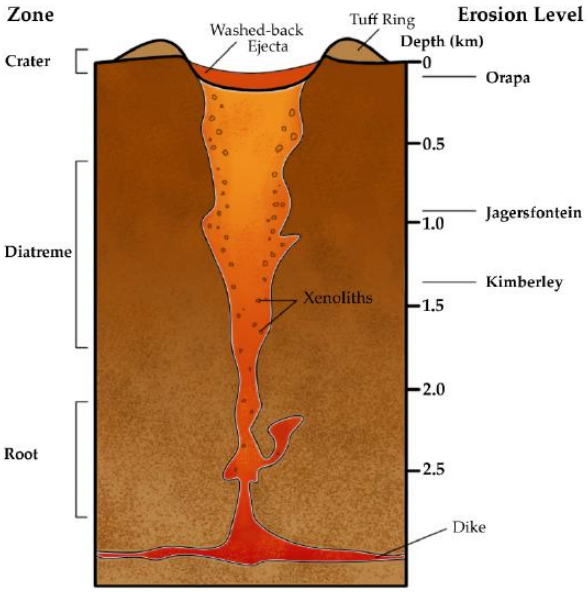


Figure 15.3.4: Schematic diagram of a kimberlite pipe.

Hydrothermal Processes

Fluids rising from crystallizing magmatic bodies or heated by the geothermal gradient cause many geochemical reactions that form various mineral deposits. The most active hydrothermal process today produces **volcanogenic massive sulfide** (VMS) **deposits**, which form from black smoker hydrothermal chimney activity near mid-ocean ridges all over the world. They commonly contain copper, zinc, lead, gold, and silver when found on the surface [34]. Evidence from around 7000 BCE in a period known as the Chalcolithic shows copper was among the earliest metals smelted by humans as means of obtaining higher temperatures were developed. The largest of these VMS deposits occur in Precambrian age rocks. The Jerome deposit in central Arizona is a good example.

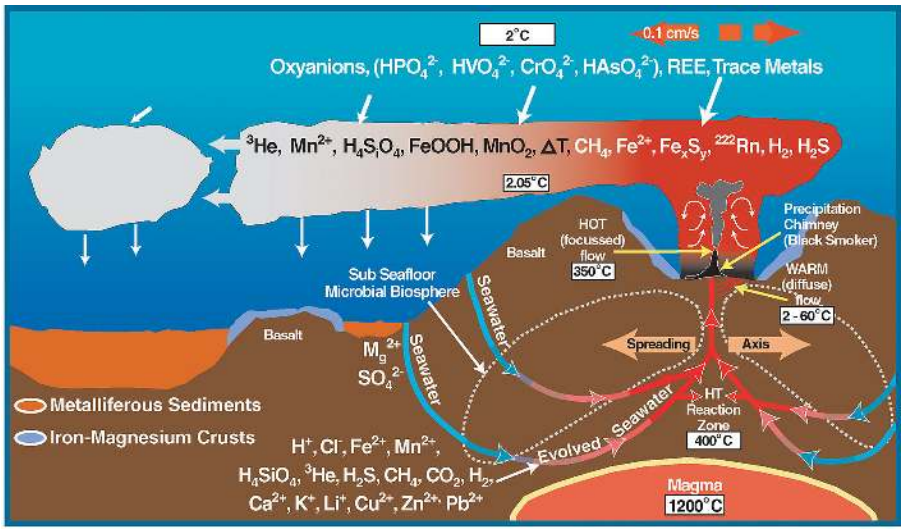


Figure 15.3.5: The complex chemistry around mid-ocean ridges.

Another deposit type that is derived from magma-heated water is a **porphyry deposit**. This is not to be confused with the igneous texture porphyritic, although the name is derived from the porphyritic texture that is nearly always present in the igneous rocks associated with a porphyry deposit. Several types of porphyry deposits exist, such as porphyry copper, porphyry molybdenum, and porphyry tin. These deposits contain low-grade disseminated ore minerals closely associated with intermediate and felsic intrusive rocks that are present over a very large area [35]. Porphyry deposits are typically the largest mines on Earth. One of the largest, richest, and possibly best-studied mine in the world is Utah's Kennecott Bingham Canyon Mine. It is an open-pit mine, which, for over 100 years, has produced several elements including copper, gold, molybdenum, and silver. Underground carbonate replacement deposits produce lead, zinc, gold, silver, and copper [36]. In the mine's past, the open pit predominately produced copper and gold from chalcopyrite and bornite. Gold only occurs in minor quantities in the copper-bearing minerals, but because the Kennecott Bingham Canyon Mine produces on such a large scale, it is one of the largest gold mines in the U.S. In the future, this mine may produce more copper and molybdenum (molybdenite) from deeper underground mines.

Most porphyry copper deposits owe their high metal content, and hence, their economic value to weathering processes called **supergene enrichment** which occurs when the deposit is uplifted, eroded, and exposed to oxidation [37]. This process occurred millions of years after the initial igneous intrusion and hydrothermal expulsion ends. When the deposit's upper pyrite-rich portion is exposed to rain, pyrite in the oxidizing zone creates an extremely acid condition that dissolves copper out of copper minerals such as chalcopyrite, and converts the chalcopyrite to iron oxides, such as hematite or goethite. The copper minerals are carried downward in water until they arrive at the groundwater table and an environment where the primary copper minerals are converted into secondary higher-copper content minerals. Chalcopyrite (35% Cu) is converted to bornite (63% Cu) and ultimately, chalcocite (80% Cu). Without this enriched zone, which is 2 to 5 times higher in copper content than the main deposit, most porphyry copper deposits would not be economic to mine.



Figure 15.3.6: The Morenci porphyry is oxidized toward its top (as seen as red rocks in the wall of the mine), creating supergene enrichment.

If limestone or other calcareous sedimentary rocks are present near the magmatic body, then another type of ore deposit called a **skarn** deposit forms. These metamorphic rocks form as magma-derived, highly saline metalliferous fluids react with carbonate rocks, creating calcium-magnesium-silicate minerals like pyroxene, amphibole, and garnet, as well as high-grade iron, copper, zinc

minerals and gold [38]. Intrusions that are genetically related to the intrusion that made the Kennecott Bingham Canyon deposit have also produced copper-gold skarns, which were mined by the early European settlers in Utah [39; 40]. When iron and/or sulfide deposits undergo metamorphism, the grain size commonly increases, which makes separating the gangue from the desired sulfide or oxide minerals much easier.

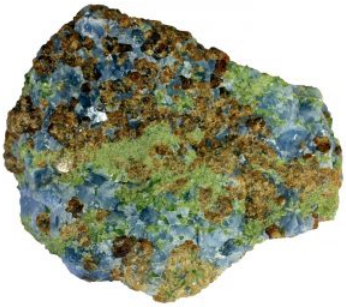


Figure 15.3.7: Garnet-augite skarn from Italy.

Sediment-hosted disseminated gold deposits consist of low concentrations of microscopic gold as inclusions and disseminated atoms in pyrite crystals. These are formed via low-level hydrothermal reactions, generally in the realm of diagenesis, that occur in certain rock types, namely muddy carbonates and limey mudstones. This hydrothermal alteration is generally far-removed from a magma source but can be found in rocks situated with a high geothermal gradient. The Mercur deposit in Utah's Oquirrh Mountains was this type's earliest locally mined deposit. There, almost one million ounces of gold were recovered between 1890 and 1917. In the 1960s, a metallurgical process using cyanide was developed for these low-grade ore types. These deposits are also called **Carlin-type deposits** because the disseminated deposit near Carlin, Nevada is where the new technology was first applied and where the first definitive scientific studies were conducted [41]. Gold was introduced into these deposits by hydrothermal fluids that reacted with silty calcareous rocks, removing carbonate, creating additional permeability, and adding silica and gold-bearing pyrite in the pore space between grains. The Betze-Post mine and the Gold Quarry mine on the Carlin Trend are two of the largest of the disseminated gold deposits in Nevada. Similar deposits, but not as large, have been found in China, Iran, and Macedonia [42].



Figure 15.3.8: In this rock, a pyrite cube has dissolved (as seen with the negative “corner” impression in the rock), leaving behind small specks of gold.

Non-Magmatic Geochemical Processes

Geochemical processes that occur at or near the surface without the aid of magma also concentrate metals, but to a lesser degree than hydrothermal processes. One of the main reactions is **redox,** short for reduction/oxidation chemistry, which has to do with the amount of available oxygen in a system. Places where oxygen is plentiful, as in the atmosphere today, are considered **oxidizing** environments, while oxygen-poor places are considered **reducing** environments. Uranium deposits are an example of where redox concentrated the metal. Uranium is soluble in oxidizing groundwater environments and precipitates as uraninite when encountering reducing conditions. Many of the deposits across the Colorado Plateau, such as in Moab, Utah, were formed by this method [43].



Figure 15.3.9: Underground uranium mine near Moab, Utah.

Redox reactions were also responsible for the creation of **banded iron formations** (BIFs), which are interbedded layers of iron oxide--hematite and magnetite, chert, and shale beds. These deposits formed early in the Earth's history as the atmosphere was becoming oxygenated. Cycles of oxygenating iron-rich waters initiated precipitation of the iron beds. Because BIFs are generally Precambrian in age, happening at the event of atmospheric oxygenation, they are only found in some of the older exposed rocks in the United States, such as in the upper peninsula of Michigan and northeastern Minnesota [44].

Deep, saline, **connate fluids** (trapped in the pore spaces), within sedimentary basins may be highly metalliferous. When expelled outward and upward as basin sediments compacted, these fluids formed lead and zinc deposits in limestone by replacing or filling open spaces, such as caves and faults, and in sandstone by filling pore spaces. The most famous are called **Mississippi Valley-type deposits** [44]. Also known as **carbonate-hosted replacement deposits**, they are large deposits of galena and sphalerite (lead and zinc ores respectively) that form from hot fluids ranging from 100°C to 200°C (212°F to 392°F). Although they are named for occurrences along the Mississippi River Valley in the United States, they are found worldwide.

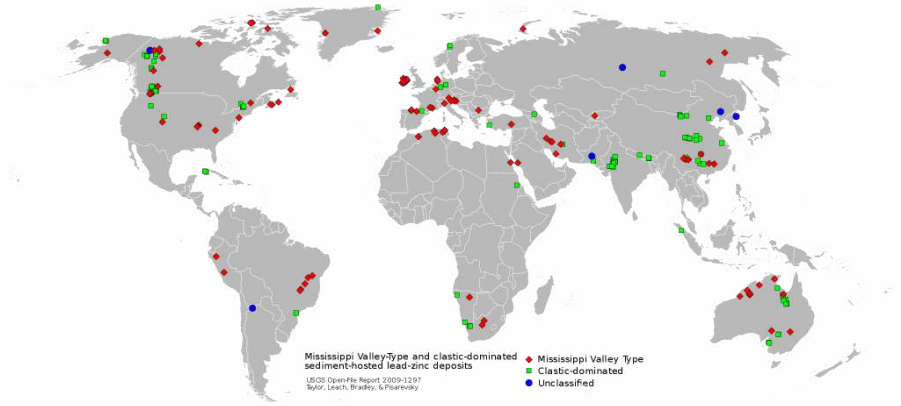


Figure 15.3.10: Map of Mississippi-Valley type ore deposits.

Sediment-hosted copper deposits occurring in sandstones, shales, and marls are enormous in size and their contained resources are comparable to porphyry copper deposits. These were most likely formed diagenetically by groundwater fluids in highly-permeable rocks [45]. Well-known examples are the Kupferschiefer in Europe, which has an areal coverage of more than 500,000 km^2, and the Zambian Copper Belt in Africa.

Soils and mineral deposits that are exposed at the surface experience deep and intense weathering, which can form surficial deposits. **Bauxite**, an ore of aluminum, is preserved in karst topography and laterites, which are soils formed in wet tropical environments [46]. Soils containing aluminum concentrate minerals, such as feldspar, and ferromagnesian minerals in igneous and metamorphic rocks, undergo chemical weathering processes that concentrate the metals. Ultramafic rocks that undergo weathering form nickel-rich soils, and when the magnetite and hematite in banded iron formations undergo weathering, they form **goethite**, a friable mineral that is easily mined for its iron content.



Figure 15.3.11: A sample of bauxite. Note the unweathered igneous rock in the center.

Surficial Physical Processes

At the Earth's surface, mass wasting and moving water can cause **hydraulic sorting**, which forces high-density minerals to concentrate. When these minerals are concentrated in streams, rivers, and beaches, they are called **placer deposits**, and occur in modern sands and ancient lithified rocks [47]. Native gold, native platinum, zircon, ilmenite, rutile, magnetite, diamonds, and other gemstones can be found in placers. Humans have mimicked this natural process to recover gold manually by gold panning and by mechanized means such as dredging.



Figure 15.3.12: Heavy mineral layers (dark) in a quartz sand beach deposit in India.

Environmental Impacts of Metallic Mineral Mining

The primary impact of metallic mineral mining comes from the mining itself, including disturbing the land surface, covering landscapes with **tailing impoundments** (piles of waste rock material after processing mined rocks), and increasing mass wasting by accelerating erosion [48]. In addition, many metal deposits contain pyrite, an uneconomic sulfide mineral, that when placed on waste dumps, generates **acid rock drainage** (ARD) during weathering. In oxygenated water, sulfides such as pyrite react and undergo complex reactions to release metal ions and hydrogen ions, which lowers pH to highly acidic levels. Mining and processing of mined materials typically increase the surface area to volume ratio in the material, causing chemical reactions to occur even faster than would occur naturally. If not managed properly, these reactions may lead to acidification of streams and groundwater plumes that carry dissolved toxic metals. In mines where limestone is a waste rock or carbonate minerals like calcite or dolomite are present, their acid-neutralizing potential helps reduce acid rock drainage. Although this is a natural process too, it is very important to isolate mine dumps and tailings from oxygenated water, both to prevent the sulfides from dissolving and subsequently percolating the sulfate-rich water into waterways. Industry has taken great strides to prevent contamination in recent decades, but earlier mining projects are still causing problems with local ecosystems.



Figure 15.3.13: Acid mine drainage in the Rio Tinto, Spain.

Nonmetallic Mineral Deposits

While receiving much less attention, nonmetallic mineral resources (also known as **industrial minerals**) are just as vital to ancient and modern society as metallic minerals. The most basic of these is building stone. Limestone, travertine, granite, slate, and marble are common building stones and have been quarried for centuries. Even today, building stones from slate roof tiles to granite countertops are very popular. Especially pure limestone is ground up, processed, and reformed as plaster, cement, and concrete. Some nonmetallic mineral resources are not mineral specific; nearly any rock or mineral can be used. This is generally called **aggregate** and is used in concrete, roads, and foundations. **Gravel** is one of the more common aggregates.



Figure 15.3.14: Carrara marble quarry in Italy, source to famous sculptures like Michelangelo's David.

Evaporites

Evaporite deposits form in restricted basins, such as the Great Salt Lake in Utah or the Dead Sea along Israel and Jordan, where water evaporates faster than it recharges [49]. As the waters evaporate, soluble minerals are concentrated and become supersaturated, at which point they precipitate from the now highly-saline waters. If these conditions persist for long stretches of time, thick deposits of rock salt, rock gypsum and other minerals can accumulate.



Figure 15.3.15: Salt-covered plain known as the Bonneville Salt Flats, Utah.

Evaporite minerals, such as **halite**, are used in our food as common table salt. Salt was a vitally important food preservative and economic resource before refrigeration was developed. While still used in food, halite is now mainly mined as a chemical agent, water softener, or a de-icer for roads. **Gypsum** is a common nonmetallic mineral used as a building material; it is the main component of drywall. It is also used as a fertilizer. Other evaporites include sylvite--potassium chloride, and bischofite--magnesium chloride, both of which are used in agriculture, medicine, food processing, and other applications. Potash, a group of highly soluble potassium-bearing evaporite minerals, is used as a fertilizer. In hyper-arid locations, even more rare and more complex evaporites, like borax, trona, ulexite, and hanksite, are mined. They can be found in places such as Searles Dry Lake and Death Valley, California, and in ancient evaporite deposits of the Green River Formation in Utah and Wyoming.



Figure 15.3.16: Hanksite, $Na_{22}K(SO_4)_9(CO_3)_2Cl$, one of the few minerals that is considered a carbonate and a sulfate.

Phosphorus

Phosphorus is an essential element that occurs in the mineral **apatite**, which is found in trace amounts in common igneous rocks. Phosphorite rock, which is formed in sedimentary environments in the ocean [50], contains abundant apatite and is mined to make fertilizer. Without phosphorus, life as we know it is not possible. Phosphorus is a major component of bone and a key component of DNA. Bone ash and guano are natural sources of phosphorus.



Figure 15.3.17: Apatite from Mexico.

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15.S: Summary

Summary

Energy and mineral resources are vital to modern society, and it is the role of a geologist to locate these resources for human benefit. As environmental concerns have become more prominent, the value of the geologist has not decreased, as they are still vital in locating and identifying the least intrusive methods of extraction.

Energy resources are generally grouped as being renewable or nonrenewable. Geologists can aid in locating the best places to exploit renewable resources (e.g. locating a dam), but are commonly tasked with finding nonrenewable fossil fuels. Mineral resources are also grouped in two categories: metallic and nonmetallic. Minerals have a wide variety of processes that concentrate them to economic levels, and are usually mined via surface or underground methods.

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CHAPTER OVERVIEW

16: Global Climate Change

Learning Objectives

- Describe the role of greenhouse gases in climate change.
- Describe the sources of greenhouse gases.
- Explain Earth's energy budget and global temperature changes.
- Explain how we know about climates in the geologic past.
- Explain how positive and negative feedback mechanisms can influence climate.
- Accurately describe which aspects of the environment are changing due to anthropogenic climate change.
- Describe the causes of recent climate change, particularly the role of humans in the overall climate balance.



Figure 16.1: The "Blue Marble," a picture of our planet from the 1972 Apollo 17 mission, shows that our planet is a finite place with may interacting systems. (By NASA; public domain via Wikimedia Commons.)

This chapter describes the Earth systems involved in climate change, the geologic evidence of past climate changes, and the human role in today's climate change. In science, a **system** is a group of interacting objects and processes. **Earth System Science** is the study of these systems: **geosphere**—rocks; **atmosphere**—gases; **hydrosphere**—water; **cryosphere**—ice; and **biosphere**—living things. Earth science studies these systems and how they interact and change in response to natural cycles and human-driven, or anthropogenic forces. Changes in one Earth system affect other systems.

It is critically important for us to be aware of the geologic context of climate change processes and how these Earth systems interact, first, for us to understand how and why human activities cause present-day climate change and, secondly, to distinguish between natural processes and human processes in the geologic past's climate record.

A significant part of this chapter introduces and discusses various processes from these Earth systems, how they influence each other, and how they impact global climate. For example, Earth's temperature and climate largely change based on atmospheric gas composition (atmosphere), ocean circulation (hydrosphere), and the land-surface characteristics of rocks (geosphere), glaciers (cryosphere), and plants (biosphere).

Also necessary to understanding climate change is to distinguish between climate and weather. **Weather** is the short-term temperature and precipitation patterns that occur in days and weeks. **Climate** is the variable range of temperature and precipitation patterns averaged over the long-term for a particular region. Thus, a single cold winter does not mean that the entire globe is cooling—indeed, the United States' cold winters of 2013 and 2014 occurred while the rest of the Earth was experiencing record warm-winter temperatures. To avoid these generalizations, many scientists use a 30-year average as a good baseline [1]. Therefore,

climate change refers to slow temperature and precipitation changes and trends over the long term for a particular area or the Earth as a whole.

16.1: Earth's Temperature
16.2: Evidence of Recent Climate Change
16.3: Prehistoric Climate Change
16.4: Anthropogenic Causes of Climate Change
16.S: Summary

Reference

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16.1: Earth's Temperature

Without an atmosphere, Earth would have huge temperature fluctuations between day and night, like the Moon. Daytime temperatures would be hundreds of degrees Celsius above normal, and nighttime temperatures would be hundreds of degrees below normal. Because the Moon doesn't have much of an atmosphere, daytime temperatures on the Moon are around 106°C (224°F) and nighttime temperatures are around -183°C (-298°F). That is an astonishing 272°C (522°F) of change between the light side and dark side of the Moon [2]. This section describes how Earth's atmosphere is involved in regulating the Earth's temperature.

Composition of Atmosphere

The composition of the atmosphere is a key component of the regulation of the planet's temperature. The atmosphere is 78% nitrogen (N_2), 21% oxygen (O_2), 1% argon (Ar), and less than 1% for all other gases known as trace components. The trace components include carbon dioxide (CO_2) water vapor (H_2O), neon, helium, and methane. Water vapor is highly variable, mostly based on region, but has been estimated to be about 1% of the atmosphere [9]. The trace gases include several important **greenhouse gases**, which are the gases responsible for warming and cooling the plant. On a geologic scale, volcanoes and the weathering process, which bury CO_2 in sediments, are the atmosphere's CO_2 sources. Biological processes both add and subtract CO_2 from the atmosphere [10].

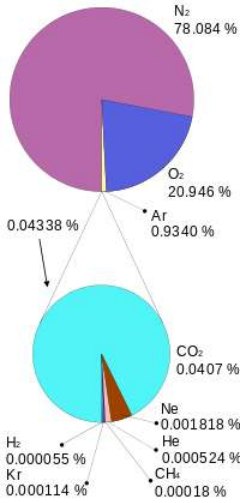


Figure 16.1.1: Composition of the atmosphere

Greenhouse gases trap heat in the atmosphere and warm the planet. They have little effect on incoming solar radiation (which is shortwave radiation) but absorb some of the outgoing infrared radiation (longwave radiation) that is emitted from Earth, thus keeping it from being lost to space. More greenhouse gases in the atmosphere absorb more longwave heat and make the planet warmer.

The most common greenhouse gases are water vapor (H_2O), carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O). Water vapor is the most abundant greenhouse gas but its abundance in the atmosphere does not change much over time. Carbon dioxide is much less abundant than water vapor, but carbon dioxide is being added to the atmosphere by human activities such as burning fossil fuels, land-use changes, and deforestation. Further, natural processes such as volcanic eruptions add carbon dioxide [3], but at an insignificant rate compared to human-based contributions.

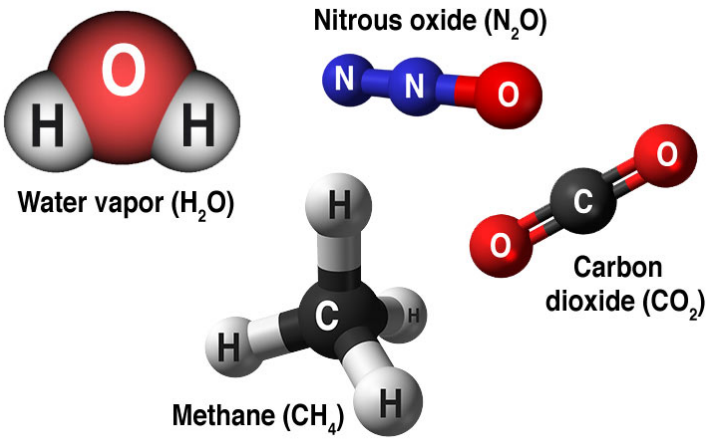


Figure 16.1.2: Common greenhouse gases.

There are two important reasons why carbon dioxide is the most important greenhouse gas. First, carbon dioxide stays in the atmosphere and does not go away for hundreds of years. Second, most of the additional carbon dioxide is “fossil” in origin, which means that it is released by burning **fossil fuels**. For example, coal and petroleum are fossil fuels. **Coal** and **oil** are made from long-dead plant material created by photosynthesis millions of years ago and stored in the ground. Photosynthesis takes sunlight plus carbon dioxide and creates the substances of plants. This transformation from plant matter to coal or oil occurs over millions of years, as a slow process accumulating fossil carbon in rocks and sediments. When we burn coal and oil, we instantaneously release the stored solar energy and the fossil carbon dioxide that took millions of years to accumulate in the first place. The rate of release is critical to comprehend current climate change.

Carbon Cycle

Critical to understanding global climate change is understanding the **carbon cycle** and how Earth’s own carbon-balancing system is being rapidly thrown off balance by human-driven activities. Earth has two important carbon cycles: the biological and the geological. In the biological cycle, living organisms—mostly plants—consume carbon dioxide from the atmosphere to make their tissues through photosynthesis. Then, after the organisms die, that carbon is released back into the atmosphere when they decay over several years or decades [11]. The following is the general equation for photosynthesis.

$$CO_2 + H_2O + \text{sunlight} \rightarrow \text{sugars} + O_2$$

In the geologic carbon cycle, a small portion of this biological-cycle carbon becomes buried in sedimentary rocks during the slow formation of coal and petroleum, as tiny fragments and molecules in organic-rich shale, and as the carbonate shells and other parts of marine organisms in limestone. This then becomes part of the geological carbon cycle, a cycle that actually involves a majority of Earth’s carbon, but one that operates only very slowly [11].

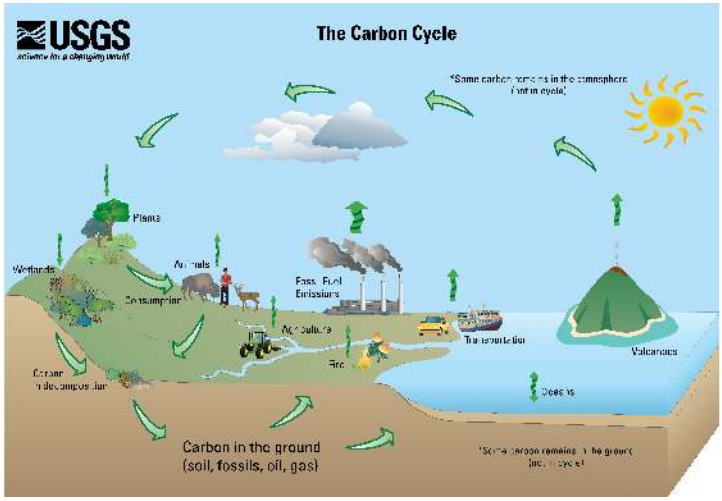


Figure 16.1.3: The carbon cycle.

The following are storage **reservoirs** for the geological carbon cycle:

- Organic matter from plants is stored in peat, coal, and permafrost for thousands to millions of years.
- Weathering of silicate minerals converts atmospheric carbon dioxide to dissolved bicarbonate, which is stored in the oceans for thousands to tens of thousands of years.
- Dissolved bicarbonate is converted by marine organisms to calcite, which is stored in carbonate rocks for tens to hundreds of millions of years.
- Carbon compounds are stored in sediments for tens to hundreds of millions of years; some end up in petroleum deposits.
- Carbon-bearing sediments are transferred by subduction to the mantle, where the carbon may be stored for tens of millions to billions of years.
- During volcanic eruptions, carbon dioxide is released back to the atmosphere, where it is stored for years to decades [11].

During much of Earth's history, the geological carbon cycle has been balanced, with carbon being released by volcanism at approximately the same rate that it is stored by the other processes. Under these conditions, the climate remains relatively stable. During some times of Earth's history, that balance has been upset. This can happen during prolonged stretches of greater than average volcanism. One example is the eruption of the Siberian Traps at around 250 million years ago, which appears to have led to strong climate warming over a few million years.

A carbon imbalance is also associated with significant mountain-building events. For example, the Himalayan Range has been forming for about 40 million years, and over that time — and still today — the rate of weathering on Earth has been enhanced because those mountains are so high and the range is so extensive that they present a greater surface area on which weathering takes place. The weathering of these rocks — most importantly the hydrolysis of feldspar — has resulted in the consumption of atmospheric carbon dioxide and transfer of the carbon to the oceans and to ocean-floor carbonate-rich sediments. The steady drop in carbon dioxide levels over the past 40 million years, which contributed to the Pliocene-Pleistocene glaciations, is partly attributable to the formation of the Himalayan Range.

Another, non-geological form of carbon-cycle imbalance is happening today on a very rapid time scale. We are in the process of extracting vast volumes of fossil fuels (coal, oil, and gas) that were stored in rocks over the past several hundred million years, and converting these fuels to energy and carbon dioxide. By doing so, we are changing the climate faster than has ever happened in the past [11]. Remember, carbon dioxide stays in the atmosphere and does not go away for hundreds of years. The more greenhouse gases in the atmosphere, the more heat is trapped and the warmer the planet becomes.

Greenhouse Effect

The **greenhouse effect** is a natural process by which the atmosphere warms surface temperatures. The greenhouse effect occurs because of the presence of greenhouse gases in the atmosphere. The greenhouse effect is named after a similar process that warms a greenhouse or a car on a hot summer day. Sunlight passes through the glass of the greenhouse or car, reaches the interior, and

changes into heat. The heat radiates upward and gets trapped by the glass windows. The greenhouse effect for the Earth can be explained in three steps.

Step 1: Solar radiation from the Sun is composed of mostly **ultraviolet** (UV), **visible light**, and **infrared** (IR) radiation. Components of solar radiation include parts with a shorter wavelength than visible light, like ultraviolet light, and parts of the spectrum with longer wavelengths, like IR and others. Some of the radiation gets absorbed, scattered, or reflected by the atmospheric gases but about half of the solar radiation eventually reaches the Earth's surface.

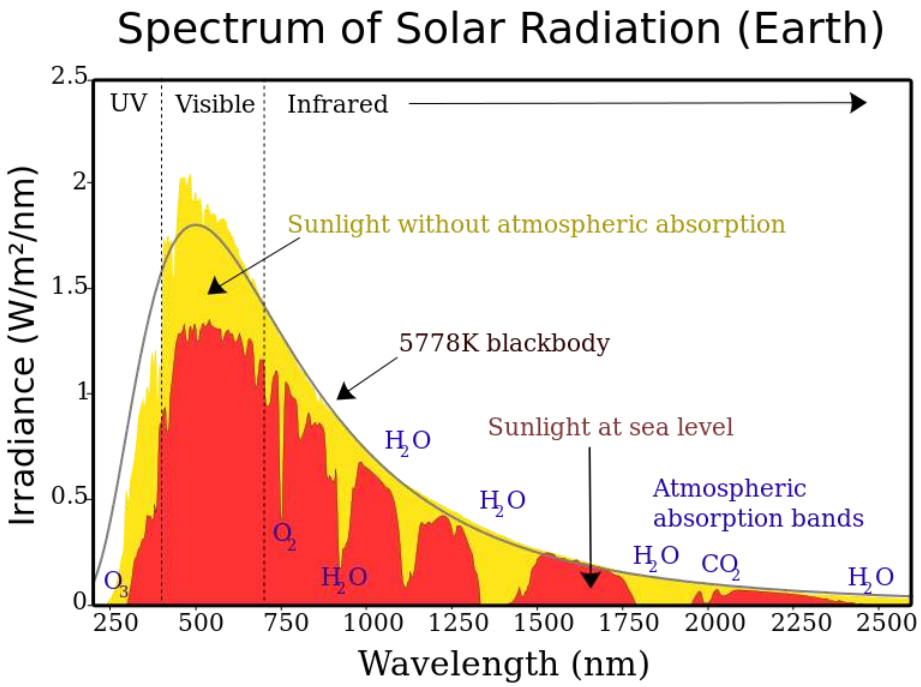


Figure 16.1.4: Incoming radiation absorbed, scattered, and reflected by atmospheric gases.

Step 2: The visible, UV, and IR radiation, that reaches the surface converts to heat energy. Most students have experienced sunlight warming a surface such as a paved surface, a patio, or deck. When this occurs, the warmer surface thus emits more **thermal radiation**, which is a type of IR radiation. So, there is a conversion from visible, UV, and IR to just thermal IR. This thermal IR is what we experience as heat. If you have ever felt the heat radiating from a fire or a hot stovetop, then you have experienced thermal IR.

Step 3: Thermal IR radiates from the Earth's surface back into the atmosphere. But since it is thermal IR instead of UV, visible, or regular IR, this thermal IR gets trapped by greenhouse gases. In other words, the Sun's energy leaves the Earth at a different wavelength than it enters, so, the Sun's energy is not absorbed in the lower atmosphere when energy is coming in, but rather when the energy is going out. The gases that typically do this blocking on Earth include carbon dioxide, water vapor, methane, and nitrous oxide. More greenhouse gases in the atmosphere result in more thermal IR being trapped.



Figure 16.1.5: Animation on the greenhouse effect. (By NASA-JPL/Caltech; public domain.)

Earth's Energy Budget

Solar radiation arriving at Earth from the Sun is relatively uniform over time. Earth is warmed, and energy (or heat) radiates from the Earth's surface and lower atmosphere back to space. This flow of incoming and outgoing energy is the **Earth's energy budget**. For Earth's temperature to be stable over long stretches of time, incoming energy and outgoing energy have to be equal on average so that the energy budget at the top of the atmosphere balances. About 29% of the incoming solar energy arriving at the top of the atmosphere is reflected back to space by clouds, atmospheric particles, or reflective ground surfaces like sea ice and snow. About 23% of incoming solar energy is absorbed in the atmosphere by water vapor, dust, and ozone. The remaining 48% passes through the atmosphere and is absorbed at the surface. Thus, about 71% of the total incoming solar energy is absorbed by the Earth system [3].

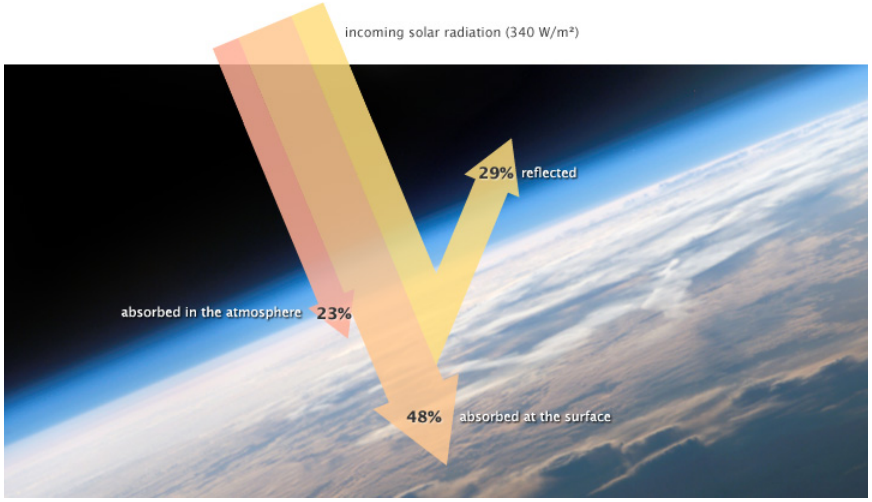


Figure 16.1.6: Incoming solar radiation filtered by the atmosphere.

When this energy reaches Earth, the atoms and molecules making up the atmosphere and surface absorb the energy and they increase in temperature. If this material could only absorb energy, then the temperature of the Earth would be like the water level in a sink with no drain where the faucet runs continuously. The sink would eventually overflow. However, the temperature does not infinitely rise because the Earth is not just absorbing sunlight. The Earth's surface is also radiating thermal energy (heat) back into the atmosphere. If the temperature of the Earth rises, the planet emits an increasing amount of heat to space and this is the primary mechanism that prevents Earth from continually heating [3].

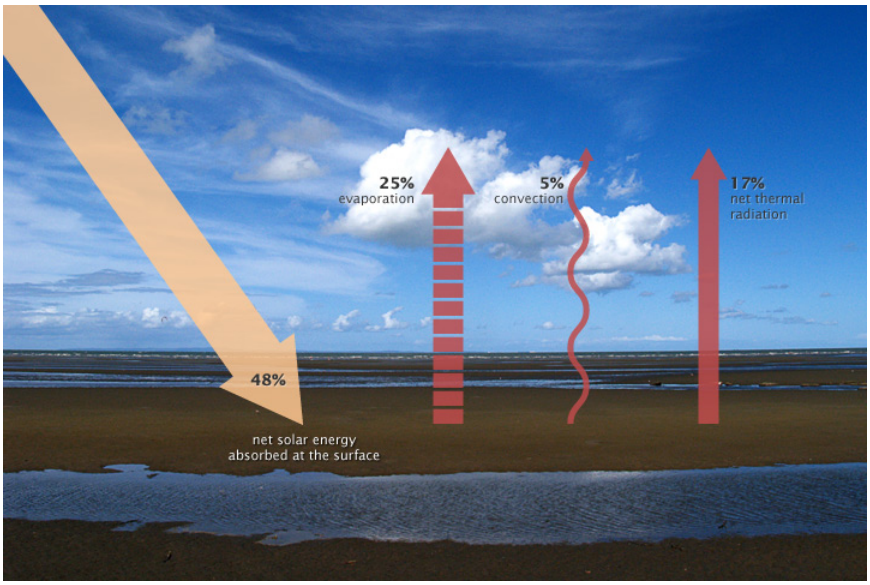


Figure 16.1.7: Some of the thermal infrared energy (heat) radiated from the surface into the atmosphere is trapped by gases in the atmosphere.

Some of the thermal infrared heat radiating from the surface is absorbed and trapped by greenhouse gases in the atmosphere. Greenhouse gases act like a giant blanket for Earth. The more greenhouse gases in the atmosphere, then the more outgoing heat will be retained by Earth and the less of this thermal infrared energy (heat) dissipates to space.

Factors that can affect the Earth's energy budget are not limited to greenhouse gases. Increases in solar energy can increase the energy received by the Earth. However, increases associated with this are very small over time [3; 4; 5]. In addition, land and water will absorb more sunlight when there is less ice and snow to reflect the sunlight back to the atmosphere. For example, the ice covering the Arctic Sea reflects sunlight back to the atmosphere. The reflectivity of the Earth's surface is called **albedo**. Furthermore, aerosols (dust particles) produced from burning coal, diesel engines, and volcanic eruptions can reflect more incoming solar radiation and actually cool the planet. The effect of anthropogenic aerosols is weak on the climate system but anthropogenic production of greenhouse gases is not weak. Thus, the net effect is warming due to more anthropogenic greenhouse gases associated with fossil fuel combustion [6; 7; 8].

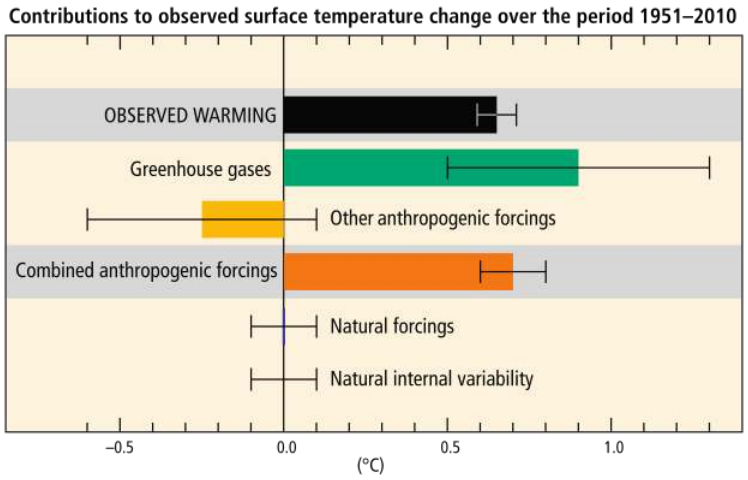


Figure 16.1.8: Net effect of factors influencing warming.

An effect that changes the planet can trigger **feedback mechanisms** that amplify or suppress the original effect. A **positive feedback mechanism** is when the output or effect enhances the original stimulus or cause. Thus, it increases the ongoing effect. For example, the loss of sea ice at the North Pole makes that area less reflective (reduced albedo). This allows the surface air and ocean to absorb more energy in an area that was once covered by sea ice [3] and lose more sea ice. Another example is the melting permafrost. **Permafrost** is permanently frozen soil located near the high latitudes, mostly in the Northern Hemisphere. As the

climate warms, more permafrost thaws and the thick deposits of organic matter are exposed to oxygen and begin to decay. This oxidation process releases carbon dioxide and methane which in turn causes more warming which melts more permafrost, and so on and on.

A **negative feedback mechanism** occurs when the output or effect reduces the original stimulus or cause [3]. For example, in the short term, more carbon dioxide (CO_2) is expected to cause forest canopies to grow which will absorb more CO_2. An example for the long term is increased carbon dioxide (CO_2) in the atmosphere is expected to cause more carbonic acid and chemical weathering, resulting in transport of dissolved bicarbonate and other ions to the oceans which then become stored in sediment.

Global warming is evidence that Earth's energy budget is not balanced. Positive effects on Earth's temperature are now greater than negative effects.

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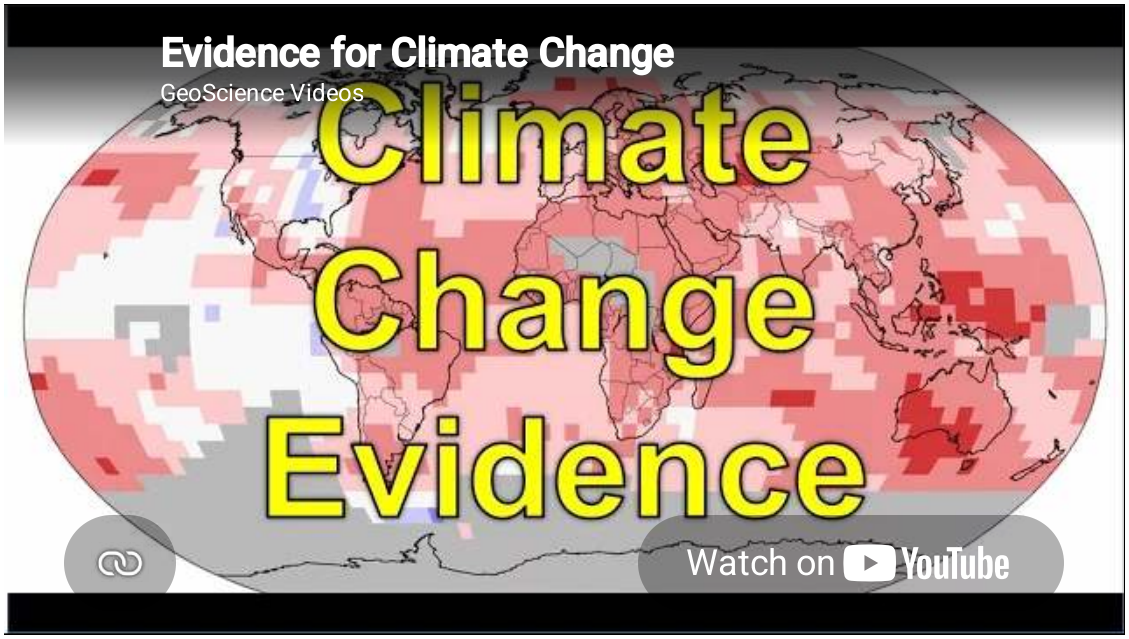
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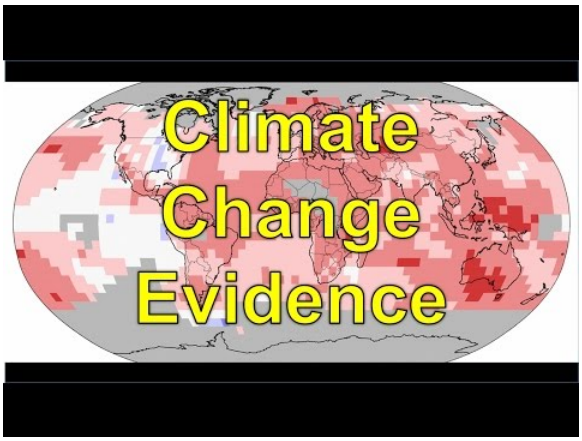
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16.2: Evidence of Recent Climate Change

While climate has changed many times in the past due to natural causes, the scientific consensus is that human activity is causing the climate to change today more rapidly [11; 7]. While this seems like a new idea, it has been suggested for more than 75 years [12]. This section describes the evidence that scientists agree is most likely a result of **anthropogenic climate change**, or, human-caused climate change. For more information, watch this video on climate change by two professors at a North Carolina State University. Although the video is from 2015 and the data is not up to date, the trends have continued and still serve as evidence of climate change.





Global Temperature Rise

Since 1880, average global surface temperatures have trended upward and most of that warming has occurred since 1970 (see this NASA animation). Since the ocean is absorbing a lot of the additional trapped heat, surface temperatures include both land surface and ocean temperatures [13]. Changes in land-surface or ocean-surface temperatures compared to a reference period from 1951 to 1980, where the long-term average remained relatively constant, are called temperature anomalies. A **temperature anomaly** thus represents the difference between the measured temperature and the average value during the reference period. Climate scientists calculate long-term average temperatures over thirty years or more such as the reference period from 1951 to 1980. Another common period is the last century (1900-2000). Therefore, an anomaly of 1.25 ℃ (2.25°F) for 2015 means that the average temperature for 2015 was 1.25 ℃ (2.25°F) greater than the 1900-2000 average. In 1950, the temperature anomaly was -0.28 ℃

(-0.5°F), so this is -0.28 ℃ (-0.5°F) lower than the 1900-2000 average [3]. These temperatures are annual average surface temperatures.

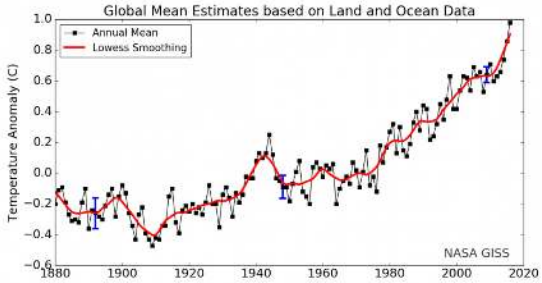


Figure 16.2.1: Land-ocean temperature index, 1880 to present, with a base time 1951-1980. The solid black line is the global annual mean and the solid red line is the five-year Lowess smoothing. The blue uncertainty bars (95% confidence limit) account only for incomplete spatial sampling. The overall trend is that Earth's temperature is rising.

In addition to a rising average surface land temperature, the ocean has absorbed a lot of the heat. With oceans covering about 70% of the Earth's surface, there is a lot of opportunities to absorb energy. The ocean has been absorbing about 80% to 90% of the additional heat added due to human activities. As a result, the top 2300 feet of the ocean has increased in temperature 0.17°C (0.3℉) since 1969 (external link to this 3-minute video by NASA JPL on heat capacity of the ocean) [3]. The reason the ocean has warmed less than the atmosphere, while still taking on most of the heat, is due to the very high **specific heat** of water, which means that water can absorb a lot of energy for a small temperature increase. In contrast, the atmosphere needs less energy to increase its temperature.

Some scientists suggest that anthropogenic greenhouse gases do not cause global warming since surface temperatures have not increased very much between 1998 and 2013, while greenhouse gas concentrations have continued to increase during that time period. However, since the oceans are absorbing most of the heat, decade-scale circulation changes in the ocean, similar to La Niña, push warmer water deeper under the surface [14; 15; 16]. Once the absorption and circulation of the ocean is accounted for and the heat added back into surface temperatures, then the temperature increases become apparent as shown in the above figure. Furthermore, this ocean heat storage is temporary, as reflected in the record-breaking warm years of 2014-2016. Indeed, with this temporary ocean storage effect, 15 of the first 16 years of the 21st century have been the hottest in recorded history.

Carbon Dioxide

Anthropogenic greenhouse gases, mostly carbon dioxide (CO_2), have increased since the Industrial Revolution when the burning of fossil fuels dramatically increased. These levels are unprecedented in the last 800,000-year Earth history as recorded in geologic sources such as ice cores. Carbon dioxide has increased by 40% since 1750 and the rate of increase has been the fastest during the last decade [3; 6]. For example, since 1750, 2040 gigatons of CO_2 have been added to the atmosphere, about 40% have remained in the atmosphere while the remaining 60% have been absorbed into the land (by plants and soil) or the oceans [6]. Indeed, during the lifetime of most young adults, the total atmospheric CO_2 has increased by 50 ppm or 15%.

Charles Keeling, an oceanographer with Scripps Institution of Oceanography in San Diego, California was the first person to make regular measurements of atmospheric CO_2. Using his methods, constant measurements of CO_2 in the atmosphere have been made at the Mauna Loa Observatory on Hawaii since 1958. These measurements are published regularly by NASA at this website: https://keelingcurve.ucsd.edu/. Go there now to see the very latest measurement. Keeling's measured values have been posted in a curve of increasing values called the **Keeling Curve**. This curve varies annually up and down from summer (when the plants in the Northern Hemisphere are using CO_2) to winter when the plants are dormant, but shows a steady increase over the past several decades. This curve increases exponentially, not linearly indicating that the rate of increase of CO_2 is accelerating!

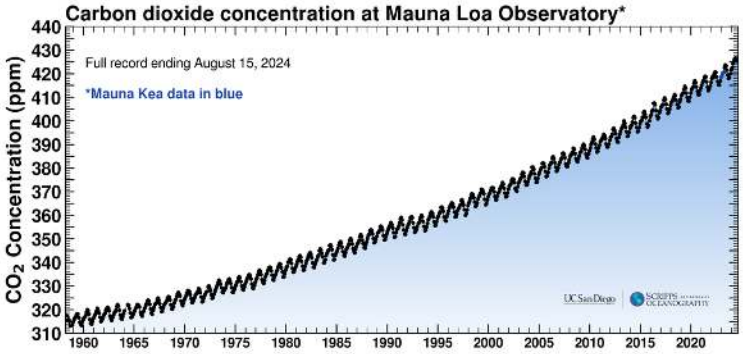
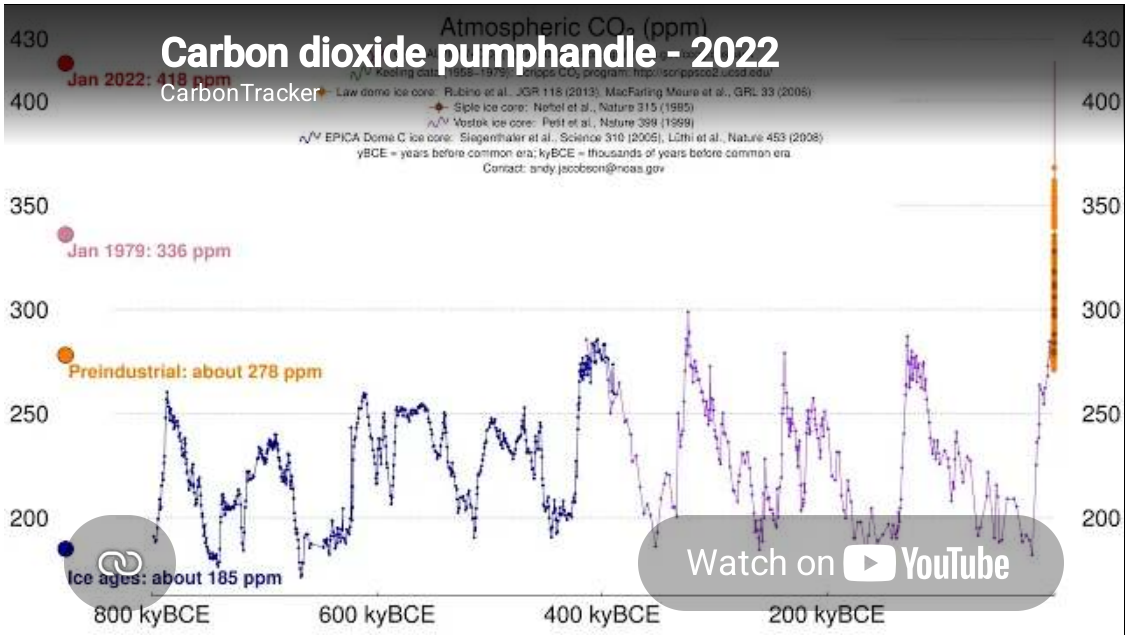
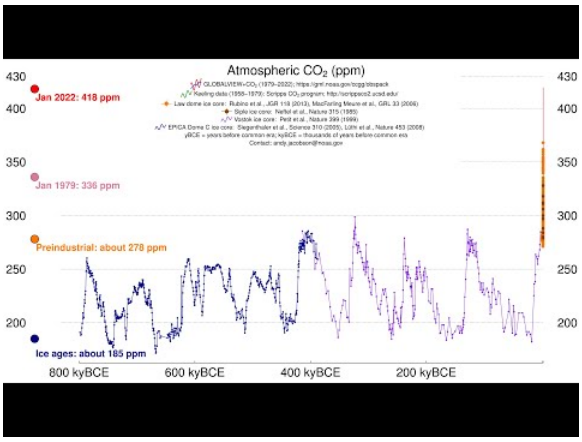


Figure 16.2.2: The Keeling Curve showing increasing atmospheric CO_2 since 1958. Note that the curve is exponential, not linear!

The following video shows how atmospheric CO_2 has varied recently and also over the last 800,000 years as determined by many CO_2 monitoring stations (shown on the insert map). It is also instructive to watch the CO_2 variation of the Keeling portion of the video by latitude. This shows that most of the human sources of CO_2 are in the Northern Hemisphere where most of the land is and where most of the developed nations are.





Melting Glaciers and Shrinking Sea Ice

Glaciers are large ice accumulations that exist year-round on the land's surface and **ice sheets** are thick glaciers that cover continents during ice ages. In contrast, **icebergs** are masses of floating sea ice, although they may have had their origin in glaciers.

Alpine glaciers, ice sheets, and sea ice are all melting. Explore melting glaciers at NASA's interactive Global Ice Viewer). Satellites have recorded that Antarctica is melting at 118 gigatons per year and Greenland is melting at 281 gigatons per year (1 gigaton is over 2 trillion pounds). Almost all major alpine glaciers are shrinking, deflating, and retreating and the rate of ice mass loss is unprecedented --never observed before-- since the 1940's when quality records for most began.

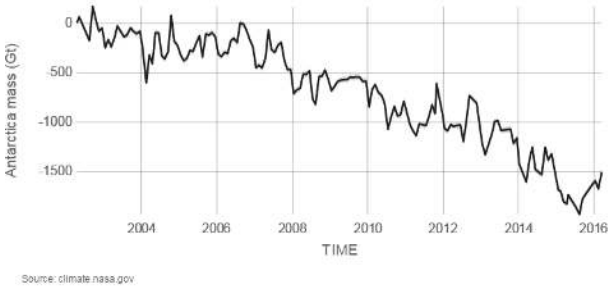
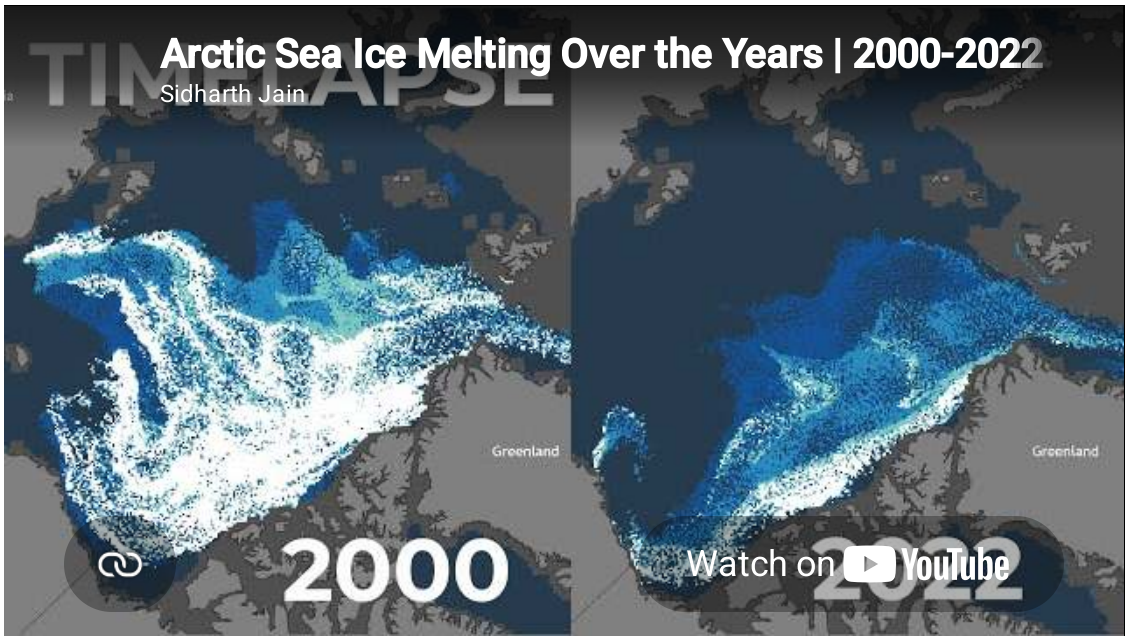
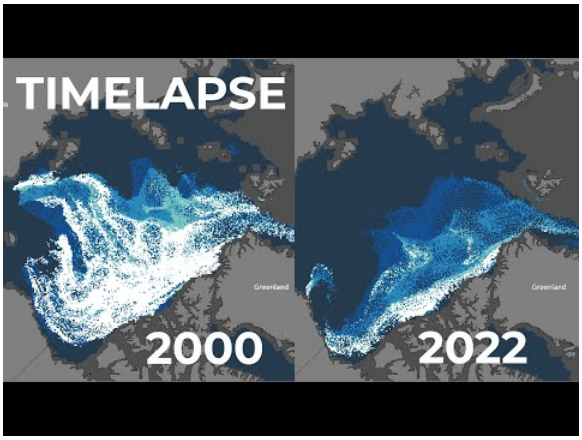


Figure 16.2.3: Decline of Antarctic ice mass from 2002 to 2016.

Before anthropogenic warming, glacial activity was variable with some retreating and some advancing [17]. Now, the extent of spring snow cover has decreased. In addition, the extent of sea ice is shrinking. Most sea ice is at the North Pole which is only occupied by the Arctic Ocean and sea ice [3; 6]. The NOAA animation shows how perennial sea ice has declined from 2000 to 2022. The oldest ice is white and the youngest (seasonal) ice is dark blue. The amount of old ice has declined from 20% in 1985 to 3% in 2015.



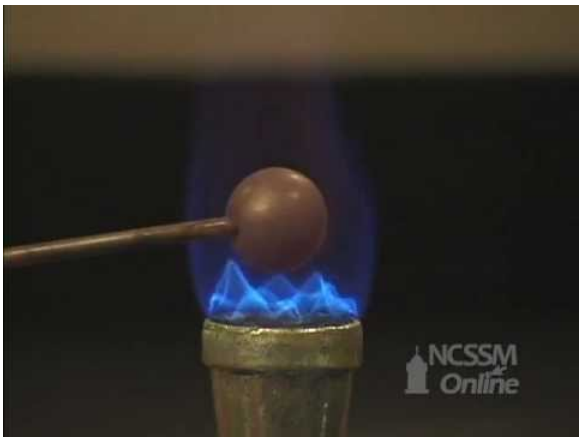


Rising Sea Level

Sea level is rising 3.4 millimeters (0.13 inches) per year and has risen 0.19 meters (7.4 inches) from 1901 to 2010. This is thought largely to be from both the melting of glaciers and thermal expansion of sea water. **Thermal expansion** means that as objects such as solids, liquids, and gases heat up, they expand in volume. Since 1970, the melting of glaciers and thermal expansion account for 75% of the sea-level rise [6].

Classic video demonstration (30 seconds) on thermal expansion with brass ball and ring (North Carolina School of Science and Mathematics).

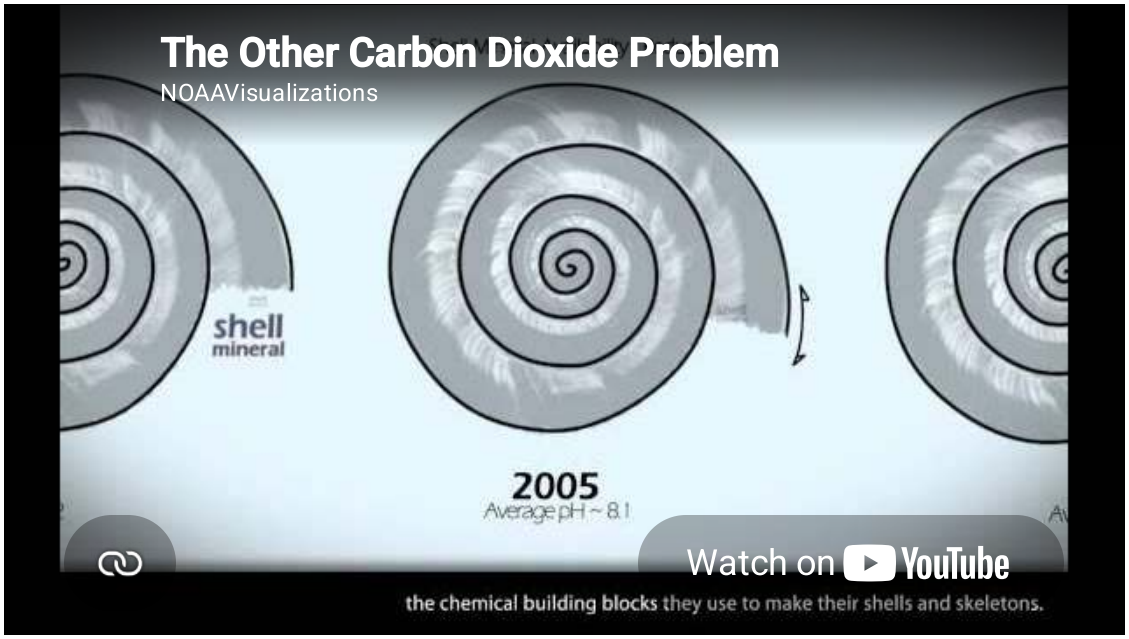


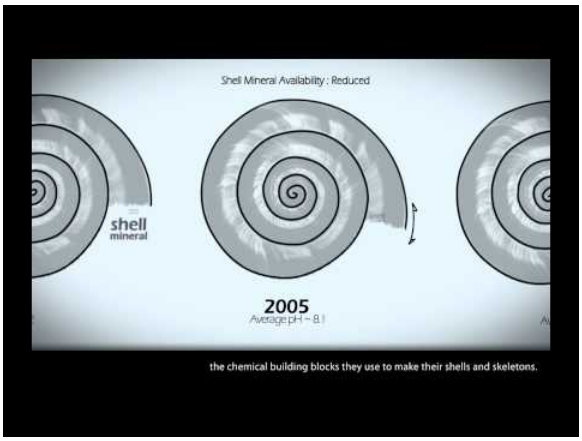


Ocean Acidification

Since 1750, about 40% of the new anthropogenic carbon dioxide has remained in the atmosphere. The remaining 60% gets absorbed by the ocean and vegetation. Therefore, the ocean has absorbed about 30% of new anthropogenic carbon dioxide. When carbon dioxide gets absorbed in the ocean, it creates **carbonic acid**. This makes the ocean more acidic, which has an impact on marine organisms that secrete calcium carbonate shells. Recall that hydrochloric acid reacts by effervescing with limestone rock

made of calcite, which is calcium carbonate. An increase in ocean acidity associated with climate change has been linked to the thinning of the carbonate shells of some sea snails (pteropods) and small protozoan zooplankton (foraminifera) and declining growth rates of coral reefs [6]. Small animals like protozoan zooplankton are an important component in the marine ecosystem. **Ocean acidification** combined with warmer temperature and lower oxygen levels is expected to have severe impacts on marine ecosystems and human-harvested fisheries, possibly affecting our ocean-derived food sources [6]. This video discusses the problem of ocean acidification.





Extreme Weather Events

Occurrence and intensity of **extreme weather events** such as hurricanes, precipitation, and heatwaves are increasing [3; 6]. Since the 1980s, hurricanes, which are generated from warm ocean water, have increased in frequency, intensity, and duration and are likely connected to a warmer climate. Since 1910, average precipitation has increased by 10% in the contiguous United States, and much of this increase is associated with heavy precipitation events like storms [18]. However, the distribution is not even and more precipitation is projected for the northern United States while less precipitation is projected for the already dry southwest [3]. Further, heatwaves have increased and rising temperatures are already affecting crop yields in northern latitudes [6]. Increased heat allows for greater moisture capacity in the atmosphere, increasing the potential for more extreme events [19].

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16.3: Prehistoric Climate Change

Over Earth's history, the climate has changed a lot. For example, during the Mesozoic Era, the age of dinosaurs, the climate was much warmer and carbon dioxide was abundant in the atmosphere. However, throughout the Cenozoic Era (65 million years ago to today), the climate has been gradually cooling. This section summarizes some of these major past climate changes.

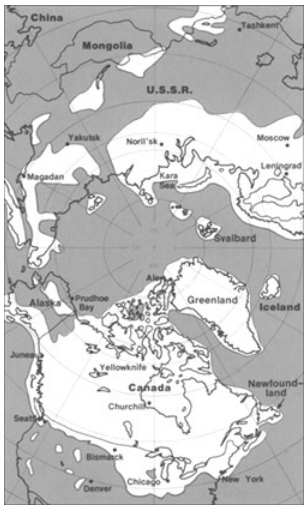


Figure 16.3.1: Maximum extent of Laurentide Ice Sheet.

Past Glaciations

Through geologic history, the climate has changed slowly over millions of years. Before the most recent Pliocene-Quaternary glaciation, there were three other major glaciations [20]. The oldest, known as the **Huronian**, occurred toward the end of the Archean Eon-early Proterozoic Eon (~2.5 billion years ago). The major event of that time, the **Great Oxygenation Event**, is most commonly associated with the cause of that glaciation. The increased oxygen is thought to have reacted with the potent greenhouse gas methane, causing cooling [21].

The end of the Proterozoic Eon (about 700 million years ago) had other glaciations, part of the **Snowball Earth hypothesis** [22]. Glacial evidence has been interpreted in widespread rock sequences globally and even has been linked to low-latitude glaciation [23]. Limestone rock (usually formed in tropical marine environments) and glacial deposits (usually formed in cold climates) are often found together from this time in regions all around the world. In Utah, Antelope Island in the Great Salt Lake has interbedded limestone and glacial deposits (diamictites) interpreted to be formed by continental glaciation [24].

The idea of the controversial Snowball Earth hypothesis is that a runaway **albedo effect** (ice and snow reflecting solar radiation) might cause the complete freezing of land and ocean surfaces and a collapse of biological activity. The ice-covered Earth would only melt from greenhouse heating when carbon dioxide from volcanoes reached high concentrations, due to the inability for carbon dioxide to enter the then-frozen ocean. Some studies estimated carbon dioxide was 350 times higher than today's concentrations [22]. The complete freezing [25] and the extent of the freezing [26] has come into question because biological activity did survive. A competing hypothesis is the **Slushball Earth hypothesis** in which some regions of the equatorial ocean remained open. Differing scientific conclusions about the stability of Earth's magnetic poles, impacts on ancient rock evidence from subsequent metamorphism, and alternate interpretations of existing evidence keep the idea of Snowball Earth controversial.

Glaciation also occurred in the Paleozoic Era, notably the **Andean-Saharan glaciation** in the late Ordovician, about 440–460 million years ago, which coincided with a major extinction event, and the **Karoo Ice Age** during the Pennsylvanian Period, 323 to 300 million years ago. This glaciation was one of the evidences cited by Wegener for his Continental Drift hypothesis as his proposed Pangea drifted into south polar latitudes. This also was caused by an increase of oxygen and a subsequent drop in carbon dioxide, most likely produced by the evolution and rise of land plants [27].

During the Cenozoic Era (the last 65 million years), the climate started out warm and gradually cooled to today. This warm time is called the **Paleocene-Eocene Thermal Maximum** and Antarctica and Greenland were ice-free during this time. Since the Eocene, tectonic events during the Cenozoic caused persistent and significant planetary cooling. For example, the collision of the Indian

Plate with the Asian Plate created the Himalaya Mountains, increasing weathering and erosion rates. An increased rate of weathering of silicate minerals, especially feldspar, consumes carbon dioxide from the atmosphere and therefore reduces the greenhouse effect, resulting in long-term cooling [28].

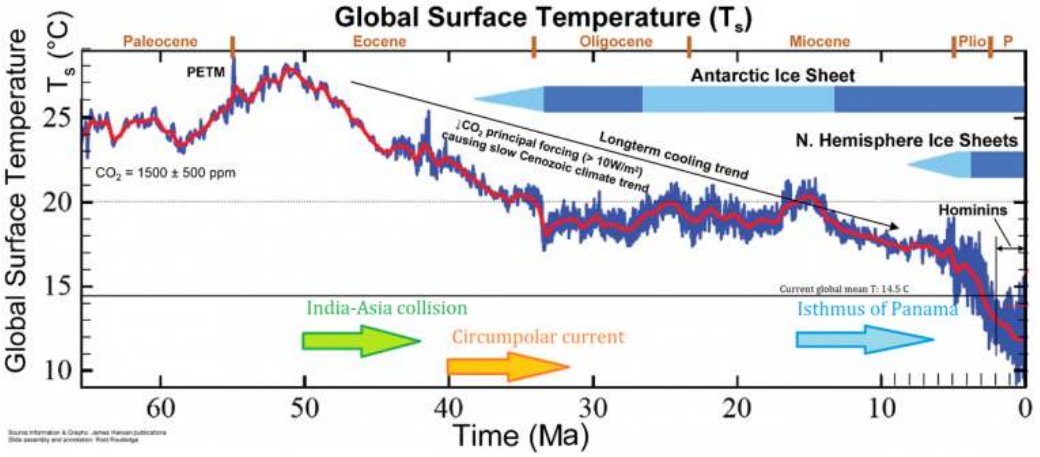


Figure 16.3.2: Global average surface temperature over the past 70 million years.

At about 40 million years ago, the narrow gap between the South American Plate and the Antarctica Plate widened, resulting in the opening of the Drake Passage. This allowed for the unrestricted west-to-east flow of water around Antarctica, the **Antarctic Circumpolar Current**, which effectively isolated the southern ocean from the warmer waters of the Pacific, Atlantic, and Indian Oceans. The region cooled significantly, and by 35 million years ago (Oligocene) glaciers had started to form on Antarctica [29].

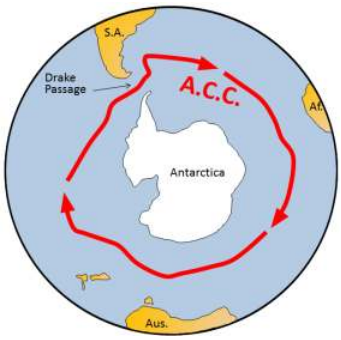
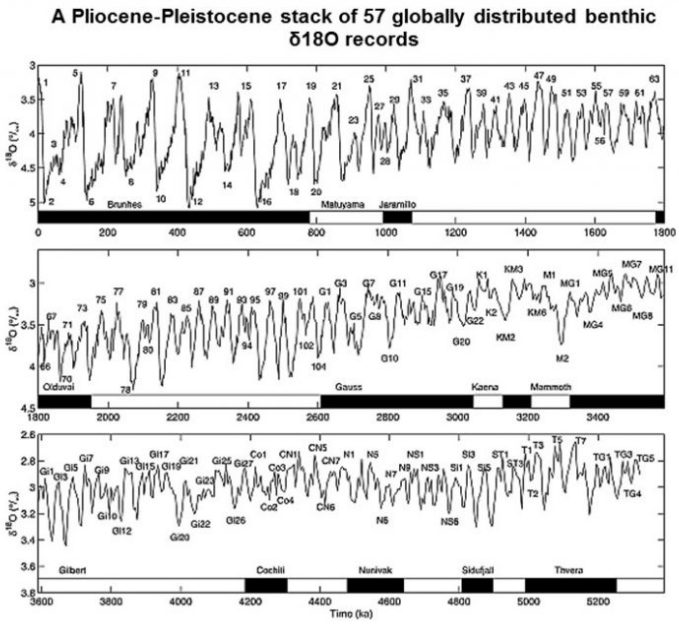


Figure 16.3.3: The Antarctic Circumpolar Current (ACC).

At around 15 million years ago, subduction-related volcanism between Central and South America created the Isthmus of Panama that connected North and South America. This prevented water from flowing between the Pacific and Atlantic Oceans and reduced heat transfer from the tropics to the poles. This reduced heat transfer and created a cooler Antarctica and larger Antarctic glaciers. The expansion of that ice sheet (on land and water) increased Earth's reflectivity of solar radiation. This enhanced the albedo effect, creating a positive feedback loop of further cooling: more reflective glacial ice, more cooling, more ice, and so on [30; 31].

By 5 million years ago (Pliocene Epoch), ice sheets had started to grow in North America and northern Europe. The most intense part of the current glaciation is the last 1 million years of the Pleistocene Epoch. The Pleistocene has significant temperature variations (through a range of almost 10°C or 18°F) on time scales of 40,000 to 100,000 years, and corresponding expansion and contraction of ice sheets. These variations are attributed to subtle changes in Earth's orbital parameters called **Milankovitch cycles** [32; 33]. Over the past million years, the glaciation cycles have been approximately every 100,000 years [34] with many glacial advances in the last 2 million years (Lisiecki and Raymo, 2005) [35].



Source: Lisiecki, L. E., & Raymo, M. E. (2005). A Pliocene-Pleistocene stack of 57 globally distributed benthic δ18O records. *Paleoceanography*, *20*(1).

Figure 16.3.4: A Pliocene-Pleistocene stack of 57 globally distributed benthic δ18O records, as a proxy for temperature. Higher values indicate cooling, and lower values indicate warming. The overall trend illustrates the glaciation cycles. The x-axis represents time in thousands of years (ka) so 200 means 200,000 years. (Source: Lisiecki and Raymo, 2005)

During an ice age, periods of warming climate are called **interglacials**; during interglacials, very brief periods of even warmer climate are called **interstadials**. These warming upticks are related to Earth's climate variations, like Milankovitch cycles, which are changes to the Earth's orbit that can fluctuate climate. In the last 500,000 years, there have been 5 or 6 interglacials, with the most recent belonging to our current time, the Holocene.

Two of the more recent climate swings demonstrate the complexity of the changes: the **Younger Dryas** and the **Holocene Climatic Optimum**. These events are more recent and yet have conflicting information. The Younger Dryas cooling is widely recognized in the Northern Hemisphere [36], though the timing of the event (about 12,000 years ago) does not appear to be equal everywhere [37]. It also is difficult to find in the Southern Hemisphere [38]. The Holocene Climatic Optimum is the warming around 6,000 years ago [39], though it was not universally warmer, and probably not as warm as current warming [40], and not at the same time everywhere [41].

Proxy Indicators of Past Climates

How do we know about past climates? Geologists use proxy indicators to understand past climate. A **proxy indicator** is a biological, chemical, or physical signature preserved in the rock, sediment, or ice record that acts like a "fingerprint" of something in the past [42]. Thus they are an *indirect* indicator of something like climate. For ancient glaciations from the Proterozoic and Paleozoic, there are rock formations of glacial sediments such as the diamictite (or tillite) of the Mineral Fork Formation in Utah. This dark rock has many fine-grained components plus some large out-sized clasts like a modern glacial till [43; 44].

Deep-sea sediment is an indirect indicator of climate change during the Cenozoic Era, about the last 65 million years. Researchers from the Ocean Drilling Program, an international research collaboration, collect deep-sea sediment cores that record continuous sediment accumulation. The sediment provides detailed chemical records of stable carbon and oxygen isotopes obtained from deep-sea benthic foraminifera shells that accumulated on the ocean floor over millions of years. Oxygen isotopes are a proxy indicator of deep-sea temperatures and continental ice volume [45].

Sediment Cores – Stable Oxygen Isotope

Oxygen isotopes are an indicator of past climate. The two main stable oxygen isotopes are ^{16}O and ^{18}O. They both occur in water (H_2O) and in the calcium carbonate ($CaCO_3$) shells of foraminifera as the oxygen component of both of those molecules. The most abundant and lighter isotope is ^{16}O. Since it is lighter, it evaporates more easily from the ocean's surface as water vapor, which later turns to clouds and precipitation on the ocean and land. This evaporation is enhanced in warmer sea water and slightly increases the concentration of ^{18}O in the surface seawater from which the plankton derives the carbonate for its shells. Thus the ratio of ^{16}O and ^{18}O in the fossilized shells in seafloor sediment is a proxy indicator of the temperature and evaporation of seawater.



Figure 16.3.5: Sediment cores from the Greenland continental slope (Source: Hannes Grobe)

Keep in mind, it is harder to evaporate the heavier water and easier to condense it. As evaporated water vapor drifts toward the poles and tiny droplets form clouds and precipitation, droplets of water with ^{18}O tend to form more readily than droplets of the lighter form and precipitate out, leaving the drifting vapor depleted in ^{18}O.

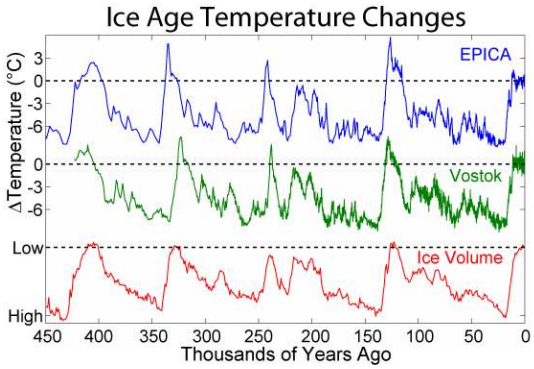


Figure 16.3.6: Antarctic temperature changes during the last few glaciations compared to global ice volume. The first two curves are based on the deuterium (heavy hydrogen) record from ice cores (EPICA Community Members 2004, Petit et al. 1999). The bottom line is ice volume based on oxygen isotopes from a composite of deep-sea sediment cores (Lisiecki and Raymo 2005).

During geologic times when the climate is cooler, more of this lighter precipitation is locked onto land in the form of glacial ice. Consider the giant ice sheets, more than a mile thick, that covered a large part of North America during the last ice age only 14,000 years ago. During glaciation, the glaciers effectively lock away more ^{16}O, thus the ocean water and foraminifera shells become enriched in ^{18}O. Therefore, a ratio of ^{18}O to ^{16}O ($\delta^{18}O$) in calcium carbonate shells of foraminifera is a proxy indicator of past climate. The sediment cores from the Ocean Drilling Program record a continuous accumulation of sediment of these fossils and the sediment provide a record of glacials, interglacials, and interstadials.

Sediment Cores – Boron Isotopes and Acidity

Ocean acidity is affected by carbonic acid and is a proxy for past atmospheric CO_2 concentrations. To estimate the ocean's pH (acidity) over the past 60 million years, researchers collected deep-sea sediment cores and examined the ancient planktonic foraminifera shells' boron-isotope ratios. Boron has two isotopes: ^{11}B and ^{10}B. In aqueous compounds of boron, the relative abundance of these two isotopes is sensitive to pH (acidity), hence CO_2 concentrations. In the early Cenozoic, around 60 million years ago, CO_2 concentrations were over 2000 ppm and started falling around 55 to 40 million years ago possibly due to reduced CO_2 outgassing from ocean ridges, volcanoes, and metamorphic belts and increased carbon burial due to uplift of the Himalaya

Mountains. By the Miocene (about 24 million years ago), CO_2 levels were below 500 ppm [46] and by 800,000 years ago CO_2 levels didn't exceed 300 ppm [47].

Carbon Dioxide Concentrations in Ice Cores

For the more recent Pleistocene Epoch's climate, there is a more detailed and direct chemical record from coring into the Antarctic and Greenland ice sheets. Snow accumulates on these ice sheets and creates yearly layers. Ice cores have been extracted from ice sheets covering the last 800,000 years. Oxygen isotopes are collected from these annual layers and the ratio of ^{18}O to ^{16}O is used to determine temperature as discussed above. In addition, the ice traps small atmospheric gas bubbles as the snow turns to ice. Analysis of these bubbles reveals the composition of the atmosphere at these previous times.

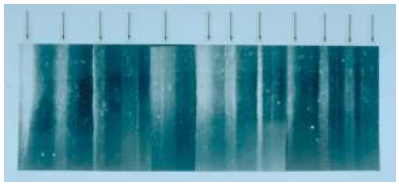


Figure 16.3.7: 19 cm long section of ice core showing 11 annual layers with summer layers (arrowed) sandwiched between darker winter layers. (Source: US Army Corps of Engineers)

Small pieces of this ice are crushed and the ancient air is extracted into a mass spectrometer that can detect the chemistry of the ancient atmosphere. Carbon dioxide levels are recreated from these measurements. Over the last 800,000 years, the maximum carbon dioxide concentration during warm times was about 300 ppm (parts per million) and the minimum during cold stretches was about 170 ppm [46; 47; 48]. The carbon dioxide content of Earth's atmosphere is currently over 410 ppm.

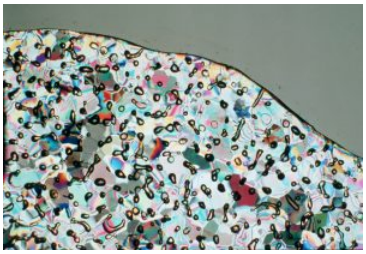


Figure 16.3.8: Antarctic ice showing hundreds of tiny trapped air bubbles from the atmosphere thousands of years ago. (Source: CSIRO)

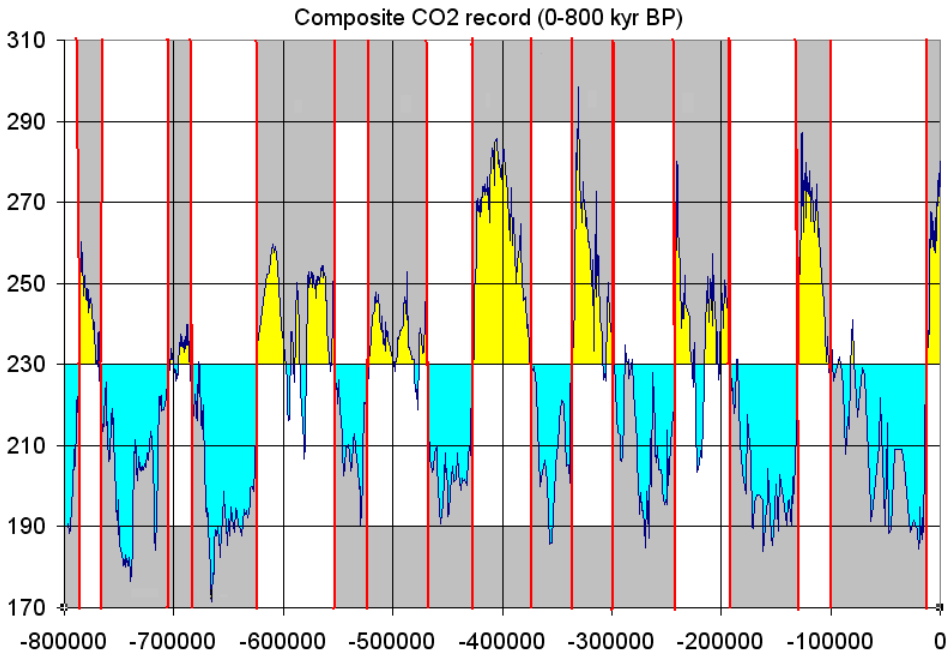


Figure 16.3.9: Composite carbon dioxide record from the last 800,000 years based on ice core data from EPICA Dome C Ice Core.

Oceanic Microfossils

Microfossils, like foraminifera, diatoms, and radiolarians, can be used to interpret past climate records. In sediment cores, different species of microfossils are found in different layers. Groups of these microfossils are called **assemblages**. One assemblage consists of species that lived in cooler ocean water (in glacial times) and another assemblage found at a different level in the same sediment core is made of warmer water species [49]. Watch this video that shows how climate data is captured in sediment cores.





Tree Rings

Tree rings, which form every year as a tree grows, are another past climate indicator. Rings that are thicker indicate wetter years, and rings that are thinner and closer together indicate drier years. Every year a tree will grow one ring with a light section and dark section. The rings vary in width. Since trees need a lot of water to survive, narrower rings indicate colder and drier climates. Since some trees can be several thousand years old, we can use their rings for regional paleoclimatic reconstructions, for example, to reconstruct past temperature, precipitation, vegetation, streamflow, sea-surface temperature, and other climate-dependent conditions. **Paleoclimatic study** means relating to a distinct past geologic climate. Further, dead trees such as those used in Puebloan ruins can be used to extend this proxy indicator, which showed long term droughts in the region and why their villages were abandoned.



Figure 16.3.10: Tree rings form every year. Rings that are farther apart are from wetter years and rings that are closer together are from drier years.

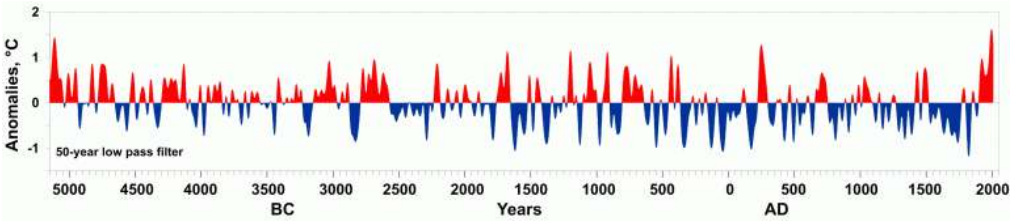


Figure 16.3.11: Summer temperature anomalies for the past 7000 years from tree ring data. (Source: R.M.Hantemirov)

Pollen

Pollen is also a proxy climate indicator. Flowering plants produce **pollen** grains. Pollen is distinctive when viewed under a microscope. Sometimes pollen can be preserved in lake sediments that accumulate in layers every year. Coring of lake sediments can reveal ancient pollen. Fossil **pollen assemblages** are groups of pollen from multiple species such as spruce, pine, and oak. Through time (via the sediment cores and radiometric age-dating techniques), the pollen assemblage will change revealing the plants that lived in the area at the time. Thus the pollen assemblages are an indicator of past climate since different plants will prefer different climates [50]. For example, in the Pacific Northwest east of the Cascades, a region close to the border of grasslands and forest, a study tracked pollen over the last 125,000 years covering the last two glaciations. Pollen assemblages with more pine tree pollen are found during glaciations and pollen assemblages with less pine tree pollen are found during interglacial times [51].

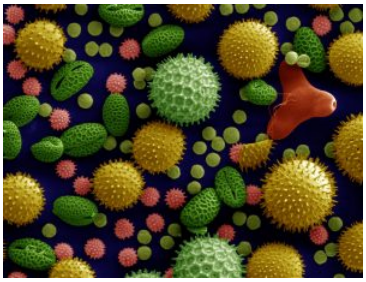


Figure 16.3.12: Scanning electron microscope image of modern pollen with false color added to distinguish plant species. (Source: Dartmouth Electron Microscope Facility, Dartmouth College)

Other Proxy Indicators

Paleoclimatologists study many other phenomena to understand past climates such as human historical accounts, human instrument record from the recent past, lake sediments, cave deposits, and corals.

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16.4: Anthropogenic Causes of Climate Change

As shown in the previous section, prehistoric changes in climate have been very slow. Climate changes typically occur slowly over many millions of years. The climate changes observed today are rapid and largely human-caused. Evidence shows that climate is changing, but what is causing that change? Scientists have suspected since the late 1800s that human-produced (anthropogenic) changes in atmospheric greenhouse gases would likely cause climate change, as changes in these gases have been the case every time in the geologic past. By the middle 1900s, systematic measurements began which confirmed that human-produced carbon dioxide was accumulating in the atmosphere and other Earth systems, like forests and the oceans. By the end of the 1900s and into the early 2000's the **Theory of Anthropogenic Climate Change** was solidified as evidence from thousands of ground-based studies and continuous satellite measurements of land and ocean mounted, revealing the expected temperature increase. Theories evolve and transform as new data and new techniques become available, but they represent the state of thinking for that field. The Theory of Anthropogenic Climate Change is that humans are causing most of the current changes to climate by burning fossil fuels such as coal, oil, and natural gas. This section summarizes the scientific understanding of anthropogenic climate change.

Scientific Consensus

The overwhelming majority of climate studies indicate that human activity is causing rapid changes to the climate, which will cause severe environmental damage. There is a strong scientific consensus on the issue. Studies published in peer-reviewed scientific journals show that 97% of climate scientists agree that climate warming is from human activities [52]. There is no alternative explanation for the observed link between human-produced greenhouse gas emissions and the changing modern climate. Most leading scientific organizations endorse this position, including the U.S. **National Academy of Science** which was established in 1863 by an act of Congress under President Lincoln. Congress charged the National Academy of Science "with providing independent, objective advice to the nation on matters related to science and technology" [53]. Therefore, the National Academy of Science is the leading authority when it comes to policy advice related to scientific issues.

One way we know that the increased greenhouse gas emissions are from human activities is with isotopic fingerprints. For example, fossil fuels have a ratio of stable carbon-13 to carbon-12 ($^{13}C/^{12}C$) that is different from today's stable carbon ratio in the atmosphere. Studies have been using isotopic carbon signatures to identify anthropogenic carbon in the atmosphere since the 1980s. Isotopic records from the Antarctic Ice Sheet show stable isotopic signature from ~1000 CE to ~1800 CE and a steady isotopic signature gradually changing since 1800 followed by rapid change after 1950 as burning of fossil fuels dilutes the CO_2 in the atmosphere. These changes show the atmosphere having a carbon isotopic signature increasingly more similar to that of fossil fuels [42; 54].

Anthropogenic Sources of Greenhouse Gases

Anthropogenic emissions of greenhouse gases have increased since pre-industrial times due to global economic growth and population growth. Atmospheric concentrations of the leading greenhouse gas, carbon dioxide, are at unprecedented levels that haven't been observed in at least the last 800,000 years [6]. The pre-industrial level of carbon dioxide was at about 278 parts per million (ppm). As of 2016, carbon dioxide was, for the first time, above 400 ppm for the entirety of the year. Measurements of atmospheric carbon at the Mauna Loa Carbon Dioxide Observatory show a continuous increase since 1957 when the observatory was established from 315 ppm to over 417 ppm in 2022. The daily reading today can be seen at Daily CO_2. Based on the ice core record over the past 800,000 years, carbon dioxide ranged from about 185 ppm during ice ages to 300 ppm during warm times [52].

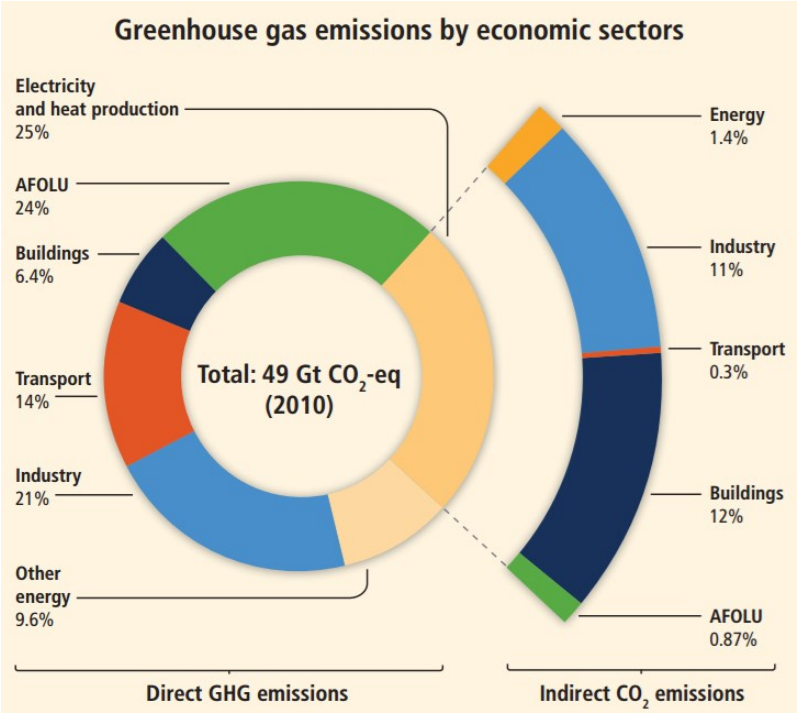


Figure 16.4.1: Total anthropogenic greenhouse gas (GHG) emissions from economic sectors in 2010. The circle shows the shares of direct GHG emissions (in % of total anthropogenic GHG emissions) from five economic sectors in 2010. The pull-out shows how shares of indirect CO_2 emissions (in % of total anthropogenic GHG emissions) from electricity and heat production are attributed to sectors of final energy use. AFOLU is agriculture, forestry, and other land use (Source: Pachauri et al. 2014).

What is the source of these anthropogenic greenhouse gas emissions? Fossil fuel combustion and industrial processes contributed 78% of all emissions since 1970. Sectors of the economy responsible for most of this include electricity and heat production (25%); agriculture, forestry, and land use (24%); industry (21%); transportation including automobiles (14%); other energy production (9.6%); and buildings (6.4%) [6]. More than half of greenhouse gas emissions have occurred in the last 40 years [6] and 40% of these emissions have stayed in the atmosphere. Unfortunately, despite scientific consensus, efforts to mitigate climate change require political action. Despite the growing amount of climate change concern, mitigation efforts, legislation, and international agreements have reduced emissions in some places, yet the continued economic growth of the less developed world has increased global greenhouse gas emissions. In fact, the time between 2000 and 2010 saw the largest increases since 1970 [6].

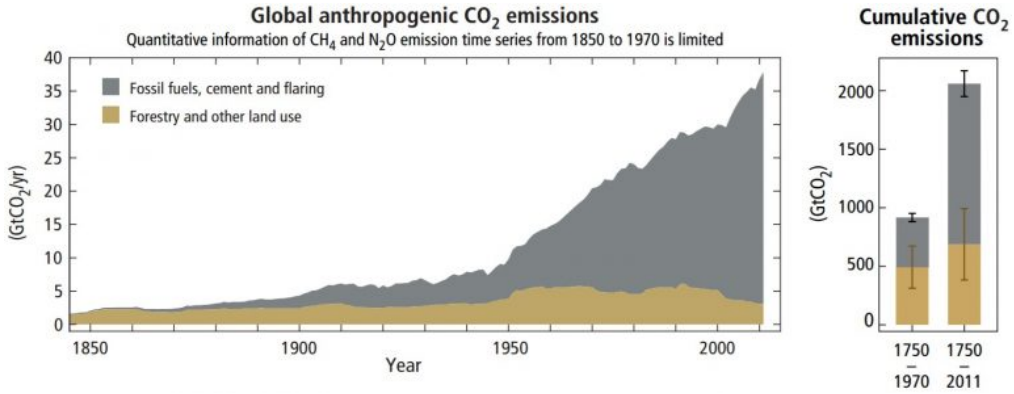


Figure 16.4.2: Annual global anthropogenic carbon dioxide (CO_2) emissions in gigatonne of CO_2-equivalent per year ($GtCO_2$/yr) from fossil fuel combustion, cement production and flaring, and forestry and other land use, 1850–2011. Cumulative emissions and their uncertainties are shown as bars and whiskers.

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16.S: Summary

Summary

Included in Earth Science is the study of the system of processes that affect surface environments and atmosphere of the Earth. Recent changes in atmospheric temperature and climate over intervals of decades have been observed. For Earth's climate to be stable, incoming radiation from the Sun and outgoing radiation from the Sun-warmed Earth must be in balance. Gases in the atmosphere called greenhouse gases absorb the infrared thermal radiation from the Earth's surface, trapping that heat and warming the atmosphere, a process called the greenhouse effect. Thus the energy budget is not now in balance and the Earth is warming. Human activity produces many greenhouse gases that have accelerated climate change. CO_2 from fossil fuel burning is one of the major ones. While atmospheric composition is mostly nitrogen and oxygen, trace components including the greenhouse gases (CO_2 and methane are the major ones and there are others) have the greatest effect on global warming.

A number of positive feedback mechanisms, processes whose results reinforce the original process, take place in the Earth system. An example of a positive feedback mechanism of great concern is permafrost melting which causes decay of melting organic material that produces CO_2 and methane (both powerful greenhouse gases) that warm the atmosphere and promote more permafrost melting. Two carbon cycles affect Earth's atmospheric CO_2 composition, the biologic carbon cycle and the geologic carbon cycle. In the biologic cycle, organisms (mostly plants and also animals that eat them) remove CO_2 from the atmosphere for energy and to build their body tissues and return it to the atmosphere when they die and decay. The biologic cycle is a rapid cycle. In the geologic cycle, some organic matter is preserved in the form of petroleum and coal while more is dissolved in seawater and captured in carbonate sediments, some of which is subducted into the mantle and returned by volcanic activity. The geologic carbon cycle is slow over geologic time.

Measurements of increasing atmospheric temperature have been made since the nineteenth century but the upward temperature trend itself increased in the mid-twentieth century showing the current trend is exponential. Because of the high specific heat of water, the oceans have absorbed most of the added heat. That this is temporary storage is revealed by the record-breaking warm years of the recent decade and the increase in intense storms and hurricanes. In 1957 the Mauna Loa CO_2 Observatory was established in Hawaii providing constant measurements of atmospheric CO_2 since 1958. The initial value was 315 ppm. The Keeling Curve, named for the observatory founder, shows that value has steadily increased, exponentially, to over 417 ppm now. Compared to proxy data from atmospheric gases trapped in ice cores that show a maximum value for CO_2 of about 300 ppm over the last 800,000 years, the Keeling increase of over 100 ppm in 50 years is dramatic evidence of human-caused CO_2 increase and climate change! As Earth's temperature rises, glaciers and ice sheets are shrinking resulting in sea level rise. Atmospheric CO_2 is also absorbed in sea water producing increased concentrations of carbonic acid which is raising the pH of the oceans making it harder for marine life to extract carbonate for their skeletal materials.

Earth's climate has changed over geologic time with periods of major glaciations. There was a high temperature period in the Mesozoic shown by fossils in high latitudes and the Western Interior Seaway covering what is now the Midwest. However, climate has been cooling during the Cenozoic culminating in the Ice Age. Since the Ice Age, several proxy indicators of ancient climate show that the rate and amount of current climate change is unique in geologic history and can only be attributed to human activity. Those who ignore the consequences of increasing global warming for our planet's future do so at the peril of our posterity!

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CHAPTER OVERVIEW

17: The Origin of the Universe and Our Solar System

Learning Objectives

- Explain the big bang theory and the origin of the elements.
- Understand the origin of our solar system.
- Describe how the objects in our solar system are identified, explored, and characterized.
- Describe the types of small bodies in our solar system, their locations, and how they formed.
- Describe the characteristics of the giant planets, terrestrial planets, and small bodies in the solar system.
- Explain what influences the temperature of a planet's surface.
- Explain why there is geological activity on some planets and not on others.
- Describe different methods for dating planets and the age of the solar system.

The universe began 13.77 billion years ago when energy, matter, and space expanded from a single point. Evidence for the big bang is the cosmic “afterglow” from when the universe was still very dense, and red-shifted light from distant galaxies, which tell us the universe is still expanding.

The big bang produced hydrogen, helium, and lithium, but heavier elements come from nuclear fusion reactions in stars. Large stars make elements such as silicon, iron, and magnesium, which are important in forming terrestrial planets. Large stars explode as supernova and scatter the elements into space.

Planetary systems begin with the collapse of a cloud of gas and dust. Material drawn to the center forms a star, and the remainder forms a disk around the star. Material within the disk clumps together to form planets. In our solar system, rocky planets are closer to the Sun, and ice and gas giants are farther away. This is because temperatures near the Sun were too high for ice to form, but silicate minerals and metals could solidify.

17.1: Origin of the Universe

17.2: Origin of the Solar System—The Nebular Hypothesis

17.3: Overview of Our Planetary System

17.4: Composition and Structure of Planets

17.S: Summary

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17.1: Origin of the Universe

The universe appears to have an infinite number of galaxies and solar systems and our solar system occupies a small section of this vast entirety. The origins of the universe and solar system set the context for conceptualizing the Earth's origin and early history.



Figure 17.1.1: The Hubble Deep Field. This image, released in 1996, is a composite long-exposure picture of one of the darkest parts of the night sky. Every light on this image that does not have diffraction spikes is believed to be an entire galaxy, with hundreds of billions of stars, demonstrating the immense size and scope of the universe.

Big Bang Theory

The mysterious details of events prior to and during the origin of the universe are subject to great scientific debate. The prevailing idea about how the universe was created is called the **big bang theory**. Although the ideas behind the big bang theory feel almost mystical, they are supported by Einstein's theory of general relativity [1]. Other scientific evidence, grounded in empirical observations, supports the big bang theory.

The big bang theory proposes the universe was formed from an infinitely dense and hot core of the material. The bang in the title suggests there was an explosive, outward expansion of all matter and space that created atoms. Spectroscopy confirms that hydrogen makes up about 74% of all matter in the universe. Since its creation, the universe has been expanding for 13.8 billion years and recent observations suggest the rate of this expansion is increasing [2].

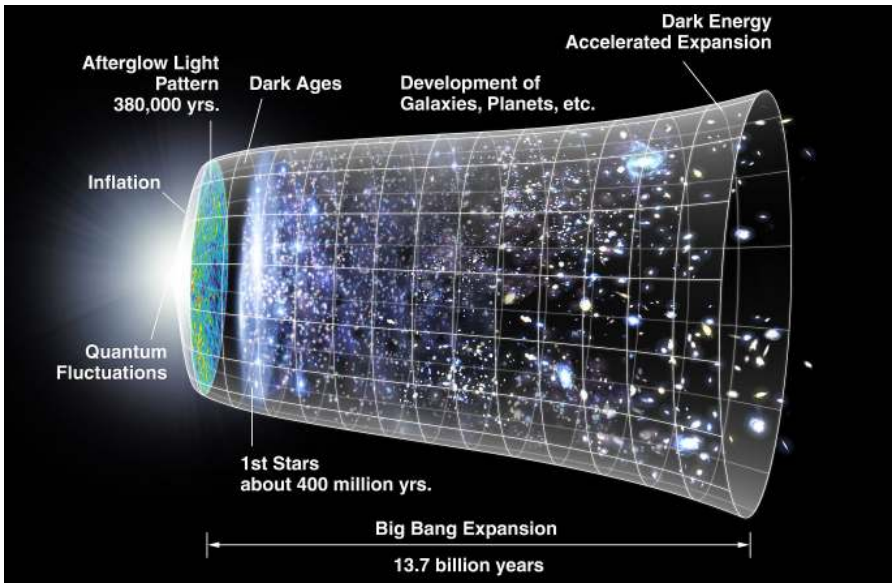


Figure 17.1.2: Timeline of expansion of the universe.

Spectroscopy

Spectroscopy is the investigation and measurement of spectra produced when materials interact with or emit electromagnetic radiation. **Spectra** is the plural for *spectrum* which is a particular wavelength from the **electromagnetic spectrum**. Common

spectra include the different colors of visible light, x-rays, ultraviolet waves, microwaves, and radio waves. Each beam of light is a unique mixture of wavelengths that combine across the spectrum to make the color we see. The light wavelengths are created or absorbed inside atoms, and each wavelength signature matches a specific element. Even white light from the Sun, which seems like an uninterrupted continuum of wavelengths, has gaps in some wavelengths. The gaps correspond to elements present in the Earth's atmosphere that act as filters for specific wavelengths. These missing wavelengths were famously observed by Joseph von Fraunhofer (1787–1826) in the early 1800s [3], but it took decades before scientists were able to relate the missing wavelengths to atmospheric filtering. Spectroscopy shows that the Sun is mostly made of hydrogen and helium. Applying this process to light from distant stars, scientists can calculate the abundance of elements in a specific star and visible universe as a whole. Also, this spectroscopic information can be used as an interstellar speedometer.

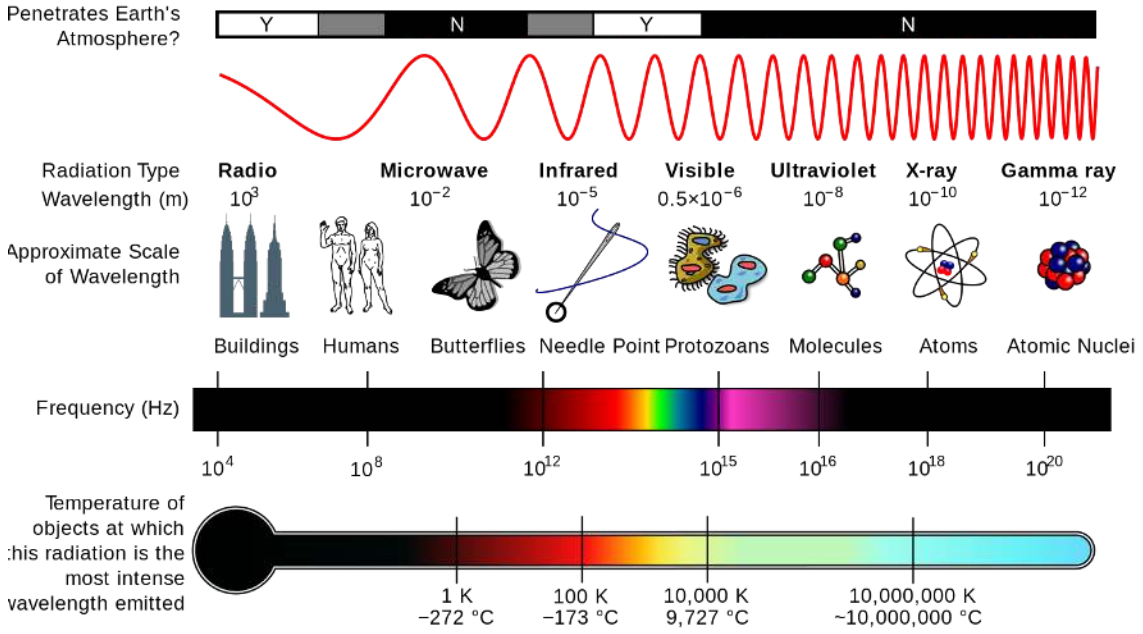


Figure 17.1.3: The electromagnetic spectrum and properties of light across the spectrum.

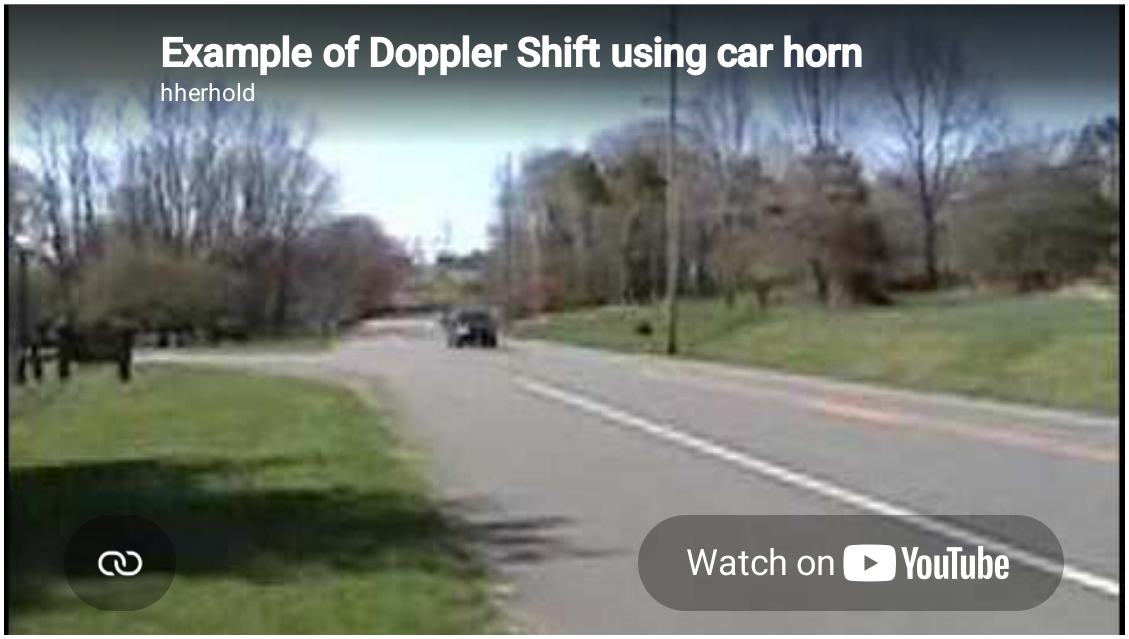
Redshift

The **Doppler effect** is the same process that changes the pitch of the sound of an approaching car or ambulance from high to low as it passes. When an object emits waves, such as light or sound, while moving toward an observer, the wavelengths get compressed. In sound, this results in a shift to a higher pitch. When an object moves away from an observer, the wavelengths are extended, producing a lower-pitched sound. The Doppler effect is used on light emitted from stars and galaxies to determine their speed and direction of travel. Scientists, including Vesto Slipher (1875–1696) [6] and Edwin Hubble (1889–1953) [7], examined galaxies both near and far and found that almost all galaxies outside of our galaxy are moving away from each other, and us. Because the light wavelengths of receding objects are extended, visible light is shifted toward the red end of the spectrum, called a **redshift**. In addition, Hubble noticed that galaxies that were farther away from Earth also had a greater amount of redshift, and thus, the faster they are traveling away from us. The only way to reconcile this information is to deduce the universe is still expanding. Hubble's observation forms the basis of the big bang theory.



Figure 17.1.4: This animation demonstrates how the Doppler effect is heard as a car moves. The waves in front of the car are compressed together, making the pitch higher. The waves in the back of the car are stretched, and the pitch gets lower.

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Cosmic Microwave Background Radiation

Another strong indication of the big bang is **cosmic microwave background radiation**. Cosmic radiation was accidentally discovered by Arno Penzias (1933–2024) and Robert Woodrow Wilson (1936–) [8] when they were trying to eliminate background noise from a communication satellite. They discovered very faint traces of energy or heat that are omnipresent across the universe. This energy was left behind from the big bang, like an echo.

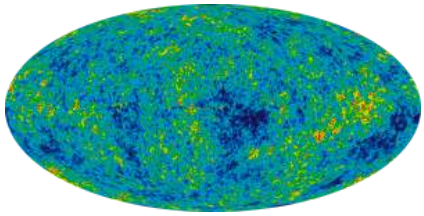


Figure 17.1.5: Heat map, showing slight variations in background heat, which is related to cosmic background radiation.

Stellar Evolution

Astronomers think the big bang created lighter elements, mostly hydrogen and smaller amounts of elements helium, lithium, and beryllium. Another process must be responsible for creating the other 90 heavier elements. The current model of stellar evolution explains the origins of these heavier elements.

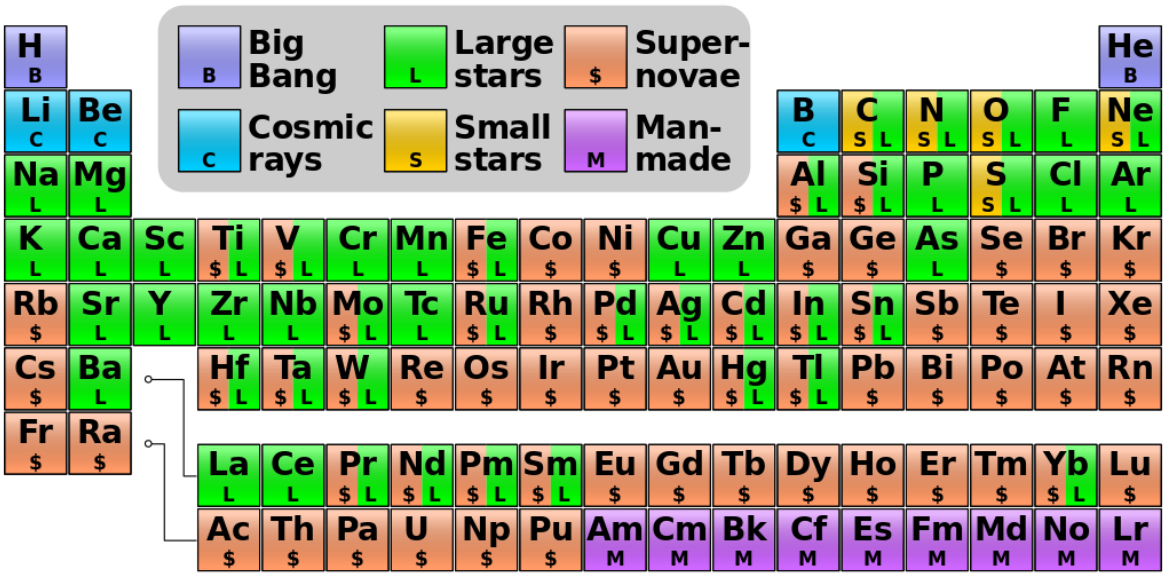


Figure 17.1.6: Origin of the elements on the periodic table, showing the important role the star life cycle plays.

Birth of a Star

Stars start their lives as elements floating in cold, spinning clouds of gas and dust known as **nebulas**. Gravitational attraction or perhaps a nearby stellar explosion causes the elements to condense and spin into a disk shape. In the center of this disk shape, a new star is born under the force of gravity. The spinning whirlpool concentrates material in the center, and the increasing gravitational forces collect even more mass. Eventually, the immensely concentrated mass of material reaches a critical point of such intense heat and pressure it initiates fusion.

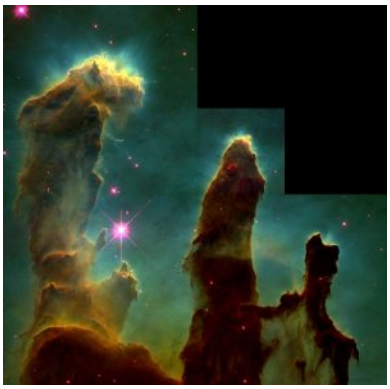


Figure 17.1.7: Section of the Eagle Nebula known as “The Pillars of Creation.”

Fusion

Fusion is not a chemical reaction. **Fusion** is a nuclear reaction in which two or more nuclei, the centers of atoms, are forced together and combine creating a new larger atom. This reaction gives off a tremendous amount of energy, usually as light and solar radiation. An element such as hydrogen combines or fuses with other hydrogen atoms in the core of a star to become a new element, in this case, helium. Another product of this process is energy, such as solar radiation that leaves the Sun and comes to the Earth as light and heat. Fusion is a steady and predictable process, which is why we call this the **main phase** of a star’s life. During its main phase, a star turns hydrogen into helium. Since most stars contain plentiful amounts of hydrogen, the main phase may last billions of years, during which their size and energy output remains relatively steady.

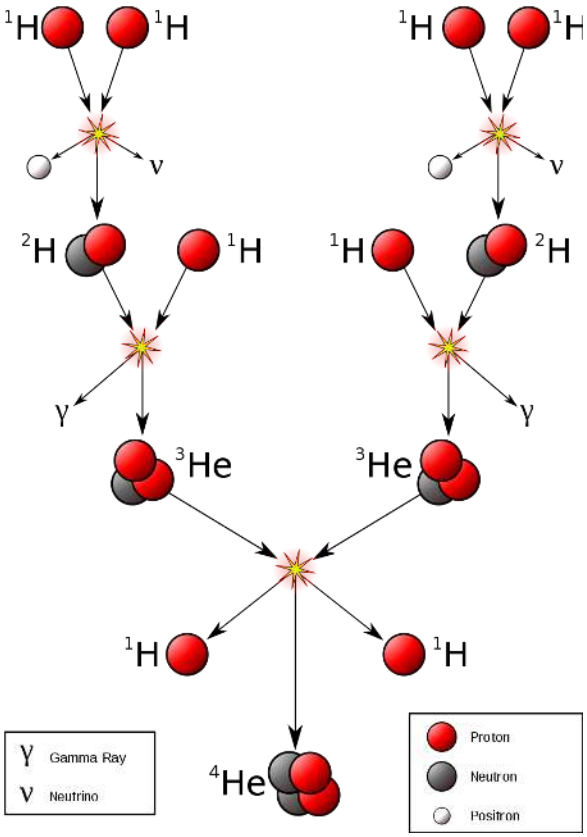


Figure 17.1.8: General diagram showing the series of fusion steps that occur in the sun.

The **giant phase** in a star's life occurs when the star runs out of hydrogen for fusion. If a star is large enough, it has sufficient heat and pressure to start fusing helium into heavier elements. This style of fusion is more energetic and the higher energy and temperature expand the star to a larger size and brightness. This giant phase is predicted to happen to our Sun in another few billion years, growing the radius of the Sun to Earth's orbit, which will render life impossible. The mass of a star during its main phase is the primary factor in determining how it will evolve. If the star has enough mass and reaches a point at which the primary fusion element, such as helium, is exhausted, fusion continues using new, heavier elements. This occurs over and over in very large stars, forming progressively heavier elements like carbon and oxygen. Eventually, fusion reaches its limit as it forms iron and nickel. This progression explains the abundance of iron and nickel in rocky objects, like Earth, within the solar system. At this point, any further fusion absorbs energy instead of giving it off, which is the beginning of the end of the star's life [9].

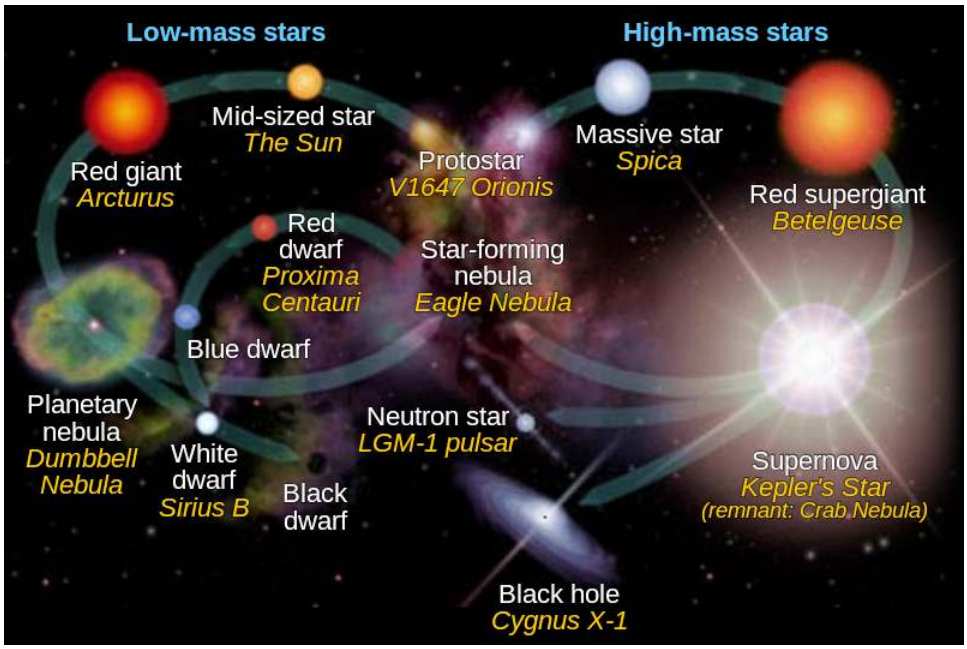


Figure 17.1.9: Two main paths of a star's life cycle, depending on mass.

Death of a Star

The death of a star can range from spectacular to other-worldly (see figure). Stars like the Sun form a planetary nebula, which comes from the collapse of the star's outer layers in an event like the implosion of a building. In the tug-of-war between gravity's inward pull and fusion's outward push, gravity instantly takes over when fusion ends, with the outer gases puffing away to form a nebula. More massive stars do this as well but with a more energetic collapse, which starts another type of energy release mixed with element creation known as a supernova. In a **supernova**, the collapse of the core suddenly halts, creating a massive outward-propagating shock wave. A supernova is the most energetic explosion in the universe short of the big bang. The energy release is so significant the ensuing fusion can make every element up through uranium [10].

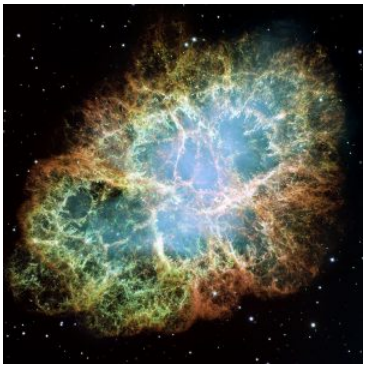


Figure 17.1.10: Hubble space telescope image of the Crab Nebula, the remnants of a supernova that occurred in 1054 CE.

The death of the star can result in the creation of white dwarfs, neutron stars, or black holes. Following their deaths, stars like the Sun turn into white dwarfs. **White dwarfs** are hot star embers, formed by packing most of a dying star's mass into a small and dense object about the size of Earth. Larger stars may explode in a supernova that packs their mass even tighter to become **neutron stars**. Neutron stars are so dense that protons combine with electrons to form neutrons. The largest stars collapse their mass even further, becoming objects so dense that light cannot escape their gravitational grasp. These are the infamous **black holes** and the details of the physics of what occurs in them are still up for debate.

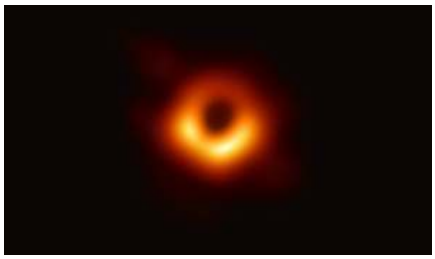


Figure 17.1.11: A black hole and its shadow have been captured in an image for the first time in 2019, a historic feat by an international network of radio telescopes called the Event Horizon Telescope (Source: NASA)

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17.2: Origin of the Solar System—The Nebular Hypothesis

Our solar system formed at the same time as our Sun as described in the nebular hypothesis. The **nebular hypothesis** is the idea that a spinning cloud of dust made of mostly light elements, called a nebula, flattened into a protoplanetary disk, and became a solar system consisting of a star with orbiting planets [12]. The spinning nebula collected the vast majority of material in its center, which is why the Sun accounts for over 99% of the mass in our solar system.

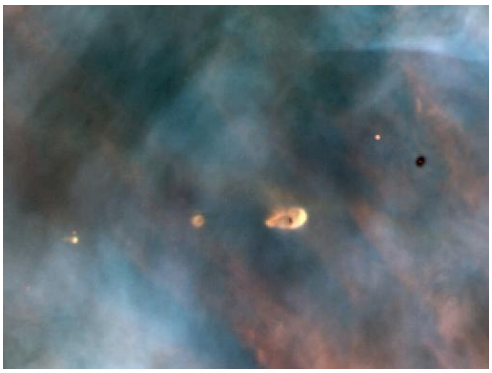


Figure 17.2.1: Small protoplanetary disks in the Orion Nebula.

Planet Arrangement and Segregation

As our solar system formed, the nebular cloud of dispersed particles developed distinct temperature zones. Temperatures were very high close to the center, only allowing condensation of metals and silicate minerals with high melting points. Farther from the Sun, the temperatures were lower, allowing the condensation of lighter gaseous molecules such as methane, ammonia, carbon dioxide, and water [13]. This temperature differentiation resulted in the inner four planets of the solar system becoming rocky, and the outer four planets becoming gas giants.

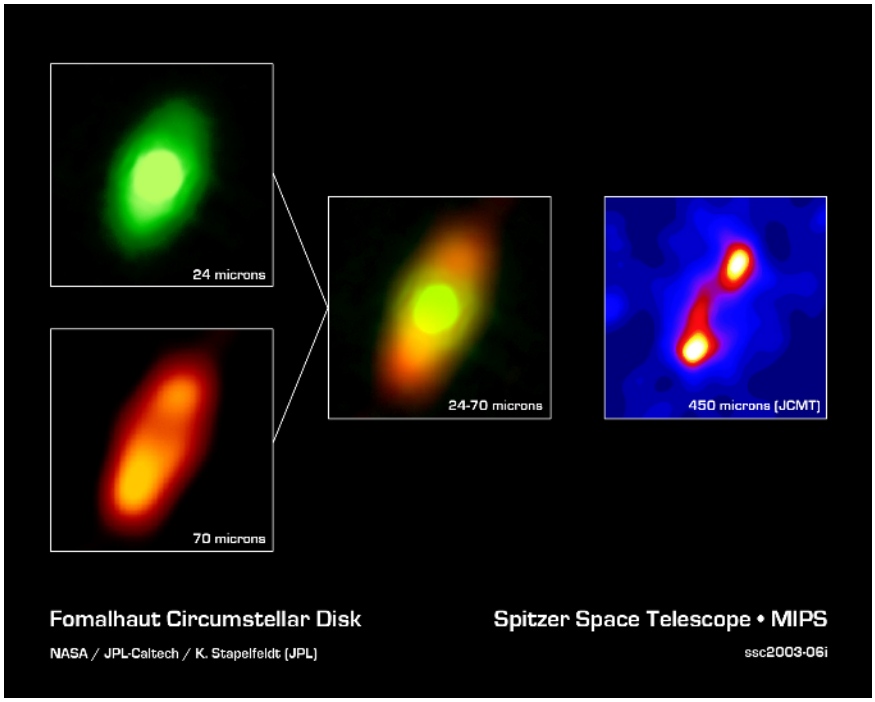


Figure 17.2.2: This disk is asymmetric, possibly because of a large gas giant planet orbiting relatively far from the star.

Both rocky and gaseous planets have a similar growth model. Particles of dust, floating in the disk were attracted to each other by static charges and eventually, gravity. As the clumps of dust became bigger, they interacted with each other—colliding, sticking, and forming proto-planets. The planets continued to grow over the course of many thousands or millions of years, as material from the protoplanetary disk was added. Both rocky and gaseous planets started with a solid core. Rocky planets built more rock on that

core, while gas planets added gas and ice. Ice giants formed later and on the furthest edges of the disk, accumulating less gas and more ice. That is why the gas-giant planets Jupiter and Saturn are composed of mostly hydrogen and helium gas, more than 90%. The ice giants Uranus and Neptune are composed of mostly methane ices and only about 20% hydrogen and helium gases.

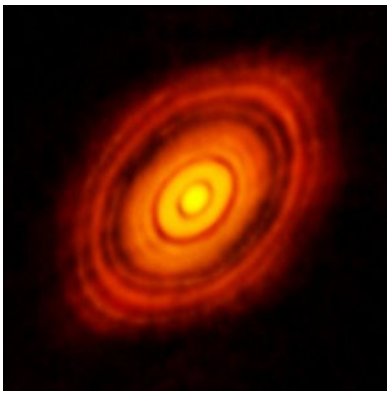


Figure 17.2.3: Image by the ALMA telescope of HL Tauri and its protoplanetary disk, showing grooves formed as planets absorb the material in the disk.

The planetary composition of the gas giants is clearly different from the rocky planets. Their size is also dramatically different for two reasons. First, the original planetary nebula contained more gases and ices than metals and rocks. There was abundant hydrogen, carbon, oxygen, nitrogen, and less silicon and iron, giving the outer planets more building material. Second, the stronger gravitational pull of these giant planets allowed them to collect large quantities of hydrogen and helium, which could not be collected by the weaker gravity of the smaller planets.

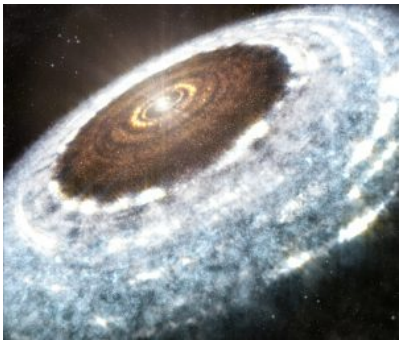


Figure 17.2.4: This artist's impression of the water snowline around the young star V883 Orionis, as detected with ALMA.

Jupiter's massive gravity further shaped the solar system and growth of the inner rocky planets. As the nebula started to coalesce into planets, Jupiter's gravity accelerated the movement of nearby materials, generating destructive collisions rather than constructively gluing material together [14]. These collisions created the asteroid belt, an unfinished planet, located between Mars and Jupiter. This asteroid belt is the source of most **meteorites** that currently impact the Earth. Study of asteroids and meteorites help geologists to determine the age of Earth and the composition of its core, mantle, and crust. Jupiter's gravity may also explain Mars' smaller mass, with the larger planet consuming material as it migrated from the inner to the outer edge of the solar system [15].



Figure 17.2.5: A polished fragment of the iron-rich Toluca Meteorite, with octahedral Widmanstätten Pattern.

Pluto and Planet Definition

The outermost part of the solar system is known as the **Kuiper belt**, which is a scattering of rocky and icy bodies. Beyond that is the **Oort cloud**, a zone filled with small and dispersed ice traces. These two locations are where most comets form and continue to orbit, and objects found here have relatively irregular orbits compared to the rest of the solar system. Pluto, formerly the ninth planet, is located in this region of space. The XXVIth General Assembly of the International Astronomical Union (IAU) stripped Pluto of planetary status in 2006 because scientists discovered an object more massive than Pluto, which they named Eris. The IAU decided against including Eris as a planet, and therefore, excluded Pluto as well. The IAU narrowed the definition of a planet to three criteria:

1. It must be big enough to have enough gravity to force it into a spherical shape.
2. It must orbit a star (the Sun).
3. It must be large enough that its gravity cleared away any other objects of a similar size near its orbit around the Sun.

Pluto passed the first two parts of the definition, but not the third. Pluto and Eris are currently classified as **dwarf planets**.

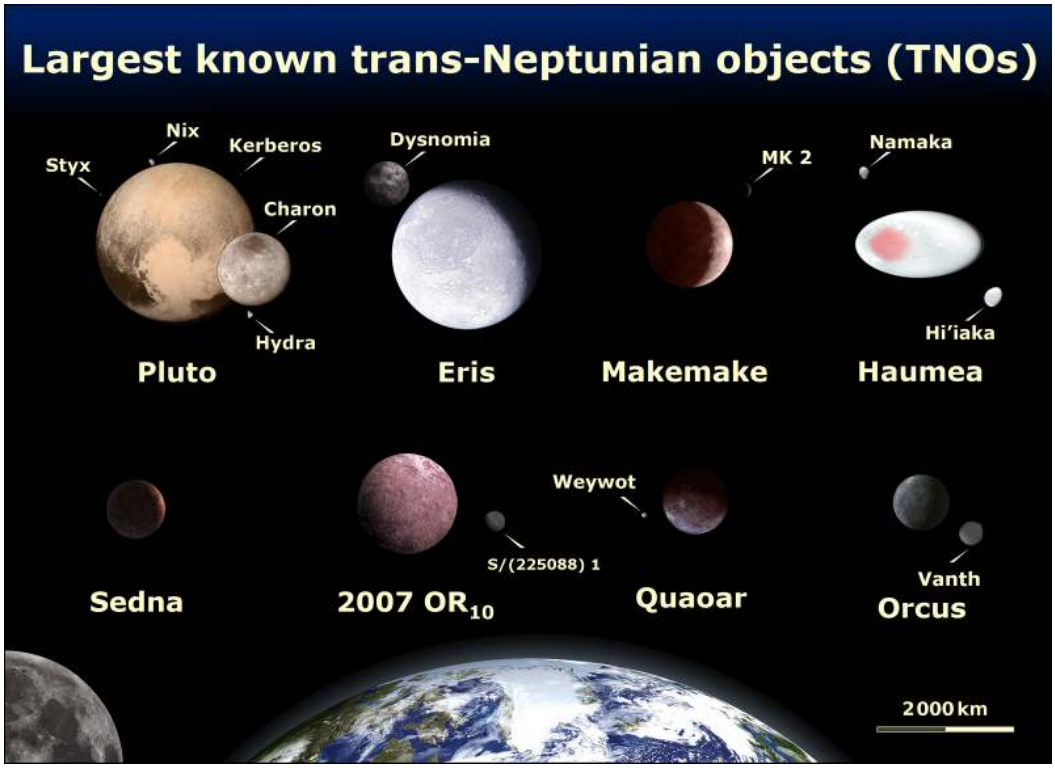


Figure 17.2.6: Eight largest objects discovered past Neptune.

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17.3: Overview of Our Planetary System

The **solar system**[1] consists of the Sun and many smaller objects: the planets, their moons and rings, and such "debris" as asteroids, comets, and dust. Decades of observation and spacecraft exploration have revealed that most of these objects formed together with the Sun about 4.5 billion years ago. They represent clumps of material that condensed from an enormous cloud of gas and dust. The central part of this cloud became the Sun, and a small fraction of the material in the outer parts eventually formed the other objects.

During the past 50 years, we have learned more about the solar system than anyone imagined before the space age. In addition to gathering information with powerful new telescopes, we have sent spacecraft directly to many members of the **planetary system**. (Planetary astronomy is the only branch of astronomy in which we can, at least vicariously, travel to the objects we want to study.) With evocative names such as *Voyager*, *Pioneer*, *Curiosity*, and *Pathfinder*, our robot explorers have flown past, orbited, or landed on every planet, returning images and data that have dazzled both astronomers and the public. In the process, we have also investigated two dwarf planets, hundreds of fascinating moons, four ring systems, a dozen asteroids, and several comets (smaller members of our solar system that we will discuss later).

Our probes have penetrated the atmosphere of Jupiter and landed on the surfaces of Venus, Mars, our Moon, Saturn's moon Titan, the asteroids Eros, Itokawa, Ryugu, and Bennu, and the comet Churyumov-Gerasimenko (usually referred to as 67P). Humans have set foot on the Moon and returned samples of its surface soil for laboratory analysis (Figure 17.3.1). We have flown a helicopter drone on Mars. We have even discovered other places in our solar system that might be able to support some kind of life.



Figure 17.3.1 : Astronauts on the Moon. The lunar lander and surface rover from the Apollo 15 mission are seen in this view of the one place beyond Earth that has been explored directly by humans. (credit: modification of work by David R. Scott, NASA)

An Inventory

The Sun, a star that is brighter than about 80% of the stars in the Galaxy, is by far the most massive member of the solar system, as shown in Table 17.3.1. It is an enormous ball about 1.4 million kilometers in diameter, with surface layers of incandescent gas and an interior temperature of millions of degrees. The Sun is our best-studied example of a star.

Table 17.3.1: Mass of Members of the Solar System

Object	Percentage of Total Mass of Solar System
Sun	99.80
Jupiter	0.10
Comets	0.0005–0.03 (estimate)
All other planets and dwarf planets	0.04
Moons and rings	0.00005
Asteroids	0.000002 (estimate)
Cosmic dust	0.0000001 (estimate)

In Table 17.3.1, note that the Sun is by far the most massive member of the solar system. Table 17.3.1 also shows that most of the material of the planets in the solar system is actually concentrated in the largest one, Jupiter, which is more massive than all the rest of the planets combined. Astronomers were able to determine the masses of the planets centuries ago using Kepler's laws of planetary motion and Newton's law of gravity to measure the planets' gravitational effects on one another or on moons that orbit them. Today, we make even more precise measurements of their masses by tracking their gravitational effects on the motion of spacecraft that pass near them.

Besides Earth, five other planets were known to the ancients—Mercury, Venus, Mars, Jupiter, and Saturn—and two were discovered after the invention of the telescope: Uranus and Neptune. The eight planets all **revolve** in the same direction around the Sun. They orbit in approximately the same plane, like cars traveling on concentric tracks on a giant, flat racecourse. Each planet stays in its own "traffic lane," following a nearly circular orbit about the Sun and obeying the "traffic" laws discovered by Galileo, Kepler, and Newton. Besides these planets, we have also been discovering smaller worlds beyond Neptune that are called **trans-Neptunian objects** or TNOs. The first to be found, in 1930, was Pluto, but others have been discovered during the twenty-first century. One of them, Eris, is about the same size as Pluto and has at least one moon (Pluto has five known moons.) The largest TNOs are also classed as **dwarf planets** as is the largest asteroid, Ceres. To date, more than 2600 of these TNOs have been discovered, and one, called Arrokoth, was explored by the New Horizons spacecraft.

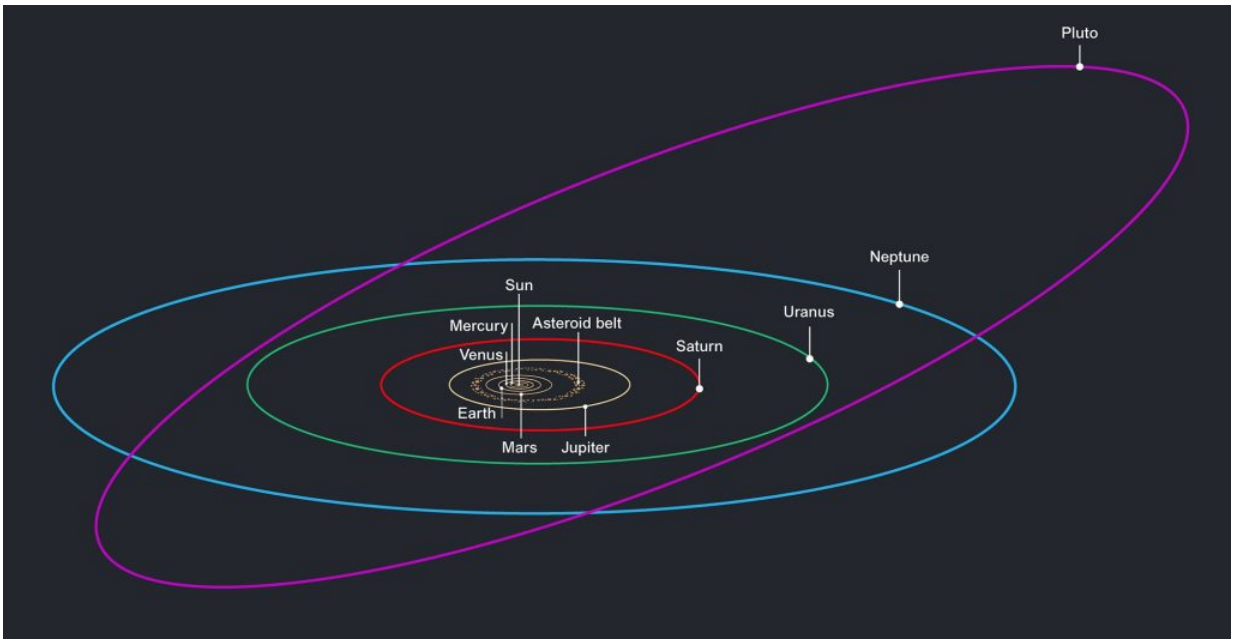


Figure 17.3.2 : Orbits of the Planets. All eight major planets orbit the Sun in roughly the same plane. Note that Pluto's orbit is not in the plane of the planets.

Each of the planets and dwarf planets also **rotates** (spins) about an axis running through it, and in most cases the direction of rotation is the same as the direction of revolution about the Sun. The exceptions are Venus, which rotates backward very slowly (that is, in a **retrograde** direction), and Uranus and Pluto, which also have strange rotations, each spinning about an axis tipped nearly on its side. We do not yet know the spin orientations of the other dwarf planets Eris, Haumea, and Makemake.

The four planets closest to the Sun (Mercury through Mars) are called the inner or terrestrial planets. Often, the Moon is also discussed as a part of this group, bringing the total of terrestrial objects to five. (We generally call Earth's satellite "the Moon," with a capital M, and the other satellites "moons," with lowercase m's.) The terrestrial planets are relatively small worlds, composed primarily of rock and metal. All of them have solid surfaces that bear the records of their geological history in the forms of craters, mountains, and volcanoes (Figure 17.3.3).

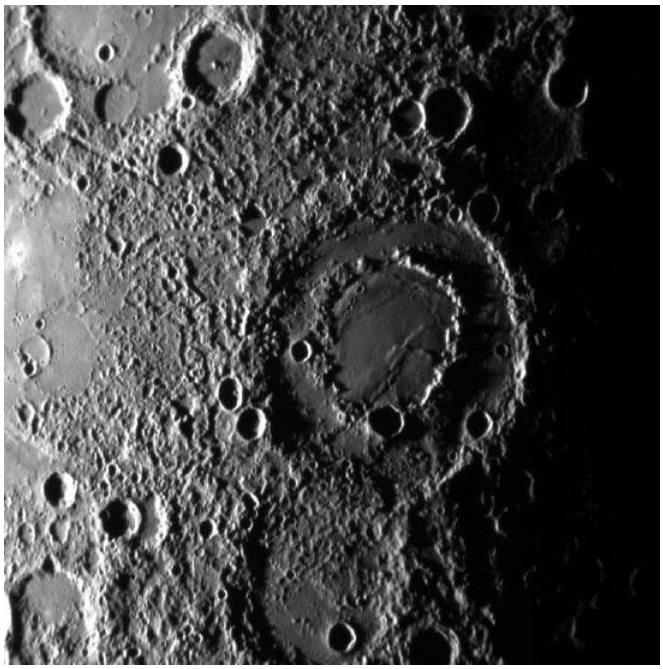


Figure 17.3.3 : Surface of Mercury. The pockmarked face of the terrestrial world of Mercury is more typical of the inner planets than the watery surface of Earth. This image shows Caravaggio, a double-ring impact basin (approximately 160 kilometers in diameter), with another large impact crater on its south-south-western side. (credit: modification of work by NASA/Johns Hopkins University Applied Physics Laboratory/Carnegie Institution of Washington)

The next four planets (Jupiter through Neptune) are much larger and are composed primarily of lighter ices, liquids, and gases. We call these four the Jovian planets (after "Jove," another name for Jupiter in mythology) or giant planets—a name they richly deserve (Figure 17.3.4). About 1300 Earths could fit inside Jupiter, for example. These planets do not have solid surfaces on which future explorers might land. They are more like vast, spherical oceans with much smaller, dense cores.

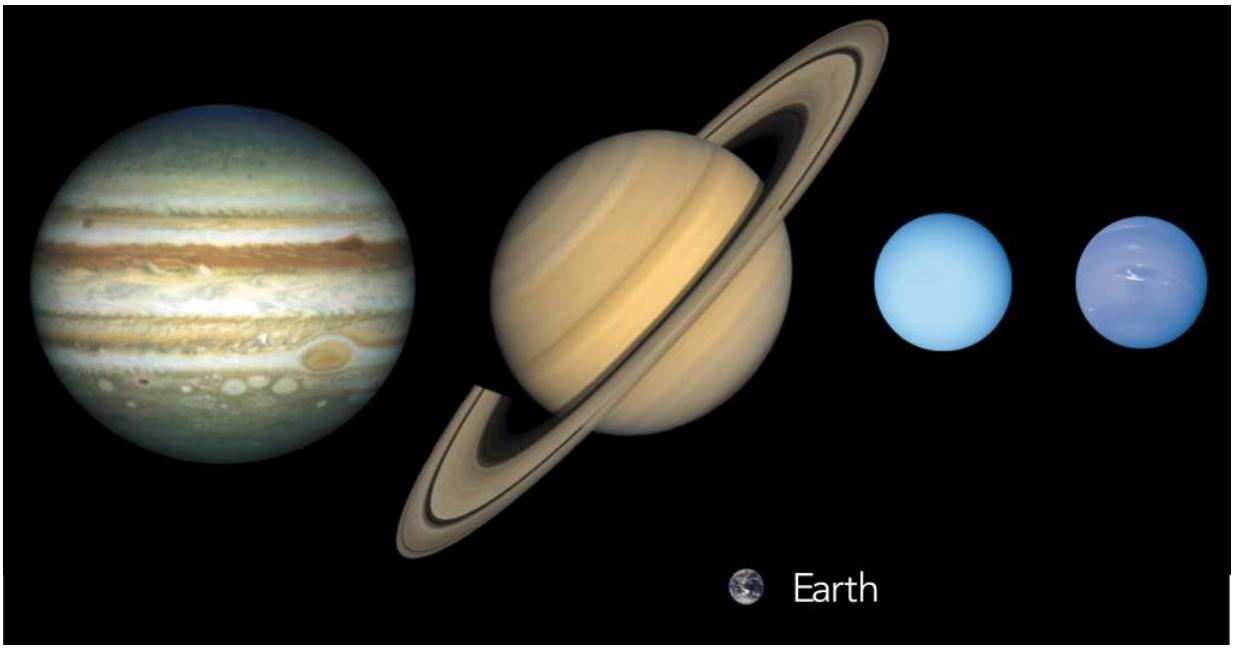


Figure 17.3.4 : The Four Giant Planets. This montage shows the four giant planets: Jupiter, Saturn, Uranus, and Neptune. Below them, Earth is shown to scale. Distance is not to scale. (credit: modification of work by NASA, Solar System Exploration)

Near the outer edge of the system lies Pluto, which was the first of the distant icy worlds to be discovered beyond Neptune (Pluto was visited by a spacecraft, the NASA New Horizons mission, in 2015 [see Figure 17.3.5]). Table 17.3.2 summarizes some of the main facts about the planets. Note that an AU (or **astronomical unit**) is the distance from Earth to the Sun.



Figure 17.3.5 : Pluto Close-up. This intriguing image from the New Horizons spacecraft, taken when it flew by the dwarf planet in July 2015, shows some of its complex surface features. The rounded white area is called the Sputnik Plain, after humanity's first spacecraft. (credit: modification of work by NASA/Johns Hopkins University Applied Physics Laboratory/Southwest Research Institute)

Table 17.3.2: The Planets

Name	Distance from Sun (AU)[2]	Revolution Period (y)	Diameter (km)	Mass (10^{23} kg)	Density (g/cm^3)
Mercury	0.39	0.24	4,878	3.3	5.4
Venus	0.72	0.62	12,120	48.7	5.2
Earth	1.00	1.00	12,756	59.8	5.5
Mars	1.52	1.88	6,787	6.4	3.9
Jupiter	5.20	11.86	142,984	18,991	1.3
Saturn	9.54	29.46	120,536	5686	0.7
Uranus	19.18	84.07	51,118	866	1.3
Neptune	30.06	164.82	49,660	1030	1.6

Smaller Members of the Solar System

Most of the planets are accompanied by one or more **moons**; only Mercury and Venus move through space alone. There are more than 210 known moons orbiting planets and dwarf planets, and undoubtedly many other small ones remain undiscovered. The largest of the moons are as big as small planets and just as interesting. In addition to our Moon, they include the four largest moons of Jupiter (called the Galilean moons, after their discoverer) and the largest moons of Saturn and Neptune (confusingly named Titan and Triton).

Each of the giant planets also has rings made up of countless small bodies ranging in size from mountains to mere grains of dust, all in orbit about the equator of the planet. The bright rings of Saturn are, by far, the easiest to see. They are among the most beautiful sights in the solar system (Figure 17.3.6). But, all four ring systems are interesting to scientists because of their complicated forms, influenced by the pull of the moons that also orbit these giant planets.

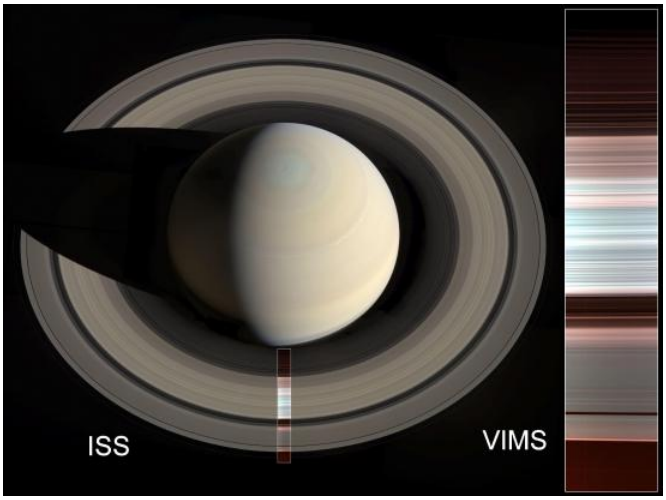


Figure 17.3.6 : Saturn and its A, B, and C rings in visible and (inset) infrared (IR) light. In the false-color IR view, greater water ice content and larger grain size lead to blue-green color, while greater non-ice content and smaller grain size yield a reddish hue. (credit: modification of work by NASA/JPL-Caltech/Space Science Institute/G. Ugarkovic (ISS), NASA/JPL-Caltech/University of Arizona/CNRS/LPG-Nantes (VIMS))

The solar system has many other less-conspicuous members. Another group is the **asteroids**, rocky bodies that orbit the Sun like miniature planets, mostly in the space between Mars and Jupiter (although some do cross the orbits of planets like Earth—see Figure 17.3.7). Most asteroids are remnants of the initial population of the solar system that existed before the planets themselves formed. Some of the smallest moons of the planets, such as the moons of Mars, are very likely captured asteroids.

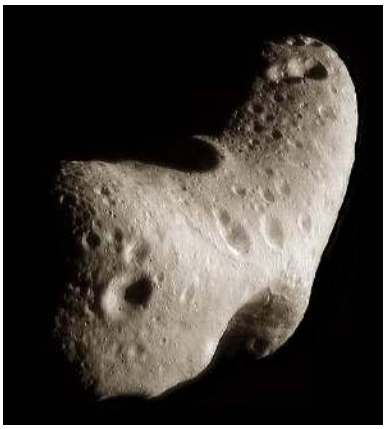


Figure 17.3.7 : Asteroid Eros. This small Earth-crossing asteroid image was taken by the NEAR-Shoemaker spacecraft from an altitude of about 100 kilometers. This view of the heavily cratered surface is about 10 kilometers wide. The spacecraft orbited Eros for a year before landing gently on its surface. (credit: modification of work by NASA/JPL/JHUAPL)

Another class of small bodies is composed mostly of ice, made of frozen gases such as water, carbon dioxide, and carbon monoxide; these objects are called **comets** (see Figure 17.3.8). Comets also are remnants from the formation of the solar system, but they were formed and continue (with rare exceptions) to orbit the Sun in distant, cooler regions—stored in a sort of cosmic deep freeze. This is also the realm of the larger icy worlds, called dwarf planets.

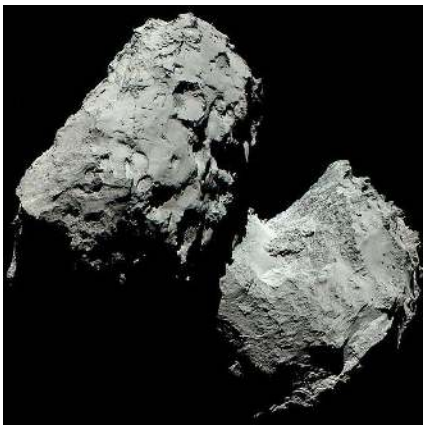


Figure 17.3.8 : Comet Churyumov-Gerasimenko (67P). This approximately true color image of the comet was taken by the Rosetta spacecraft on August 6, 2014, at a distance of 120 kilometers. There is surprisingly little color variation across the surface of the comet. (credit: modification of work by ESA/Rosetta/MPS for OSIRIS Team/UPD/LAM/IAA/SSO/INTA/UPM/DASP/IDA)

Finally, there are countless grains of broken rock, which we call cosmic dust, scattered throughout the solar system. When these particles enter Earth's atmosphere (as millions do each day) they burn up, producing a brief flash of light in the night sky known as a **meteor** (meteors are often referred to as shooting stars). Occasionally, some larger chunk of rocky or metallic material survives its passage through the atmosphere and lands on Earth. Any piece that strikes the ground is known as a **meteorite**. You can see meteorites on display in many natural history museums and can sometimes even purchase pieces of them from gem and mineral dealers.

A Scale Model of the Solar System

Astronomy often deals with dimensions and distances that far exceed our ordinary experience. What does 1.4 billion kilometers—the distance from the Sun to Saturn—really mean to anyone? It can be helpful to visualize such large systems in terms of a scale model.

In our imaginations, let us build a scale model of the solar system, adopting a scale factor of 1 billion (10^9)—that is, reducing the actual solar system by dividing every dimension by a factor of 10^9. Earth, then, has a diameter of 1.3 centimeters, about the size of a grape. The Moon is a pea orbiting this at a distance of 40 centimeters, or a little more than a foot away. The Earth-Moon system fits into a standard backpack.

In this model, the Sun is nearly 1.5 meters in diameter, about the average height of an adult, and our Earth is at a distance of 150 meters—about one city block—from the Sun. Jupiter is five blocks away from the Sun, and its diameter is 15 centimeters, about the size of a very large grapefruit. Saturn is 10 blocks from the Sun; Uranus, 20 blocks; and Neptune, 30 blocks. Pluto, with a distance that varies quite a bit during its 249-year orbit, is currently just beyond 30 blocks and getting farther with time. Most of the moons of the outer solar system are the sizes of various kinds of seeds orbiting the grapefruit, oranges, and lemons that represent the outer planets.

In our scale model, a human is reduced to the dimensions of a single atom, and cars and spacecraft to the size of molecules. Sending the Voyager spacecraft to Neptune involves navigating a single molecule from the Earth–grape toward a lemon 5 kilometers away with an accuracy equivalent to the width of a thread in a spider's web.

If that model represents the solar system, where would the nearest stars be? If we keep the same scale, the closest stars would be tens of thousands of kilometers away. If you built this scale model in the city where you live, you would have to place the representations of these stars on the other side of Earth or beyond.

By the way, model solar systems like the one we just presented have been built in cities throughout the world. In Sweden, for example, Stockholm's huge Globe Arena has become a model for the Sun, and Pluto is represented by a 12-centimeter sculpture in the small town of Delsbo, 300 kilometers away. Another model solar system is in Washington, D.C. on the Mall between the White House and Congress (perhaps proving they are worlds apart?).

Footnotes

- 1 The generic term for a group of planets and other bodies circling a star is *planetary system*. Ours is called the *solar system* because our Sun is sometimes called *Sol*. Strictly speaking, then, there is only one solar system; planets orbiting other stars are in planetary systems.
- 2 An AU (or astronomical unit) is the distance from Earth to the Sun.
- 3 We give densities in units where the density of water is 1 g/cm^3. To get densities in units of kg/m^3, multiply the given value by 1000.

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17.4: Composition and Structure of Planets

The fact that there are two distinct kinds of planets—the rocky terrestrial planets and the gas-rich Jovian planets—leads us to believe that they formed under different conditions. Certainly their compositions are dominated by different elements. Let us look at each type in more detail.

The Giant Planets

The two largest planets, Jupiter and Saturn, have nearly the same chemical makeup as the Sun; they are composed primarily of the two **elements** hydrogen and helium, with 75% of their mass being hydrogen and 25% helium. On Earth, both hydrogen and helium are gases, so Jupiter and Saturn are sometimes called gas planets. But, this name is misleading. Jupiter and Saturn are so large that the gas is compressed in their interior until the hydrogen becomes a liquid. Because the bulk of both planets consists of compressed, liquefied hydrogen, we should really call them liquid planets.

Under the force of gravity, the heavier elements sink toward the inner parts of a liquid or gaseous planet. Both Jupiter and Saturn, therefore, have cores composed of heavier rock, metal, and ice, but we cannot see these regions directly. In fact, when we look down from above, all we see is the atmosphere with its swirling clouds (Figure 17.4.1). We must infer the existence of the denser core inside these planets from studies of each planet's gravity.

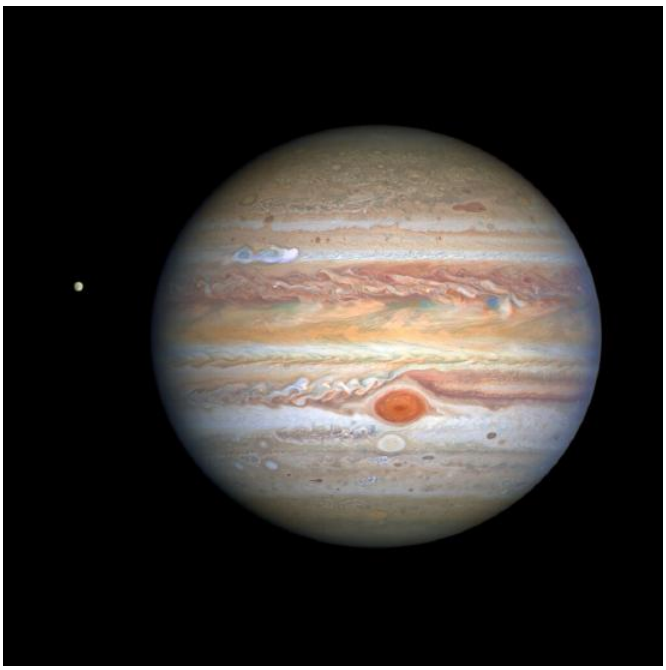


Figure 17.4.1 : Jupiter with its moon Europa on the left. Earth's diameter is 11 times smaller than Jupiter, and 4 times larger than Europa. (credit: NASA/ESA/STScI/A. Simon (Goddard Space Flight Center)/MH Wong (University of California, Berkeley)/OPAL team)

Uranus and Neptune are much smaller than Jupiter and Saturn, but each also has a core of rock, metal, and ice. Uranus and Neptune were less efficient at attracting hydrogen and helium gas, so they have much smaller atmospheres in proportion to their cores.

Chemically, each giant planet is dominated by hydrogen and its many compounds. Nearly all the oxygen present is combined chemically with hydrogen to form water (H_2O). Chemists call such a hydrogen-dominated composition ***reduced***. Throughout the outer solar system, we find abundant water (mostly in the form of ice) and reducing chemistry.

The Terrestrial Planets

The terrestrial planets are quite different from the giants. In addition to being much smaller, they are composed primarily of rocks and metals. These, in turn, are made of elements that are less common in the universe as a whole. The most abundant rocks, called **silicates**, are made of silicon and oxygen, and the most common metal is iron. We can tell from their densities that Mercury has the greatest proportion of metals (which are denser) and the Moon has the lowest. Earth, Venus, and Mars all have roughly similar bulk

compositions: about one third of their mass consists of iron-nickel or iron-sulfur combinations; two thirds is made of silicates. Because these planets are largely composed of oxygen compounds (such as the silicate minerals of their crusts), their chemistry is said to be ***oxidized***.

When we look at the internal structure of each of the terrestrial planets, we find that the densest metals are in a central core, with the lighter silicates near the surface. If these planets were liquid, like the giant planets, we could understand this effect as the result the sinking of heavier elements due to the pull of gravity. This leads us to conclude that, although the terrestrial planets are solid today, at one time they must have been hot enough to melt.

Differentiation is the process by which gravity helps separate a planet's interior into layers of different compositions and densities. The heavier metals sink to form a core, while the lightest minerals float to the surface to form a crust. Later, when the planet cools, this layered structure is preserved. In order for a rocky planet to differentiate, it must be heated to the melting point of rocks, which is typically more than 1300 K.

Moons, Asteroids, and Comets

Chemically and structurally, Earth's Moon is like the terrestrial planets, but most moons are in the outer solar system, and they have compositions similar to the cores of the giant planets around which they orbit. The three largest moons—Ganymede and Callisto in the Jovian system, and Titan in the Saturnian system—are composed half of frozen water, and half of rocks and metals. Most of these moons differentiated during formation, and today they have cores of rock and metal, with upper layers and crusts of very cold and—thus very hard—ice (Figure 17.4.2).



Figure 17.4.2 : Ganymede. This view of Jupiter's moon Ganymede was taken in June 1996 by the Galileo spacecraft. The brownish gray color of the surface indicates a dusty mixture of rocky material and ice. The bright spots are places where recent impacts have uncovered fresh ice from underneath. (credit: modification of work by NASA/JPL)

Most of the asteroids and comets, as well as the smallest moons, were probably never heated to the melting point. However, some of the largest asteroids, such as Vesta, appear to be differentiated; others are fragments from differentiated bodies. Many of the smaller objects seem to be fragments or rubble piles that are the result of collisions. Because most asteroids and comets retain their original composition, they represent relatively unmodified material dating back to the time of the formation of the solar system. In a sense, they act as chemical fossils, helping us to learn about a time long ago whose traces have been erased on larger worlds.

Temperatures: Going to Extremes

Generally speaking, the farther a planet or moon is from the Sun, the cooler its surface. The planets are heated by the radiant energy of the Sun, which gets weaker with the square of the distance. You know how rapidly the heating effect of a fireplace or an outdoor radiant heater diminishes as you walk away from it; the same effect applies to the Sun. Mercury, the closest planet to the Sun, has a blistering surface temperature that ranges from 280–430 °C on its sunlit side, whereas the surface temperature on Pluto is only about –220 °C, colder than liquid air.

Mathematically, the temperatures decrease approximately in proportion to the square root of the distance from the Sun. Pluto is about 30 AU at its closest to the Sun (or 100 times the distance of Mercury) and about 49 AU at its farthest from the Sun. Thus, Pluto's temperature is less than that of Mercury by the square root of 100, or a factor of 10: from 500 K to 50 K.

In addition to its distance from the Sun, the surface temperature of a planet can be influenced strongly by its atmosphere. Without our atmospheric insulation (the greenhouse effect, which keeps the heat in), the oceans of Earth would be permanently frozen. Conversely, if Mars once had a larger atmosphere in the past, it could have supported a more temperate climate than it has today. Venus is an even more extreme example, where its thick atmosphere of carbon dioxide acts as insulation, reducing the escape of heat built up at the surface, resulting in temperatures greater than those on Mercury. Today, Earth is the only planet where surface temperatures generally lie between the freezing and boiling points of water. As far as we know, Earth is the only planet to support life.

Dating Planetary Surfaces

How do we know the age of the surfaces we see on planets and moons? If a world has a surface (as opposed to being mostly gas and liquid), astronomers have developed some techniques for estimating how long ago that surface solidified. Note that the age of these surfaces is not necessarily the age of the planet as a whole. On geologically active objects (including Earth), vast outpourings of molten rock or the erosive effects of water and ice, which we call planet **weathering**, have erased evidence of earlier epochs and present us with only a relatively young surface for investigation.

One way to estimate the age of a surface is by counting the number of impact craters. This technique works because the rate at which impacts have occurred in the solar system has been roughly constant for several billion years. Thus, in the absence of forces to eliminate craters, the number of craters is simply proportional to the length of time the surface has been exposed. This technique has been applied successfully to many solid planets and moons (Figure 17.4.3).

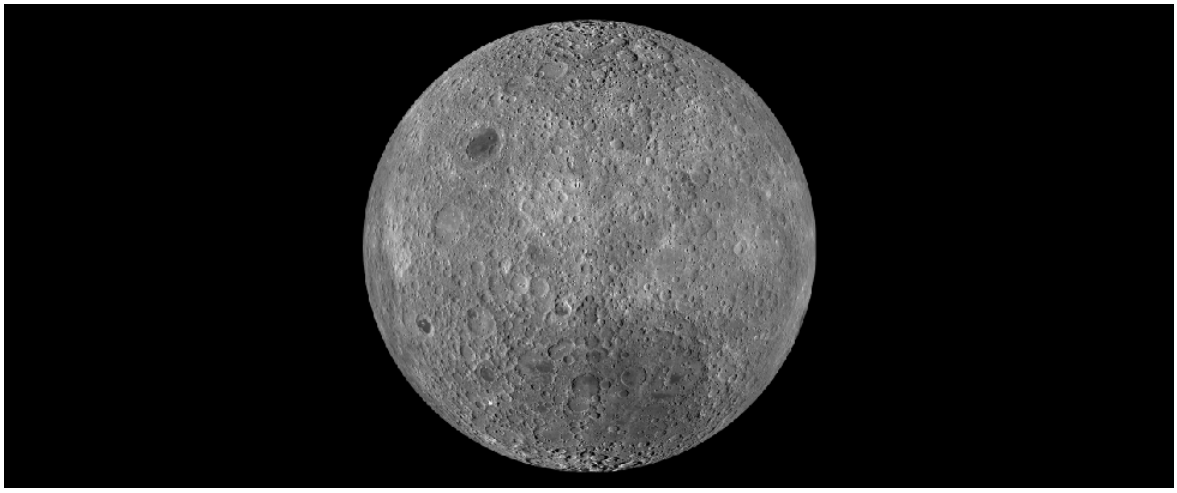


Figure 17.4.3 : Our Cratered Moon. This composite image of the Moon's surface was made from many smaller images taken between November 2009 and February 2011 by the Lunar Reconnaissance Orbiter (LRO) and shows craters of many different sizes. (credit: modification of work by NASA/GSFC/Arizona State University)

Bear in mind that crater counts can tell us only the time since the surface experienced a major change that could modify or erase preexisting craters. Estimating ages from crater counts is a little like walking along a sidewalk in a snowstorm after the snow has been falling steadily for a day or more. You may notice that in front of one house the snow is deep, while next door the sidewalk may be almost clear. Do you conclude that less snow has fallen in front of Ms. Jones' house than Mr. Smith's? More likely, you conclude that Jones has recently swept the walk clean and Smith has not. Similarly, the numbers of craters indicate how long it has been since a planetary surface was last "swept clean" by ongoing lava flows or by molten materials ejected when a large impact happened nearby.

Still, astronomers can use the numbers of craters on different parts of the same world to provide important clues about how regions on that world evolved. On a given planet or moon, the more heavily cratered terrain will generally be older (that is, more time will have elapsed there since something swept the region clean).

Radioactive Rocks

Another way to trace the history of a solid world is to measure the age of individual rocks. After samples were brought back from the Moon by Apollo astronauts, the techniques that had been developed to date rocks on Earth were applied to rock samples from

the Moon to establish a geological chronology for the Moon. Furthermore, a few samples of material from the Moon, Mars, and the large asteroid Vesta have fallen to Earth as meteorites and can be examined directly.

Scientists measure the age of rocks using the properties of natural **radioactivity**. Around the beginning of the twentieth century, physicists began to understand that some atomic nuclei are not stable but can split apart (decay) spontaneously into smaller nuclei. The process of radioactive decay involves the emission of particles such as electrons, or of radiation in the form of gamma rays.

For any one radioactive nucleus, it is not possible to predict when the decay process will happen. Such decay is random in nature, like the throw of dice: as gamblers have found all too often, it is impossible to say just when the dice will come up 7 or 11. But, for a very large number of dice tosses, we can calculate the odds that 7 or 11 will come up. Similarly, if we have a very large number of radioactive atoms of one type (say, uranium), there is a specific time period, called its **half-life**, during which the chances are fifty-fifty that decay will occur for any of the nuclei.

A particular nucleus may last a shorter or longer time than its half-life, but in a large sample, almost exactly half of the nuclei will have decayed after a time equal to one half-life. Half of the remaining nuclei will have decayed after two half-lives pass, leaving only one half of a half—or one quarter—of the original sample (Figure 17.4.4).

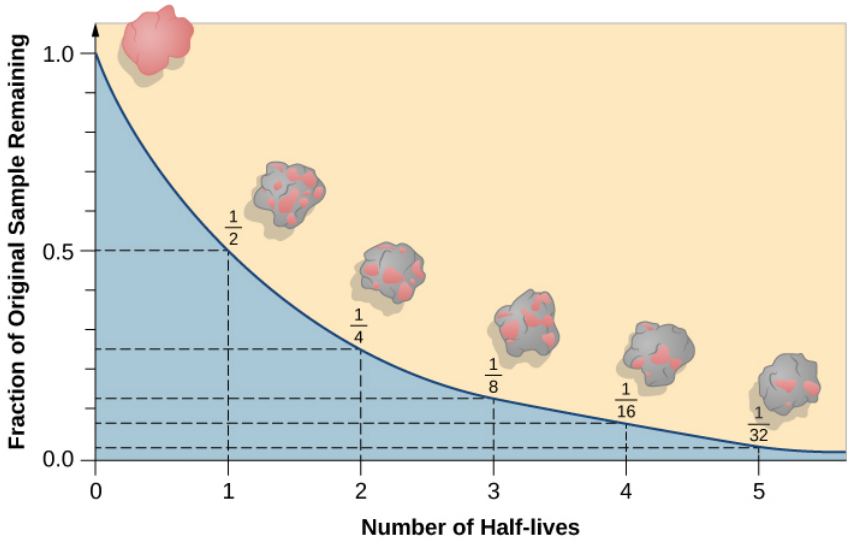


Figure 17.4.4 : Radioactive Decay. This graph shows (in pink) the amount of a radioactive sample that remains after several half-lives have passed. After one half-life, half the sample is left; after two half-lives, one half of the remainder (or one quarter) is left; and after three half-lives, one half of that (or one eighth) is left. Note that, in reality, the decay of radioactive elements in a rock sample would not cause any visible change in the appearance of the rock; the splashes of color are shown here for conceptual purposes only.

If you had 1 gram of pure radioactive nuclei with a half-life of 100 years, then after 100 years you would have 1/2 gram; after 200 years, 1/4 gram; after 300 years, only 1/8 gram; and so forth. However, the material does not disappear. Instead, the radioactive atoms are replaced with their decay products. Sometimes the radioactive atoms are called ***parents*** and the decay products are called ***daughter*** elements.

In this way, radioactive elements with half-lives we have determined can provide accurate nuclear clocks. By comparing how much of a radioactive parent element is left in a rock to how much of its daughter products have accumulated, we can learn how long the decay process has been going on and hence how long ago the rock formed. Table 17.4.1 summarizes the decay reactions used most often to date lunar and terrestrial rocks.

Table 17.4.1: Radioactive Decay Reaction Used to Date Rocks[4]

Parent	Daughter	Half-Life (billions of years)
Samarium-147	Neodymium-143	106
Rubidium-87	Strontium-87	48.8
Thorium-232	Lead-208	14.0
Uranium-238	Lead-206	4.47
Potassium-40	Argon-40	1.31

When astronauts first flew to the Moon, one of their most important tasks was to bring back lunar rocks for radioactive age-dating. Until then, astronomers and geologists had no reliable way to measure the age of the lunar surface. Counting craters had let us calculate relative ages (for example, the heavily cratered lunar highlands were older than the dark lava plains), but scientists could not measure the actual age in years. Some thought that the ages were as young as those of Earth's surface, which has been resurfaced by many geological events. For the Moon's surface to be so young would imply active geology on our satellite. Only in 1969, when the first Apollo samples were dated, did we learn that the Moon is an ancient, geologically dead world. Using such dating techniques, we have been able to determine the ages of both Earth and the Moon: each was formed about 4.5 billion years ago (although, as we shall see, Earth probably formed earlier than the Moon).

We should also note that the decay of radioactive nuclei generally releases energy in the form of heat. Although the energy from a single nucleus is not very large (in human terms), the enormous numbers of radioactive nuclei in a planet or moon (especially early in its existence) can be a significant source of internal energy for that world. Geologists estimate that about half of Earth's current internal heat budget comes from the decay of radioactive isotopes in its interior.

Footnotes

- 4 The number after each element is its atomic weight, equal to the number of protons plus neutrons in its nucleus. This specifies the *isotope* of the element; different isotopes of the same element differ in the number of neutrons.

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17.S: Summary

Origin of the Universe

The universe is an incredibly vast and complex expanse, containing an infinite number of galaxies and solar systems, with our solar system occupying just a small part of it. The origins of the universe, often explained by the big bang theory, provide the context for understanding the Earth's formation and early history. The big bang theory, supported by Einstein's theory of general relativity and various empirical observations, suggests that the universe began from an infinitely dense and hot core, followed by an explosive expansion that has been ongoing for 13.8 billion years. Key evidence supporting this theory includes the redshift of galaxies, cosmic microwave background radiation, and the process of stellar evolution, which explains the formation of elements and the life cycles of stars. The universe's expansion and the death of stars lead to the creation of celestial phenomena such as white dwarfs, neutron stars, and black holes, highlighting the dynamic and ever-changing nature of the cosmos.

Origin of the Solar System--The Nebular Hypothesis

The formation of our solar system is explained by the nebular hypothesis, which suggests that a spinning cloud of dust, or nebula, composed mostly of light elements, flattened into a protoplanetary disk, eventually forming a star with orbiting planets. As this nebula condensed, the majority of its material accumulated at the center, creating the Sun, which accounts for over 99% of the solar system's mass. Temperature variations within the nebula led to the differentiation of planets: the inner, rocky planets formed in hotter regions, while the outer, gas giants formed in cooler areas. The planets grew as particles collided and merged, with gas giants like Jupiter and Saturn accumulating more hydrogen and helium, while ice giants like Uranus and Neptune developed from a mix of gases and ices. Jupiter's immense gravity further influenced the solar system's structure, contributing to the formation of the asteroid belt and impacting the development of nearby planets. The outermost regions of the solar system, including the Kuiper belt and Oort cloud, contain icy bodies and dwarf planets like Pluto, which was reclassified in 2006 due to new criteria defining what constitutes a planet.

Overview of Our Planetary System

Our solar system currently consists of the Sun, eight planets, five dwarf planets, nearly 200 known moons, and a host of smaller objects. The planets can be divided into two groups: the inner terrestrial planets and the outer giant planets. Pluto, Eris, Haumea, and Makemake do not fit into either category; as icy dwarf planets, they exist in an ice realm on the fringes of the main planetary system. The giant planets are composed mostly of liquids and gases. Smaller members of the solar system include asteroids (including the dwarf planet Ceres), which are rocky and metallic objects found mostly between Mars and Jupiter; comets, which are made mostly of frozen gases and generally orbit far from the Sun; and countless smaller grains of cosmic dust. When a meteor survives its passage through our atmosphere and falls to Earth, we call it a meteorite.

Composition and Structure of Planets

The giant planets have dense cores roughly 10 times the mass of Earth, surrounded by layers of hydrogen and helium. The terrestrial planets consist mostly of rocks and metals. They were once molten, which allowed their structures to differentiate (that is, their denser materials sank to the center). The Moon resembles the terrestrial planets in composition, but most of the other moons—which orbit the giant planets—have larger quantities of frozen ice within them. In general, worlds closer to the Sun have higher surface temperatures. The surfaces of terrestrial planets have been modified by impacts from space and by varying degrees of geological activity.

The ages of the surfaces of objects in the solar system can be estimated by counting craters: on a given world, a more heavily cratered region will generally be older than one that is less cratered. We can also use samples of rocks with radioactive elements in them to obtain the time since the layer in which the rock formed last solidified. The half-life of a radioactive element is the time it takes for half the sample to decay; we determine how many half-lives have passed by how much of a sample remains the radioactive element and how much has become the decay product. In this way, we have estimated the age of the Moon and Earth to be roughly 4.5 billion years.

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