AN INTRODUCTION TO GEOLOGY



Kelly Ruppert
Coastline College

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An Introduction To Geology

Kelly Ruppert

Coastline College

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CHAPTER OVERVIEW

1: Understanding Science

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, geology can teach people how to develop credible conclusions, as well as identify and stop misinformation.

1.1: What is Science?

1.2: The Scientific Method

1.3: The Study of Geology

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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: This is Grand Canyon of the 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." An inference would be "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 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

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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 1,000 years ago. At the forefront of the scientific method are conclusions based on objective evidence, not opinion or hearsay [4].

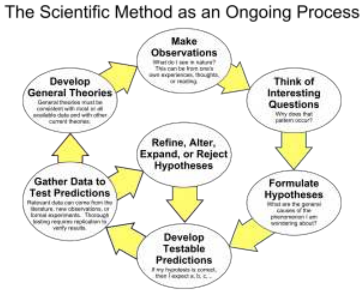


Figure 1.2.1: Diagram of the cyclical nature of the scientific method.

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. These 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 since 1927. A petroleum product called pitch, which is highly viscous, drips out of a funnel about once per decade.

Step 4: Peer Review, Publication, and Publication

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: 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 1.3.1: 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 1.3.2: 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 1.3.3: 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 1.3.4: 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 1.3.5: 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.

References

- none

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CHAPTER OVERVIEW

2: Plate Tectonics

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.

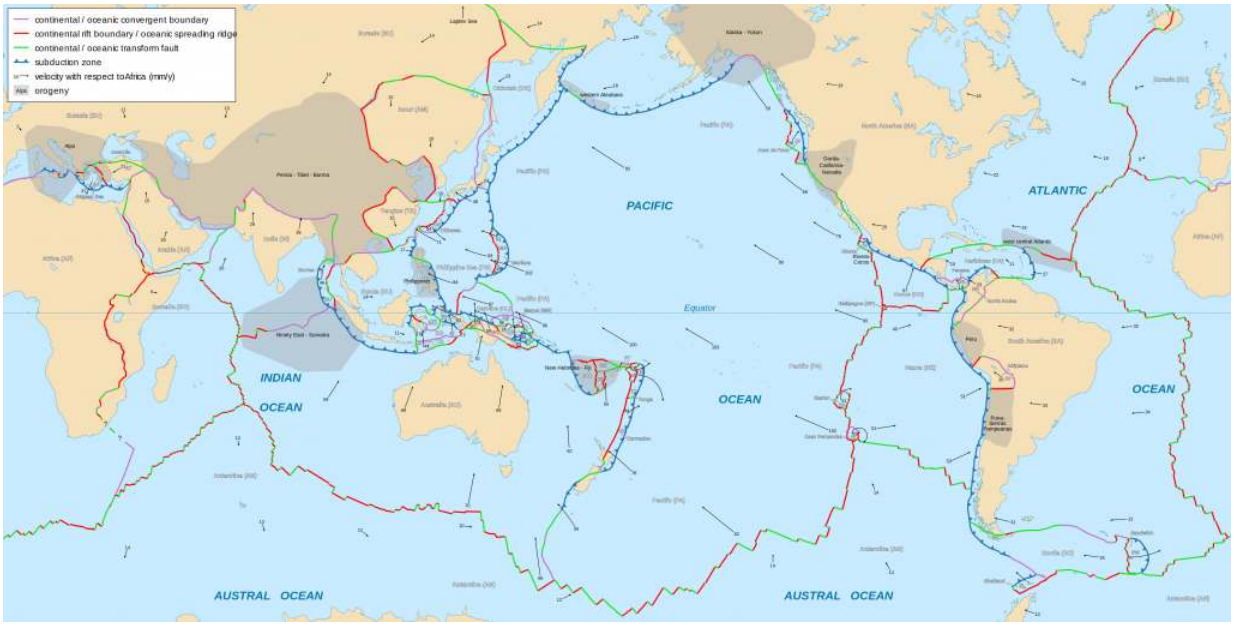


Figure 2.1: Detailed map of all known plates, their boundaries, and movements.

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 Boudaries
2.6: Hotspots

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.

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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 that connected Asia and North America across the Pacific Ocean 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.

Early Evidence for Continental Drift Hypothesis

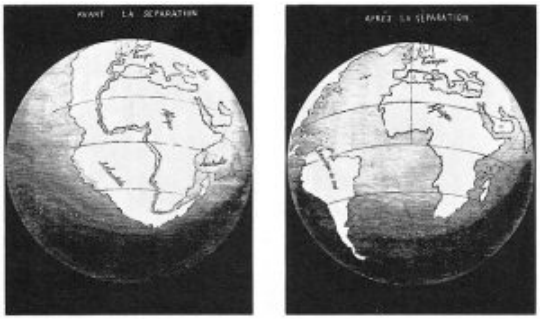


Figure 2.1.2: Snider-Pellegrini's map showing the continental fit and separation, 1858.

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].

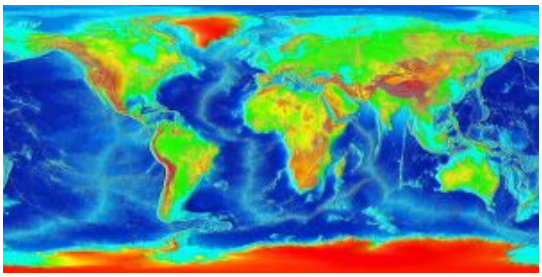


Figure 2.1.3: Map of world elevations. Note the light blue, which are continental shelves flooded by shallow ocean water. These show the true shapes of the continents.

What Wegener did differently was synthesizing 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].

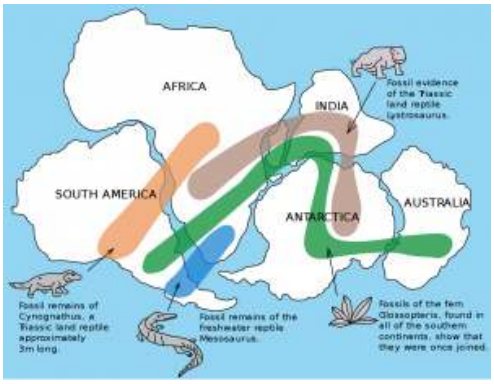


Figure 2.1.4: Image showing fossils that connect the continents of Gondwana (the southern continents of Pangea).

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 were land-dwelling creatures could not have swum across an entire ocean.

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.

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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. In general, the Earth can be divided into layers based on chemical composition and physical characteristics.

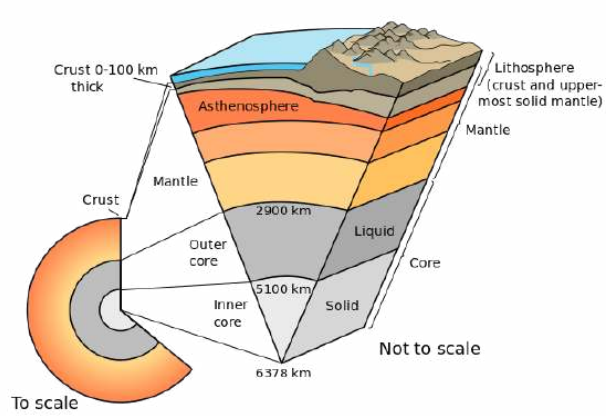


Figure 2.2.1: The layers of the Earth. Physical layers include the lithosphere and asthenosphere; chemical layers are crust, mantle, and core.

Chemical Composition 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 layers have distinct chemical compositions.

Crust

The outermost chemical layer and the one we walk on is the crust. Its composition is such that it has the highest amounts of silicon and oxygen of any of the layers. Since there is so much silicon and oxygen there is not much room for other element, most importantly iron or magnesium. There are two types of crust. Continental crust is thick has a relatively low density (is lighter). Oceanic crust is thin and has a relatively high density (is heavier), especially when cold and old. The surface levels of crust are relatively brittle which means they behave like solid glass or pottery and break under force. The deeper parts of the crust are subjected to higher temperatures and pressure, which makes them more ductile which means they behave like soft plastics or putty, and move under force.

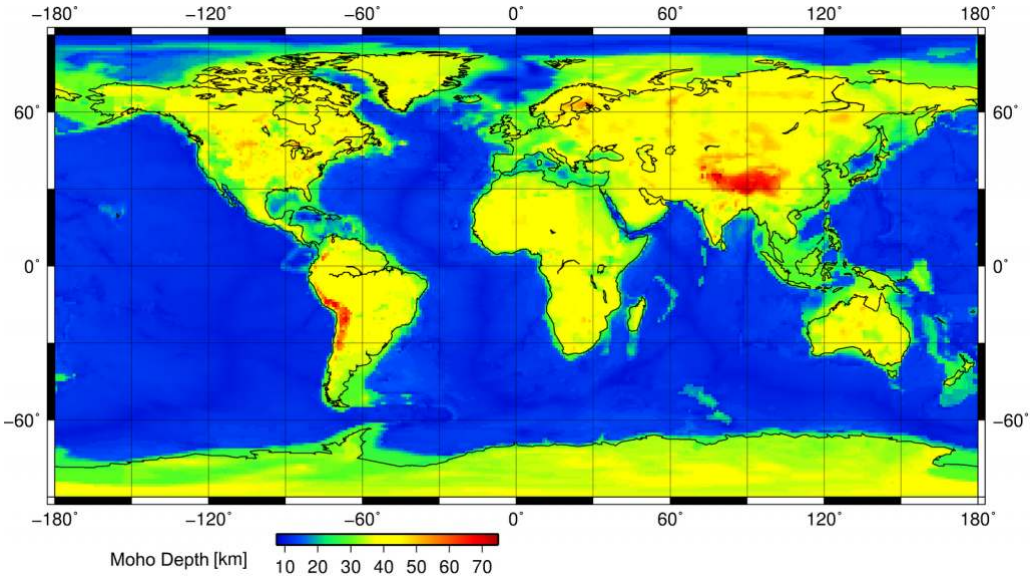


Figure 2.2.2: The global map of the depth of the moho.

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The base of the crust is characterized by a large increase in seismic velocity, which measures 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) 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 next layer (the mantle) and showed that the earth was not the same substance all the way through!

Mantle



Figure 2.2.3: This mantle rock containing olivine (green) contains a lot of iron.

The mantle sits below the crust. 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 very iron-rich igneous rock we will learn about in Chapter 4. Because of this, scientists hypothesize most of the mantle is made of peridotite which is much lower in silicon and oxygen, and much higher in iron and magnesium [30].

Core



Figure 2.2.4: A polished fragment of the iron-rich Toluca Meteorite, with octahedral Widmanstätten Pattern.

Below the mantle, at the center of the earth is the core. The core, has both liquid and solid layers, and consists mostly mostly metallic iron [33]. Scientists looking at seismic data first discovered this innermost chemical layer in 1906 [32].

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

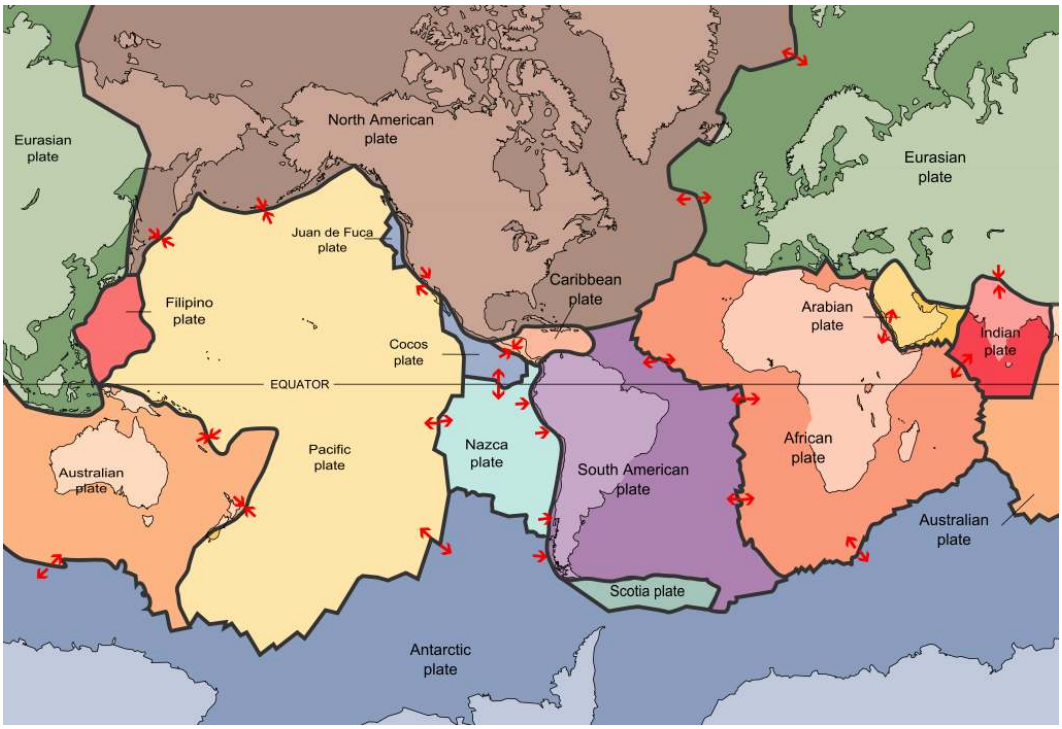


Figure 2.2.5: Map of the major plates and their motions along boundaries.

Lithos is Greek for stone, and the lithosphere is the outermost physical layer of the Earth made up of the crust and solid upper mantle. It is grouped into two types: oceanic and continental. Oceanic lithosphere is thin and relatively rigid. Continental lithosphere is generally thicker and considerably more plastic, especially at the deeper levels. 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—mountain building, triggering earthquakes, and generating volcanic activity. Some plates are made up of only oceanic lithosphere but most have a combination of both types.

Asthenosphere

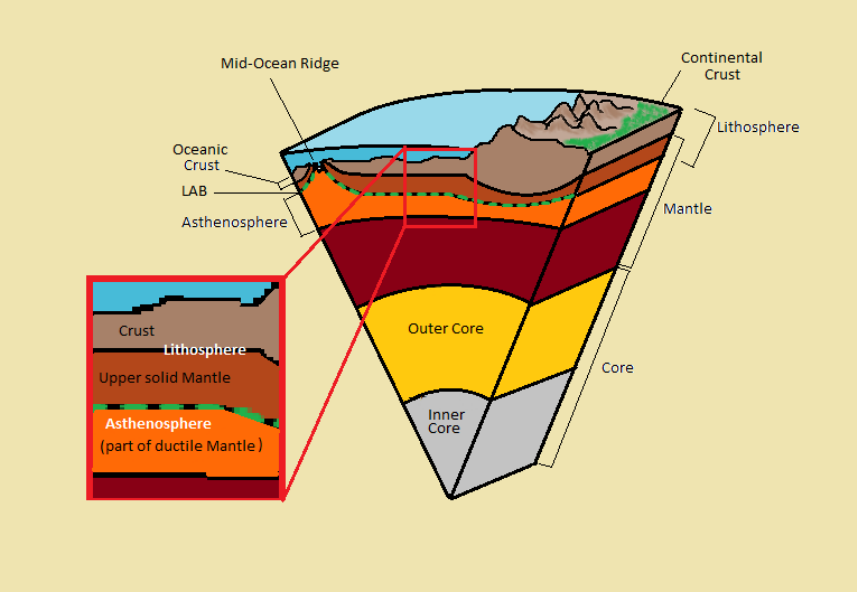


Figure 2.2.5: The lithosphere-asthenosphere boundary changes with certain tectonic situations.

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].

Mesosphere

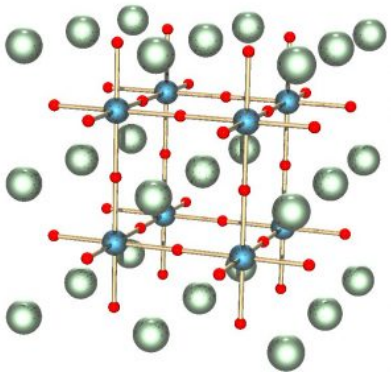


Figure 2.2.6: General perovskite structure. Perovskite silicates (e.g. Bridgmenite, $(Mg, Fe)SiO_3$) are thought to be the main component of the lower mantle, making it the most common mineral in or on Earth.

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.

Inner and Outer Core



Figure 2.2.7: Lehmann in 1932

The outer core is the only entirely liquid layer within the Earth. 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 1,220 km thick, and the outer core is about 2,300 km thick [40].

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].

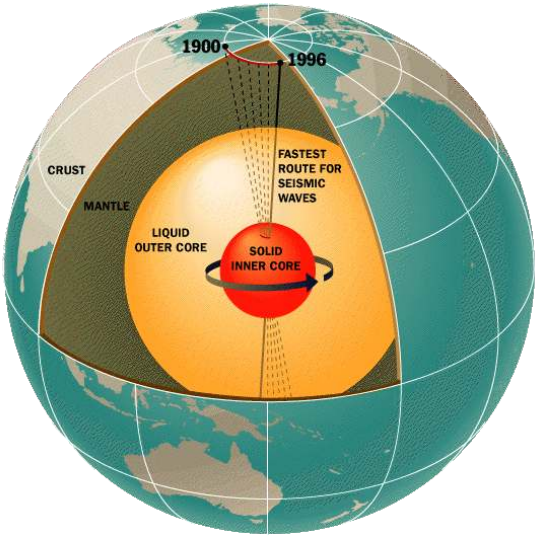


Figure 2.2.8: The outer core's spin most likely causes our protective magnetic field.

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.

Plate Tectonic Boundaries

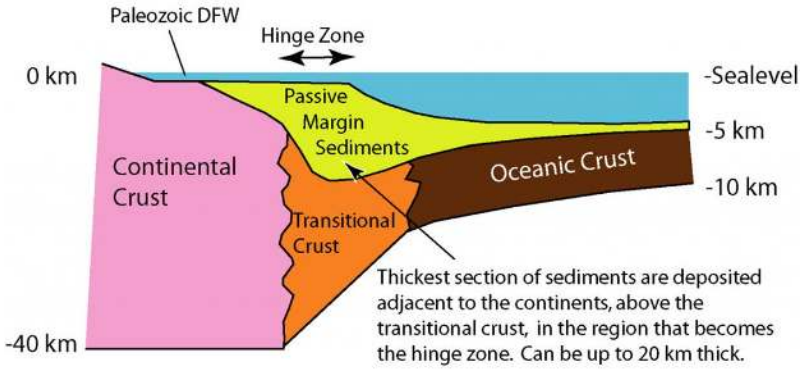


Figure 2.2.9: Passive Margin

At passive margins, there is no plate boundary—the continental lithosphere transitions into the oceanic lithosphere and forms plates made of both types. 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. 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, where two plates meet.

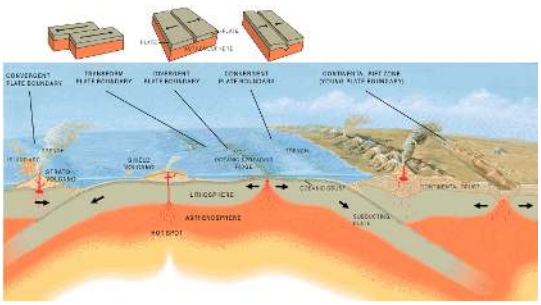


Figure 2.2.10: Schematic of plate boundary types.

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.

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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.

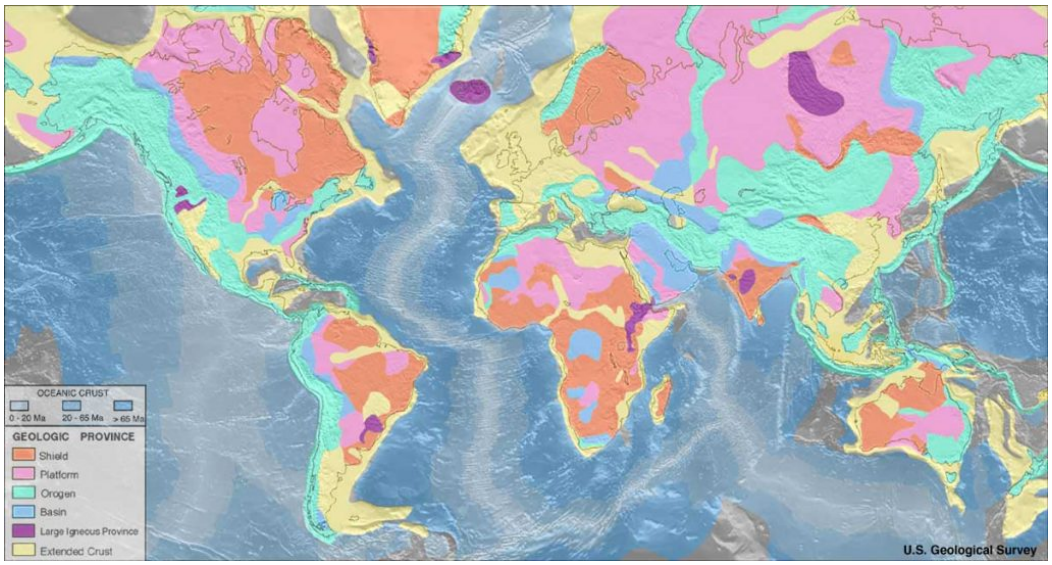


Figure 2.3.1: Geologic provinces of Earth. Orogenies are labeled light blue.

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, 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 three to four km. The Mariana Trench, for example, approaches a staggering 11 km [47].

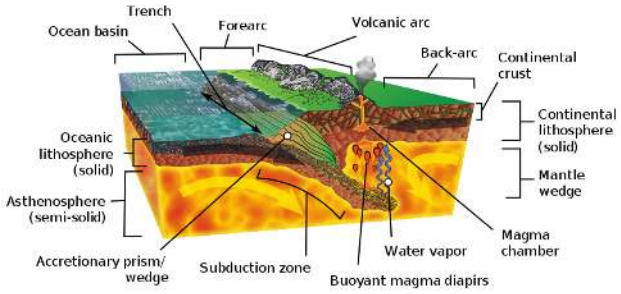


Figure 2.3.2: Diagram of ocean-continent subduction.

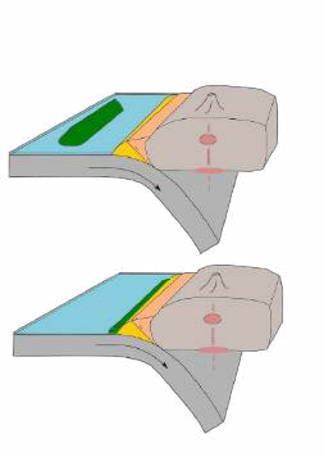


Figure 2.3.3: Microcontinents can become part of the accretionary prism of a subduction zone.

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].

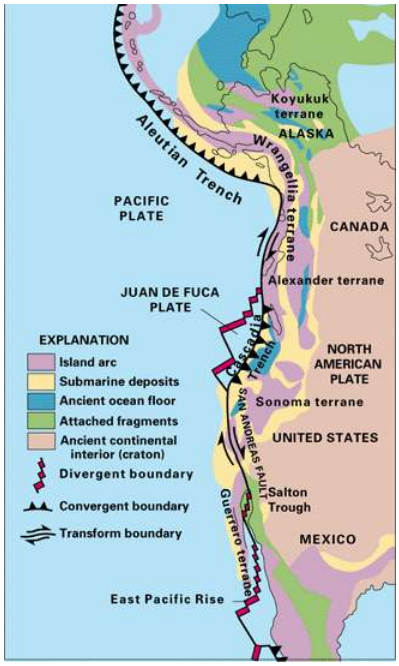


Figure 2.3.4: Accreted terranes of western North America. Everything that is not the “Ancient continental interior (craton)” has been smeared onto the side of the continent by accretion from subduction.

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. Through a process called fluid-induced melting covered in Chapter 4, magma is produced. The magma is more buoyant than the lithospheric plate above it and migrates to the Earth’s surface where it erupts as lava. The resulting volcanoes frequently appear as curved mountain chains, volcanic arcs, due to the curvature of the earth.

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.

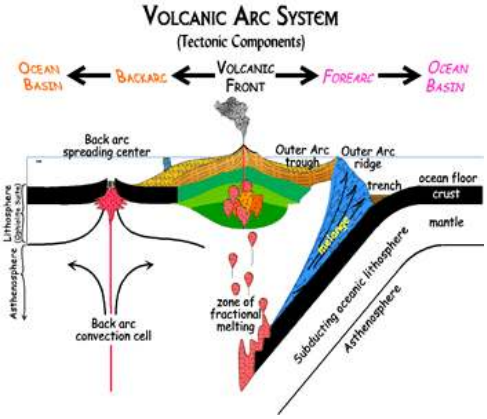


Figure 2.3.5: Various parts of a subduction zone. This subduction zone is ocean-ocean subduction, though the same features can apply to continent-ocean subduction.

Oceanic-Continental Subduction

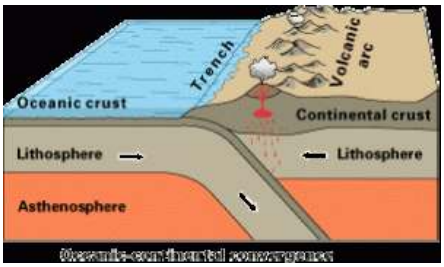


Figure 2.3.6: Subduction of an oceanic plate beneath a continental plate, forming a trench and volcanic arc.

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].

Oceanic-Oceanic Subduction

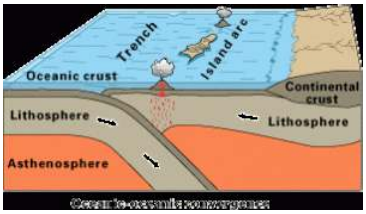


Figure 2.3.7: Subduction of an oceanic plate beneath another oceanic plate, forming a trench and an island arc.

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].

Collisions

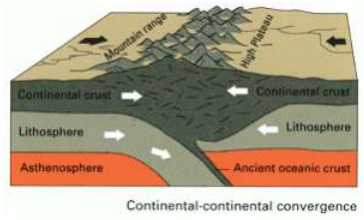


Figure 2.3.8: Two continental plates colliding.

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.

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.



Figure 2.3.9: A reconstruction of the supercontinent Pangaea, showing approximate positions of modern continents.

The process of collision created Pangea, the supercontinent envisioned by Wegener. 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 an 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.

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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.

Continental Rifting

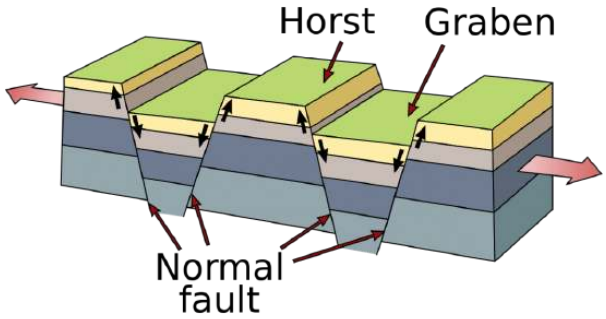


Figure 2.4.1: Faulting that occurs in divergent boundaries.

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 normal faults in the diagram above. 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 2.4.2: 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.

While seemingly occurring at random, rifting is dictated by two factors. 1) Rifting does not occur in continents with older and more stable interiors, known as cratons. 2) When continental rifting does occur, the break-up pattern resembles the seams of a soccer ball (in three directions from a central point). 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. 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 2.4.3: NASA image of the Basin and Range horsts and grabens across central Nevada.

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.

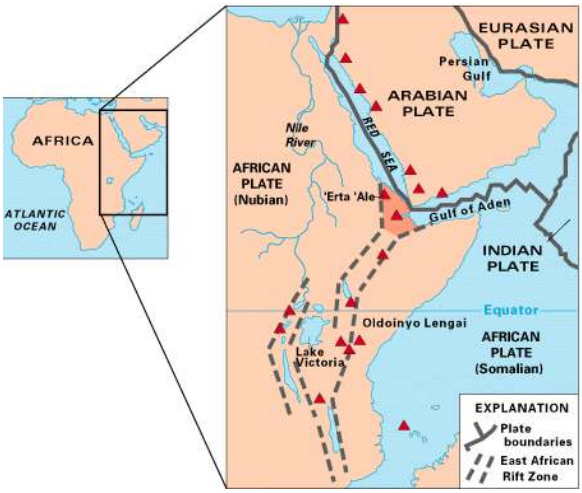


Figure 2.4.4: The narrow East African Rift.

Rifts have earthquakes, although not of the magnitude and frequency of other boundaries. They may also exhibit volcanism.

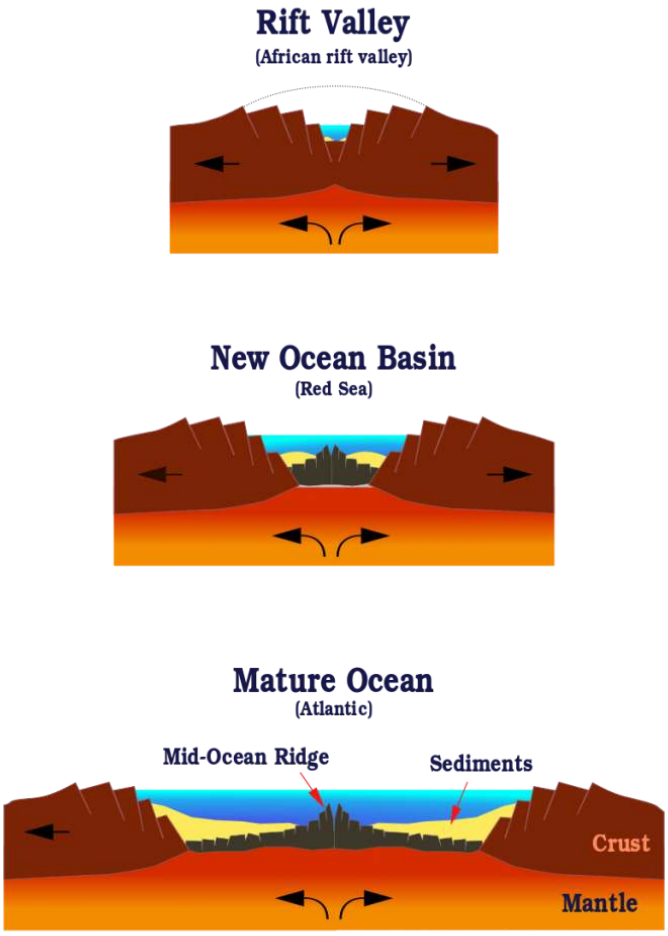


Figure 2.4.5: Progression from rift to mid-ocean ridge.

Mid-ocean ridges, also known as spreading centers, have several distinctive features. They are the only places on earth that create new oceanic lithosphere. Melting in the rift zone changes asthenosphere material into the new lithosphere, 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.

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

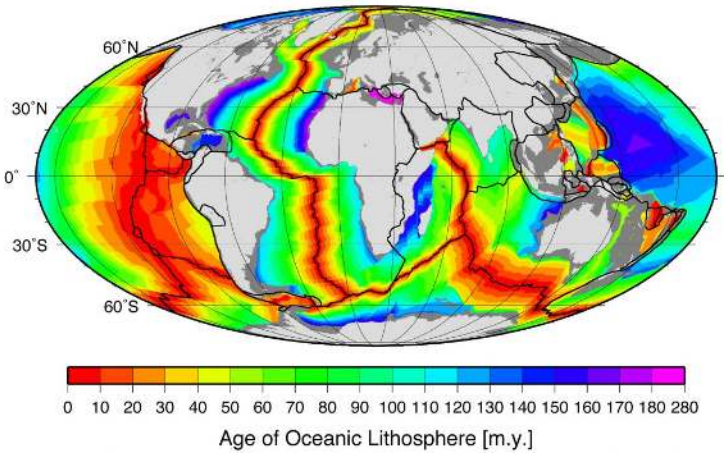


Figure 2.4.6: Age of oceanic lithosphere, in millions of years. Notice the differences in the Atlantic Ocean along the coasts of the continents.

As the ridge continues to spread, the lithosphere material is pulled away from the area of volcanism and becomes colder and denser. This means the sections of the lithosphere furthest away from the mid-ocean ridges will be the oldest. Sediment accumulation patterns also confirm the idea of the ocean being younger near the mid-ocean ridges as sediment layers tend to be thinner near them, indicating it has had less time to build up.

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2.5: Transform Boudaries

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, also known as right-lateral, movement describes the opposing plate moving to the right. Sinistral, also known as left lateral, movement describes the opposing plate moving to the left.

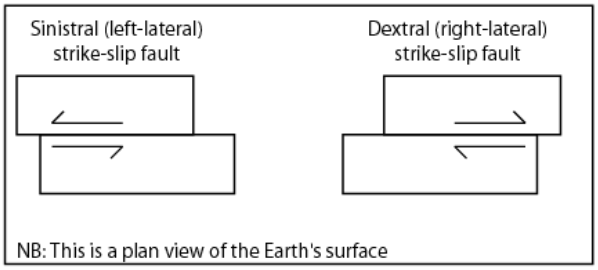


Figure 2.5.1: The two types of transform/strike-slip faults.

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



Figure 2.5.2: Map of the San Andreas fault, showing relative motion.

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.

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].

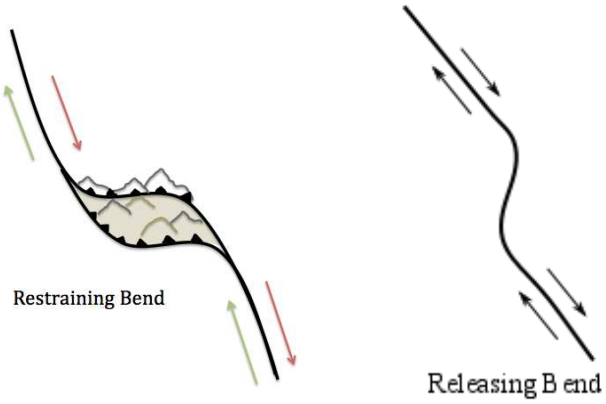


Figure 2.5.3: A transpressional strike-slip fault, causing uplift called a restraining bend. A transtensional strike-slip fault, resulting in a releasing bend.

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.

Piercing Points



Figure 2.5.4: Wallace (dry) Creek on the Carrizo Plain, California. Note 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.

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 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.

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2.6: Hotspots

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, islands in the ocean and mountains on land [82]. As the plate moves across the hotspot, the volcano 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].

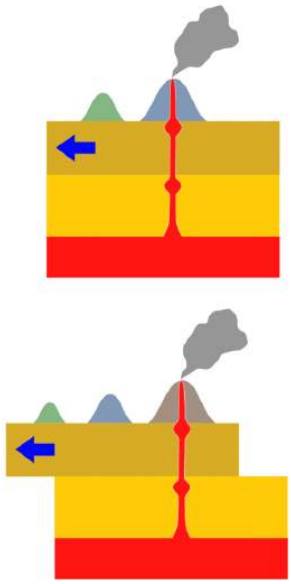


Figure 2.6.1: Diagram showing a non-moving source of magma (mantle plume) and a moving overriding plate.

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 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.

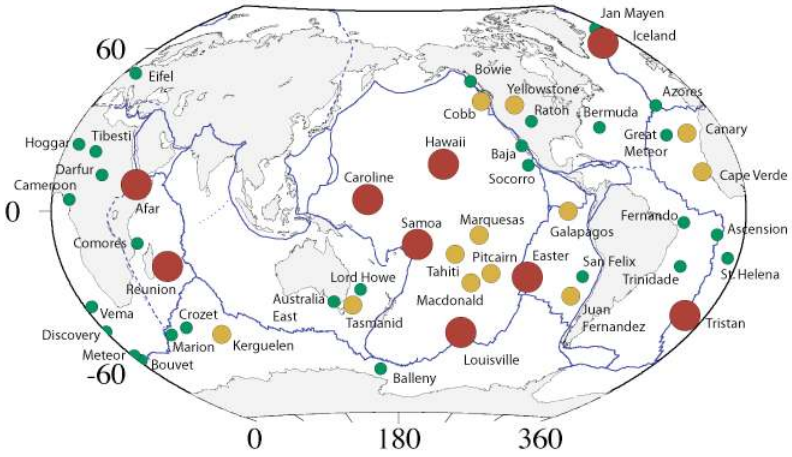


Figure 2.6.2: Map of world hotspots. Larger circles indicate more active hotspots.

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].

The United States is home to two of the largest and best-studied hotspots: Hawaii and Yellowstone.

Hawaiian Hotspot

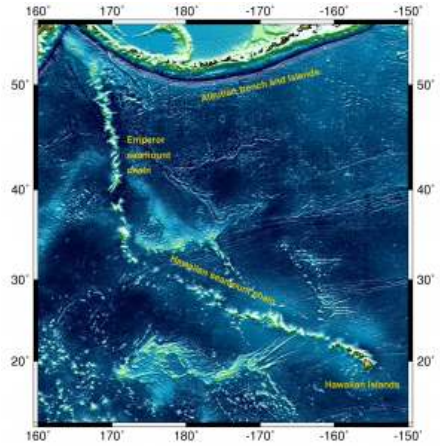


Figure 2.6.3: The Hawaii-Emperor seamount and island chain.

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. The Kilauea volcano is the main vent for this hotspot and has been actively erupting since 1983. So the youngest islands are above the active hotspot and the oldest islands in the chain are the furthest away from the active hotspot.

This enormous volcanic island chain, much of which is underwater, stretches across the Pacific for almost 6,000 km. The seamount chain's most striking feature is a sharp 60-degree 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].

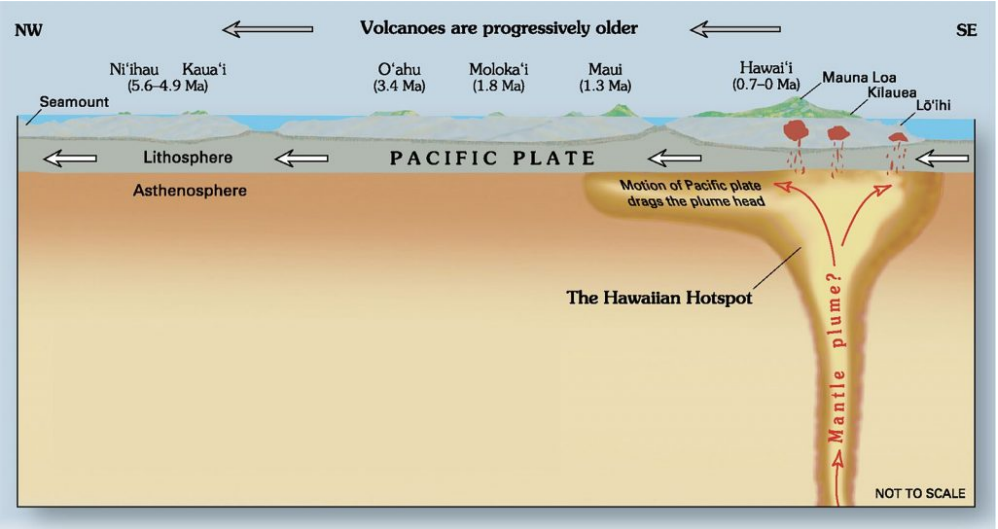


Figure 2.6.4:: Diagram of the Hawaiian hotspot and islands that it formed.

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.

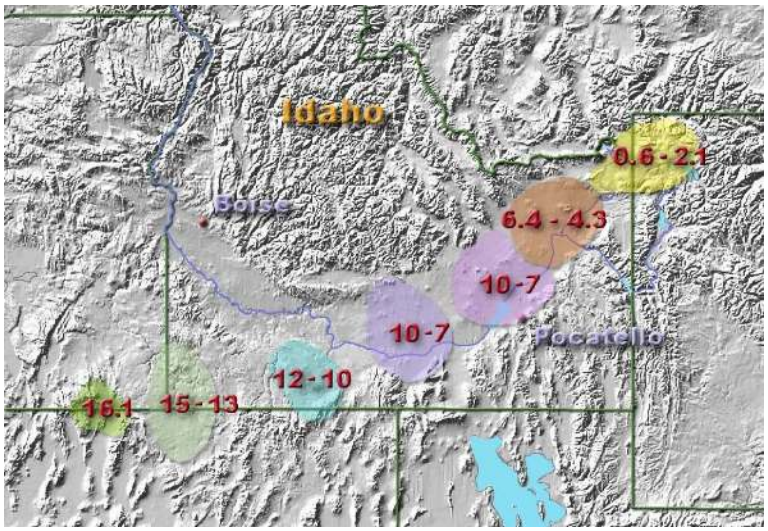


Figure 2.6.5:: The track of the Yellowstone hotspot which shows the age of different eruptions millions of years ago.

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].

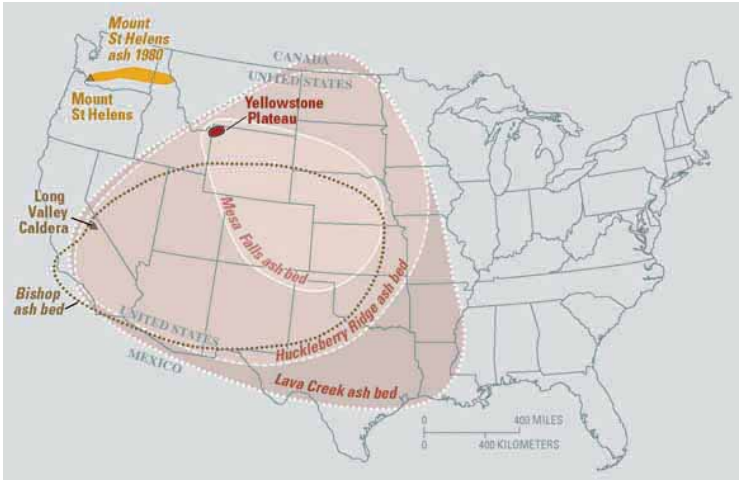


Figure 2.6.6:: Several prominent ash beds found in North America, including three Yellowstone eruptions which are shaded pink (Mesa Falls, Huckleberry Ridge, and Lava Creek), the Bisho Tuff ash bed (brown dashed line), and the modern May 18th, 1980 ash fall (yellow).

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 1,000 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.

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CHAPTER OVERVIEW

3: Minerals

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. Ordered atomic structure (crystalline)
3. Defined chemical composition
4. Inorganic
5. Solid at room temperature
6. Homogeneous

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. Typically, substances like coal, pearl, opal, or obsidian that do not fit the definition of a mineral are called mineraloids.



Figure 3.1: These selenite (gypsum) crystals, found in The Cave of the Crystals in Naica, Mexico, has some of the largest minerals ever found. The largest crystal found here is 39 feet (12 meters) and 55 tones.

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 sand, gravel, things precipitated from solution, etc.), 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

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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.

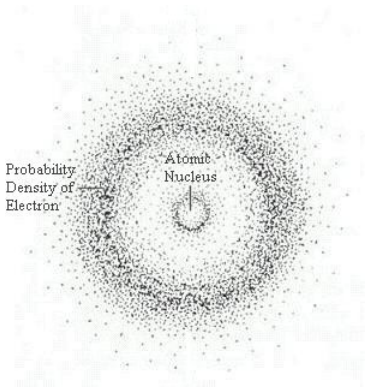


Figure 3.1.1: Electron cloud model of the atom

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

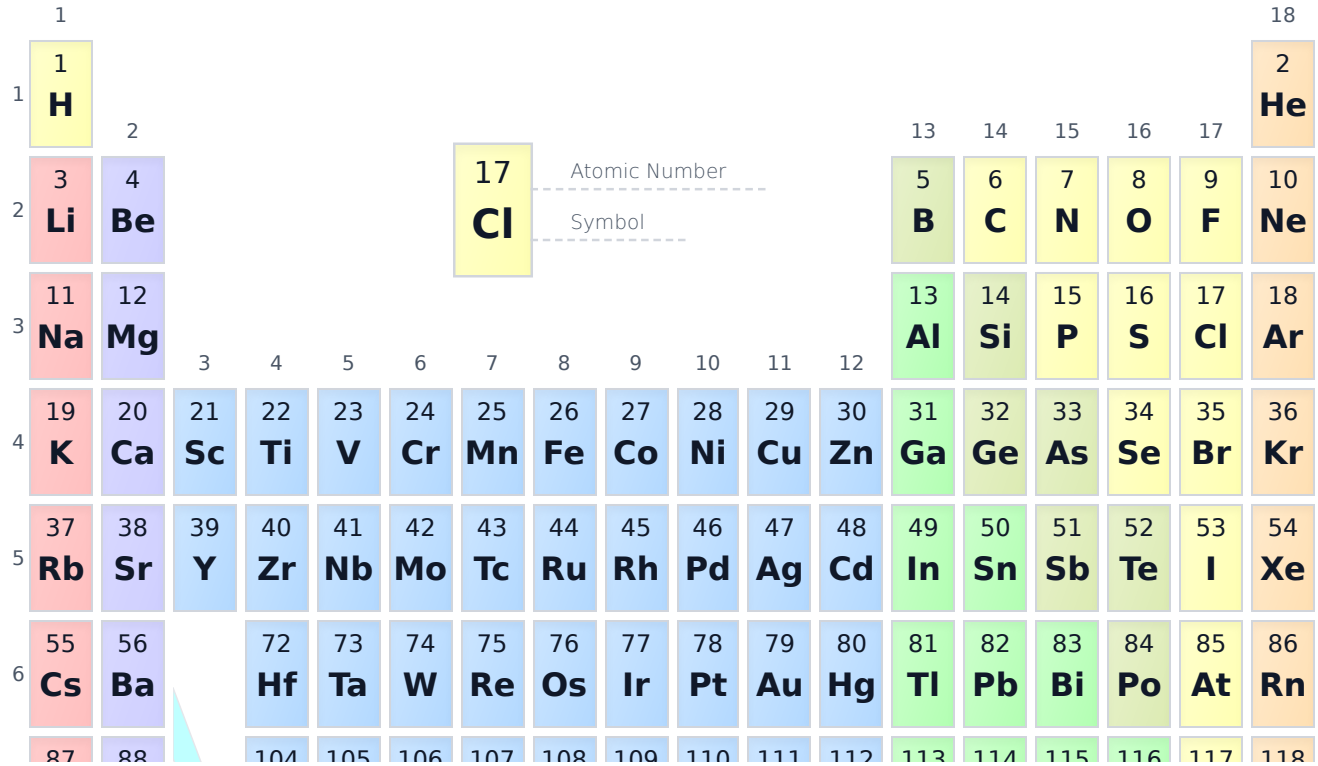


Figure 3.1.2: The Periodic Table of the Elements

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.

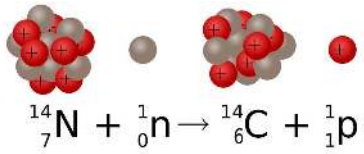


Figure 3.1.3: Formation of Carbon 14 from Nitrogen 14

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. The heaviest naturally occurring

element is uranium, 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.

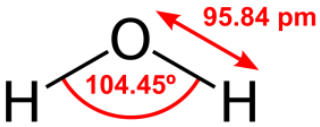


Figure 3.1.4: A model of a water molecule, showing the bonds between the hydrogen and oxygen.

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.

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 substitute for one another. Some olivine has all magnesium, some has all iron, and some has a 50/50 mix or any combination in between.

Charge

The number of electrons, which carry a negative charge, are compared with the number of protons, which carry a positive charge, to determine an atom's overall charge. An atom with a positive or negative charge is called an **ion**.

Ionic Bonding

Ionic 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 transfers both atoms thus achieve stability. For example one atom of sodium (Na^{+1}) and one atom of chlorine (Cl^{-1}) 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^{-1}) 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. This is the weaker of the two bond types presented here.

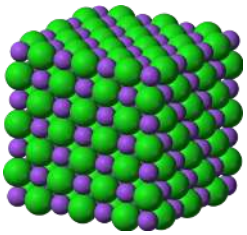


Figure 3.1.5: Cubic arrangement of Na and Cl ions in Halite

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 achieve stability. For example, oxygen (atomic number 8) has 8 electrons. In the case of oxygen, two atoms attach to each other and share 2 electrons 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 to become stable. Each hydrogen shares its electron with the carbon to form a molecule as shown in the figure below. This is the stronger of the two bond types presented here.

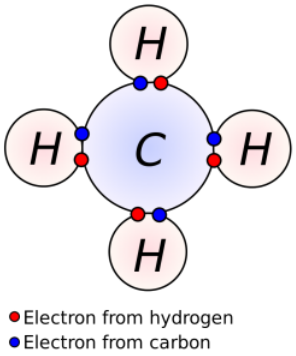


Figure 3.1.6: Methane molecule

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3.2: Formation of Minerals

Minerals form when atoms bond together in a crystalline arrangement (ordered atomic structure). 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 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^{+1} and Cl^{-1} ions separate and disperse into the solution.



Figure 3.2.1: Calcium carbonate deposits from hard water on a faucet

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 ions or molecules 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 daily life is when water evaporates and leaves behind a rind of minerals on faucets, showerheads, and drinking glasses.

In nature, changes in environmental conditions may cause the minerals dissolved in water to form bonds and grow into crystals or cement grains of sediment together. The travertine terraces at Mammoth Hot Springs in Yellowstone Park are an example of precipitation where at the edges of the shallow spring-fed ponds, water flows over those edges and calcite forms.



Figure 3.2.2: The Bonneville Salt Flats of Utah

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. When magma,

which is hot and has freely moving ions, is emplaced at depth or extruded onto the surface as lava, it starts to cool, the ions vibrate less, and mineral crystals form.



Figure 3.2.3: Lava, magma at the earth's surface

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 ($CaCO_3$). Marine invertebrates such as corals and clams precipitate 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 called limestone. Though limestone can form inorganically, the vast majority is formed by this biological process.

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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 four thousand known minerals on Earth, most are rare. There are only a few that make up most of the rocks likely to be encountered by us. These are generally called the **rock-forming minerals**.

The silicon-oxygen tetrahedron (SiO_4) consists of a single silicon atom at the center and four oxygen atoms located at the four corners of the tetrahedron. Each oxygen ion has a -2 charge and the silicon ion has a +4 charge. The silicon ion shares one of its four electrons with each of the four oxygen ions in a covalent bond to create a symmetrical geometric four-sided pyramid figure. Only half of the oxygen's electrons are shared, giving the silicon-oxygen tetrahedron an ionic charge of -4. This silicon-oxygen tetrahedron forms bonds with many other combinations of ions to form the large group of silicate minerals.



Figure 3.3.1: Ping pong ball model of a tetrahedron: balls are oxygen, small space in the center is silicon

The silicon ion is much smaller than the oxygen ions (see the figures) and fits into a small space in the center of the four large oxygen ions. The metal ball represents the silicon ion in the figure. Because only one of the electrons of the corner oxygens is shared, the silicon-oxygen tetrahedron has chemically active corners available to form bonds with other silica tetrahedra or other positively charged ions such as Al_{+3}, $Fe_{+2,+3}$, Mg_{+2}, K_{+1}, Na_{+1}, and Ca_{+2}. Depending on many factors, such as the original magma chemistry, silica-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 chemical structure in which positively charged ions can be inserted for unique chemical compositions forming silicate mineral groups.



Figure 3.3.2: The silicon atom in the center of the tetrahedron (with the top oxygen removed) represented by a metal ball

The Dark Ferromagnesian Silicates



Figure 3.3.3: Green olivine in basalt

The Olivine Family

Olivine is the primary mineral component in mantle rock. 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 can freely substitute for each other in the same location in the crystal structure.

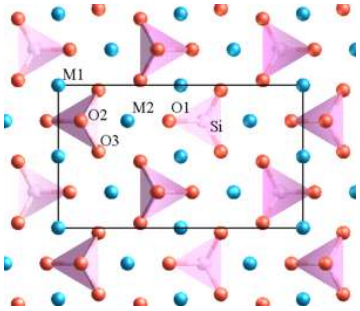


Figure 3.3.4: Tetrahedral structure of olivine

The crystal structure of olivine is built from independent silica tetrahedra. This arrangement, where the tetrahedra are not connected to one another, is also often referred to as having isolated tetrahedra.

Pyroxene Family



Figure 3.3.5: Crystals of diopside, a member of the pyroxene family

Pyroxenes are typically black or dark green in color and have a complex chemical composition that includes iron, magnesium, aluminum, and other elements bonded to the silica tetrahedra. They are built from long, single chains of silica tetrahedra formed as tetrahedra share two corner oxygens. The chains are bonded together into the crystal structures by metal positively charge ions.



Figure 3.3.6: Single chain

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.

Amphibole Family



Figure 3.3.7: Hornblende crystals



Figure 3.3.8: Elongated crystals of hornblende in orthoclase

Amphibole minerals are built from double chains of silica tetrahedra. Imagine two pyroxene chains that connect together by sharing the third oxygen on each tetrahedron.

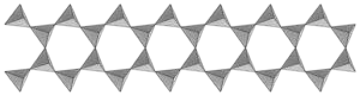


Figure 3.3.9: Double chain structure

Amphiboles are composed of iron, magnesium, aluminum, and other positive ions bonded with silica tetrahedra. Their chemical formula is very complex and generally written as $(RSi_4O_{11})_2$, where R represents many different ions. 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 equals Li, Na, Mg, Fe, Mn, or Ca; and Z is 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.

Sheet Silicates



Figure 3.3.10: (left) Sheet crystals of biotite mica. (right) A stack of sheets of muscovite mica

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.

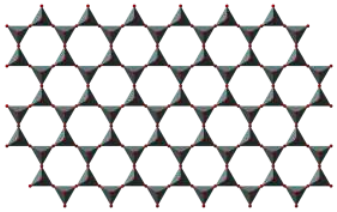


Figure 3.3.11: Sheet structure of mica, view perpendicular to the sheets

Chemically, sheet silicates usually contain silicon and oxygen in a 2:5 ratio (Si_4O_{10}). Micas contain mostly silica, aluminum, and potassium.

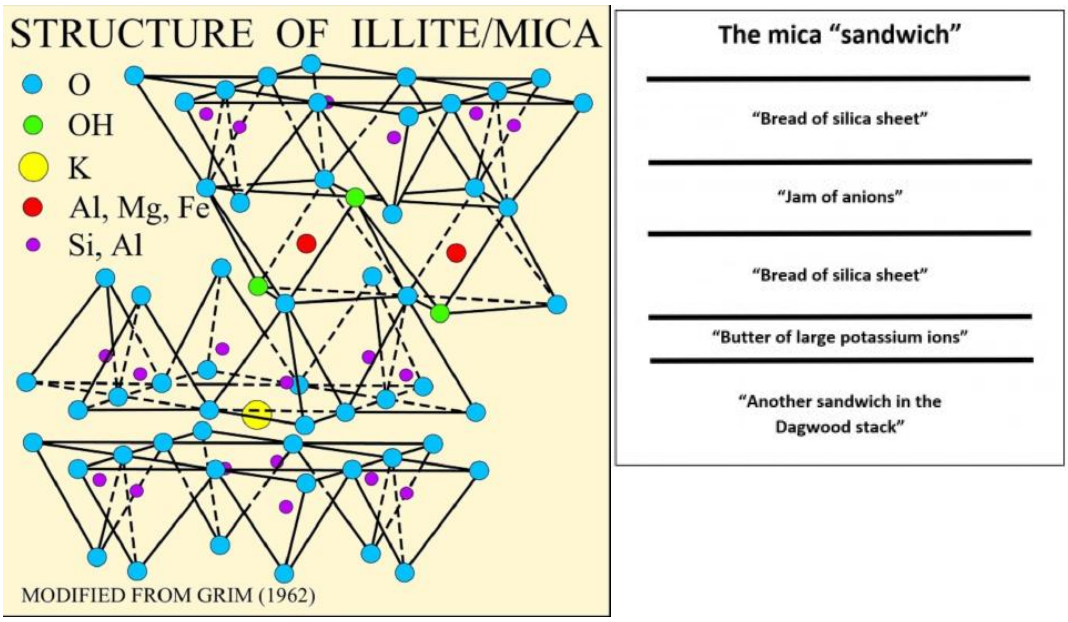


Figure 3.3.12: (left) Crystal structure of a mica, view parallel to the sheets. (right) Mica "silica sandwich" structure related to layers in illite structure.

The illustration of the crystalline structure of mica shows the corner O atoms 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 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 breaking into sheets.

Framework Silicates



Figure 3.3.13: Quartz crystals

Quartz and feldspar are the two most abundant minerals in the continental crust. In fact, feldspar itself is the single most abundant mineral in the Earth's crust. These minerals are classified as framework silicates. They are built with a three-dimensional framework of silica tetrahedra in which all four corner oxygens are shared with adjacent tetrahedra. Within these frameworks in feldspar, are 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.



Figure 3.3.14: Pink orthoclase crystals

Quartz is composed of pure silica, SiO_2. and lacks the holes and spaces that feldspar has. Impurities consisting of atoms within this framework give rise to many varieties of quartz among which are gemstones like amethyst (purple), rose quartz (pink), and citrine (yellow).

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3.4: Non-Silicate Minerals

The crystal structure of non-silicate minerals (see table) does not contain silica-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



Figure 3.4.1: Calcite crystal in shape of rhomb. Note the double-refracted word “calcite” in the center of the figure due to birefringence.

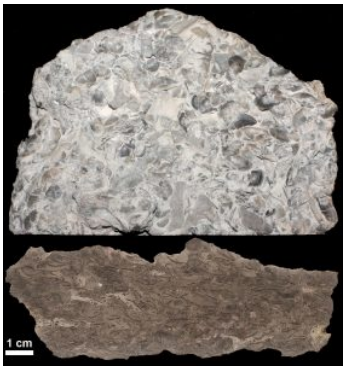


Figure 3.4.2: Limestone full of small fossils

Calcite ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$) are the two most frequently occurring carbonate minerals (contain CO_3), and usually occur in sedimentary rocks which we will learn about later.

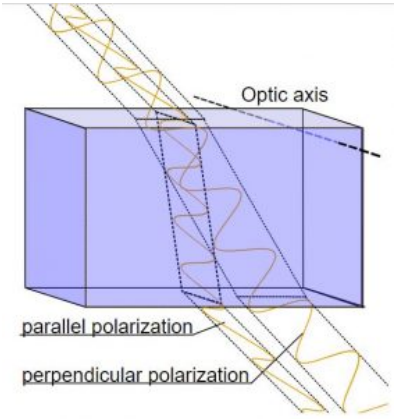


Figure 3.4.3: Birefringence in calcite crystals

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.

Oxides, Halides, and Sulfides



Figure 3.4.4: Limonite, hydrated oxide of iron

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. 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.

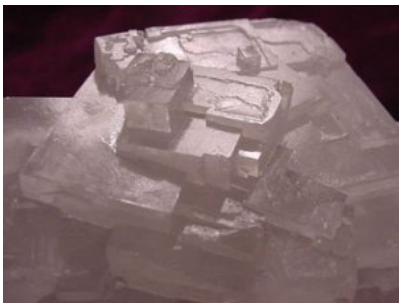


Figure 3.4.5: Halite crystal showing cubic habit

The **halides** consist of halogens, usually fluorine or chlorine, ionically bonded with sodium or other positive ions. 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: Salt crystals at the Bonneville Salt Flats

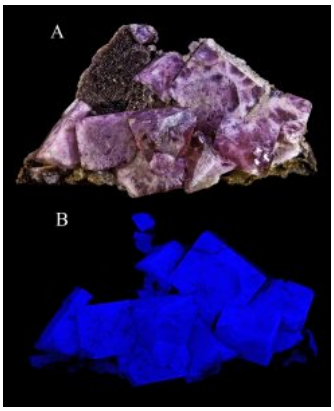


Figure 3.4.7: Fluorite. B shows fluorescence of fluorite under UV light

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 3.4.8: Cubic crystals of pyrite

Many important metal ores are **sulfides,** in which metals are bonded to sulfur. Significant examples include galena (lead sulfide), sphalerite (zinc sulfide), pyrite (iron sulfide, sometimes called “fool’s gold”), and chalcopyrite (iron-copper sulfide). 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.

Sulfates



Figure 3.4.9: Gypsum crystal

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.

Phosphates



Figure 3.4.10: Apatite crystal

Phosphate minerals have a tetrahedral phosphate unit (PO_4^{-3}) combined with various ions. 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.

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. (right) Native sulfur deposited around a volcanic fumarole

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3.5: Identifying Minerals

Geologists identify minerals by their physical 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.

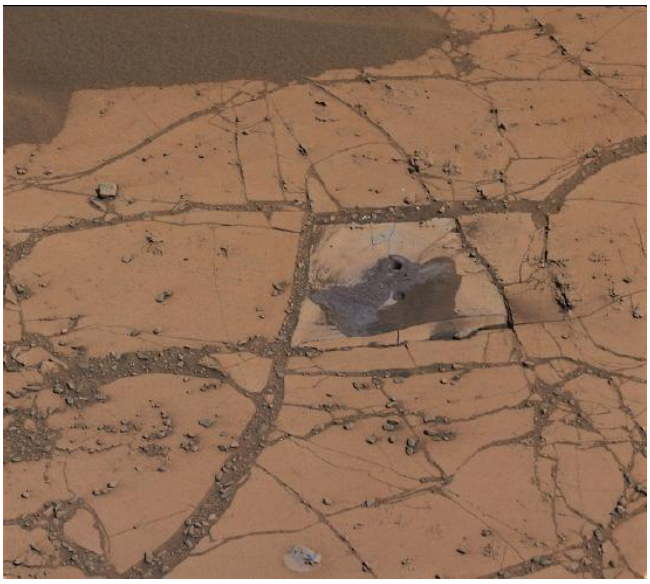


Figure 3.5.1: The rover Curiosity drilled a hole in this rock from Mars, and confirmed the mineral Hematite, as mapped from satellites.

Luster and Color



Figure 3.5.2: 15 mm metallic hexagonal molybdenite crystal from Quebec.

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 3.5.3: Submetallic luster shown on an antique pewter plate.

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.

Luster	Image	Description
Vitreous/glassy	 Quartz crystal	Surface is shiny like glass
Earthy/dull	 Kaolin specimen showing dull or earthy luster	Dull, like dried mud or clay
Silky	 Specimen showing silky luster	Soft shine like silk fabric
Pearly	 Specimen showing pearly luster	Like the inside of a clam shell or mother-of-pearl
Submetallic	 Submetallic luster on sphalerite	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.

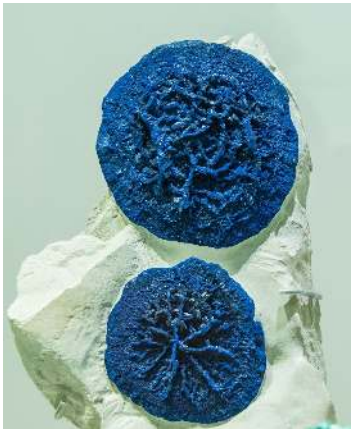


Figure 3.5.4: Azurite is ALWAYS a dark blue color and has been used for centuries for blue pigment.

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. 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.

Streak



Figure 3.5.5: Some minerals have different streaks than their visual color

Streak examines the color of a powdered mineral and can be seen when a mineral sample is scratched or scraped on an unglazed porcelain plate. 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.

Hardness

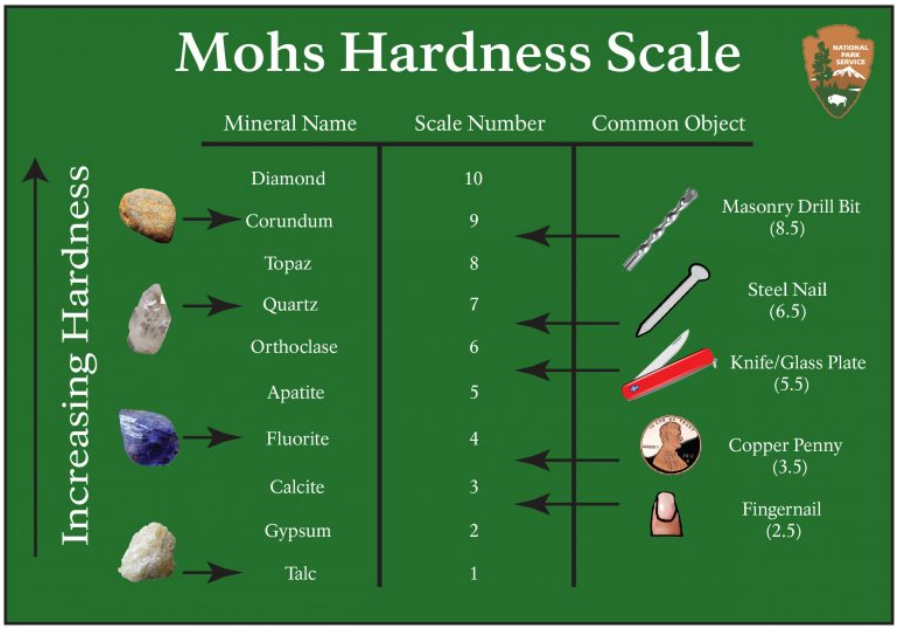


Figure 3.5.6: Mohs Hardness Scale

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. 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.

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. The table lists typical crystal habits of various minerals.

Habit	Image	Examples
Bladed long and flat crystals	 Bladed kyanite	kyanite, amphibole, gypsum
Botryoidal/mammillary blobby, circular crystals	 Malachite from the Congo	hematite, malachite, smithsonite
Coating/laminae/druse crystals that are small and coat surfaces	 Quartz (var. amethyst) in a geode	quartz, calcite, malachite, azurite
Cubic cube-shaped crystals	 Cubic crystals of galena	pyrite, galena, halite
Dodecahedral 12-sided polygon shapes	 Pyrite crystals with dodecahedral habit	garnet, pyrite
Dendritic branching crystals	 Manganese dendrites, scale in mm.	Mn-oxides, copper, gold
Equant crystals that do not have a long direction	 Olivine crystal	olivine, garnet, pyroxene
Fibrous thin, very long crystals	 Tremolite, a type of amphibole	serpentine, amphibole, zeolite
Layered, sheets stacked, very thin, flat crystals	 Sheet crystals of muscovite	mica (biotite, muscovite, etc.)

Lenticular/platy crystals that are plate-like	 Orange wulfenite	selenite roses, wulfenite, calcite
Hexagonal crystals with six sides	 Hexagonal hanksite	quartz, hanksite, corundum
Massive/granular Crystals with no obvious shape, microscopic crystals	 Limonite, a hydrated oxide of iron	limonite, pyrite, azurite, bornite
Octahedral 4-sided double pyramid crystals	 Octahedral fluorite	diamond, fluorite, magnetite, pyrite
Prismatic/columnar very long, cylindrical crystals	 Columnar tourmaline	tourmaline, beryl, barite
Radiating crystals that grow from a point and fan-out	 Pyrophyllite	pyrite “suns”, pyrophyllite
Rhombohedral crystals shaped like slanted cubes	 Calcite crystal in shape of rhomb	calcite, dolomite
Tabular/blocky/stubby sharp-sided crystals with no long direction		feldspar, pyroxene, calcite

	Crystals of diopside, a member of the pyroxene family	
Tetrahedral three-sided, pyramid-shaped crystals	 Tetrahedrite	magnetite, spinel, tetrahedrite

Another crystal habit that may be used to identify minerals is striations, which are dark and light parallel lines on a crystal face. Twinning is another, which occurs when the crystal structure replicates in mirror images along certain directions in the crystal.

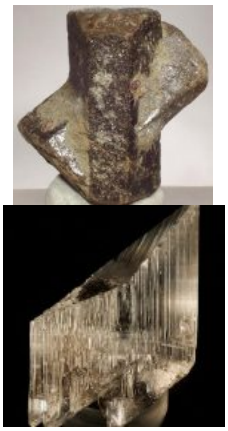


Figure 3.5.7: Twinned staurolite Gypsum with striations

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.

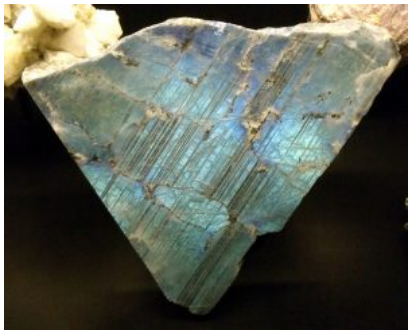


Figure 3.5.8: Striations on plagioclase

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. Some minerals have a strong cleavage, some minerals only have weak cleavage or do not typically demonstrate cleavage.



Figure 3.5.9: Citrine, a variety of quartz showing conchoidal fracture

For example, quartz and olivine rarely show cleavage and typically break into conchoidal fracture patterns.

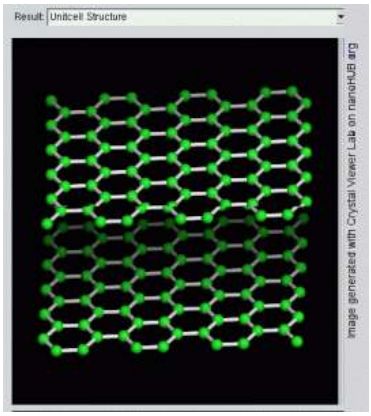


Figure 3.5.10: Graphite showing layers of carbon atoms separated by a gap with weak bonds holding the layers together.

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.

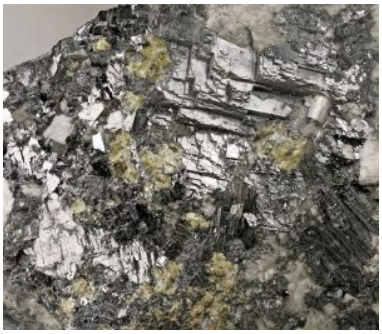


Figure 3.5.11: Cubic cleavage of galena; note how the cleavage surfaces show up as different but parallel layers in the crystal.

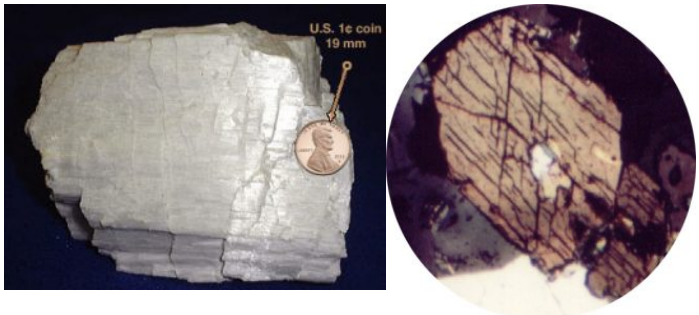


Figure 3.5.12: (left) Steps of cleavage along the same cleavage direction. (right) Photomicrograph showing 120/60 degree cleavage within a grain of amphibole

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 90°.

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 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 (see figure). 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 3.5.13: A demonstration of ulexite's image projection

Another property geologists may use to identify minerals is a property related to density called **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.



Figure 3.5.14: Native gold has one of the highest specific gravities.

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.



Figure 3.5.15: Nails attracted to lodestone (magnetite).

Some iron-oxide minerals are magnetic and are attracted to magnets. Magnetite ($Fe3O_4$) is strongly attracted to magnets and can be magnetized.



Figure 3.5.16: Iridescence on plagioclase; also showing striations on the cleavage surface

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.

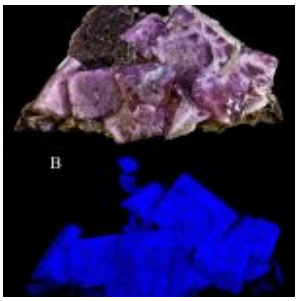


Figure 3.5.17: Fluorite. Lower image shows fluorescence of fluorite under UV light

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CHAPTER OVERVIEW

4: Igneous Processes and Volcanoes

Igneous rock is formed when molten material freezes into solid rock. This molten material is called **magma** when it is underground and **lava** when it is on the surface. 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.

Igneous rock is divided into two major groups: extrusive rock formed from lava that erupts and cools on the surface, and intrusive rock that solidifies from underground magma.



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 mass of 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: Partial Melting and Crystallization

4.5: Volcanism

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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 material from which it came. Composition refers to the rock's specific mineralogy and chemical composition.

Texture



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.

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 like you see in the image above.

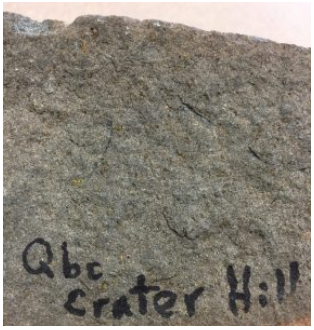


Figure 4.1.2: Basalt is a classic fine-grained (aphanitic) extrusive igneous rock. This sample is mostly fine groundmass with a few small green phenocrysts that are the mineral olivine.

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 like you see in the image above. The fine-grained texture indicates the quickly cooling lava did not have time to grow large crystals.



Figure 4.1.3: Obsidian (volcanic glass). Note conchoidal fracture.

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.

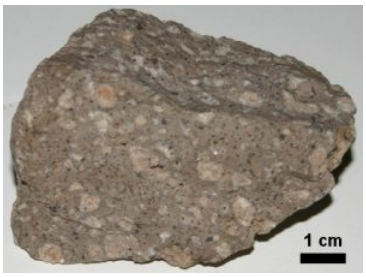


Figure 4.1.4: Porphyritic texture

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** or **matrix**. 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 4.1.5: Pegmatitic texture

Residual molten material expelled from igneous intrusions may form masses containing very large crystals of minerals. 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. In North Dakota, there is a single feldspar crystal the size of a small car!



Figure 4.1.6: Scoria, a vesicular extrusive igneous rock

All magmas contain dissolved gases. As the magma rises to the surface, the drop in pressure causes gasses 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. The type of volcanic rock with common vesicles is called **scoria**.



Figure 4.1.7: Pumice

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

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). Rock formed from large deposits of tephra fragments is called **tuff**.

Composition

Composition refers to a rock's chemical and mineral make-up. For igneous rock, the 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. 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 (which we will not address in this text). 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.

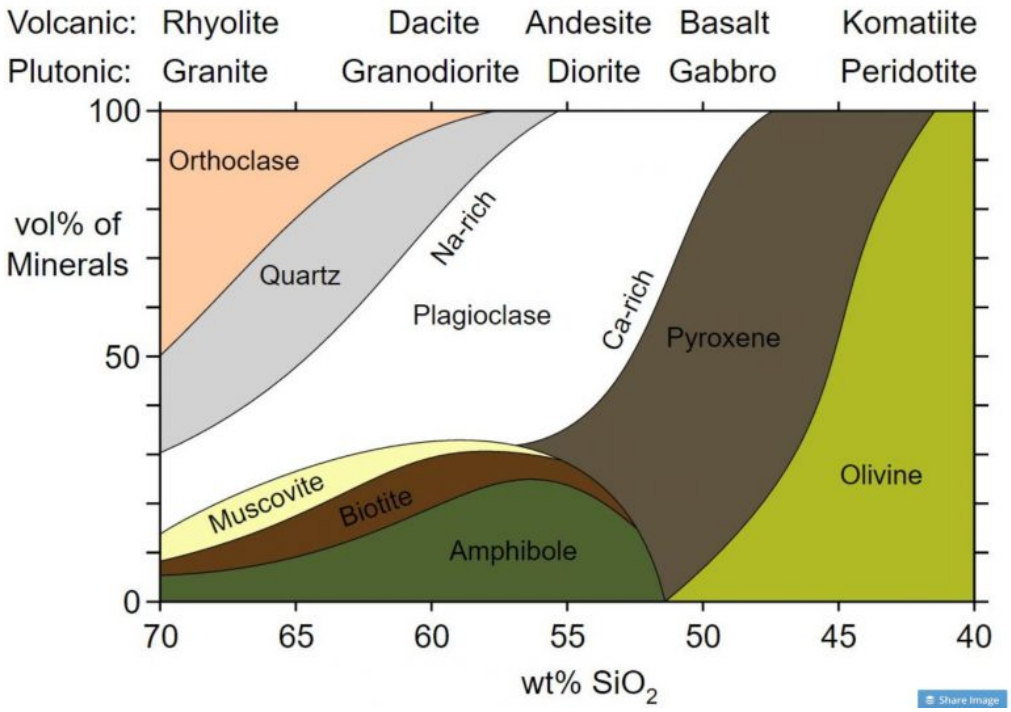


Figure 4.1.9: Mineral composition of common igneous rocks. The percentage of minerals is shown on the vertical axis. The percentage of silica is shown on the horizontal axis. Rock names at the top include a continuous spectrum of compositions grading from one into another.

- **Felsic** refers to a predominance of the light-colored (felsic) minerals **fel**dspar and **si**lica 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 **M**g and **F**e) 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 rocks [2].

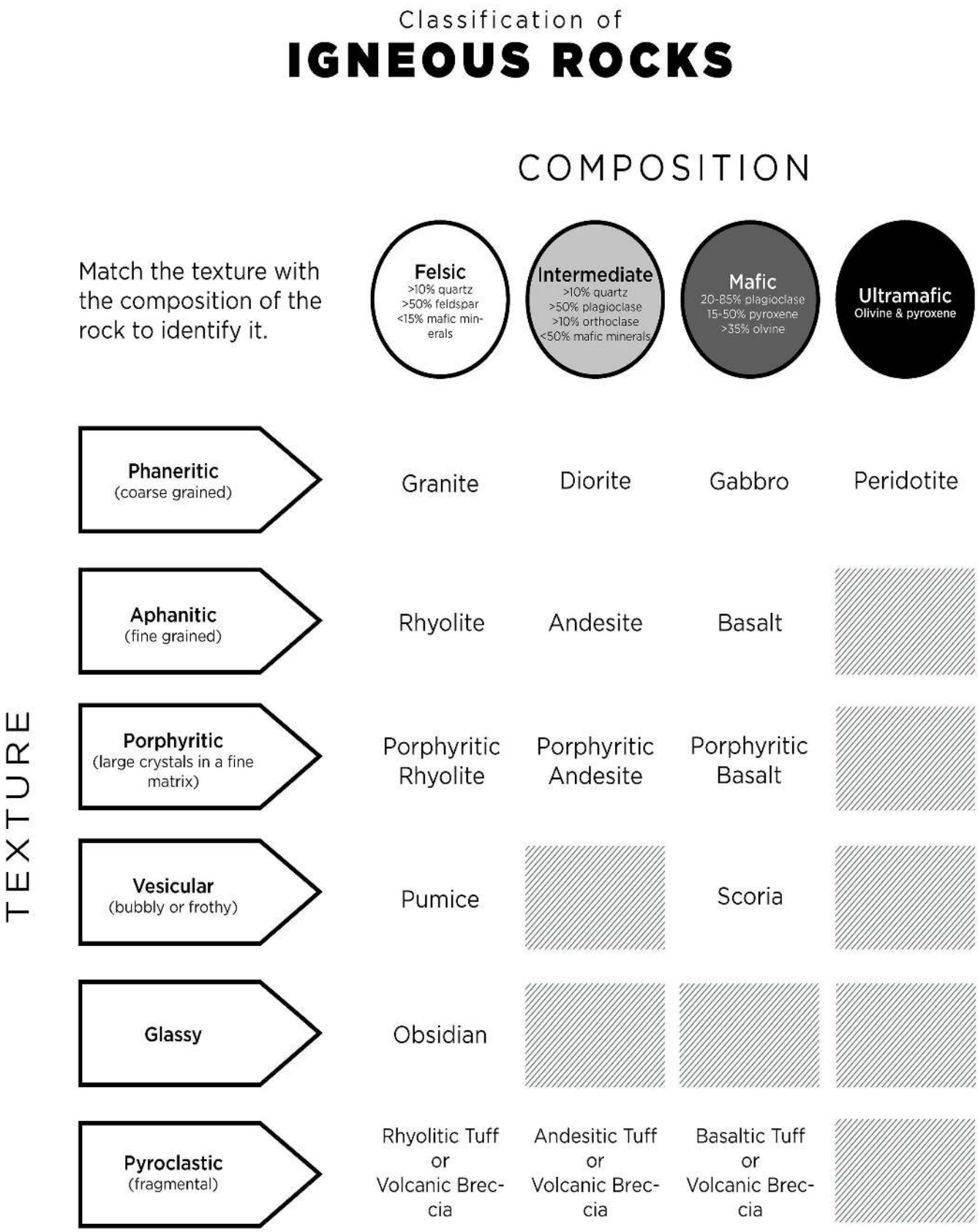
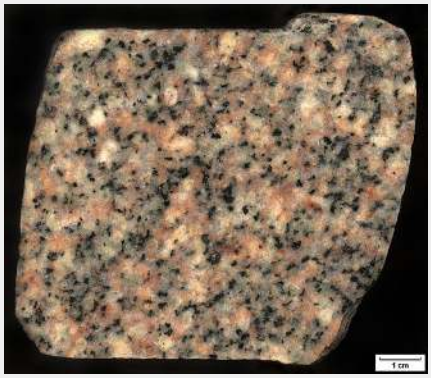


Figure 4.1.10: Igneous rock classification table with composition as vertical columns and texture as horizontal rows.

Aphanitic/Phaneritic Rock Types with Images

Felsic Composition



Granite from Cape Cod, Massachusetts.

Granite is a course-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 (source: Michael C. Rygel via Wikimedia Commons)

Rhyolite is a fine-crystalline felsic extrusive rock. Rhyolite is commonly pink and will often have glassy quartz phenocrysts.

Intermediate Composition



Diorite

Diorite is a coarse-crystalline intermediate intrusive igneous rock. Diorite is identifiable by it's Dalmatian-like appearance of black hornblende and biotite and white plagioclase feldspar.



Andesite

Andesite is a fine crystalline intermediate extrusive rock. It is commonly grey and porphyritic. It can be found in the Andes Mountains and in some island arcs (see Chapter 2). 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. Gabbro is a major component of the lower oceanic crust.	Basalt is a fine-grained mafic igneous rock. It is commonly vesicular and aphanitic. 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 are 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.



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

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.



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

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.

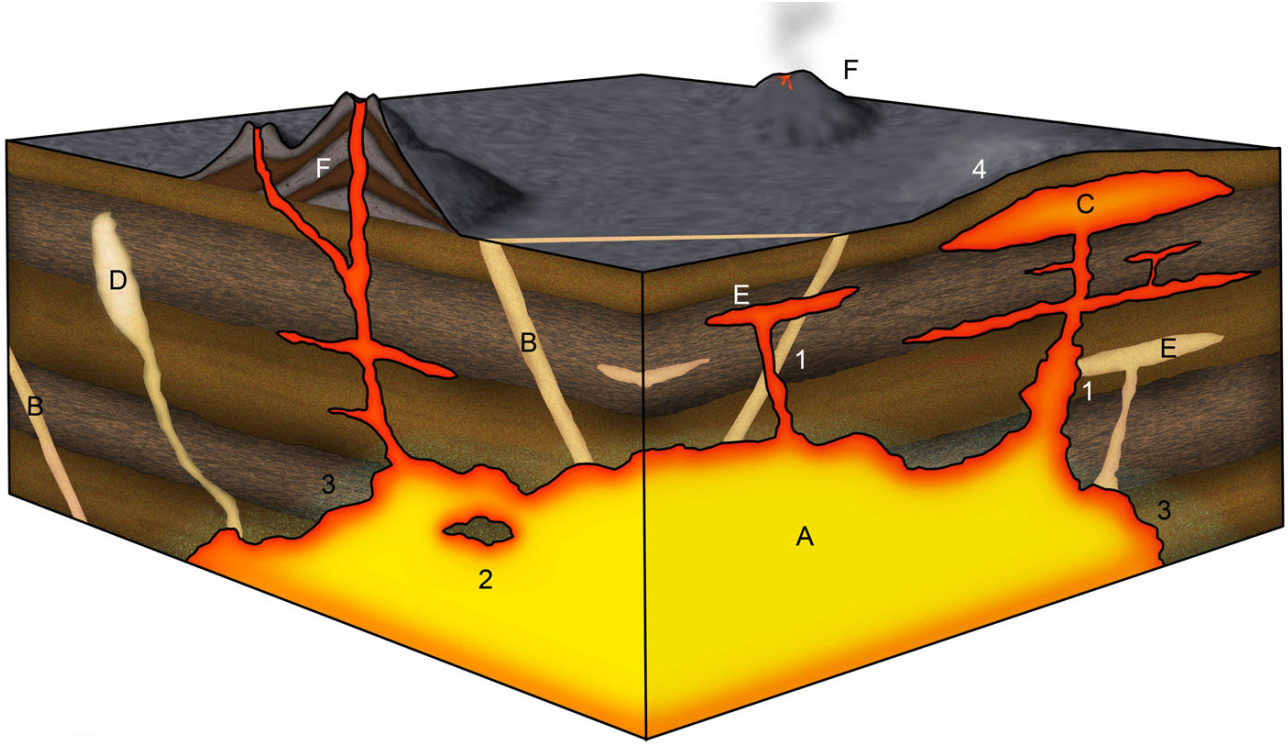


Figure 4.1.13: Cottonwood Stock, a quartz monzonite pluton exposed at the mouth of Little Cottonwood Canyon, Utah

A magma chamber is a large underground reservoir of molten rock. The processes by which a magma 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 body intrudes. The cooking magma forms a mass of intrusive rock called a **pluton**. Plutons can have irregular shapes, but can often be somewhat round.



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.



Guide for Igneous Structures Image (shown above)

Number/Letter	Description
1	Young, emerging subvolcanic intrusion cutting through older one
2 (not addressed in this text)	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 (not addressed in this text yet)	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 dykes/dikes
C (not addressed in this text)	Emerging laccolith
D	Old pegmatite (late-magmatic dyke formed by aggressive and highly mobile residual melts of a magma chamber)
E	Old and emerging magmatic sills
F	Stratovolcano

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4.2: Bowen's Reaction Series

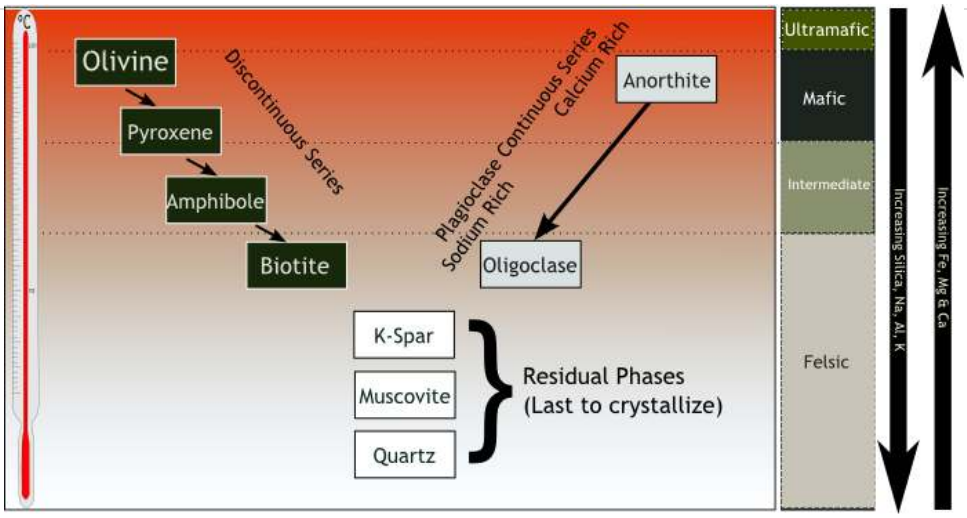


Figure 4.2.1: Bowen's Reaction Series. Minerals that crystallize at higher temperatures are at the top (olivine) and minerals that crystallize at lower temperatures are at the bottom (quartz). (Source Colivine, modified from Bowen, 1922)



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

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 (158°F). The upper end of the range where all minerals exist in a molten state is approximately 1,250°C (2,282°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 (see Chapter 4.4). However, the order and relationships are maintained.

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, crystalize at lower temperatures.



Figure 4.2.3: Normal L. Bowen

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.

The results of his 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.

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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. **Volatiles** are gaseous components—such as water vapor, and carbon dioxide—dissolved in the magma [6]. The presence and amount of these three components affect the physical behavior of the magma.

Geothermal Gradient

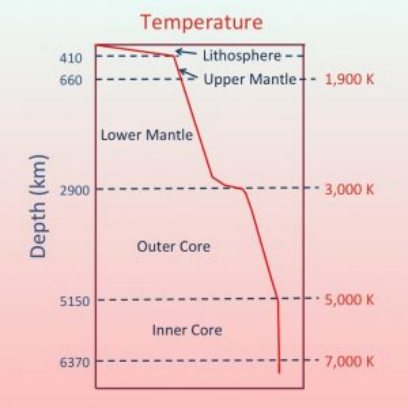


Figure 4.3.1: Geothermal gradient

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**. For every kilometer of depth, the temperature increases by about 25°C.

An important concept to know is pressure affects the temperature at which a material melts. If a material is solid at a high temperature and high pressure, you can either raise the temperature to get it to melt OR lower the pressure. The lower pressure allows the ions in the minerals to vibrate and the material will melt. This is called decompression melting.

There are three principal ways rock converts to molten magma: 1) decompression melting caused by lowering the pressure, 2) fluid-induced melting caused by adding volatiles (see more below), 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 temperature at which it cools 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.



Figure 4.3.2: Progression from rift to the mid-ocean ridge, the divergent boundary types. Note the rising material in the center.

Magma is created at mid-ocean ridges via **decompression melting**. Strong convection currents cause the solid asthenosphere to slowly flow beneath the lithosphere. The upper part of the lithosphere (crust) is a poor heat conductor, so the temperature remains about the same throughout the underlying mantle material. Where the convection currents cause mantle material to rise, the pressure decreases, which causes the melting point to drop, and partial melting starts. As this magma continues to rise, it cools and crystallizes to form new lithospheric crust.

Fluid-Induced Melting

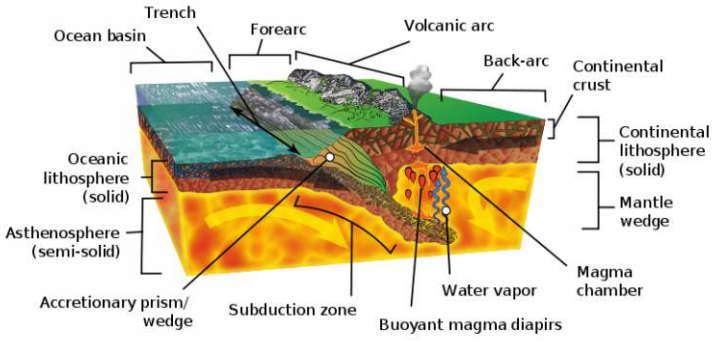


Figure 4.3.3: Diagram of ocean-continent subduction. Note water vapor driven out of hydrated minerals in the descending oceanic slab.

Fluid-induced melting occurs in island arcs and subduction zones when volatile gases are added to mantle material. Fluid-induced melting magma produces many of the volcanoes in the circum-Pacific subduction zones, also known as the Ring of Fire. The subducting slab contains oceanic lithosphere and hydrated minerals. As covered in Chapter 2, these hydrated forms are created when water ions bond with the crystal structure of silicate minerals. As the slab descends into the hot mantle, the increased temperature causes the hydrated minerals to emit water vapor and other volatile gases, which are expelled from the slab-like water being squeezed out of a sponge. The volatiles dissolve into the overlying asthenospheric mantle and decrease its melting point. In this situation the applied pressure and temperature have not changed, the mantle's melting point has been lowered by the addition of volatile substances. This is analogous to adding salt to an icy roadway. The salt lowers the freezing temperature of the solid ice so it turns into liquid water.

Heat-Induced Melting



Figure 4.3.4: Migmatite is a partially molten metamorphic rock. (Source: Peter Davis)

Heat-induced melting, transforming solid mantle into liquid magma by simply applying heat, is the least common process for generating magma. Heat-induced melting occurs at the mantle plumes or hotspots. The rock surrounding the plume is exposed to higher temperatures, and the rock begins to melt. The mantle plume includes rising mantle material, meaning the pressure drops and some decompression melting is occurring as well.

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4.4: Partial Melting and Crystallization

Even though all magmas originate from similar mantle rocks, and start out as similar magma, other things, like partial melting and crystallization processes like magmatic differentiation, can change the chemistry of the magma. This explains the wide variety of resulting igneous rocks that are found all over Earth.

Magmatic Differentiation

Liquid magma is less dense than the surrounding solid rock, so it rises through the mantle and crust. As magma begins to cool and crystallize, a process known as **magmatic differentiation** changes the chemistry of the resultant rock towards a more felsic composition. This happens via two main methods: assimilation and partial melting/fractional crystallization [8].



Figure 4.4.1: Xenoliths in Little Cottonwood Stock, Utah

Assimilation

During **assimilation**, pieces of country rock with a different, often more felsic, composition are added to the magma. These solid pieces may melt, which changes the composition of the original magma. At times, the solid fragments may remain intact within the cooling magma and only partially melt. The unmelted country rocks within an igneous rock mass are called xenoliths (see figure above).

Much of the continental lithosphere is felsic (i.e. granitic), and normally more buoyant than the underlying mafic/ultramafic mantle. When mafic magma rises through thick continental crust, it does so slowly, more slowly than when magma rises through oceanic plates. This gives the magma lots of time to react with the surrounding country rock. The mafic magma tends to assimilate felsic rock, becoming more silica-rich as it migrates through the lithosphere and changing into intermediate or felsic magma by the time it reaches the surface. This is why felsic magmas are much more common within continents.

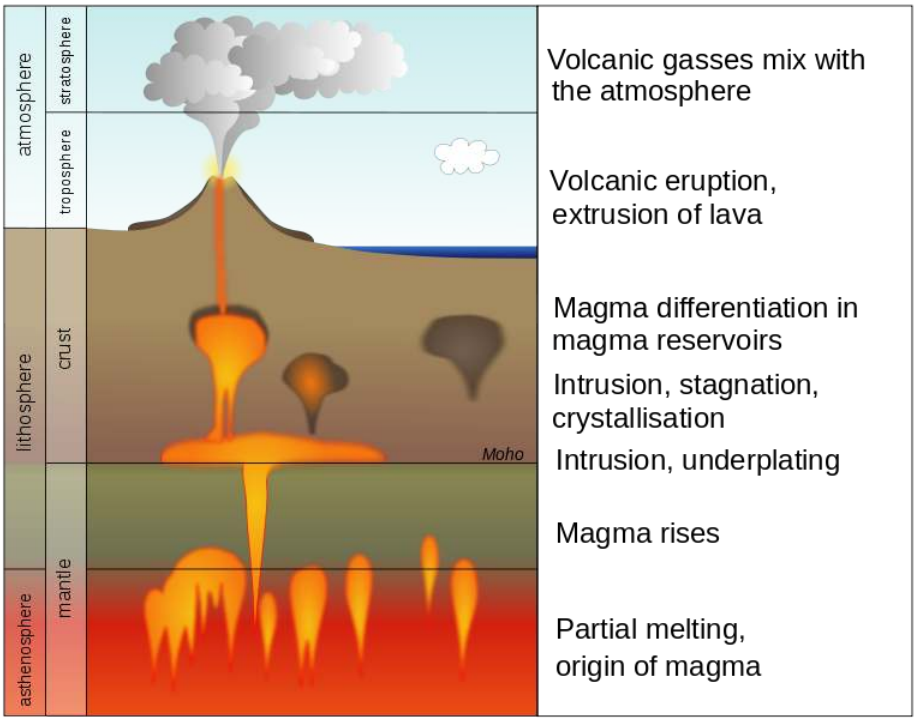


Figure 4.4.2: Rising magma diapirs in mantle and crust. Fractional crystallization occurs in the diapirs in the crust. (Source: Woudloper)

Partial Melting/Fractional Crystallization

According to Bowen's Reaction Series (Section 4.2), each mineral has a unique melting and crystallization temperature. Since most rocks are made of many different minerals, when they start to melt, some minerals begin melting sooner than others. This is known as partial melting and creates magma with a different composition than the original mantle material.

The most important example occurs as magma is generated from mantle rocks (as discussed in Section 4.3). The chemistry of mantle rock (peridotite) is ultramafic, low in silicates and high in iron and magnesium. When peridotite begins to melt, the silica-rich portions melt first due to their lower melting point. If this continues, the magma becomes increasingly silica-rich, turning ultramafic mantle into mafic magma, and mafic mantle into intermediate magma. The ultramafic elements are left behind as solids. The now intermediate magma rises to the surface because it is more buoyant than the mantle.

Partial melting also occurs as existing crustal rocks melt in the presence of heat from magmas. In this process, existing rocks melt, allowing the magma formed to be more felsic and less mafic than the pre-existing rock.

Fractional crystallization is another process that increases the magma silica content, making it more felsic [9]. As the temperature drops within a magma rising through the crust, some minerals will crystallize and settle to the bottom of the magma chamber, leaving the remaining melt depleted of those ions. Olivine is a mafic mineral at the top of the Bowen's Reaction series with a high melting point and a smaller percentage of silica versus other common igneous minerals. When ultramafic magma cools, the olivine crystallizes first and settles to the bottom of the magma chamber (see figure). This means the remaining melt becomes more silica-rich and felsic. As the mafic magma further cools, the next minerals on Bowen's Reaction Series (plagioclase and pyroxene) crystallize next, removing even more low-silica components from the magma, making it even more felsic.

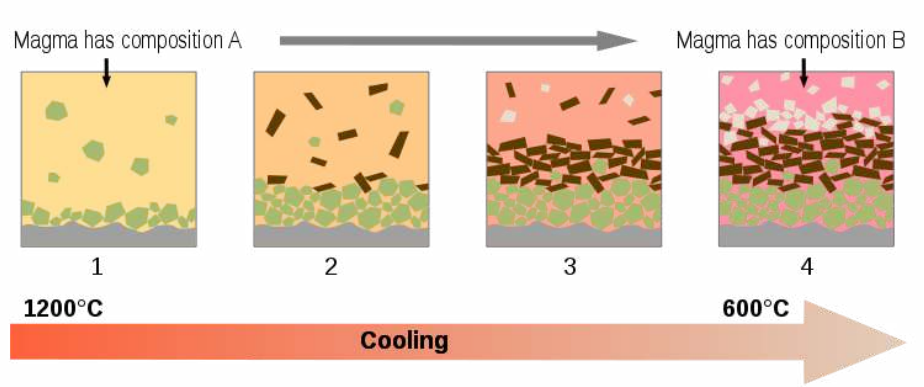


Figure 4.4.3: Schematic diagram illustrating fractional crystallization. If magma at composition A is ultramafic, as the magma cools it changes composition as different minerals crystallize from the melt and settle to the bottom of the magma chamber. In section 1, olivine crystallizes; section 2: olivine and pyroxene crystallize; section 3: pyroxene and plagioclase crystallize; and section 4: plagioclase crystallizes. The crystals are separated from the melt and the remaining magma (composition B) is more silica-rich. (Source: Woudloper)

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4.5: Volcanism

A **volcano** is a type of land formation created when lava solidifies into rock. This section describes volcano location, type, hazards, and monitoring.

Distribution and Tectonics

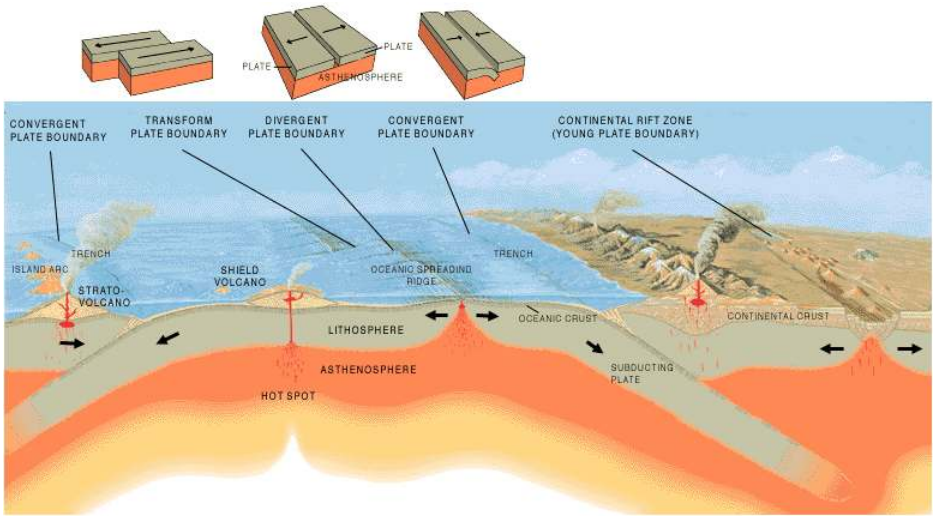


Figure 4.5.1: Association of volcanoes with plate boundaries

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. 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.

Volcanoes at Mid-Ocean Ridges

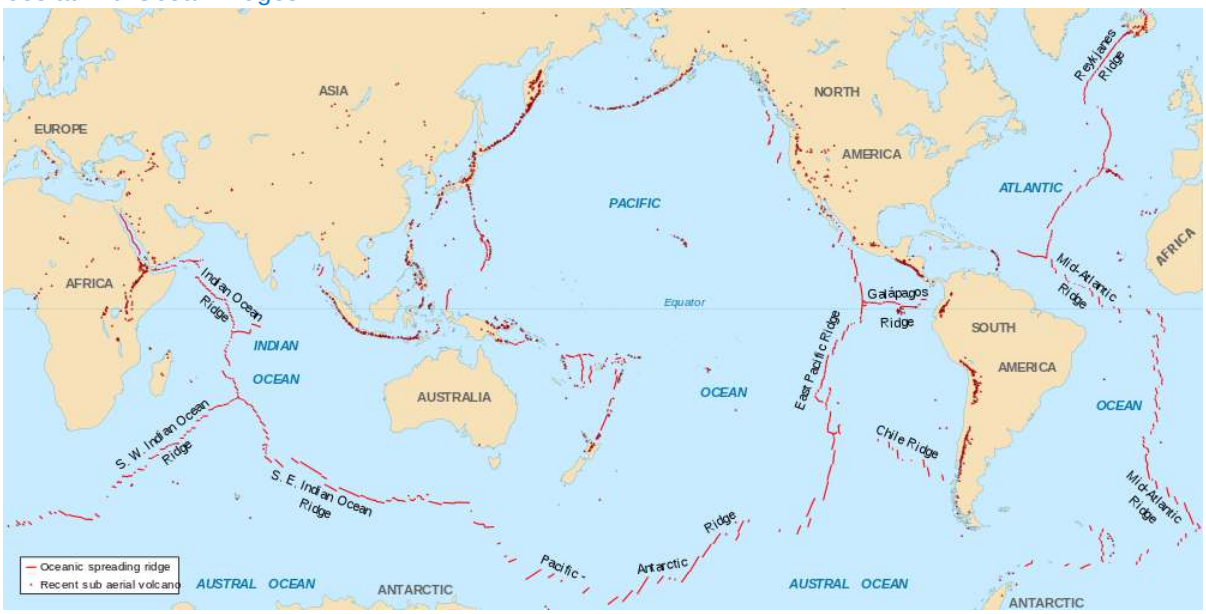


Figure 4.5.2: Map of mid-ocean ridges throughout the world.

Most volcanism on Earth occurs on the ocean floor along mid-ocean ridges, a type of divergent plate boundary (see Chapter 2). These interplate volcanoes are also the least observed and famous since most of them are located under 3,000-4,500 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 since that is at the surface. The diverging and thinning oceanic plates allow hot mantle rock to rise, releasing pressure and causing

decompression melting as described earlier in this chapter. 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 and oceanic lithosphere are basaltic.



Figure 4.5.3: Pillow basalt on the seafloor near Hawaii.

When basaltic lava erupts underwater it emerges in small explosions and/or forms pillow-shaped structures called pillow basalts.

Volcanoes at Subduction Zones

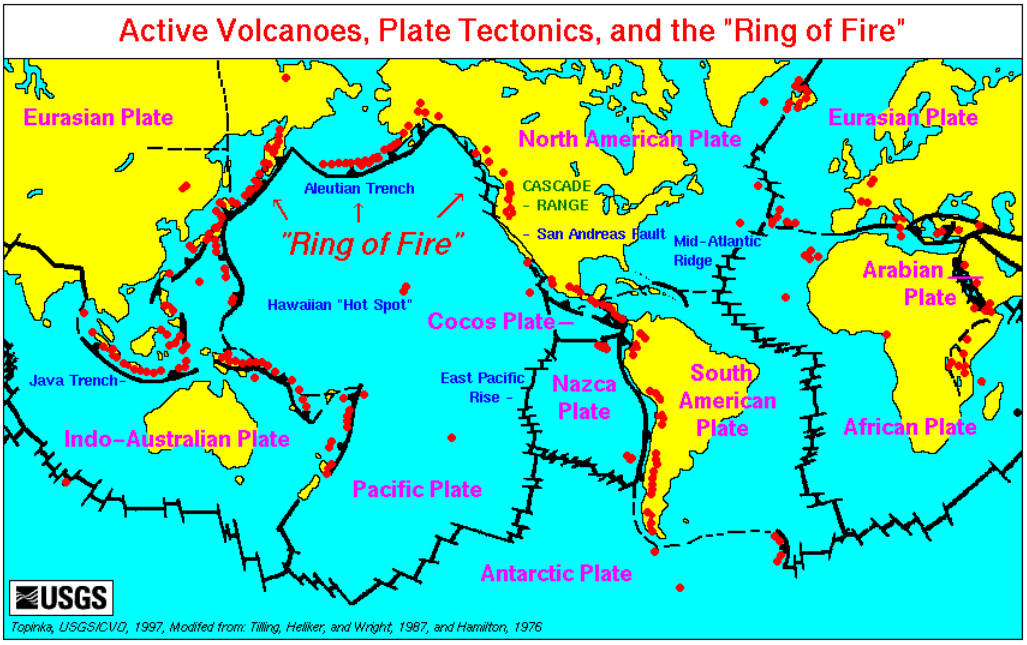


Figure 4.5.4: Distribution of volcanoes on the planet. Click here for an interactive map of volcano distributions.

The second most commonly found location for volcanism is adjacent to subduction zones, a type of convergent plate boundary (see Chapter 2). The process of subduction causes flux melting in the overlying mantle rock as described earlier in this chapter. 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 and rhyolite.

Volcanoes at Continental Rifts



Figure 4.5.5: Basaltic cinder cones of the Black Rock Desert near Beaver, Utah (part of the Basin and Range Province).

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.

Hotspots

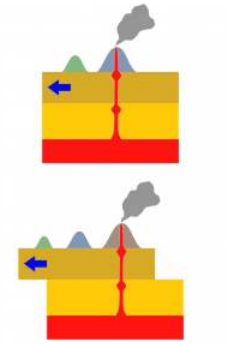


Figure 4.5.6: Diagram showing a non-moving source of magma (mantle plume) and a moving overriding plate.

Hotspots, already covered in the plate tectonics chapter of this book, 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.

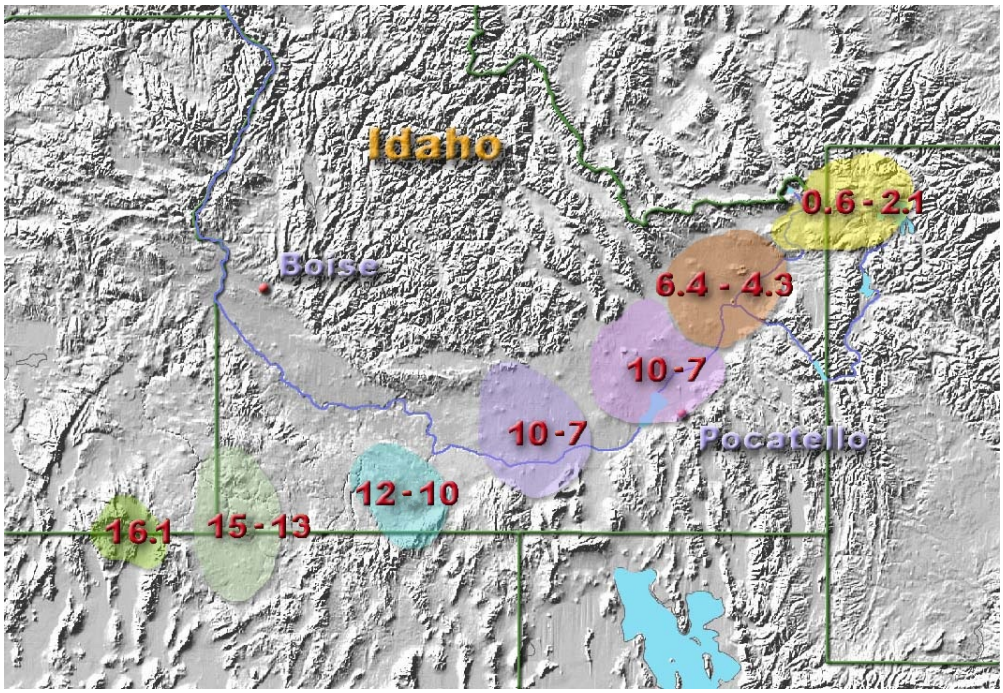


Figure 4.5.7: The track of the Yellowstone hotspot, which shows the age of different eruptions millions of years ago.

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.

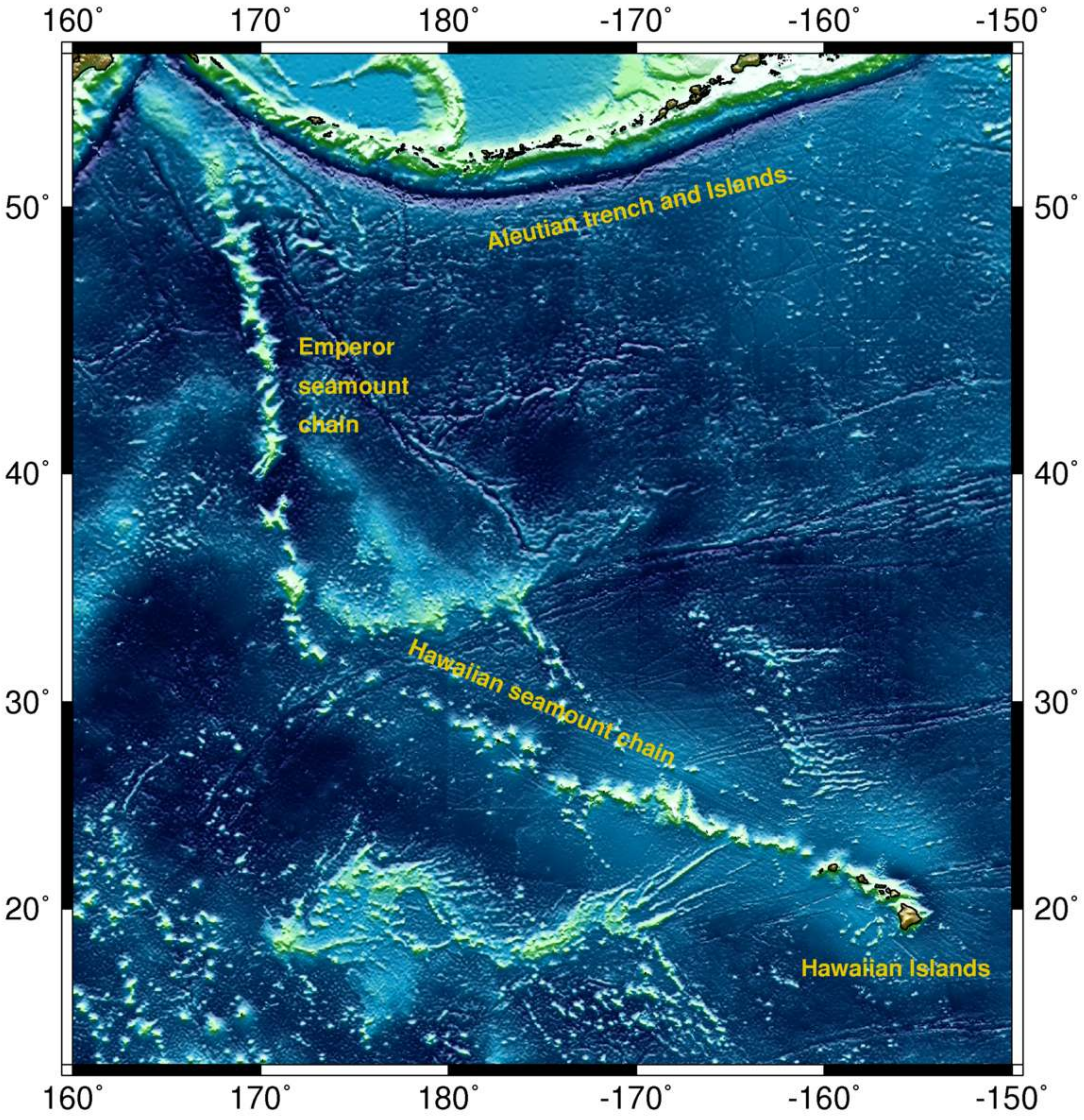
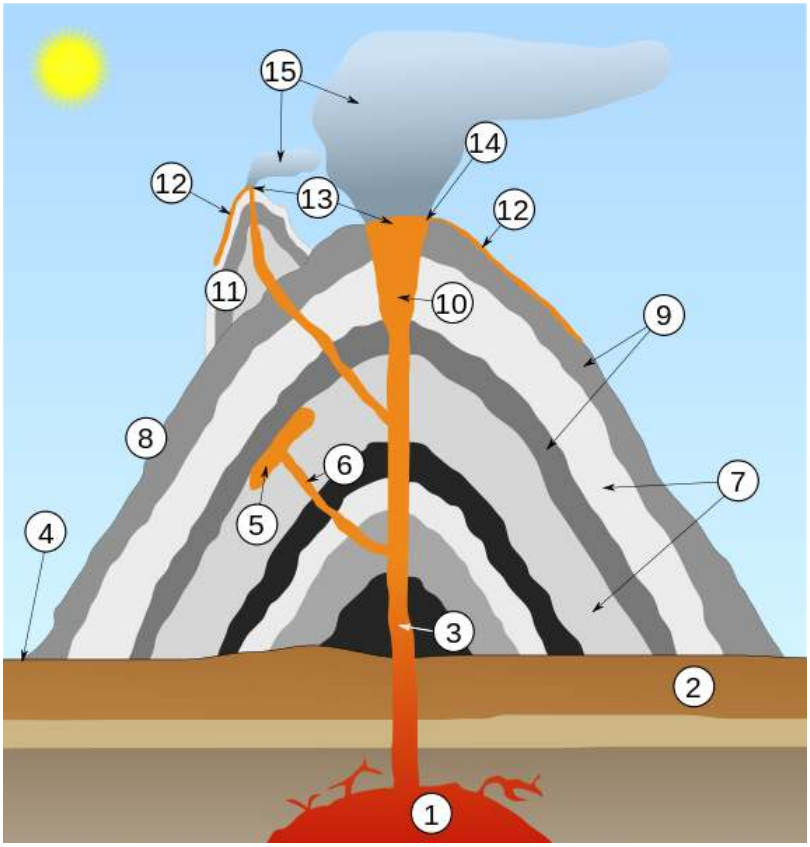


Figure 4.5.8: The Hawaii-Emperor seamount and island chain.

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) upper layers of lithosphere, 3) the 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) feeder tube to the sill, 7) layers of **tephra** (ash) from previous eruptions, 8) & 9) layers of lava erupting from the vent and flowing down the sides of the volcano, 10) the **crater** at the top of the volcano (formed as an eruption blows off the top of the volcano), 11) layers of lava and tephra on 12), a parasitic cone. A **parasitic cone** is 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]. 13) the vents of the parasite and the main volcano, 14) the rim of the crater, 15) clouds of ash blown into the sky by the eruption; this settles back onto the volcano and surrounding land.



Figure 4.5.9: Mt. Shasta in California with Shastina, its parasitic cone



Figure 4.5.10: Oregon's Crater Lake was formed about 7700 years ago after the eruption of Mount Mazama.

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

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

Shield Volcano



Figure 4.5.11: Kilauea in Hawaii.

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 eruptions are small, localized, and predictable.



Figure 4.5.12: 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.

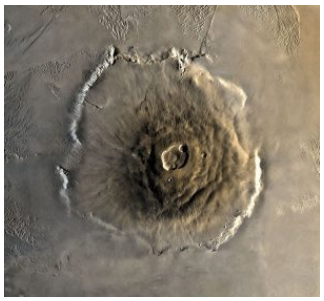


Figure 4.5.13: Olympus Mons, an enormous shield volcano on Mars, the largest volcano in the solar system.

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 4.5.14: Volcanic fissure and flow, which could eventually form a lava tube.

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 Kiluaea eruption included fissures associated with the lava flows. Some fissures are caused by the 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 4.5.15: Devils Tower in Wyoming has columnar jointing.



Figure 4.5.16: Columnar jointing on Giant's Causeway in Ireland.

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.

Stratovolcano



Figure 4.5.17: Mount Rainier towers over Tacoma, Washington.

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 4.5.18: 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].

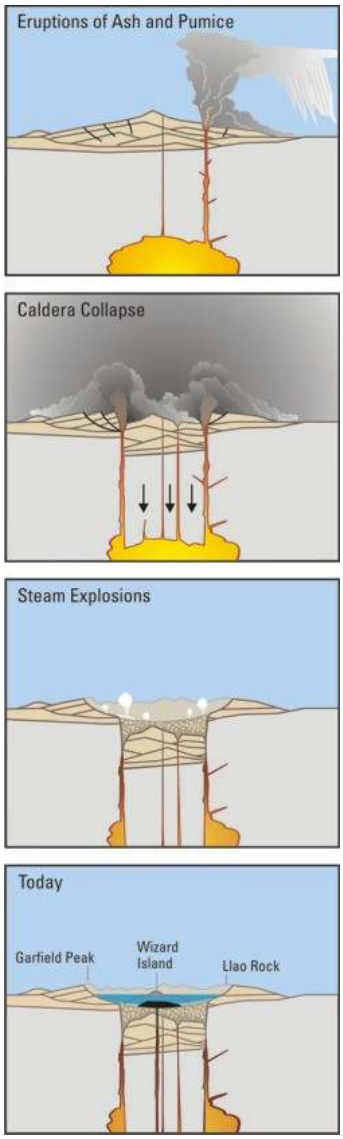


Figure 4.5.19: Timeline of events at Mount Mazama.



Figure 4.5.20: Wizard Island sits in the caldera at Crater Lake.

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 volatiles content.

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 6,800 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].

Cinder Cone



Figure 4.5.21: Sunset Crater, Arizona is a 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 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.



Figure 4.5.22: Soon after the birth of Parícutin in 1943.



Figure 4.5.23: Lava from Parícutin covered the local church and destroyed the town of San Juan, Mexico

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 (1,200 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.

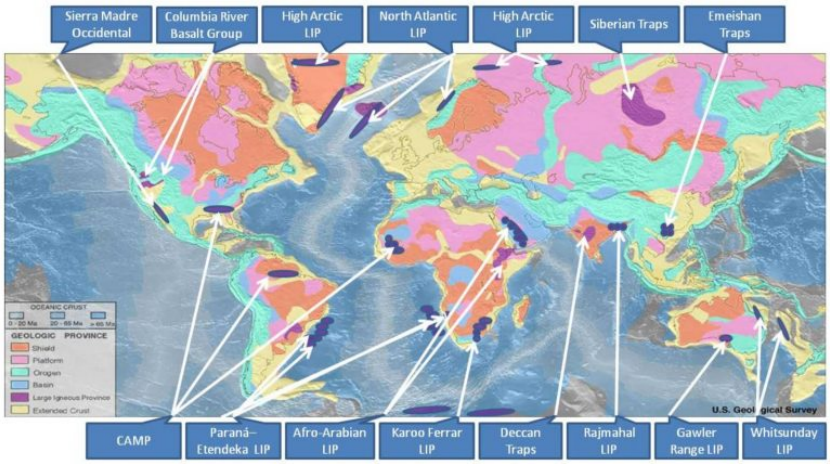
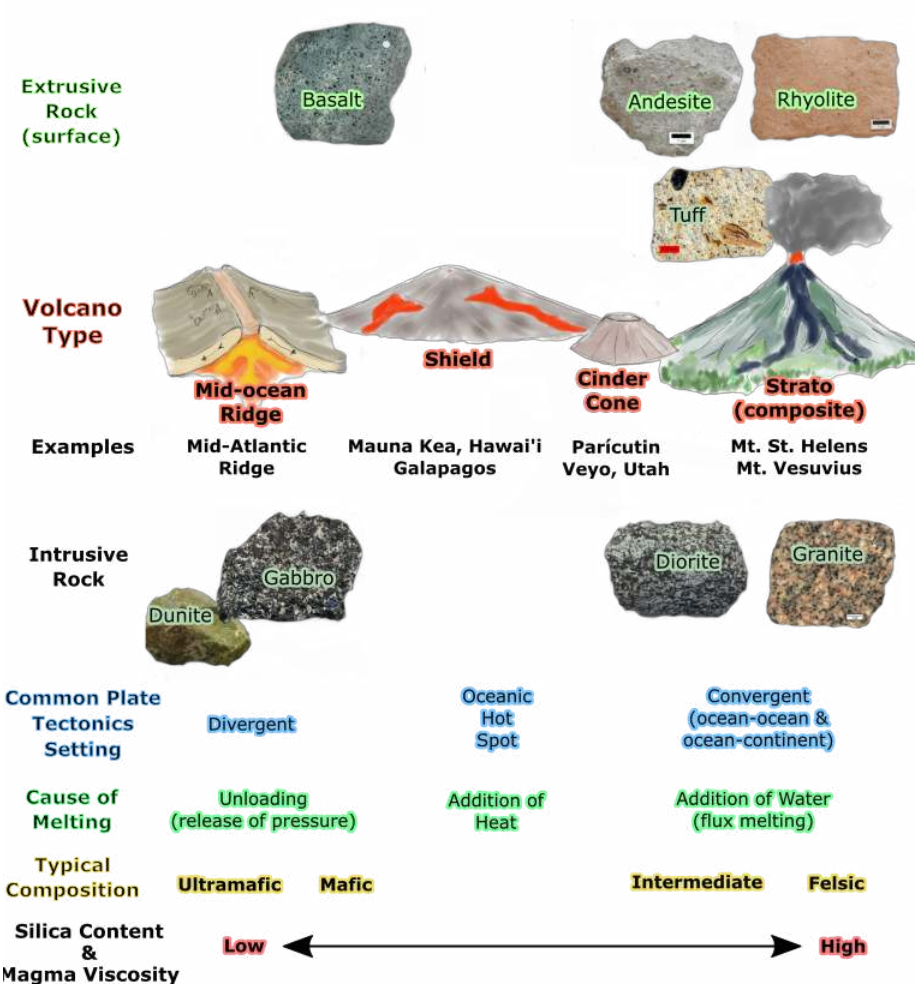


Figure 4.5.24: World map of flood basalts. Note the largest is the Siberian Traps

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 (see chapter 8).



Igneous rock types and related volcano types. Mid-ocean ridges and shield volcanoes represent more mafic compositions, and strato (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.

Volcanic Hazards and Monitoring

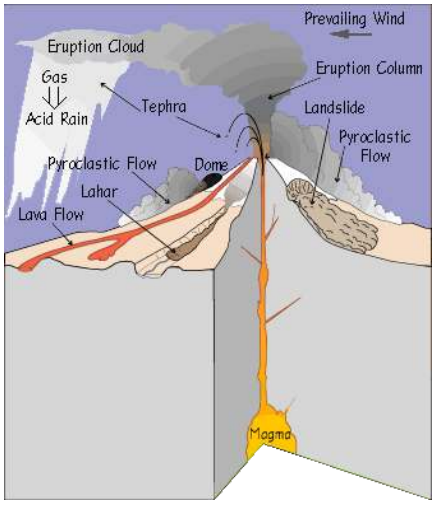


Figure 4.5.25: General diagram of volcanic hazards.

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 Saint Helens (Washington, United States) erupted with an explosion and landslide that removed the upper 400 m (1,300 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 kph (50-80 mph), flattening trees and ejecting clouds of ash into the air. The explosive eruption killed 57 people [24].



Figure 4.5.26: Human remains from the 79 CE eruption of Vesuvius.

In 79 AD, 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 4.5.27: Mount St. Helens, the day before the May 18th, 1980 eruption.



Figure 4.5.28: Picture of 4 months after the major eruption of Mount St. Helens.

Pyroclastic Flows



Figure 4.5.29: The material coming down from the eruption column onto the flanks is a pyroclastic flow.

The most dangerous volcanic hazard are **pyroclastic flows**. These flows are a mix of lava blocks, pumice, ash, and hot gases between 200°C-700°C (400°F-1,300°F). The turbulent cloud of ash and gas races down the steep flanks at high speeds up to 193 kph (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.



Figure 4.5.30: The remains of St. Pierre.

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]. 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 [28].

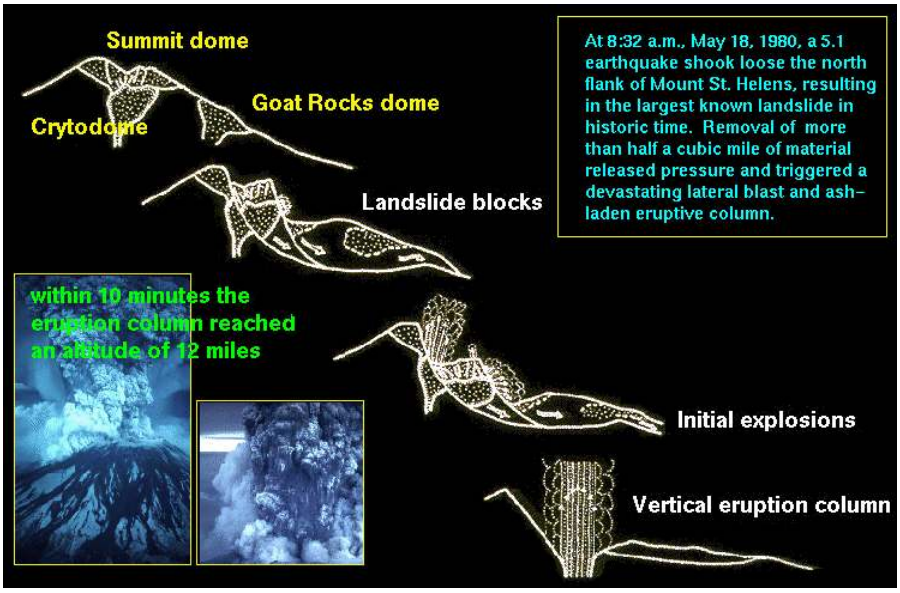


Figure 4.5.31: Sequence of events for Mount St. Helens, May 18th, 1980. Note that an earthquake caused a landslide, which caused the "uncorking" of the mountain and started the eruption.

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 kph (100-180 mph).

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 (see USGS page). 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



Figure 4.5.32: Aman sweeps ash from an eruption of Kelud, Indonesia.

Volcanoes, especially composite volcanoes, eject large amounts of **tephra** (ejected rock materials), most notably **ash** (tephra fragments less than 0.08 inches [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.

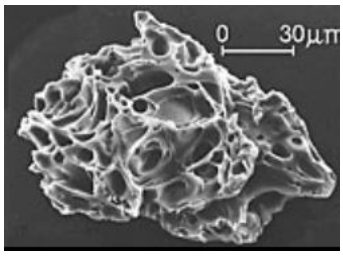


Figure 4.5.33: Micrograph of silica particle in volcanic ash. A cloud of these is capable of destroying an aircraft or automobile engine.

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].

Volcanic Ashes

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.

Lahars



Figure 4.5.34: Mud line shows the extent of lahars around Mount St. Helens

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 kph (50 mph).

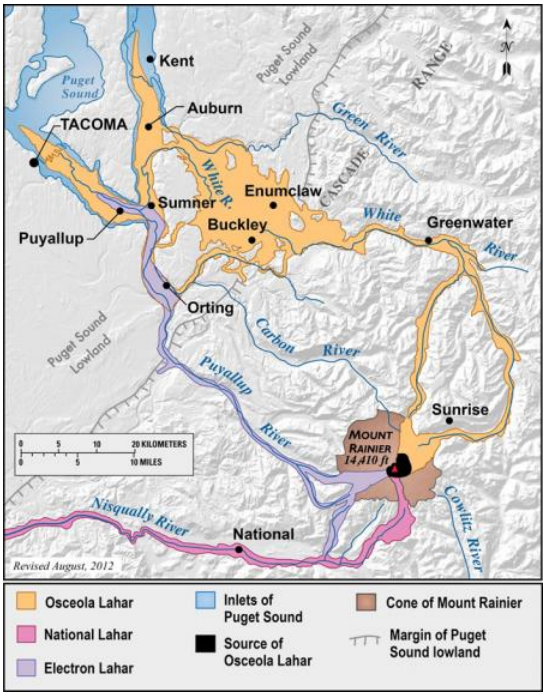


Figure 4.5.35: Old lahars around Tacoma, Washington.

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 (see map) [24]. A tragic scenario played out recently, 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.

Monitoring

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

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CHAPTER OVERVIEW

5: Weathering, Erosion, and Sedimentary Rocks

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.

5.1: Weathering, Erosion, and Soil

5.2: Sedimentary Rocks

5.3: Sedimentary Structures

5.4: Depositional Environments

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5.1: Weathering, Erosion, and Soil

Weathering is a process that turns solid rock into smaller particles, called **sediment** or **soil**. The two types of weathering are mechanical (or physical) and chemical weathering, each of which will be discussed below, though their names give clues as to what they entail.

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. 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.

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



Figure 5.1.1: The outer layer of this granite is fractured and eroding away, known as exfoliation.

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. The "cracking" is the mechanical weathering.

Frost Wedging

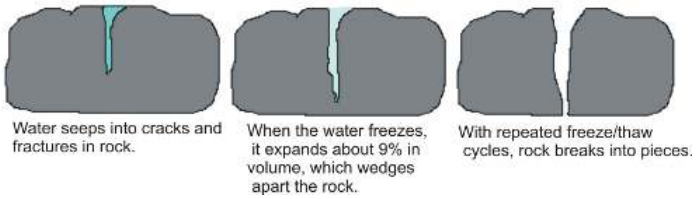


Figure 5.1.2: The process of frost wedging

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.

Root Wedging



Figure 5.1.3: 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.

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. Tunneling organisms such as earthworms, termites, and ants are biological agents that induce weathering similar to root wedging.

Salt Expansion



Figure 5.1.4: Tafoni from Salt Point, California.

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 in the cracks in the rock, forcing them apart.

Chemical Weathering

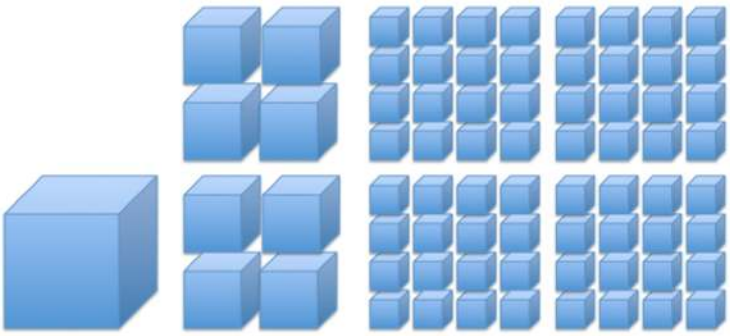


Figure 5.1.5: Each of these three groups of cubes has an equal volume. However, their surface areas are vastly different. On the left, the single cube has a length, width, and height of 4 units, giving it a surface area of $6(4 \times 4) = 48$ and a volume of $4^3 = 64$. The middle eight cubes have a length, width, and height of 2, meaning a surface area of $8(6(2 \times 2)) = 8 \times 24 = 96$. 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 $\frac{48}{64} = 0.75$ or 3:4. The center has a SA/V of $\frac{96}{64} = 1.5$, or 3:2. On the right, the SA:V is $\frac{384}{64} = 6$, or 6:1.

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 rock causing a change in composition. 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.

Carbonic Acid and Hydrolysis



Figure 5.1.6: Generic hydrolysis diagram, where the bonds in mineral in question would represent the left side of the diagram.

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.

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 (see Chapter 3, Minerals) 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



Figure 5.1.7: In this rock, a pyrite cube has dissolved (as seen with the negative “corner” impression in the rock), leaving behind small specks of gold.

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]. Regions with high humidity (airborne moisture) and precipitation experience more dissolution due to greater contact time between rocks and water.



Figure 5.1.8: Eroded karst topography in Minevre, France.



Figure 5.1.9: A formation called The Great Heart of Timpanogos in Timpanogos Cave National Monument

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

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.

Oxidation



Figure 5.1.10: 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.

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 lines 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 or hollows on exposed surfaces.

The most commonly found mineral produced by iron-oxidation reactions is red or grey **hematite**. This iron oxide 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.

Erosion



Figure 5.1.11: A hoodoo near Moab, Utah. The more resistant cap has protected the less resistant underlying layers.

Erosion is a mechanical process, usually driven by water, gravity, (see Chapter 10), wind, or ice (see Chapter 14) that removes sediment from the place of weathering. Liquid water is the main agent of erosion.



Figure 5.1.12: Grand Canyon from Mather Point.

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 resistant also create a unique-looking features called hoodoos in Bryce Canyon National Park and Goblin Valley State Park in Utah.

Soil

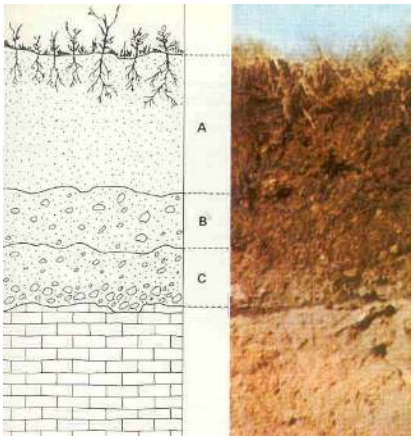


Figure 5.1.13: Sketch and picture of soil.

Soil is a combination of air, water, minerals, and organic matter that is made when weathering breaks down rock or sediment and creates weathering products in place. If the newly created material does not erode away, organisms can access the mineral content of the material. These organisms turn minerals, water, and atmospheric gases into organic substances that contribute to the soil.

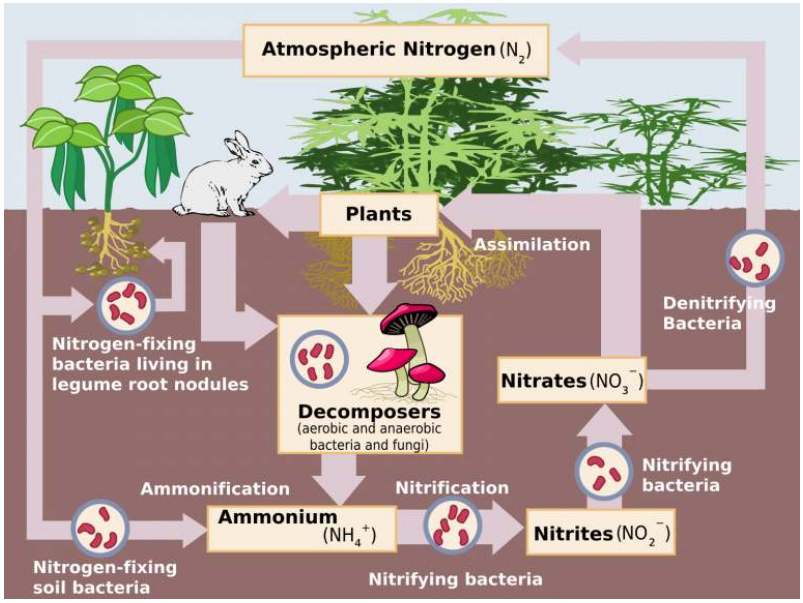


Figure 5.1.14: Schematic of the nitrogen cycle.



Figure 5.1.15: Agricultural terracing, as made by the Inca culture from the Andes, helps reduce erosion and promote soil formation, leading to better farming practices.

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

1. The mineralogy of the parent material
2. Relief or slope
3. Amount of time that it has to form
4. Climate
5. The 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 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].

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. Each soil horizon reflects its organic material and mineral sediment composition. The horizons are assigned names and letters. The figure shows a simplified soil profile that uses commonly designated names and letters.

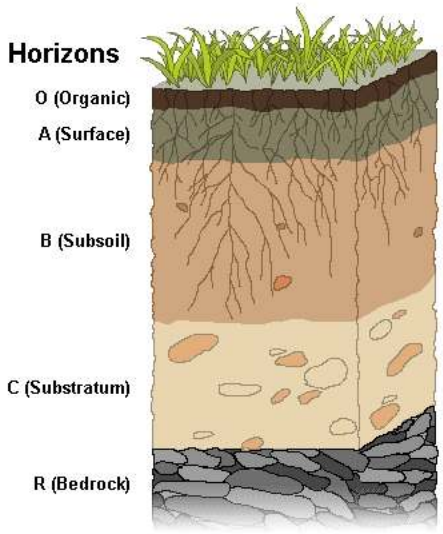


Figure 5.1.16: A simplified soil profile, showing labeled layers.

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 consists of humus (dark, organic material that forms in soil when plant and animal matter decays) 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: This layer consists of particles mixed with humus (dark, organic material that forms in soil when plant and animal matter decays); chemicals are removed from the upper layers and collect here. The subsoil is where mineral sediment is chemically weathered. The amount of organic material and the degree of weathering decrease with depth.

C Horizon: This is substratum and is a zone of weathering. Here, the parent material has just begun to weather. This layer contains no organic material.

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



Figure 5.1.17: A dust storm approaches Stratford, Texas in 1935.

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.

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5.2: Sedimentary Rocks

Sedimentary rock is classified into two main categories: clastic and chemical. **Clastic** (also named detrital) sedimentary rocks are made as sediment, derived from mechanical and chemical weathering, is collected and cemented together. 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.



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 sediment stops moving and accumulates. **Compaction** occurs when material continues to accumulate on top of the sediment layer, squeezing the grains together and driving out the water. **Cementation** is the process of cementing minerals coating the sediment grains and gluing them together into a fused rock. The "glue" often comes from groundwater which typically carries cementing agents, such as calcite, silica, or oxides, into the sediment. These cement may therefore have a different composition than the sediment grains.



Figure 5.2.1: Permineralization in petrified wood

Diagenesis is an accompanying process of lithification and is a low-temperature form of rock metamorphism (see Chapter 6, Metamorphic Rock). During diagenesis, sediments are slightly chemically altered by heat and pressure. Diagenesis may reduce the pore space, or open volume, 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.

Clastic Sedimentary Rocks

Clastic sedimentary rocks consist of preexisting sediment pieces that come from weathered rock. Most of this is mechanically weathered sediment, although some clasts may be pieces of chemical rocks. Clastic rocks are classified by grain shape, grain size, and sorting.

Grain Size

Detrital 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. 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. Clay is so fine it is smooth when you feel it with your fingers or teeth! Since silt and clay are so fine, they are commonly grouped together and called mud.

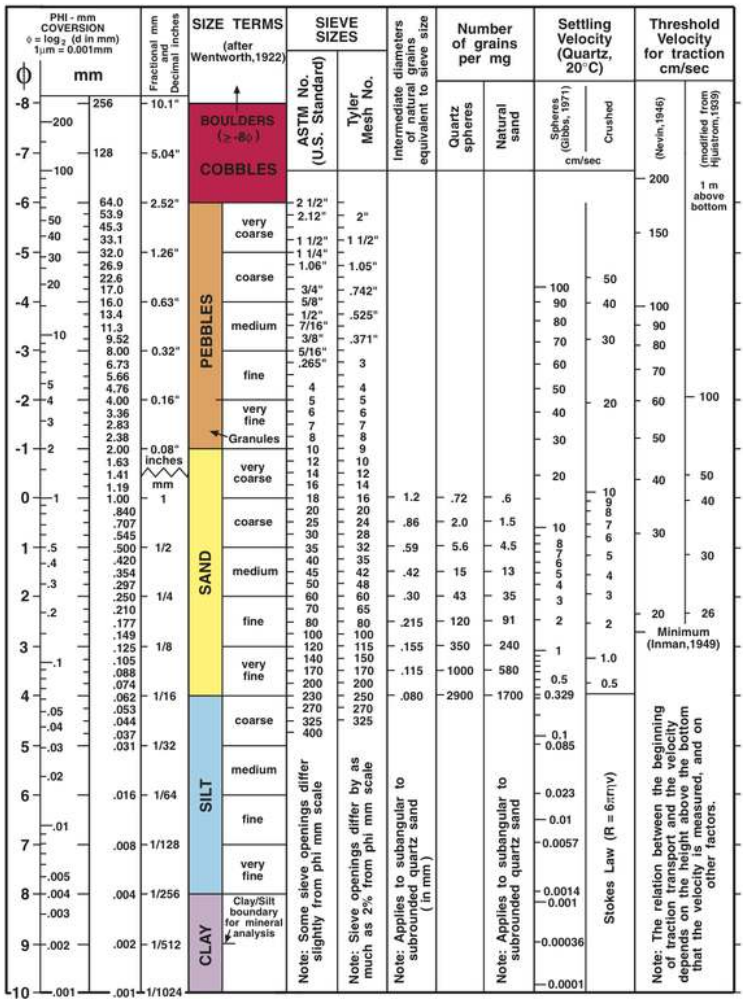


Figure 5.2.2: Size categories of sediments, known as the Wentworth scale.

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 (all grains are of similar size), and "poorly sorted" for a wide range of grain sizes (grains come in many sizes) (see figure) [11].

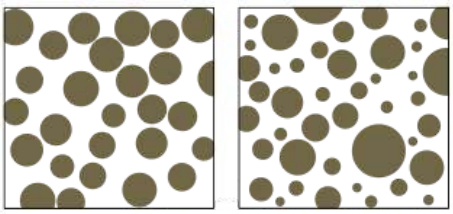


Figure 5.2.3: Well-sorted sediment (left) and Poorly-sorted sediment (right).

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 the 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.

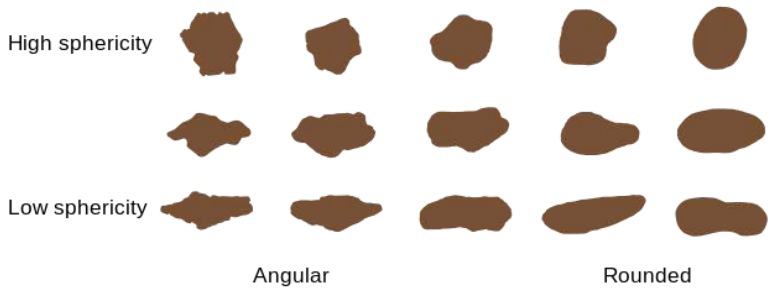


Figure 5.2.4: Degree of rounding in sediments. Sphericity refers to the spherical nature of an object, a completely different measurement unrelated to rounding.

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.

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 (see Chapter 3, Minerals). The most commonly found sediment mineral is quartz because of its low chemical reactivity and high hardness, making it resistant to weathering, and its common occurrence in continental rock. Other commonly found sediment grains include feldspar and lithic fragments. Lithic fragments are pieces of fine-grained bedrock [12], and include mud chips.

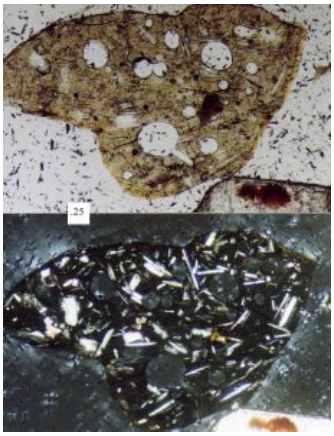


Figure 5.2.5: A sand grain made of basalt, known as a microlithic volcanic lithic fragment. The box is 0.25 mm. Top picture is plane-polarized light; bottom is cross-polarized light.

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.



Figure 5.2.6: Hawaiian beach composed of green olivine sand from weathering of nearby basaltic rock.

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], unraveling an area's geologic history, or reconstructing past supercontinents [16].

Classification of Clastic Rocks



Figure 5.2.7: Megabreccia in Titus Canyon, Death Valley National Park, California.

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 (see figure above). Both conglomerates and breccias are usually poorly sorted.

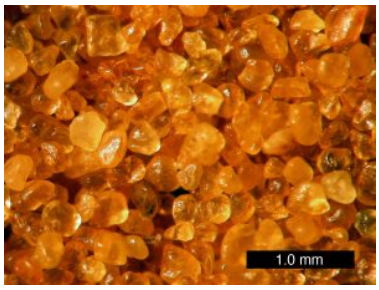


Figure 5.2.8: Enlarged image of frosted and rounded windblown sand grains

Medium-grained rocks composed mainly of sand are called **sandstone**. Sediment grains in sandstone can having 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.



Figure 5.2.9: The Rochester Shale, New York. Note the thin fissility in the layers.

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). Mudstones 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 which contain both silt and clay.



Figure 5.2.10: Claystone laminations from Glacial Lake Missoula.

Chemical 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. They are broken into three groups:

- 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 gasses, temperatures, or pressures are changing, which causes 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.
- 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

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.



<a">

Figure 5.2.11: Limestone tufa towers along the shores of Mono Lake, California.

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 5.2.12: Travertine terraces of Mammoth Hot Springs, Yellowstone National Park, USA

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

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. 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.

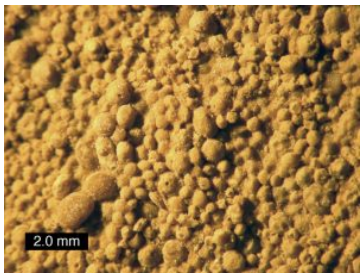


Figure 5.2.14: Ooids forming an oolite.

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.

Biochemical

Biochemical sedimentary rocks are not that different from chemical sedimentary rocks; they are also formed 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 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 5.2.15: 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 ($CaCO_3$). Solid calcite reacts with hydrochloric acid by effervescing or fizzing.



Figure 5.2.16: Close-up on coquina.

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 (see figure). **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.

Biogenetic chert forms on the deep ocean floor, created from biochemical sediment made of microscopic organic shells. This sediment, called ooze, may be 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

Under the right conditions, intact pieces of organic material or material derived from organic sources are preserved in the geologic record and are called **organic sedimentary rocks**. 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 fuels develop in areas where organic material collects in large quantities. Lush swamplands can create conditions conducive to the coal formation. Shallow-water, organic material-rich marine sediment can become highly productive petroleum and natural gas deposits.



Figure 5.2.17: Anthracite coal, the highest grade of coal.

Classification of Chemical Sedimentary Rocks

In contrast to clastic sedimentary rocks, 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) is an exception, having elaborate subclassifications.

CLASSIFICATION OF

Sedimentary Rocks

CLASTIC

Composed of pieces of rocks and minerals.

STEP 1 Determine makeup	STEP 2 Determine Grain Size	STEP 3 Rock Description	STEP 4 Rock Name
CLASTIC	Granule (> 2 mm)	Rounded rock or mineral fragments; usually poorly sorted.	*Conglomerate*
		Angular rock or minerals fragments; usually poorly sorted.	*Breccia*
	Sand (0.06 mm to 2 mm)	Mostly quartz grains; sorting and rounding variable.	*Quartz Sandstone*
		Mostly quartz with at least 25% feldspar; rock fragments common; usually poorly sorted with angular grains.	*Arkose*
		Clay, quartz, feldspars, and rock fragments; usually poorly sorted with angular grains; often dark color.	*Graywacke*
	Silt (< 0.06 mm)	Silt-sized particles are too small to identify; no layering making it not fissile.	*Siltstone*
		Silt and clay-sized particles are too small to identify; layering makes rock break in planes making it fissile.	*Shale*
	Clay (< 0.004 mm)	Clay-sized particles are too small to see.	*Claystone*

CHEMICAL, BIOCHEMICAL & ORGANIC

Made of minerals that have crystallized together, or biological fragments of shells or plants.

STEP 1 Determine makeup	STEP 2 Composition		STEP 3 Rock Description	STEP 4 Rock Name
CHEMICAL, BIOCHEMICAL & ORGANIC	Calcite	All fizz with acid	Abundant fossils. Possibly in micrite.	*Fossilliferous Limestone*
			Abundant ooids - coarse, sand-sized spheres with concentric internal layers	*Oolitic Limestone*
			Microcrystalline*; breaks with conchoidal fracture.	*Micrite*
			Shell fragments loosely cemented with a high porosity.	*Coqina*
			Microscopic fossil fragments; chalky; white color; soft.	*Chalk*
			Microcrystalline*; color banding of browns, grays, whites, and blacks.	*Travertine*
			Coarse crystals easily visible.	*Crystalline Limestone*
	Dolomite		Fizzes with acid only when powdered; fine to coarsely crystalline; may contain fossils.	*Dolostone*
	Microcrystalline Quartz		Microcrystalline*; conchoidal fracture; hardness of 7 so steel nail leaves a metal streak on the chert surface; sharp edges used for pre historic spearheads and knives; synonymous with flint.	*Chert*
	Gypsum		Crystalline; very soft with a hardness < 2, white, gray, or pink color.	*Rock Gypsum*
	Halite		Crystalline; salty taste, white or gray color; fairly soft hardness of 2.5.	*Rock Salt*
	Plant Material		Plant fragments; low density; brown to black color; often crumbly.	*Bituminous Coal*

* Microcrystalline - crystals that are visible only through a high-powered microscope.

Figure 5.2.18: Sedimentary rock identification chart

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5.3: 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 use uniformitarianism to usually 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



Figure 5.3.1: Horizontal strata in southern Utah.

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.



Figure 5.3.2: Students from the University of Wooster examine beds of Ordovician limestone in central Tennessee.

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.

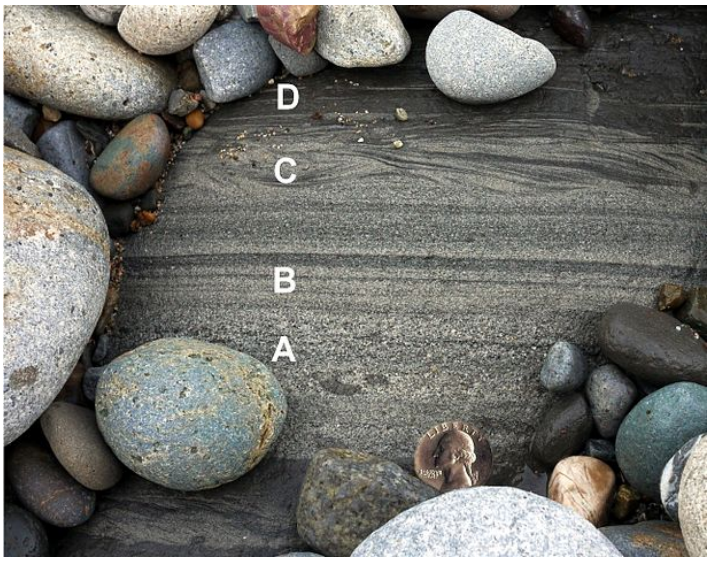


Figure 5.3.3: Image of the classic Bouma sequence. A=coarse- to fine-grained sandstone, possibly with an erosive base. B=laminated medium- to fine-grained sandstone. C=rippled fine-grained sandstone. D=laminated siltstone grading to mudstone.

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 (see figure).

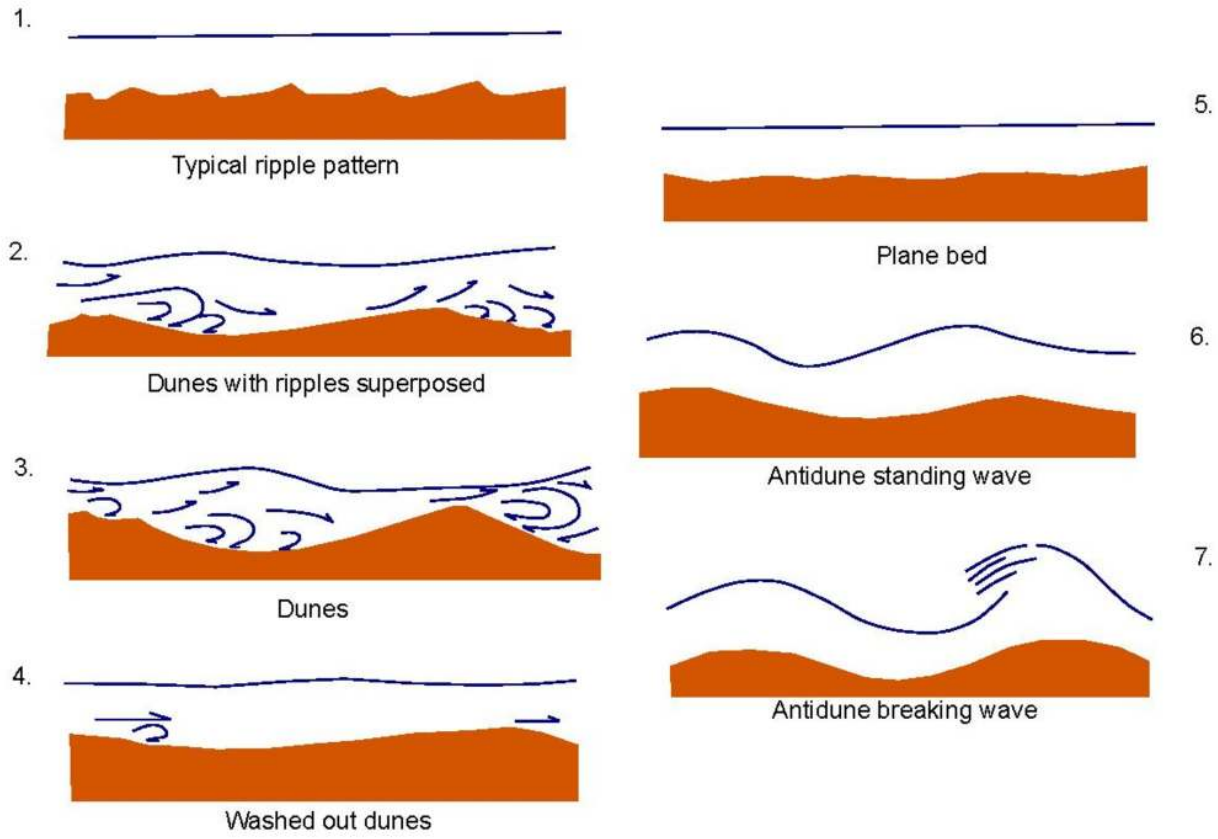


Figure 5.3.4: Bedforms from under increasing flow velocities.

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.

Ripples



Figure 5.3.5: 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.

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 5.3.6: A bidirectional flow creates this symmetrical wave ripple. From rocks in 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 (see figure).



Figure 5.3.7: Climbing ripple deposit from India.

Dunes



Figure 5.3.8: 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 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 5.3.9: The modern sand dune in Morocco.

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].

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

Bioturbation



Figure 5.3.11: Bioturbated dolomitic siltstone from Kentucky.

Bioturbation is the result of organisms burrowing through soft sediment, which disrupts the bedding layers. These tunnels are back filled 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].

Mudcracks



Figure 5.3.12: Lithified mud cracks from Maryland.

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.

Raindrop Impressions

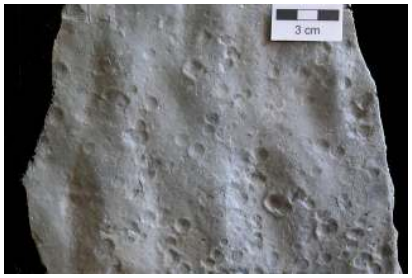


Figure 5.3.13: Mississippian raindrop impressions over wave ripples from Nova Scotia.

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].

Imbrication



Figure 5.3.14: 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.

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 (see figure). 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.

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5.4: Depositional Environments

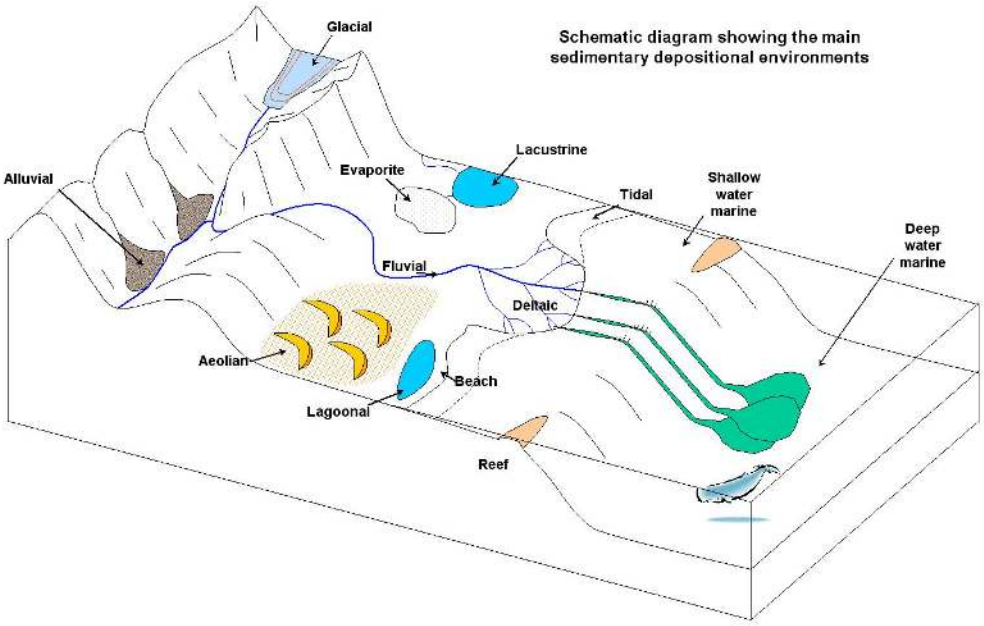


Figure 5.4.1: A representation of common depositional environments.

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 study of depositional environments is a complex endeavor; the table shows a simplified version of what to look for in the rock record. Don't feel that you need to memorize this, this is just to give you a feel for what this looks like. Read a few entries to get an idea of how this works.

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, countourites	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
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

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.

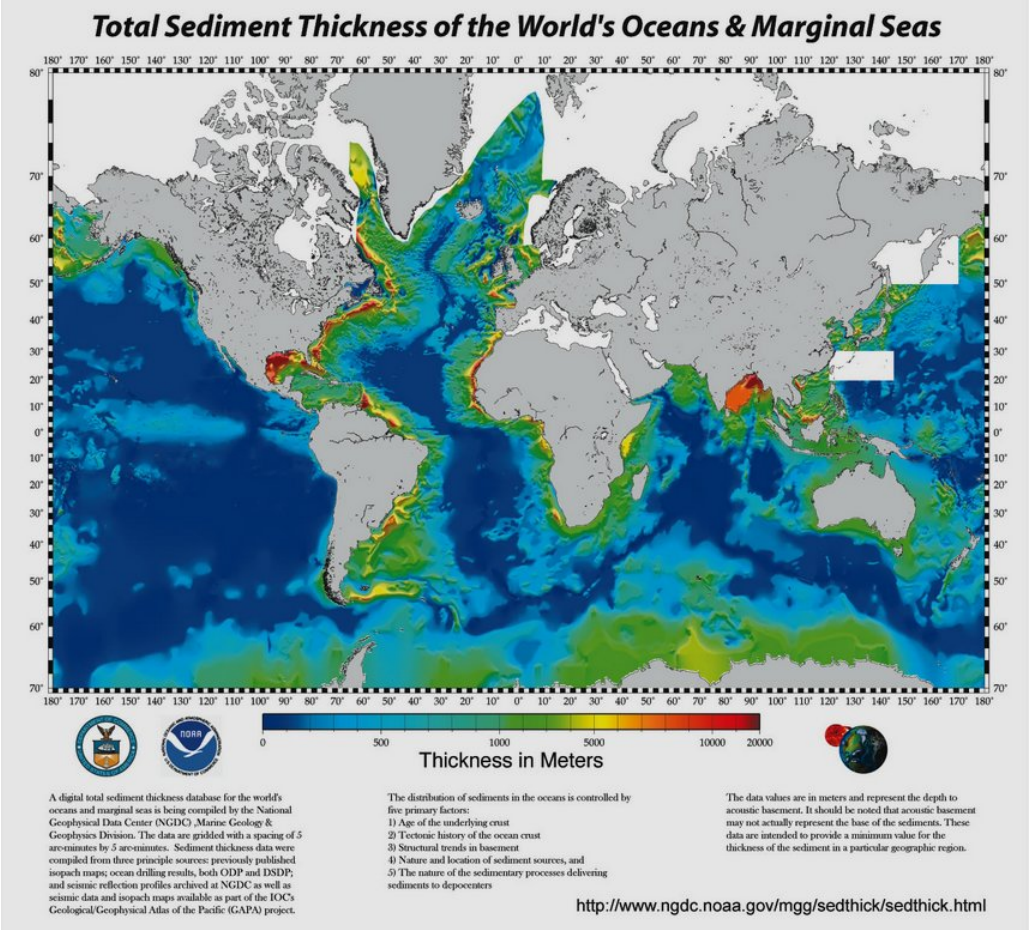


Figure 5.4.2: Marine sediment thickness. Note the lack of sediment away from the continents.

Abyssal sedimentary rocks form on the **abyssal plain** which is the vast flat ocean floor. Most abyssal plains do not experience significant fluid movement, so sedimentary rocks formed there are very fine-grained [39]. This includes things like chalk, chert and clay-based rocks like mudstone or claystone. These rocks are discussed earlier in this chapter.

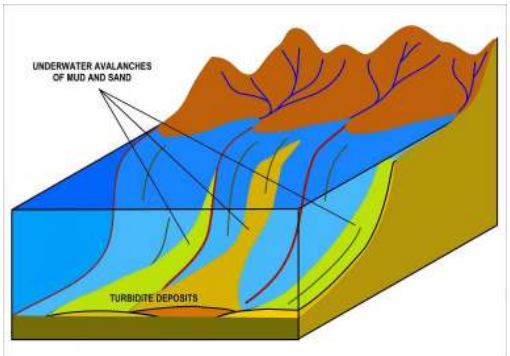


Figure 5.4.3: Turbidites inter-deposited within submarine fans.

A notable exception to the fine-grained nature of abyssal sediment are **turbidite** deposits [40]. These 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. They deposit their sediment loads as the slope decreases. This sudden flushing transports coarser sediment to the ocean floor where they are otherwise uncommon.

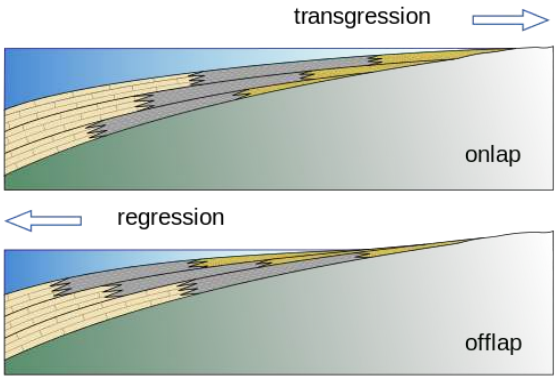


Figure 5.4.4: The rising sea levels of transgressions create onlapping sediments, regressions create offlapping.

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 (the shallow area underwater just off the coastline).

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. For a more in-depth look at these environments, see the chapter on Coastlines.

Littoral



Figure 5.4.5: Lithified heavy mineral sand (dark layers) from a beach deposit in India.

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. 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.

Tidal Flats

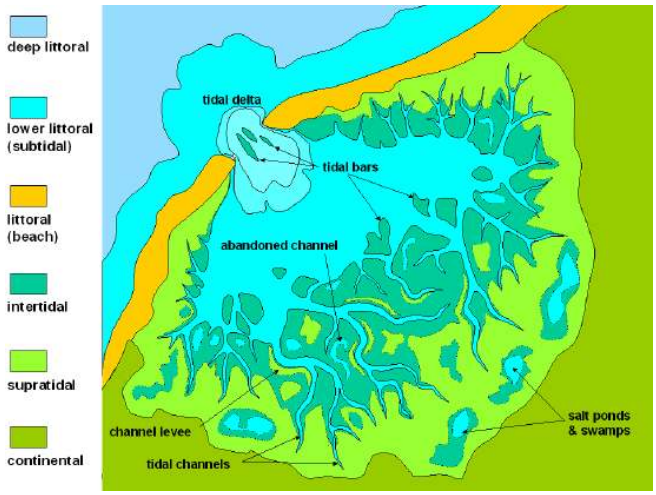


Figure 5.4.6: General diagram of a tidal flat and associated features.

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].

Reefs



Figure 5.4.7: Waterpocket fold, Capitol Reef National Park, Utah

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 (atop the continental crust) features. Capitol Reef National Park in Utah contains a topographic barrier, a reef, called the Waterpocket Fold.



Figure 5.4.8: A modern coral reef.

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 5.4.9: 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.

Lagoon



Figure 5.4.10: Kara-Bogaz Gol lagoon, Turkmenistan.

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].



Figure 5.4.11: The Nile delta, in Egypt.



Figure 5.4.12: Birdfoot river-dominated delta of the Mississippi River.

Deltas form where rivers enter lakes or oceans and are of three basic shapes: river-dominated deltas, wave-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.

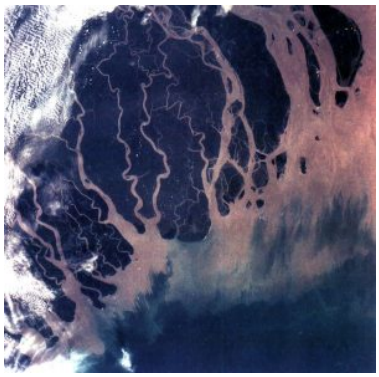


Figure 5.4.13: Tidal delta of the Ganges River.

Deltas are organized by the dominant process that controls their shape: tide-dominated, wave-dominated, or river-dominated. 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. (see figure).

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.

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.

Terrestrial

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).

Fluvial



Figure 5.4.14: The Cauto River in Cuba. Note the sinuosity in the river, which is meandering.

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 floodplain sediment away from the channel is mostly fine grained material that only gets deposited during floods.



Figure 5.4.15: The braided Waimakariri river in New Zealand.

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

Alluvial



Figure 5.4.16: An alluvial fan spreads out into a broad alluvial plain. From Red Rock Canyon State Park, California.

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.

Lacustrine



Figure 5.4.17: Oregon's Crater Lake was formed about 7700 years ago after the eruption of Mount Mazama.

Lake systems and deposits, called **lacustrine**, form via processes somewhat similar to marine deposits, but on a much smaller scale. 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 setting during a climate that was relatively wetter and cooler than that of modern Utah. Oxbow lakes, named for their curved shape, originated next to rivers.

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.

Aeolian

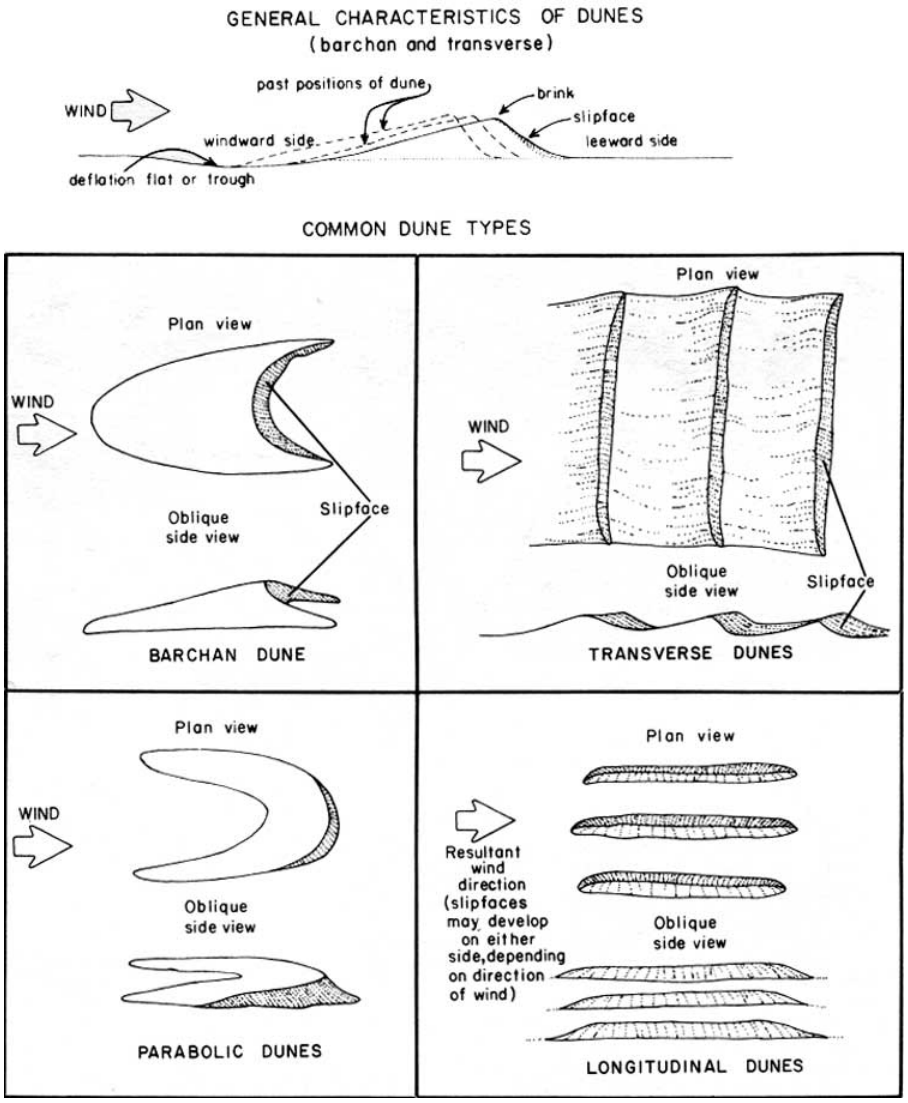


Figure 5.4.18: Formation and types of dunes.

Aeolian, sometimes spelled eolian, 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.



Figure 5.4.19: Wide range of sediments near Athabaska Glacier, Jasper National Park, Alberta, Canada.

Glacial sedimentation is very diverse and generally consists of the most poorly-sorted sediment deposits found in nature. Many glacial sediments, include very finely-pulverized rock flour along with giant boulders showing the power of ice to crush up rock and transport giant 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 (see the chapter on Glaciers).

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CHAPTER OVERVIEW

6: Metamorphic Rocks

Metamorphic rocks, *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.



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.

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 motion, metamorphic rock provides geologists with a history book of how past tectonic processes shaped our planet [1].

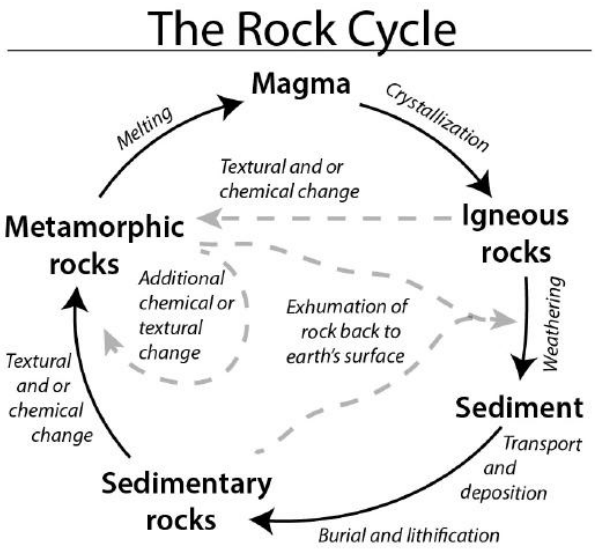


Figure 6.2: Rock cycle showing the five materials (such as igneous rocks and sediment) and the processes by which one changes into another (such as weathering). (Source: Peter Davis)

Reference

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6.1: Metamorphic Processes
6.2: Metamorphic Textures
6.3: Metamorphic Grade

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6.1: Metamorphic Processes

Metamorphism occurs when solid rock changes in composition and/or texture without the mineral crystals melting. When melt occurs and rock forms from the melt, that forms igneous rock. 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 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 metamorphic rock lie in between surficial processes (as in sedimentary rock) and magma in the rock cycle. So hot enough to cause a change, but not so hot that melt forms. Temperature increases with increasing depth in the Earth along a geothermal gradient (see Chapter 4) 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.

Confining Pressure

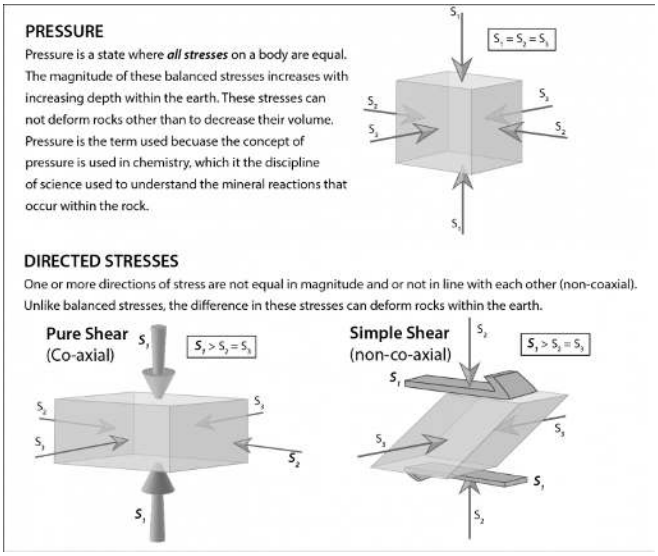


Figure 6.1.1: Difference between pressure and stress and how they deform rocks. Pressure (or confining pressure) has equal stress (forces) in all directions and increases with depth under the Earth's surface. Under directed stress, some stress directions (forces) are stronger than others, and this can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

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 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.

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].



Figure 6.1.2: Pebbles (that used to be spherical or close to spherical) in quartzite deformed by directed stress

Directed Stress

Directed stress, also called 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 at plate boundaries. 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 (note the banding you see in the rock on the right).

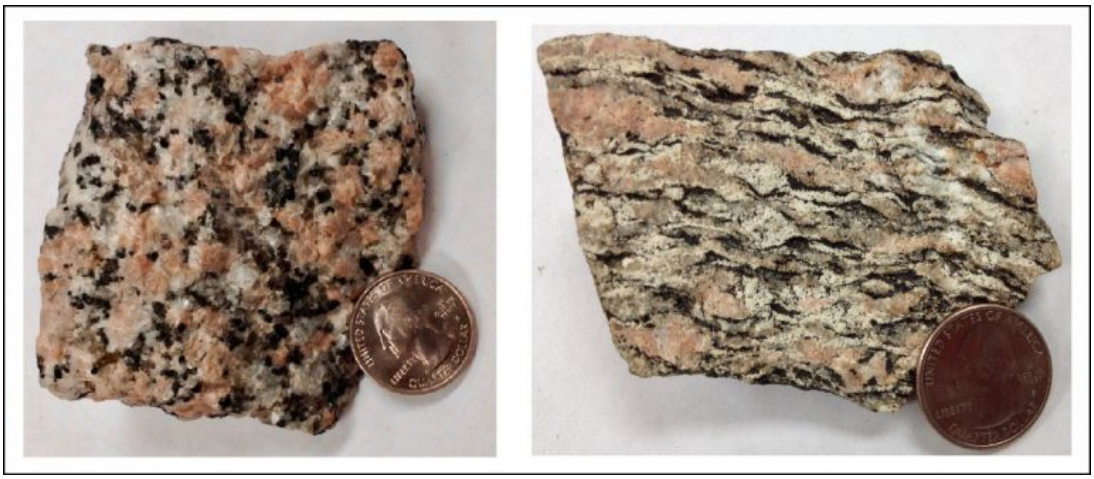


Figure 6.1.3: An igneous rock granite (left) and foliated high-temperature and high-pressure metamorphic rock gneiss (right) illustrating a metamorphic texture. (Source: Peter Davis)

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**. Water actively participates in chemical reactions and allows extra mobility of the components.

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.

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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.

METAMORPHIC ROCK IDENTIFICATION

1 IDENTIFY ROCK'S FOLIATION		2 TEXTURAL FEATURES	3 MINERAL COMPOSITION	4 ROCK NAME	5 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 texture.	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-colored. Fine or microcrystalline texture. Very hard. Color can range from gray, gray-green to black.	Microscopic dark silicates.	HORNFELS	Many rock types
		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

Created by Belinda C. Madsen for Salt Lake Community College

Metamorphic rock identification table. (Source: Belinda Madsen)

Foliation and Lineation

Foliation is a term used that describes minerals lined up in planes formed in response to directed stress. 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.

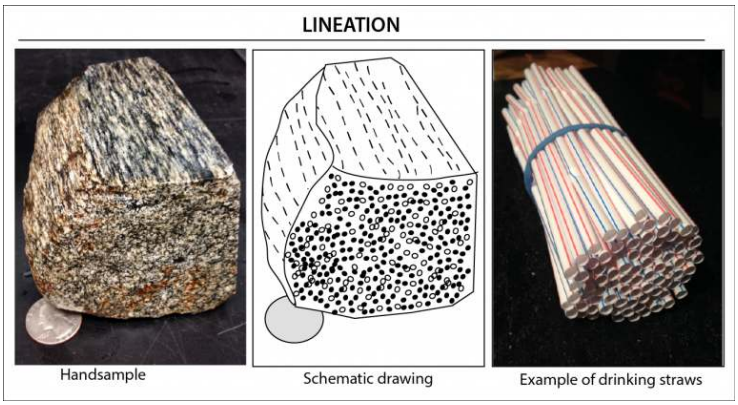


Figure 6.2.1: Example of lineation where minerals are aligned like a stack of straws or pencils. (Source: Peter Davis)

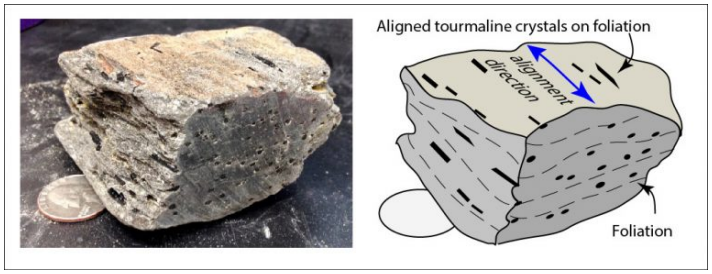


Figure 6.2.2: An example of foliation WITH lineation. (Source: Peter Davis)

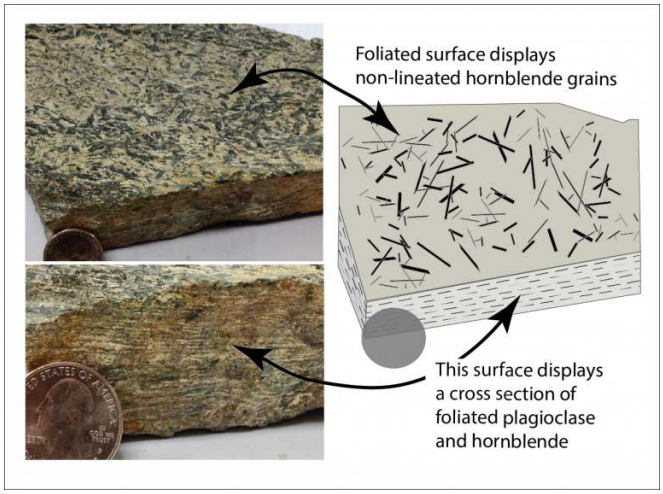


Figure 6.2.3: An example of foliation WITHOUT lineation. (Source: Peter Davis)

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 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 sedimentary layering. 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 cleavage.

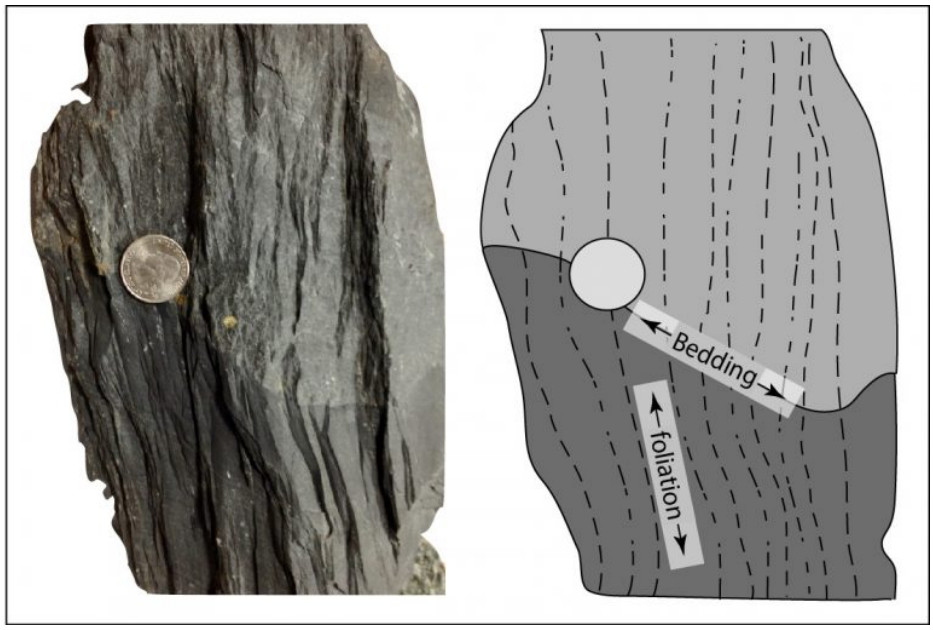


Figure 6.2.5: Foliation vs. bedding. Foliation is caused by metamorphism. Bedding is a result of sedimentary processes. They do not have to align. (Source: Peter Davis)



Figure 6.2.6: Phyllite with a small fold. (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. 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 6.2.7: Schist

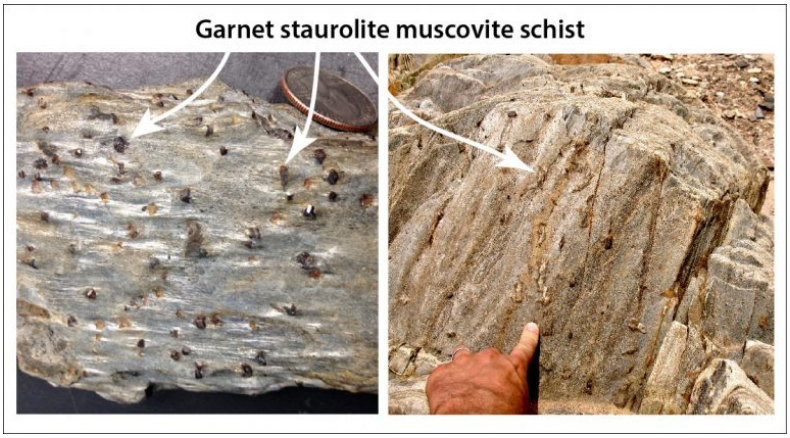


Figure 6.2.8: Garnet staurolite muscovite schist.



Figure 6.2.9: Gneiss

Gneissic banding is a metamorphic foliation in which visible silicate minerals separate into dark and light bands. These grains tend to be coarse and often folded. A rock with this texture is called **gneiss**. Since gneisses form 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 6.2.10: Migmatite

Migmatites appear as dark and light banded gneiss that may be swirled or twisted some since some minerals started to melt. Thin accumulations of light-colored rock layers can occur in a darker rock that is 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.

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 quartzites, the quartz grains from the original sandstone are enlarged and interlocked by recrystallization.



Figure 6.2.11: Two Marbles on left, Quartzite on right

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.

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 grow smaller with metamorphism and become 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.



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

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6.3: Metamorphic Grade

Metamorphic grade refers to the range of metamorphic change a rock undergoes, progressing from low (little metamorphic change) grade to high (significant metamorphic change) grade. 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 created by different pressure and temperature conditions. If a rock forms at 400 degrees and 6 kbars, it would have kyanite in it. If a rock forms at 400 degrees and 1 kbar, it would have andalusite in it. If a rock forms at 800 degrees and 6 kbars, it would have sillimanite in it. These index minerals are therefore really important!

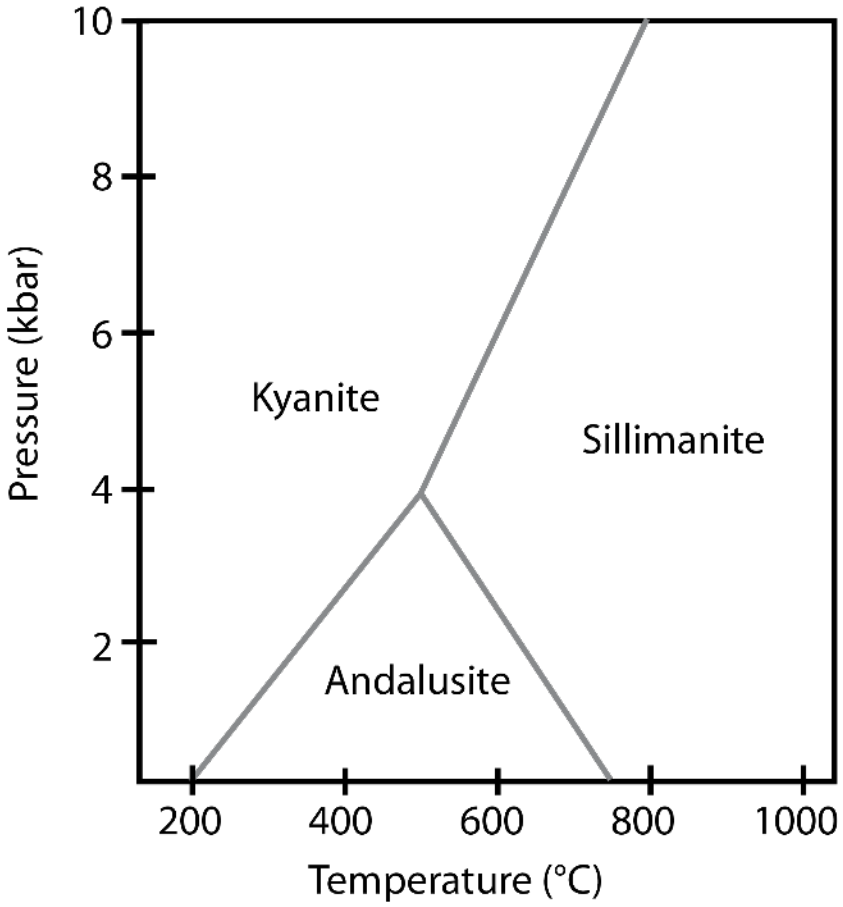


Figure 6.3.2:Al2SiO5 stability diagram. Equilibrium Temperature (T)-Pressure (P) stability of three crystalline phases (polymorphs) with the chemical composition Al2SiO5: andalusite, kyanite, and sillimanite. Notice that only one Al2SiO5 phase is stable at a time, except for the special conditions indicated by the black reaction curves for which two phases are stable. All three phases are stable together only at the special "triple point".

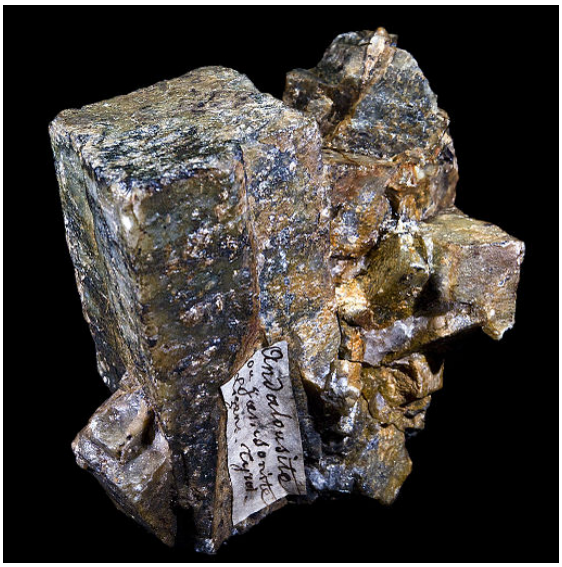


Figure 6.3.3:Andalusite



Figure 6.3.4: Kyanite



Figure 6.3.5: 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

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 (Chapter 5), burial metamorphism can cause clay minerals in shales to change to another clay mineral. Or it can cause quartz sandstone to metamorphose into the 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 (see Chapter 7), deeply buried and metamorphosed to quartzite, folded, and later exposed at the surface in the Wasatch Range today.

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.

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 6.4.1: Contact metamorphism in outcrop.

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. The lowest grade of metamorphism occurs 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



Figure 6.4.2: Blueschist

Subduction zone metamorphism is a type of regional metamorphism that occurs when a slab of oceanic crust is subducted under continental crust (see Chapter 2). 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.

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CHAPTER OVERVIEW

7: Geologic Time

The geologic time scale and basic outline of Earth's history were worked out long before we had any scientific means of assigning ages in 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]. Nineteenth-century scientists developed a time scale using Steno's principles, showing the order in which many rock units formed. Using this time scale, geologists can place all events of Earth history in order without ever knowing their numerical ages.



Figure 7.1: Nicolas Steno, c. 1670

7.1: Relative Dating
7.2: Absolute Dating
7.3: Fossils
7.4: Correlation and the Geologic Time Scale

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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. This contrasts with the other type of dating, **absolute** (or numerical) **dating**, covered in the next section in which a number age is assigned to the sediment, rock, or event—i.e., the rock is 8 million years old.

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.

EON	ERA	PERIOD	MILLIONS OF YEARS AGO
Phanerozoic	Cenozoic	Quaternary	1.6
		Tertiary	66
	Mesozoic	Cretaceous	138
		Jurassic	205
		Triassic	240
	Paleozoic	Permian	290
		Pennsylvanian	330
		Mississippian	360
		Devonian	410
		Silurian	435
		Ordovician	500
		Cambrian	570
Proterozoic	Late Proterozoic Middle Proterozoic Early Proterozoic		2500
Archean	Late Archean Middle Archean Early Archean		3800?
Pre-Archean			

Figure 7.1.1: A Geologic Time Scale

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.



Figure 7.1.2: Lower strata are older than those lying on top of them.

Principle of Superposition: In an undisturbed sequence of sedimentary layers, the layers on the bottom are the oldest and layers above them are younger.

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.



Figure 7.1.3: Lateral continuity

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. Strata that are cut by a canyon later remain continuous on either side of the canyon.



Figure 7.1.4: Dark dike cutting across older rocks, the lighter of which is younger than the grey rock.

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.

Principle of Inclusions: When one rock formation contains pieces or inclusions of another rock, the included rock is older than the host rock.

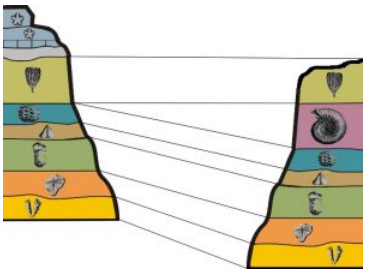


Figure 7.1.5: Fossil succession showing the correlation among strata.

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: The Grand Canyon of Arizona

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.

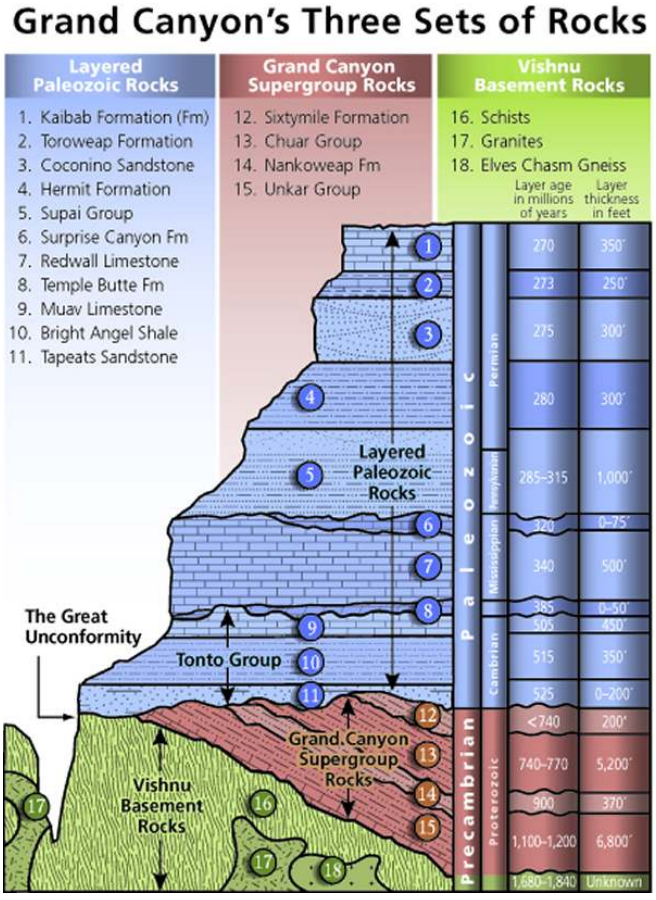


Figure 7.1.7: The rocks of the Grand Canyon

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 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.

The photo of the Grand Canyon here show 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.

Unconformities



Figure 7.1.9: 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.

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. It should be noted that the layers above and below the disconformity are parallel to one another.



Figure 7.1.10: 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 (Source: Doug Dolde).

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.

Here are three graphical illustrations of the three types of unconformity.

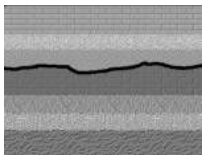


Figure 7.1.11: Disconformity

Disconformity, where is a break or stratigraphic absence between strata in an otherwise parallel sequence of strata.

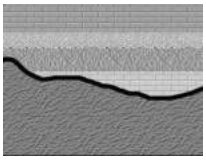


Figure 7.1.12: Nonconformity (the lower rocks are igneous or metamorphic)

Nonconformity, where sedimentary strata are deposited on crystalline (igneous or metamorphic) rocks.

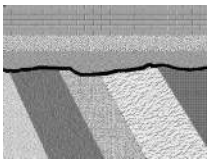


Figure 7.1.13: Angular unconformity

Angular unconformity, where sedimentary strata are deposited on a terrain developed on sedimentary strata that have been deformed by tilting, folding, and/or faulting. so that they are no longer horizontal.

Applying Relative Dating Principles

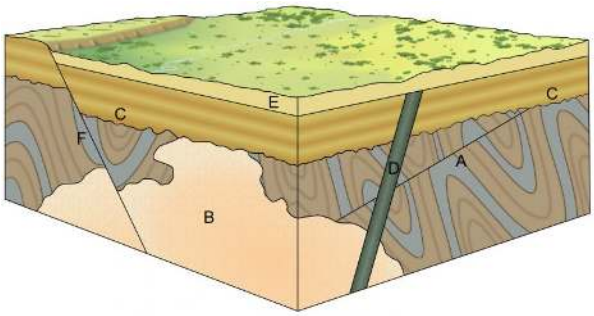


Figure 7.1.14: Block diagram to apply relative dating principles. The wavy rock is an old metamorphic gneiss, A and F are faults, B is an igneous granite, D is a basaltic dike, and C and E are sedimentary strata.

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 (see Chapter 4, Chapter 5, and Chapter 6). 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.

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7.2: Absolute Dating



Figure 7.2.1: Canada's Nuvvuagittuq Greenstone Belt may have the oldest rocks and oldest evidence life on Earth, according to recent studies.

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 (see Chapter 1), early geologists surmised geological processes work slowly and the Earth is very old. Relative dating principles was 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 radioisotopic 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 radioisotopic dating is the most common and therefore is the focus on this section.

Radioactive Decay

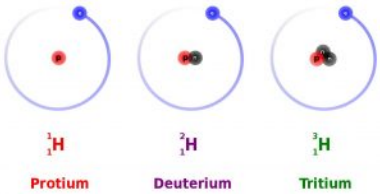


Figure 7.2.2: Three isotopes of hydrogen

All elements on the Periodic Table of Elements (see Chapter 3) contain isotopes. An isotope is an atom of an element with a different number of neutrons. For example, hydrogen (H) always has 1 proton in its nucleus (the atomic number), but the number of neutrons can vary among the isotopes (0, 1, 2). Recall that the number of neutrons added to the atomic number gives the atomic mass. When hydrogen has 1 proton and 0 neutrons it is sometimes called protium (^{1}H), when hydrogen has 1 proton and 1 neutron it is called deuterium (^{2}H), and when hydrogen has 1 proton and 2 neutrons it is called tritium (^{3}H).

Many elements have both stable and unstable isotopes. For the hydrogen example, ^{1}H and ^{2}H are stable, but ^{3}H is unstable. Unstable isotopes, called radioactive isotopes, spontaneously decay over time releasing subatomic particles or energy in a process called **radioactive decay**. When this occurs, an unstable isotope becomes a more stable isotope of another element. For example, carbon-14 (^{14}C) decays to nitrogen-14 (^{14}N).

The radioactive decay of any individual atom is a completely unpredictable and random event. However, some rock specimens have an enormous number of radioactive isotopes, perhaps trillions of atoms and this large group of radioactive isotopes do have a predictable pattern of radioactive decay. The radioactive decay of *half* of the radioactive isotopes in this group takes a specific amount of time. The time it takes for half of the atoms in a substance to decay is called the **half-life**. In other words, the half-life of an isotope is the amount of time it takes for half of a group of unstable isotopes to decay to a stable isotope. The half-life is

constant and measurable for a given radioactive isotope, so it can be used to calculate the age of a rock. For example, the half-life uranium-238 (^{238}U) is 4.5 billion years and the half-life of ^{14}C is 5,730 years.

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.



Figure 7.2.3: 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.

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 radioisotopic dating. Igneous pyroclastic layers and lava flows within a sedimentary sequence can be used to date the sequence.

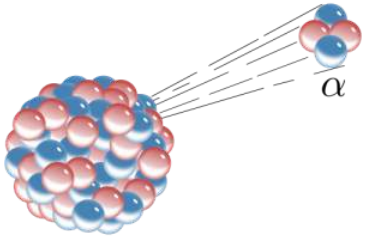


Figure 7.2.4: An alpha decay: Two protons and two neutrons leave the nucleus.

There are several ways radioactive atoms decay. Alpha decay is when an alpha particle, which consists of two protons and two neutrons, is emitted from the nucleus of an atom. The loss of four particles, in this case, two neutrons and two protons, lowers the mass of the atom by four. For example, alpha decay takes place in the unstable isotope ^{238}U, which has an atomic number of 92 (92 protons) and a mass number of 238 (total of all protons and neutrons). When ^{238}U spontaneously emits an alpha particle, it becomes thorium-234 (^{234}Th). The radioactive decay product of an element is called its **daughter isotope** and the original element is called the **parent isotope**. In this case, ^{238}U is the parent isotope and ^{234}Th is the daughter isotope. The half-life of ^{238}U is 4.5 billion years, i.e., the time it takes for half of the parent isotope atoms to decay into the daughter isotope. This isotope of uranium, ^{238}U, can be used for absolute dating the oldest materials found on Earth, and even meteorites and materials from the earliest events in our solar system.

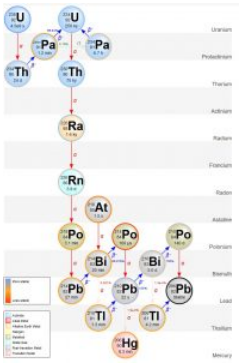


Figure 7.2.5: Decay chain of U-238 to stable Pb-206 through a series of alpha and beta decays.

It is important to note that ^{234}Th is also not stable. It continues to decay as seen in the figure above until it forms the stable daughter isotope, lead-206 (^{206}Pb). Below is a table of some of the more commonly-used radioactive dating isotopes and their half-lives.

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	5,730 years

Some common isotopes used for radioisotopic dating [7; 8].

Radioisotopic Dating

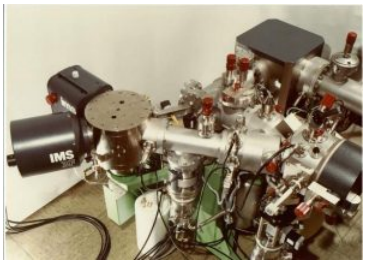


Figure 7.2.6: Mass spectrometer instrument

For a given sample of rock, how is the dating procedure carried out? The parent and daughter isotopes are separated out of the mineral using chemical extraction. In the case of uranium, ^{238}U and ^{235}U isotopes are separated out together, as are the ^{206}Pb and ^{207}Pb with an instrument called a mass spectrometer [9].

Here is a simple example of age calculation using the daughter-to-parent ratio of isotopes. When the mineral initially forms, it consists of 0% daughter and 100% parent isotope, so the daughter-to-parent ratio (D/P) is 0. After one half-life, half the parent has decayed so there is 50% daughter and 50% parent, a 50/50 ratio, with D/P = 1. After two half-lives, there is 75% daughter and 25% parent (75/25 ratio) and D/P = 3. This can be further calculated for a series of half-lives as shown in the table. The table does not show more than 10 half-lives because, after about 10 half-lives, the amount of remaining parent is so small it becomes too difficult to accurately measure via chemical analysis. Modern applications of this method have achieved remarkable accuracies of plus or minus two million years in 2.5 billion years (that's ±0.055%) [10]. Applying the uranium/lead technique in any given sample analysis provides two separate clocks running at the same time, ^{238}U and ^{235}U. The existence of these two clocks in the same sample gives a cross-check between the two. Many geological samples contain multiple parent/daughter pairs, so cross-checking the clocks confirms that radioisotopic dating is highly reliable.

Half-lives (#)	Parent present (%)	Daughter present (%)	Daughter/ Parent ratio	Parent/ Daughter ratio
Start the clock	100	0	0	infinite
1	50	50	1	1
2	25	75	3	0.33
3	12.5	87.5	7	0.143
4	6.25	93.75	15	0.0667
5	3.125	96.875	31	0.0325
10	0.098	99.9	1023	0.00098

The ratio of parent to a daughter in terms of half-life.

Another radioisotopic dating method involves carbon and is useful for dating archaeologically important samples containing organic substances like wood or bone. **Radiocarbon dating**, also called carbon dating, uses the unstable isotope carbon-14 (^{14}C). While an organism is alive, the ratio of ^{14}C amounts don't really change. However, when it dies, the radiocarbon clock starts ticking as the ^{14}C decays back to ^{14}N by beta decay, which has a half-life of 5,730 years. The radiocarbon dating technique is thus useful for 57,300 years or so, about 10 half-lives back. Since rock or sediment were never living, you are not dating the rock or sediment itself but organic material found layered in with the rock or sediment. You can then infer the age of the surrounding material.

Age of the Earth

The work of Hutton and other scientists gained attention after the Renaissance (see Chapter 1), 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]. Kelvin's numbers were not reliable due to miscalculations related to unequal heat transfer within the Earth. Kelvin failed to consider pliability and convection in the Earth's mantle as a heat transfer mechanism as well. 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].

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.

Dating Geological Events

As stated above igneous rock is the best for radioisotopic dating with most elements. 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 radioisotopic dating. Igneous pyroclastic layers and lava flows within a sedimentary sequence can be used to date the sequence.

Sediment that contains carbon may be dated using radiocarbon dating, as long as the sediment is less than about 60,000 years old due to Carbon's short half life discussed above.

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7.3: Fossils



Figure 7.3.1: Archaeopteryx lithographica, specimen displayed at the Museum für Naturkunde in Berlin.

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.

Types of Preservation



Figure 7.3.2: Trilobites had a hard exoskeleton and are often preserved by permineralization.

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.

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



Figure 7.3.4: Permineralization in petrified wood

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.

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 precipitate, the organism's external shape is preserved as a cast. Sometimes internal cavities of organisms, such 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.

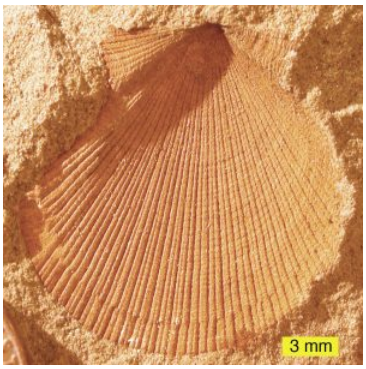


Figure 7.3.5: External mold of a clam

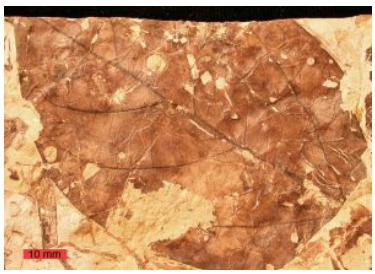


Figure 7.3.6: Carbonized leaf

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].

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: Footprints of the early crocodile Chirotherium



Figure 7.3.8: Fossil animal droppings (coprolite)

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7.4: Correlation and the Geologic Time Scale

Correlation is the process of establishing which sedimentary strata are of the same age but geographically separated. Correlation can be determined by using rock types, unique rock sequences, or index fossils. There are four main types of correlation: stratigraphic, lithostratigraphic, chronostratigraphic, and biostratigraphic.

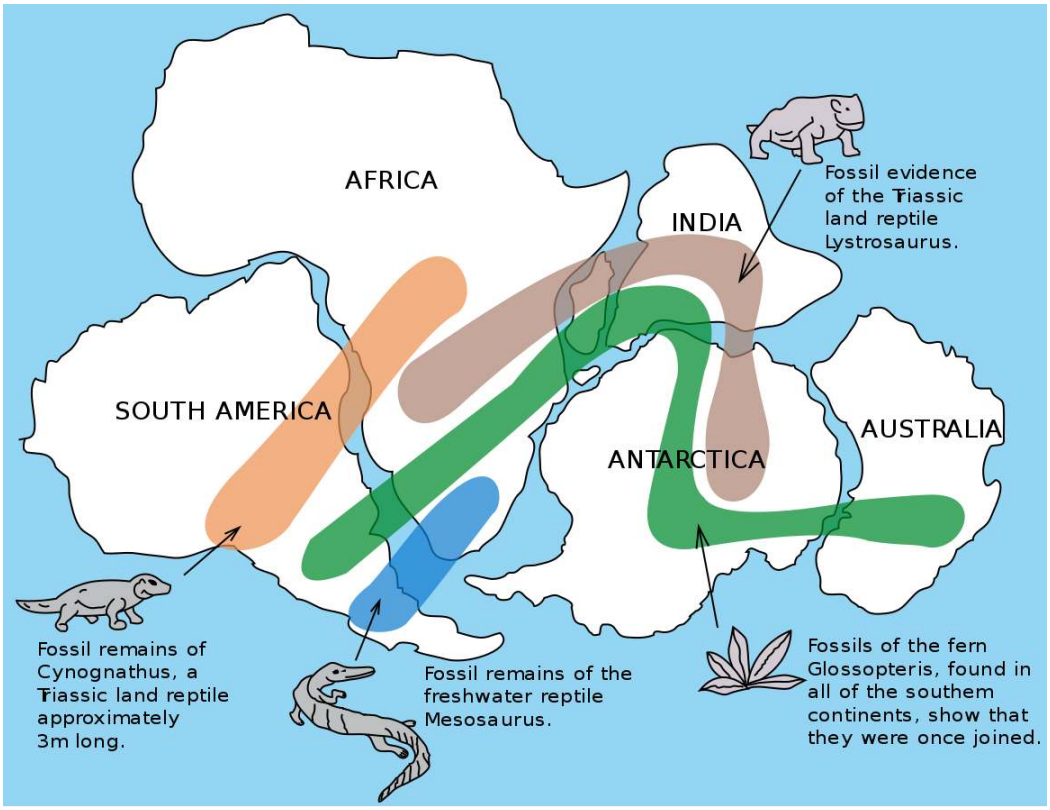


Figure 7.4.1: Image showing fossils that connect the continents of Gondwana (the southern continents of Pangea). Wegener used correlation to help develop the idea of continental drift.

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 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 is slightly different. Other strata in the Colorado Plateau and their sequence can be recognized and correlated over thousands of square miles.

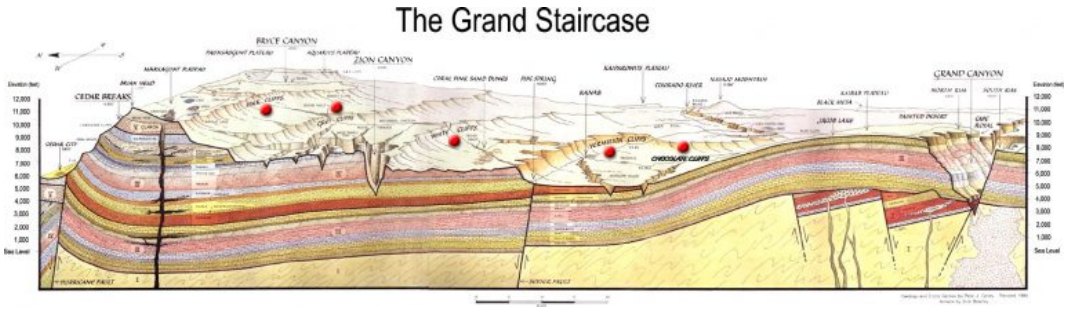


Figure 7.4.2: Correlation of Paleozoic and Mesozoic strata along the Grand Staircase from the Grand Canyon to Zion Canyon, Bryce Canyon, and Cedar Breaks. (Source: National Park Service)

Lithostratigraphic Correlation

Lithostratigraphic correlation establishes a similar age of strata based on **lithology**, which is the composition and physical properties of that strata.

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.

Chronostratigraphic Correlation

Chronostratigraphic correlation matches rocks of the same age, even though they are made of different lithologies or rocks. 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 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 light blue, 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.

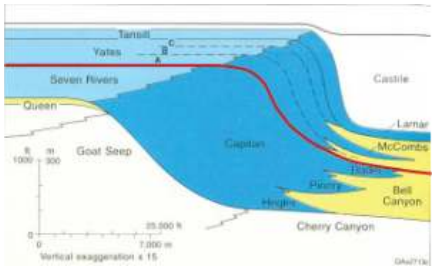


Figure 7.4.3: Cross-section of the Permian El Capitan Reef at Guadalupe National Monument, Texas. The red line shows a chronostratigraphic timeline that represents a snapshot in time in which the shallow marine lagoon/back reef area (light blue), main Capitan reef (dark blue), and deep marine siltstones (yellow) were all being deposited at the same time.

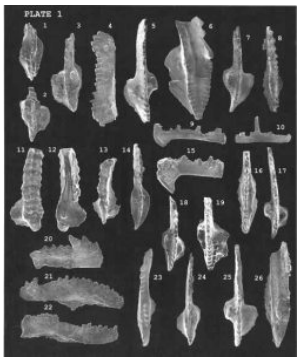


Figure 7.4.4: Conodonts

Biostratigraphic correlation uses index fossils to determine strata ages. **Index fossils** represent assemblages or groups of fossils that were uniquely present during specific intervals of geologic time. 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].

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.

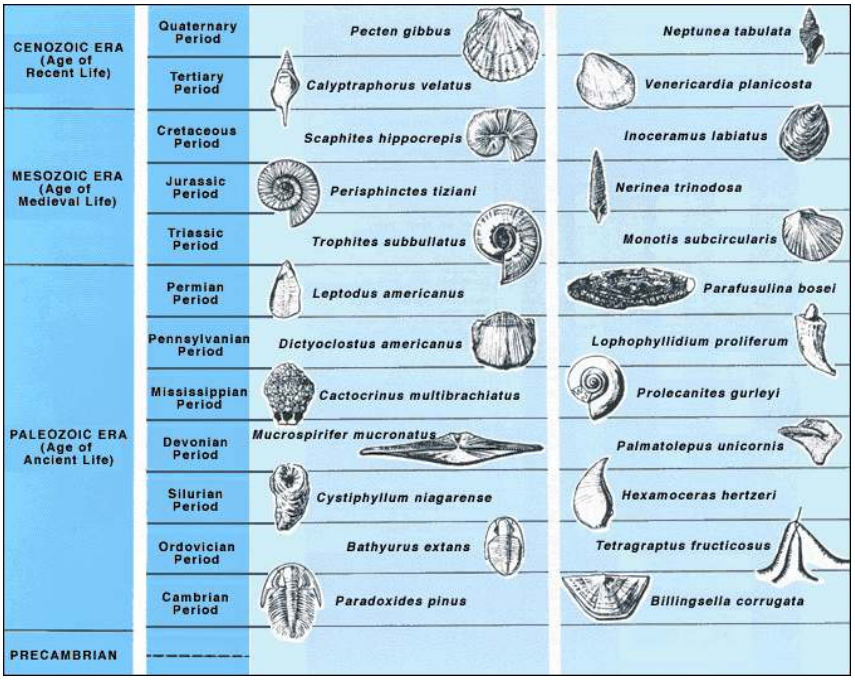


Figure 7.4.5: Index fossils used for biostratigraphic correlation

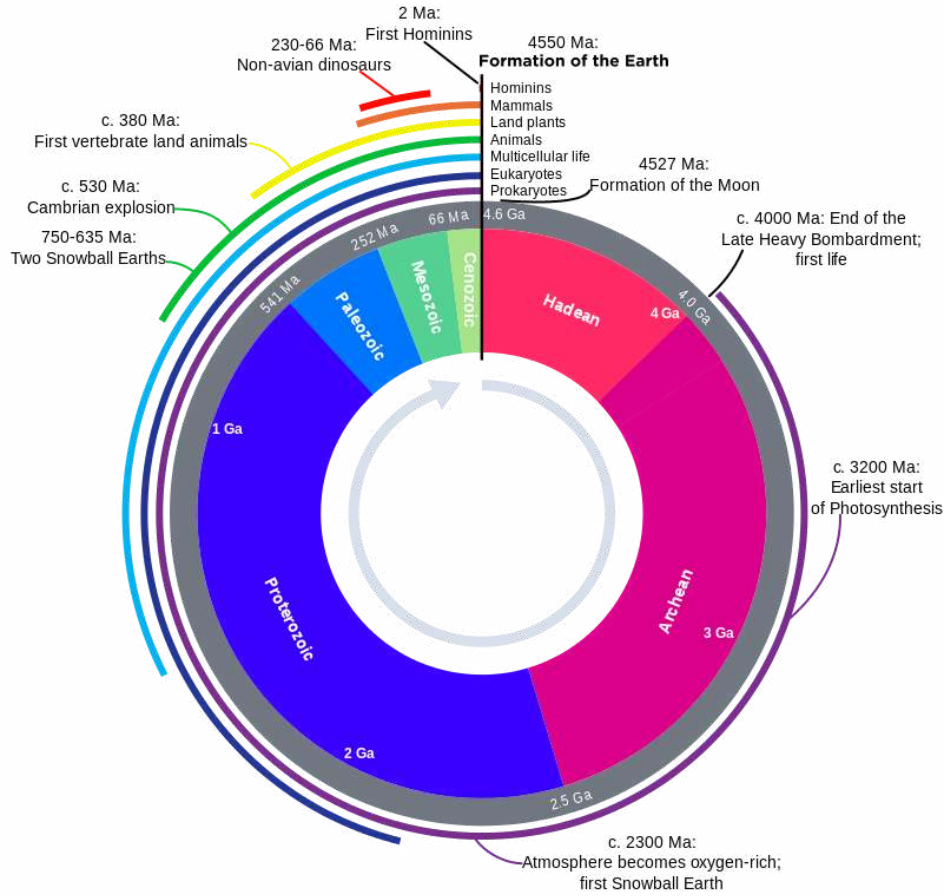


Figure 7.4.6: Geologic time on Earth, represented circularly, to show the individual time divisions and important events. Ga=billion years ago, Ma=million years ago.

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. But that does not mean there are rocks available for study for all of these time units.

The divisions are not based on regular intervals (e.g. every 100 million years). The divisions are placed at significant events in Earth's history like the extinction of the dinosaurs or the dramatic warming or cooling of the earth.

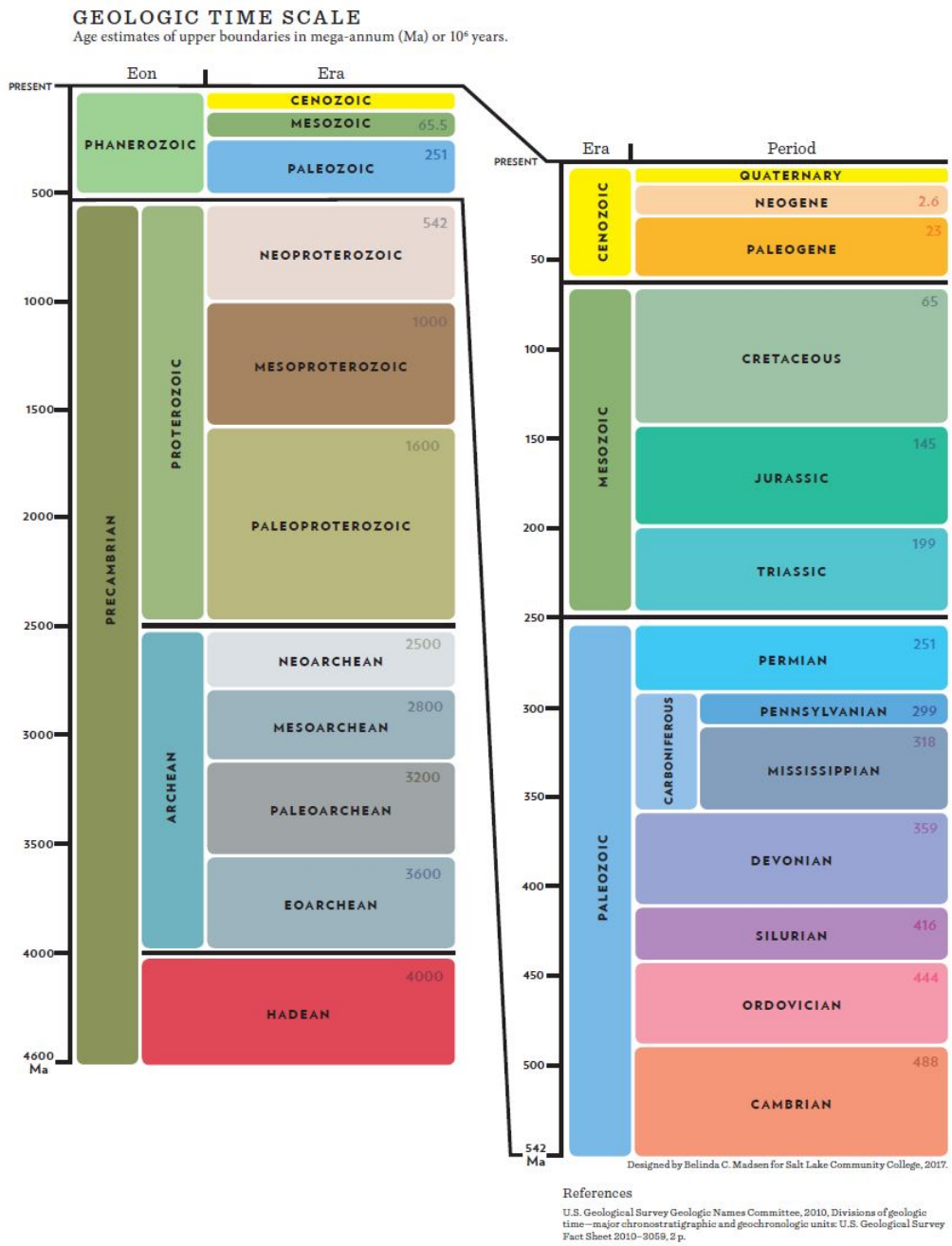


Figure 7.4.7: Geologic Time Scale with ages shown in millions of years ago (Ma).

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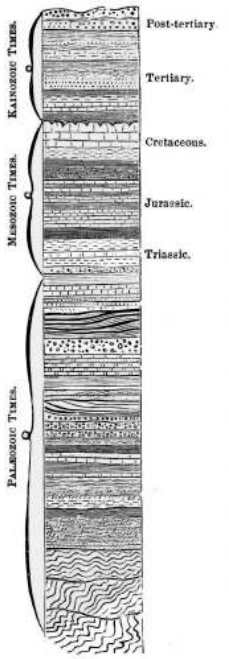


Figure 7.4.8: Names from the Geologic Time Scale applied to strata in a region

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 using 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].

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CHAPTER OVERVIEW

8: Crustal Deformation and Earthquakes

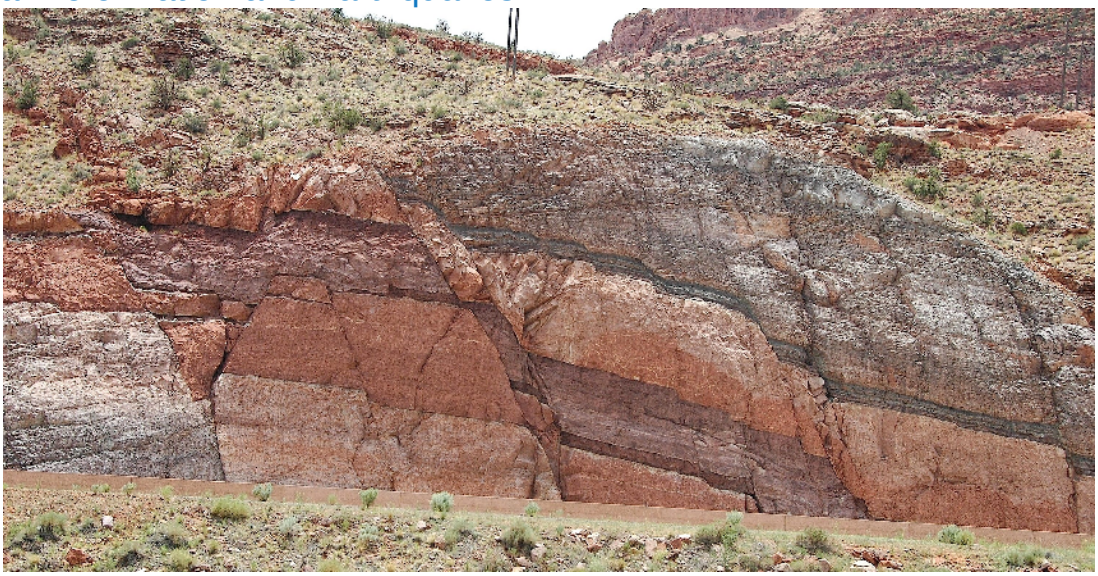


Figure 8.1: Example of normal faulting in an outcrop of the Pennsylvanian Honaker Trail Formation near Moab, Utah.

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 stress and breaks, energy is released in the form of seismic waves, commonly known as an earthquake.

8.1: Stress and Strain

8.2: Deformation

8.3: Geological Maps

8.4: Folds

8.5: Faults

8.6: Earthquake Essentials

8.7: Measuring Earthquakes

8.8: Earthquake Risk

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8.1: Stress and Strain

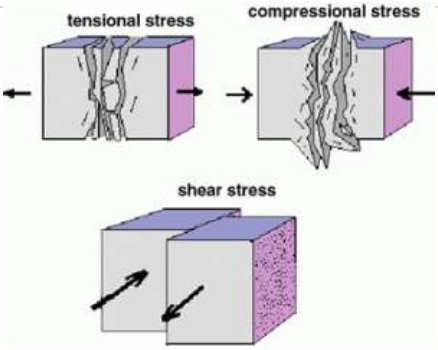


Figure 8.1.1: Types of stress. Clockwise from top left: tensional stress, compressional stress, and shear stress, and some examples of resulting strain.

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.

Table showing types of stress and resulting strain:

Type of Stress	Associated Plate Boundary type (see Ch. 2)	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

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8.2: 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**. **Brittle deformation** is another critical point of no return when rock integrity fails and the rock fractures under increasing stress.

The type of deformation a rock undergoes depends on pore pressure, strain rate, rock strength, temperature, stress intensity, time, and confining pressure. 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. Decreasing the temperature makes materials more rigid and susceptible to brittle deformation. On the other hand, heating materials makes them more ductile and less brittle. For example, heated glass can be bent and stretched.

Table showing the 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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8.3: Geological Maps

Geologic maps are two dimensional (2D) representations of geologic formations and structures at the Earth's surface, including formations, faults, folds, inclined strata, and rock types. Formations are recognizable rock units. Each formation is indicated by a color and a label.

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.

Strike and Dip

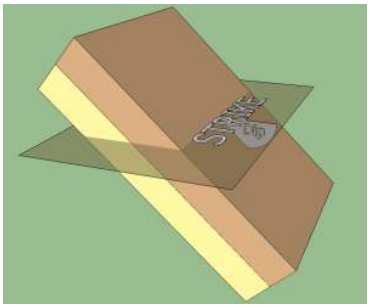


Figure 8.3.1: "Strike" and "dip" are words used to describe the orientation of rock layers with respect to North/South and Horizontal.

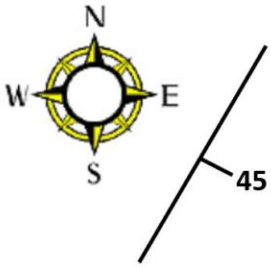


Figure 8.3.2: Attitude symbol on a geologic map (with compass directions for reference) showing strike of N30°E and dip of 45 to the SE.

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. **Strike** is the direction the bed travels. **Dip** has two components. Dip direction is the direction a ball would roll if set on the layer and released. Dip amount is how inclined from horizontal the layer is in degrees. A number next to the symbol represents the dip amount. A horizontal rock bed has a dip of 0° and a vertical bed has a dip of 90°.

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8.4: Folds

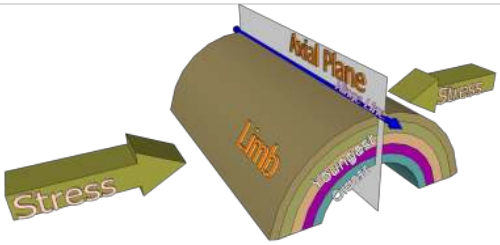


Figure 8.4.1: Model of anticline. Oldest beds are in the center and youngest on the outside. The axial plane intersects the center angle of bend. The hinge line follows the line of greatest bend, where the axial plane intersects the outside of the fold.

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 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**.

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. When the axis of the fold plunges into the ground, the fold is called a **plunging fold**. Folds are classified into several categories. The two most common are anticlines and synclines.

Anticline

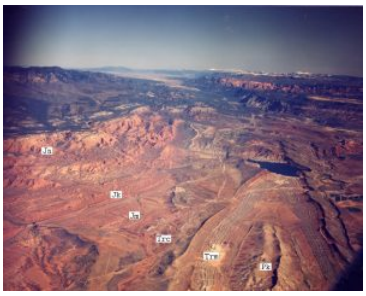


Figure 8.4.2: Oblique view of the Virgin Anticline (bottom right of photo) looking north. The anticline is plunging into the ground to the north. Units from youngest to oldest Jn = Jurassic Navajo Sandstone; Jk = Jurassic Kayenta Formation; Trc = Triassic Chinle Formation; Trm = Triassic Moenkopi; Pk = Permian Kaibab Formation.

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.

Syncline



Figure 8.4.3: Syncline in Rainbow Basin, Mojave Desert, California.

Synclines are trough-like, or U shaped, folds that are concave-downward 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.

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8.5: Faults

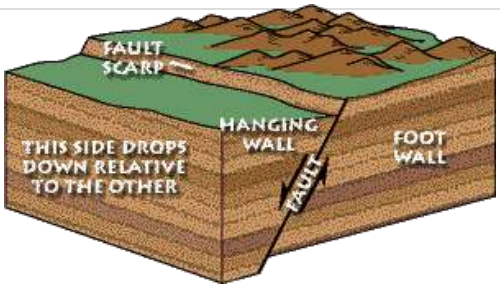


Figure 8.5.1: Common terms used for normal faults. Normal faults form when the hanging wall moves down relative to the footwall.

Faults are the places in the crust where brittle deformation occurs as two blocks of rocks move relative to one another. This breakage happens when stress is placed on cold rock such that it breaks instead of bends. These ruptures often occur at plate boundaries but can also occur in plate interiors as well.

The surface where the slip occurs is called the fault plane. The fault scarp is the offset of the surface produced where the fault breaks through to the surface (see figure above).

Normal and reverse faults, discussed in more detail below, 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 standing on the fault plane; the hanging wall would be hanging above you and your feet would be on the footwall.

A joint or fracture is a plane of brittle deformation in the rock created by the movement that is not offset. Joints can result from many processes, such as cooling, depressurizing, or folding.

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.

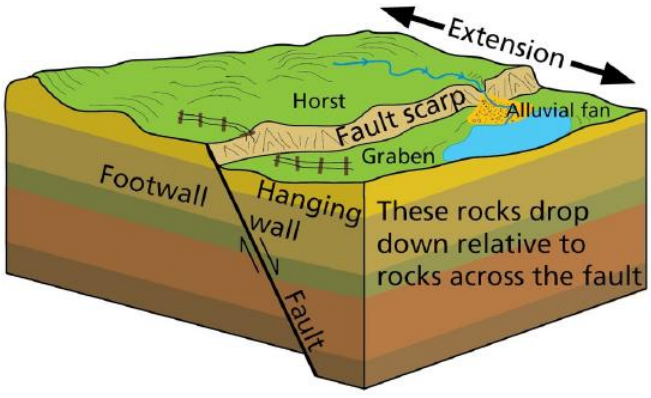


Figure 8.5.2: Block diagram from the National Parks Service showing a normal fault with the hanging wall sliding down relative to the footwall due to extension.

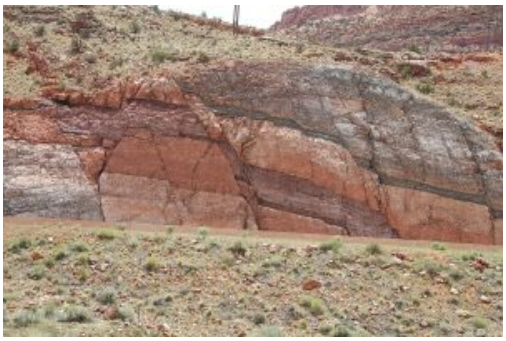


Figure 8.5.3: Example of a normal fault in an outcrop of the Pennsylvanian Honaker Trail Formation near Moab, Utah. Notice the thin gray bed on the left side of the fault is higher than it is on the right side of the fault. This is because the fault block on the right (hanging wall) has slid down.

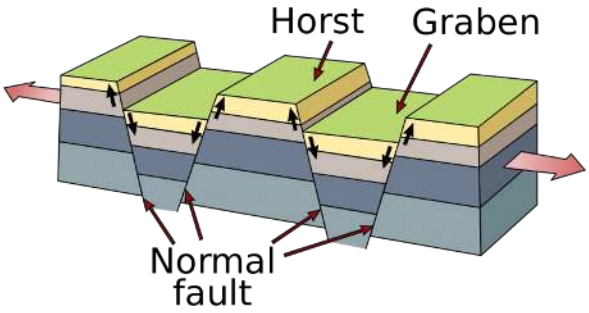


Figure 8.5.1: Faulting that occurs in the crust under tensional stress.

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 are left over 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.

Reverse Faults

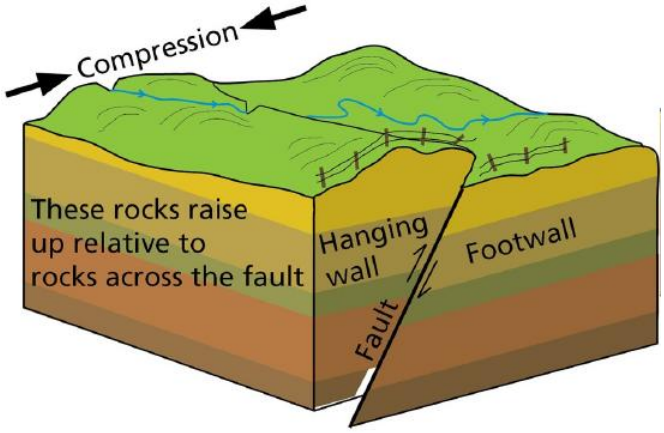


Figure 8.5.4: Block diagram from the National Parks Service showing a reverse fault with the hanging wall sliding up relative to the footwall due to compression.

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 as you can see in the diagram below where the older gray rock appears on top of the younger white rock.

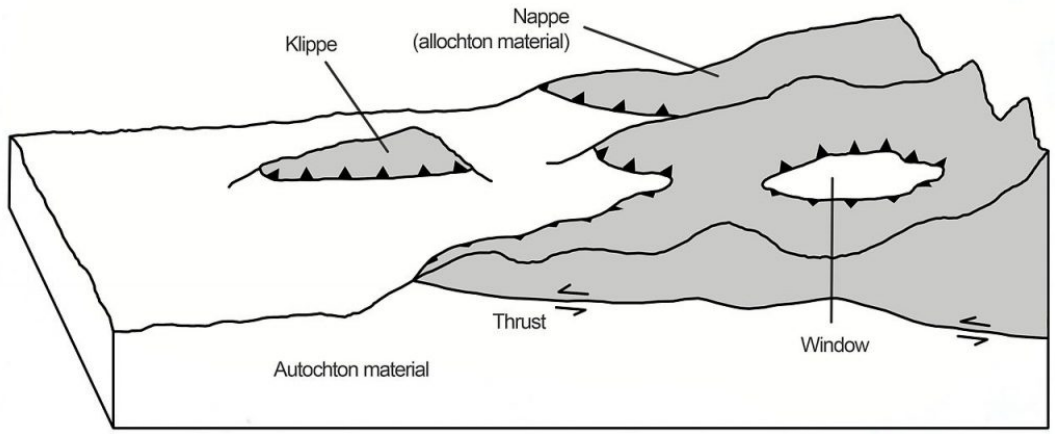


Figure 8.5.5: Terminology of thrust faults (low-angle reverse faults). A klippe is the remnant of the hangingwall (aka nappe), where the surrounding material has been eroded away. A window is where part of the hangingwall has been eroded away to expose the footwall (autochthon). Note the symbol shows flags on the overlying thrust plate.



Figure \(\PageIndex{6}\): 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. 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, or side to side.

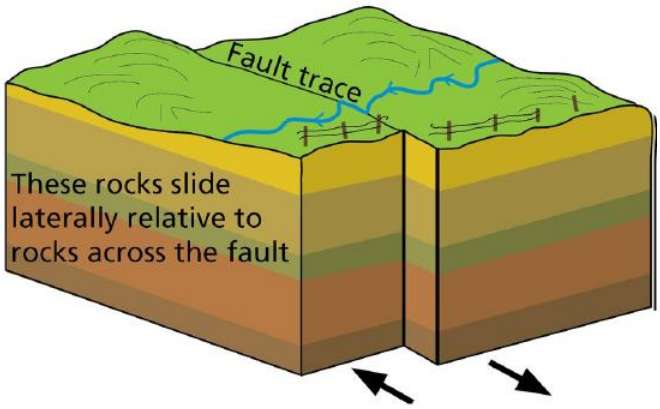


Figure 8.5.7: Block diagram from the National Parks Service showing a strike slip fault where the two blocks slide past one another.

Bends along strike-slip faults create areas of compression or tension between the sliding blocks. Releasing bends are created by tensional stresses which create features like normal faults and basins, such as the Salton Sea in California. Restraining bends are created by compressional stresses which create features like reverse faults causing mountain building, such as the San Gabriel Mountains in California.

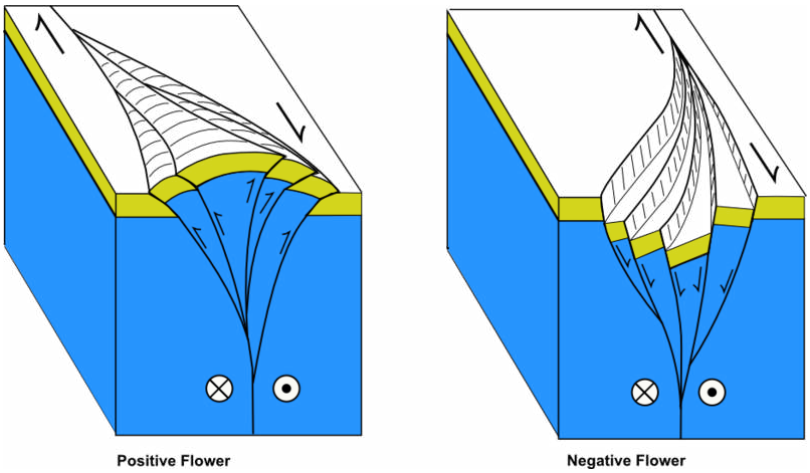


Figure 8.5.8: Structures created by bends in strike-slip faults. Depending on the relative movement in relation to the bend in the fault, these structures can create basins or mountains. On the left is a restraining bend where compression causes uplift. On the right is a releasing bend where tension causes blocks to drop down.

An example of a strike-slip fault is the San Andreas Fault, which denotes a transform boundary between the North American and Pacific plates.

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8.6: 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. The seismic energy released travels through the Earth in the form of seismic waves. Most earthquakes occur along active plate boundaries though some do occur in the middle of plates. These intraplate earthquakes (not along plate boundaries) are still poorly understood.

How Earthquakes Happen

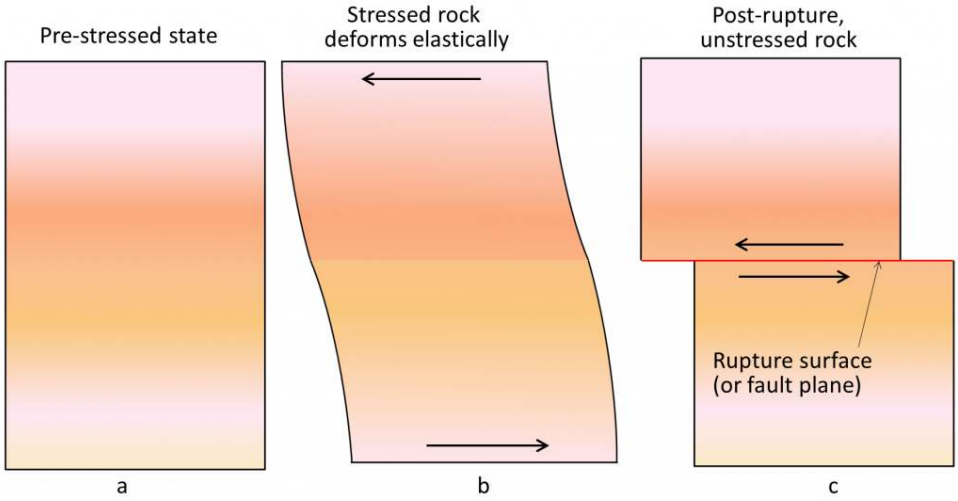


Figure 8.6.1: Process of elastic rebound: a) Undeformed state, b) accumulation of elastic strain, and c) brittle failure and release of elastic strain.

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 takes place between the fault blocks is called the **focus**. This offset propagates along the fault, which is known as the fault plane.

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 predicting them. In locations where the fault is not locked, seismic stress causes continuous, the gradual displacement between the fault blocks called fault creep. Fault creep occurs along some parts of the San Andreas Fault in 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 generally decrease over time.

Focus and Epicenter

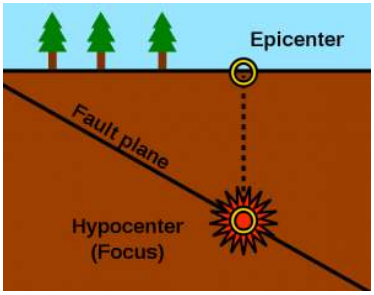
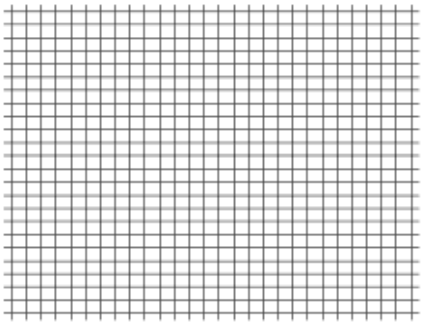


Figure 8.6.2: The hypocenter is the point along the fault plane in the subsurface from which seismic energy emanates. The epicenter is the point on a land surface vertically above the hypocenter.

The earthquake focus, also called the **hypocenter**, is the initial point of rupture. 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 just above the focus. This is the location that most news reports give because it is where the waves first hit the surface and is where people are most affected.

Seismic Waves

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.



P-waves are compressional.

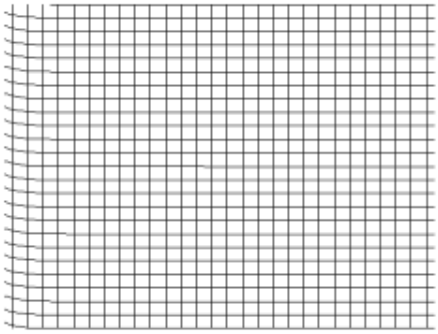


Figure 8.6.3: S waves are shear.

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**. 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. 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.

Induced Seismicity

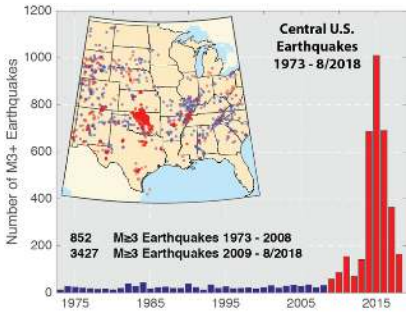


Figure 8.6.4: Frequency of earthquakes in the central United States. Note the sharp increase in the number of earthquakes from 2010 to 2015.

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, in a human-induced activity known as **induced seismicity** [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.

Reference

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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8.7: Measuring Earthquakes

Seismographs

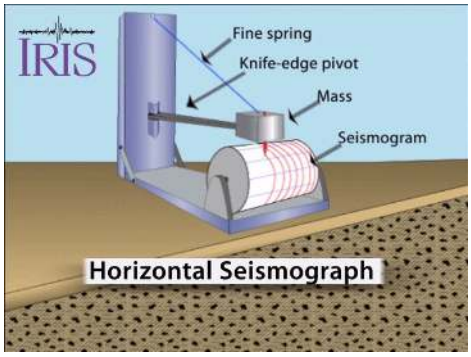


Figure 8.7.1: Animation of a horizontal seismograph.

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 has a real-time map showing the location and magnitude of recent earthquakes around the world.

To accurately study seismic waves, geologists use **seismographs** that can measure even the slightest ground vibrations. Early 20th-century seismograms use 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 moves 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.

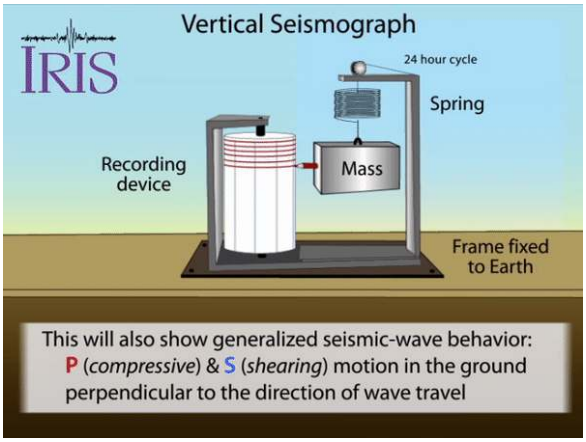


Figure 8.7.2: Animation of how a basic vertical seismograph records seismic waves.

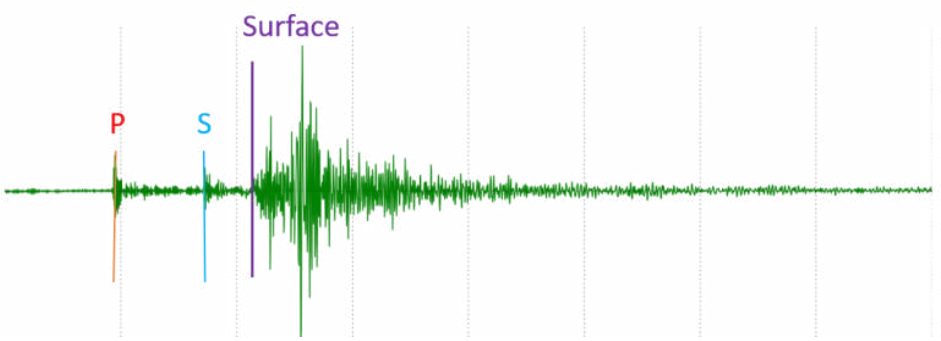


Figure 8.7.3: A seismogram showing the arrivals of the P, S, and surface waves

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 because they are the fastest, followed by S waves which are slower, and finally surface waves, which have the largest amplitude (height). 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 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 demarked where the circles intersect. This is much easier to show in a video; one is provided in your course materials.

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. Used 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.

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, even though scientific calculations now use moment magnitudes.

Mercalli Scale

The **Mercalli scale** measures the intensity of an earthquake by observing its effect on people, the environment, and the earth's surface. The effects experienced are influenced by many factors such as distance to the epicenter, quality of the construction, and the geologic materials underfoot. While it is not objective like the Richter and Moment Magnitude scales, it does give meaningful information to non scientists who want to get an idea of the size of the event. It uses an intensity rating from I-XII where I is the smallest event and XII is the largest.

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.

Table. Abridged Mercalli Scale from USGS General Interest Publication 1989-288-913.

ShakeMaps

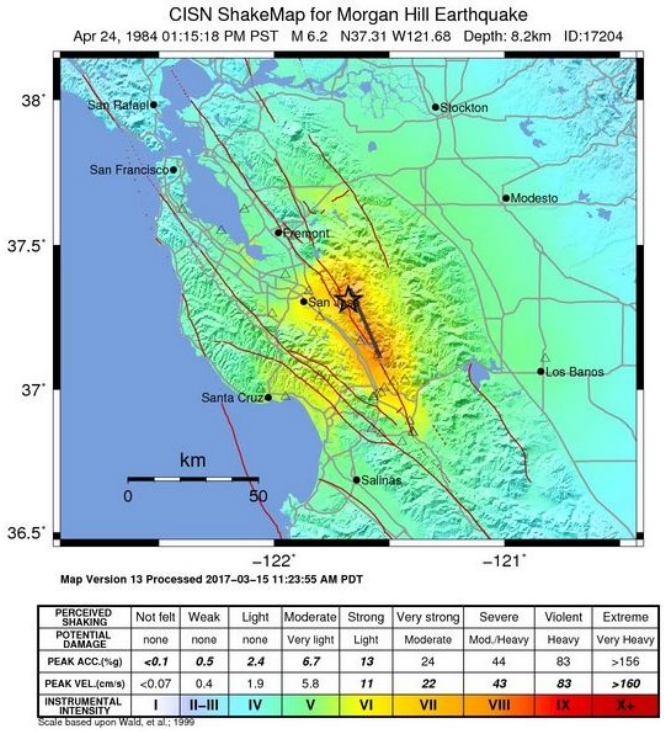


Figure 8.7.4: Example of a shake map.

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.

Reference

7. Richter CF (1935) An instrumental earthquake magnitude scale. Bulletin of the Seismological Society of America 25:1–32

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8.8: 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.

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 – IX**	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 a meatloaf, one will jiggle much more when hit by waves of the same amplitude. Solid sedimentary, igneous, or metamorphic bedrock shakes less than unconsolidated sediments.

Seismic waves move fastest through consolidated bedrock, slower through unconsolidated sediments, and slowest through unconsolidated sediments with a high water content. 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 8.8.1: 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 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.

Earthquake Distribution

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. Transform fault boundaries create moderate and large earthquakes, usually having a maximum magnitude of about 8. The San Andreas fault (California) is an example of a transform-boundary earthquake zone.

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).

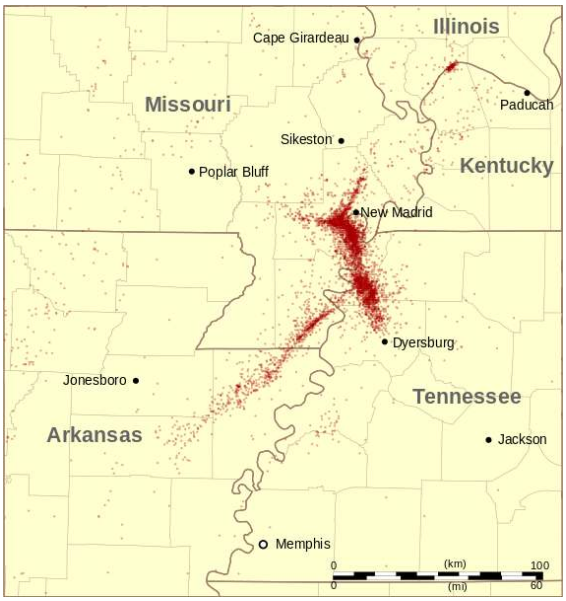


Figure 8.8.2: High density of earthquakes in the New Madrid seismic zone.

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.

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.



Figure 8.8.3: Buildings toppled from liquefaction during a 7.5 magnitude earthquake in Japan.

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 liquified sediment, which looks very much like quicksand in the movies.

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. 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.

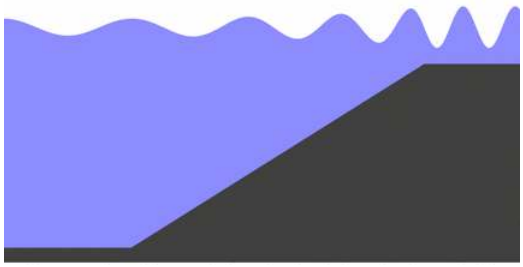


Figure 8.8.4: As the ocean depth becomes shallower, the wave slows down and pile up on top of itself, making large, high-amplitude waves.



Figure 8.8.5: House in Springdale, Utah destroyed by an earthquake-triggered landslide.

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].

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CHAPTER OVERVIEW

9: Mass Wasting

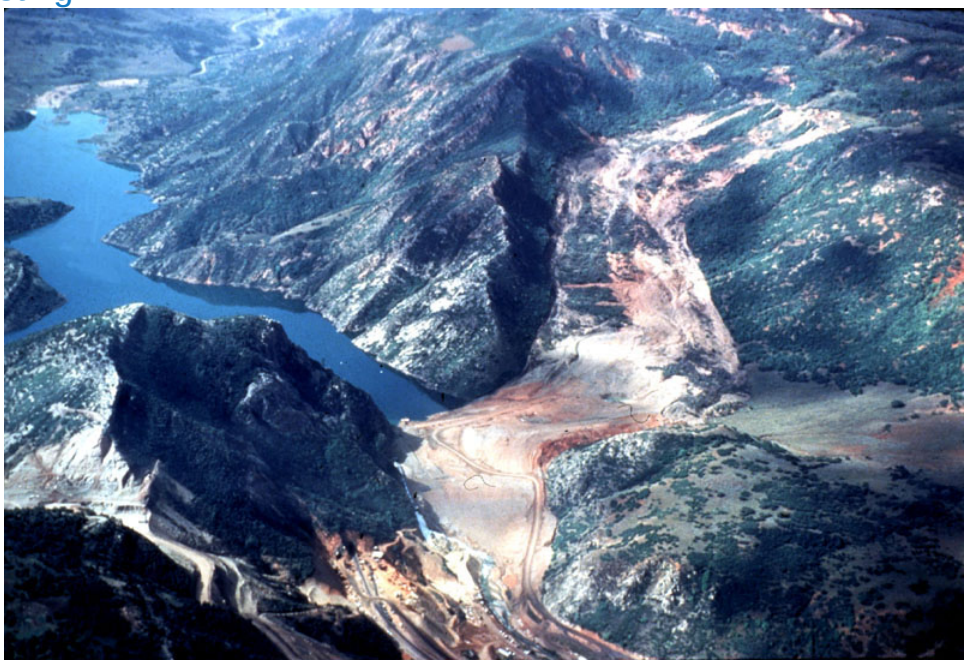


Figure 9.1: The 1983 Thistle landslide (foreground) dammed the Spanish Fork River creating a lake that covered the town of Thistle, Utah. The slide covered Hwy 6 and the main railroad between Salt Lake and Denver.

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.

9.1: Slope Strength

9.2: Mass-Wasting Triggers and Mitigation

9.3: Landslide Classification and Identification

9.4: Examples of Landslides

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9.1: Slope Strength

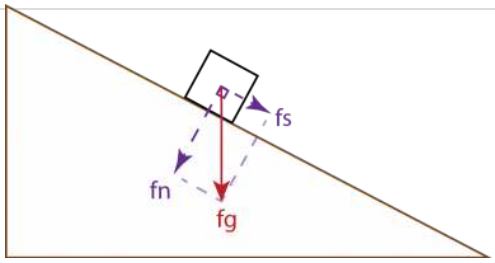


Figure 9.1.1: Forces on a block on an inclined plane (fg = force of gravity; fn = normal force; fs = shear force).

Mass wasting occurs when a slope fails. A slope fails 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. In the figure, 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.

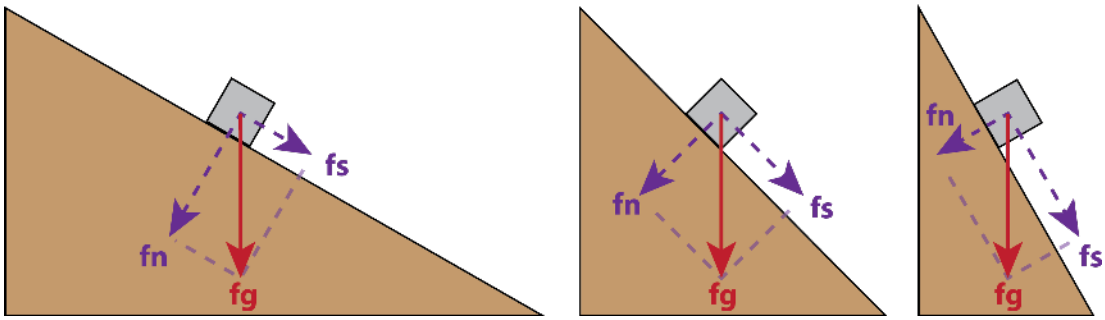


Figure 9.1.2: As slope increases, the force of gravity (fg) stays the same and the normal force (fn) decreases while the shear force (fs) proportionately increases. You can see this represented in the length of the arrows.

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. 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.

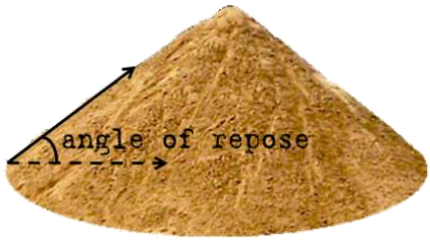


Figure 9.1.3: Angle of repose in a pile of sand.

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 in sediments or rocks between the grains. For example, assume a dry sand pile has an angle of repose of 30 degrees. If water is added to the sand, the angle of repose will increase, possibly to 60 degrees or even 90 degrees, 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 are planes of weakness in sedimentary rocks. Bedding planes (see Chapter 5) 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]. 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].

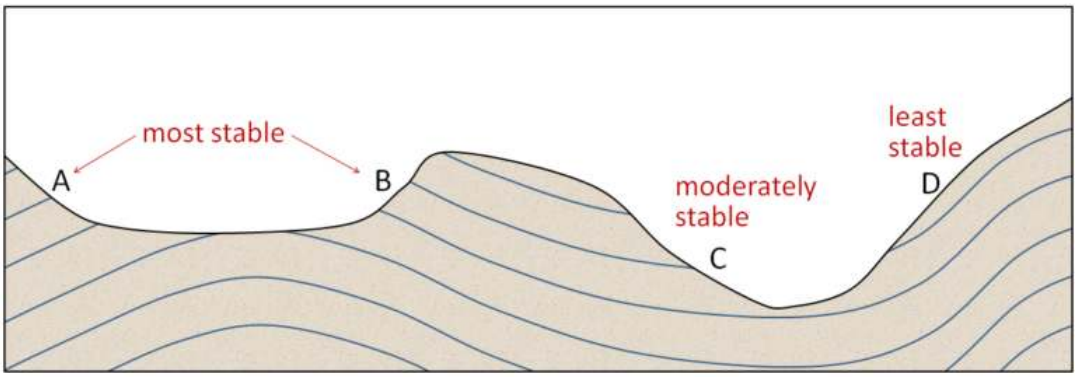


Figure 9.1.4: Locations A and B have bedding nearly perpendicular to the slope, making for a relatively stable slope. Location D has bedding nearly parallel to the slope, increasing the risk of slope failure. Location C has bedding nearly horizontal and the stability is relatively intermediate.

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9.2: Mass-Wasting Triggers and Mitigation

Mass-wasting events often have a **trigger**: something changes that cause an event to occur at a specific time. It could be rapid snowmelt, intense rainfall, earthquake shaking, a 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. Most landslide mitigation diverts and drains 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 oversteepened 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. Shockcrete, a reinforced spray-on form of concrete, can strengthen a slope face when applied properly. Buttressing a slide by adding weight at the base of the slide and removing weight from the top of the slide, can stabilize a landslide. Terracing, which creates a stairstep landscape, can be applied to help with slope stabilization.

A different approach in reducing landslide hazard is to shield, catch, and divert the runout material. Special stretchable fencing can be applied in areas where rockfall is common to protect pedestrians and vehicles. 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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9.3: Landslide Classification and Identification

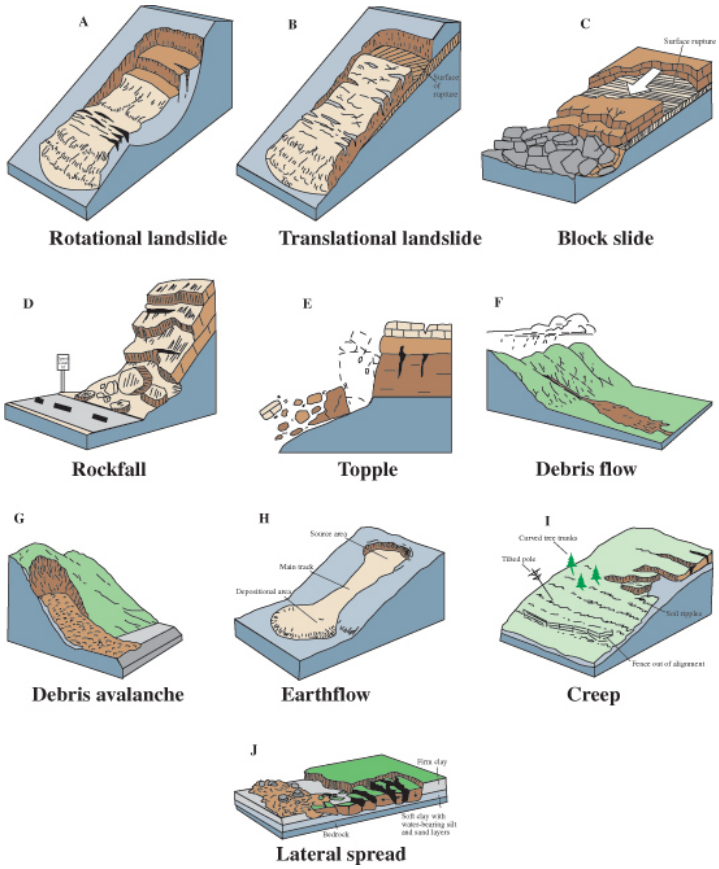
Mass-wasting events are classified by type of movement and type of material. The figure and table show the terms used. In addition, mass-wasting types often share common morphological features observed on the surface, such as the head scarp at the top of the failure—commonly seen as crescent shapes on a cliff face; uneven surfaces; accumulations of talus—loose rocky material falling from above; and toe (bottom) of the slope, which covers existing surface material.

Types of Mass Wasting

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 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).

Type of Movement	Primary Material Type and Common Name of Slide		
	Bedrock	Soil Types	
		Mostly Coarse-Grained	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	Earth flow
Soil Creep	—	Creep	Creep

Table of Mass Wasting Types. Mass wasting movement type and primary earth material. Modified from [2; 3].



Examples of some of the types of landslides.

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.

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.

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 have 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.

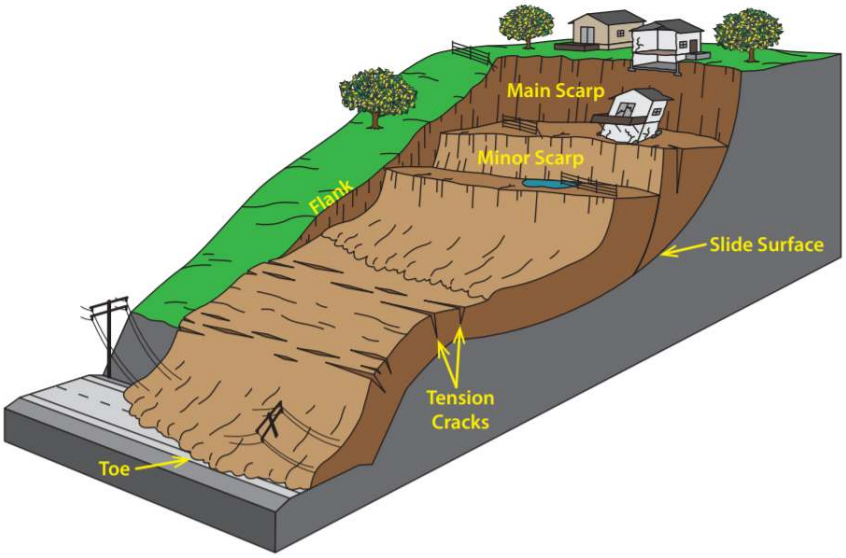


Diagram of a rotational landslide

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.
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.
Toe	The downhill end of the slide. The most distant part of the slide from the main scarp.

The **slide surface** is the lower boundary of the body that moved. 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. 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, and hummocky terrain. 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 [4].

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9.4: Examples of Landslides



Figure 9.4.1: Scar of the Gros Ventre landslide in the background with landslide deposits in the foreground.

1925, Gros Ventre, Wyoming: On June 23, 1925, a 38 million cubic meter (50 million cu yd) 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].

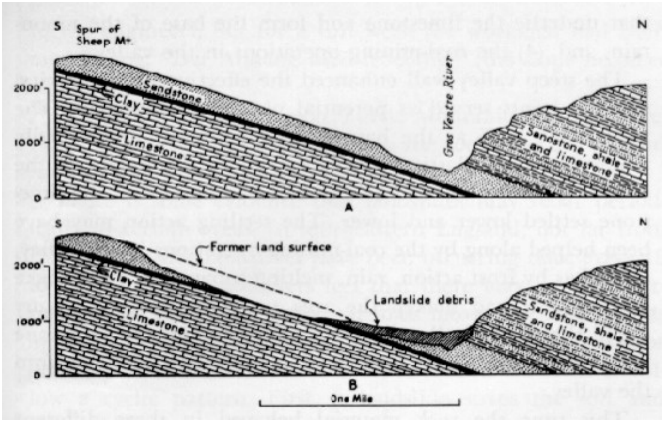


Figure 9.4.2: Cross-section of 1925 Gros Ventre slide showing sedimentary layers parallel with the surface and undercutting (oversteepening) of the slope by the river.

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.

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 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 9.4.3: 1959 Madison Canyon landslide scar. Photo taken from landslide material.

1980, Mount Saint 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 Saint 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 9.4.4: 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

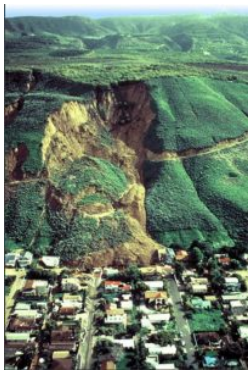


Figure 9.4.5: 1995 La Conchita slide. Source: USGS.

La Conchita Landslide



Figure 9.4.6: 2014 Oso slide in Washington killed 43 people and buried many homes (source: USGS, public domain).

2014, Oso Landslide, Washington: On March 22, 2014, a landslide of approximately 18 million tons (10 million yd^3) traveled at 64 kph (40 mph), extended for nearly a 1.6 km (1 m), 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].

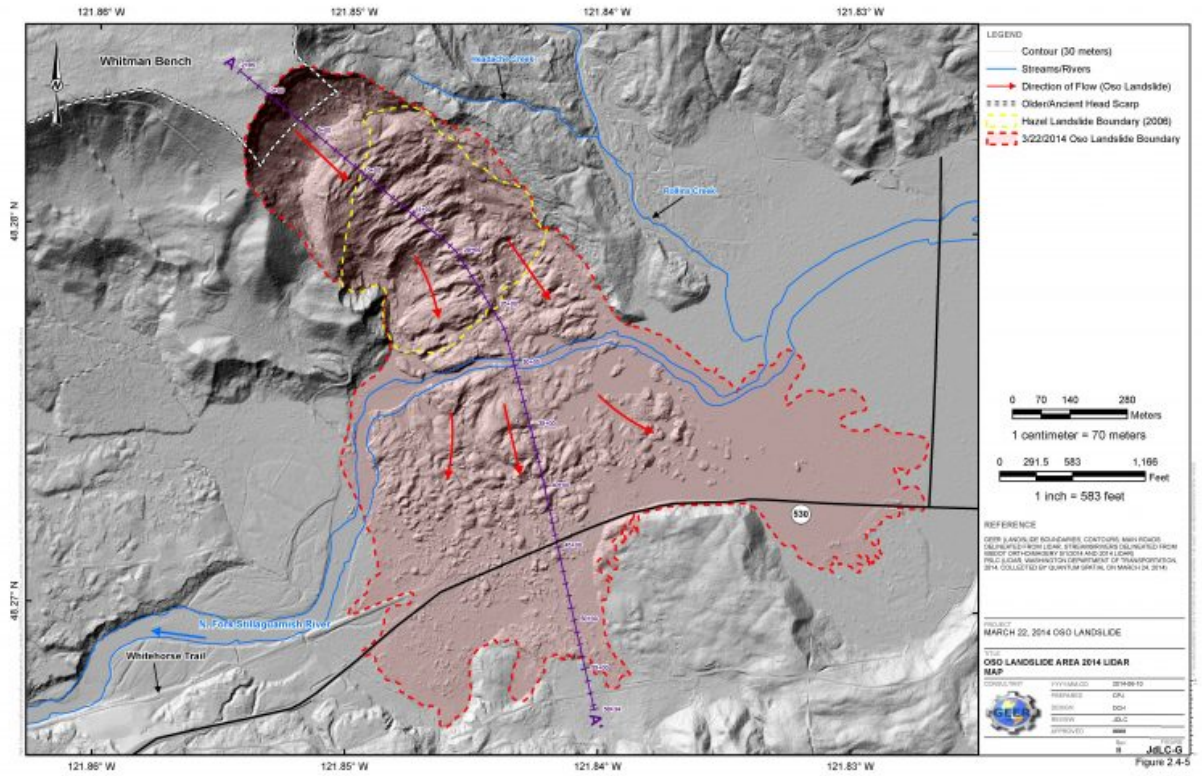


Figure 9.4.7: Annotated LiDAR map of 2014 Oso slide in Washington.

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CHAPTER OVERVIEW

10: Water

All life requires water. The hydrosphere (Earth's water) is an important agent of geologic change. It 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 melting in the upper mantle material. Communities rely on suitable water sources for consumption, power generation, crop production, and many other things.

10.1: Properties of Water
10.2: Water Cycle
10.3: Water Basin and Budgets
10.4: Water Use and Distribution
10.5: Water Law
10.6: Surface Water
10.7: Groundwater
10.8: Karst
10.9: Water Contamination

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10.1: Properties of Water

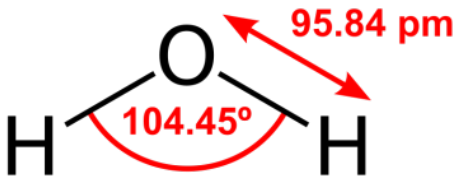


Figure 10.1.1: A model of a water molecule, showing the bonds between the hydrogen and oxygen.

The physical and chemical properties of water are what make it essential to life and useful to civilization. Water is a molecule made of one negatively charged (-2) oxygen ion and two positively-charged (+1) hydrogen ions, giving it the chemical formula H^2O, with strong covalent bonds between the oxygen and two hydrogen ions. The shape of the water molecule allows for an uneven distribution of charge, where one side is slightly positive and one side is slightly negative. Because of this polarity, water molecules form hydrogen bonds with each other. Hydrogen bonds are electrostatic intermolecular bonds that are weaker than ionic and covalent bonds. Water is amphoteric, that is it can self-ionize, breaking down into an acidic hydrogen ion (H^+) and a hydroxyl ion (OH^-), chemically a base. Because of its polarity and its ability to be amphoteric, water is a **universal solvent**—a chemical that can dissolve a wide range of other chemicals.

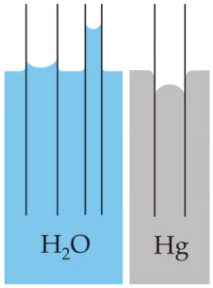


Figure 10.1.2: Capillary action by water in a narrow tube. Unlike water, mercury does not rise up in the tube.

Other side-effects of water's polarity are **cohesion** (water likes to stick to itself) and **adhesion** (water likes to stick to other things). Water has the highest cohesion of all nonmetallic liquids. Cohesion gives water surface tension, allowing water glider insects to float on the water surface. Surface tension is what gives raindrops a spherical shape. Capillary action occurs when a combination of adhesive and cohesive forces causes water to move up narrow passages and tubes, rising higher than surrounding liquid. Paper towels have small pores that use capillary forces to clean up water spills. Plants use capillary forces to pump water into their tissues.

Water has a high specific heat capacity. Specific-heat is the amount of heat required to raise the temperature of a substance. Compared to many other substances, water requires a large amount of heat to raise its temperature. The high specific heat of water allows it to act as an energy buffer to extreme changes in air temperature. It also allows the oceans to soak up solar heat without changing temperature much and distribute that heat over the Earth by currents thus making the Earth habitable.

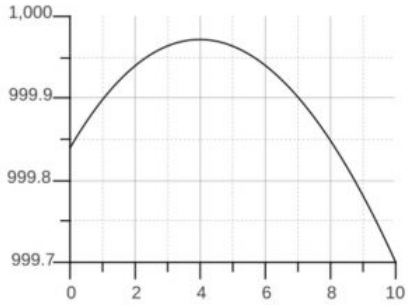


Figure 10.1.3: Density curve of water showing the greatest density at 3.98 degrees Celcius. Density is shown as kg/m_3.

The density curve of water shows that as water is cooled, it becomes more dense, as do most other substances, but its greatest density occurs at about 4 degrees Celsius while most other substances continue to increase in density until they freeze. This unique

density curve means that water is most dense just above its freezing point and sinks. Thus the oceans remain liquid. If water behaved like other substances, the oceans would be frozen.

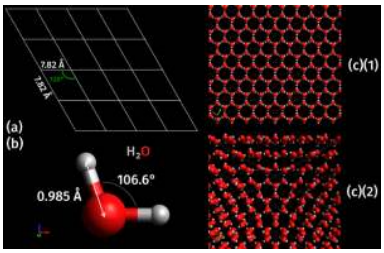


Figure 10.1.4: Molecular arrangement of hydrogen dioxide (water) molecules in ice.

When water freezes, the molecules arrange themselves in a well-ordered crystal structure, creating a spacing between molecules that is greater than if the water is in liquid form. The difference in molecular spacing causes ice to be less dense than water, making it more buoyant than liquid water, causing it to float on water. Because of its high specific heat capacity, ice floating on a lake's surface insulates the liquid water beneath and keeps it from freezing.

Because of its hydrogen bonds, water also has a high heat of vaporization. A significant amount of energy is required to evaporate water. As water evaporates, energy is absorbed by the breaking of hydrogen bonds and the air around the evaporating water is cooled. This energy is stored in the water vapor.

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10.2: Water Cycle

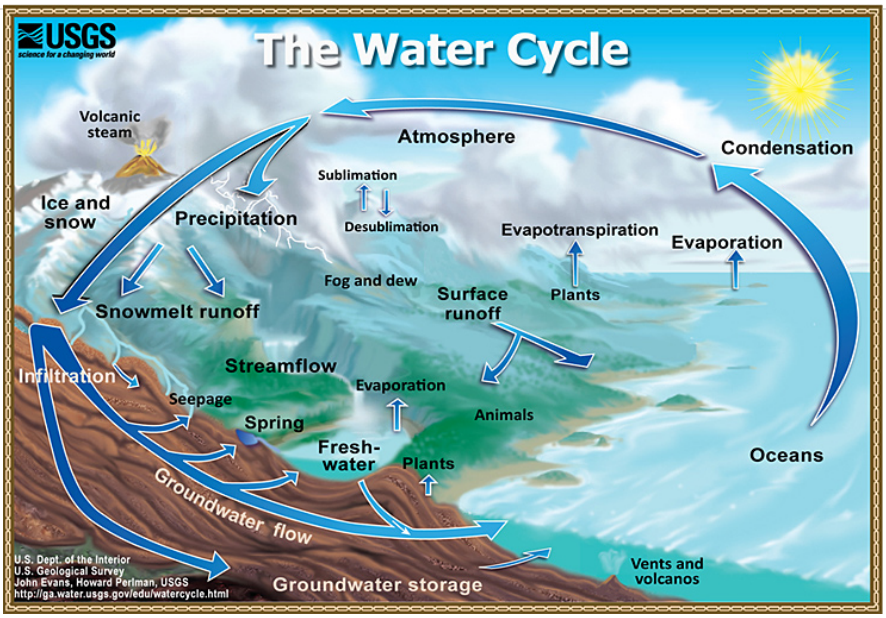


Figure 10.2.1: The water cycle.

The water cycle describes how water changes between solid, liquid, and gas (water vapor) phases and changes location. Water can be evaporated, which is the process where a liquid is converted to a gas. Solar energy warms the water sufficiently to excite the water molecules to the point of vaporization. Evaporation occurs from surface water bodies such as oceans, lakes, and streams and the land surface. Plants contribute significant amounts of water vapor as a byproduct of photosynthesis in a process called transpiration. Geologists commonly combine these two sources of water entering the atmosphere in a term called evapotranspiration.

Water vapor in the atmosphere can migrate long-distance from ocean to over land by way of winds. Over the ocean or land, the air can cool and cause the water to condense back into liquid water. These small water droplets are visible as a cloud. Clouds build and once the water droplets are big enough, they fall to earth as precipitation. Precipitation can take the form of rain, snow, hail, or sleet.

Once it has reached the surface it does two important things relevant to the geology of this chapter. At the surface, liquid water can flow as **runoff** into streams, lakes, and eventually back to the oceans (in most cases). Water in streams and lakes is called surface water. In addition, water can also **infiltrate** into the soil and finally fill the pore spaces in the rock or sediment deep underground to become **groundwater**, the name given to all subsurface water. Groundwater slowly moves through rock and unconsolidated materials and some of it eventually reaches 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.

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10.3: Water Basin and Budgets

Drainage Basins

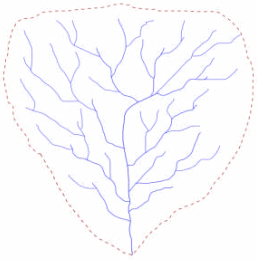


Figure 10.3.1: Map view of a drainage basin with main trunk streams and many tributaries with drainage divide in the dashed red line.

The basic unit of division of the landscape is the drainage basin. A **drainage basin**, is the area of land that captures precipitation and contributes runoff to a stream or stream segment (Figure). **Drainage divides** are local topographic high points that separate one drainage basin from another [5]. If water falls on one side of the divide, that water goes to one stream, and if it falls on the other side of the divide, then the water goes to a different stream.

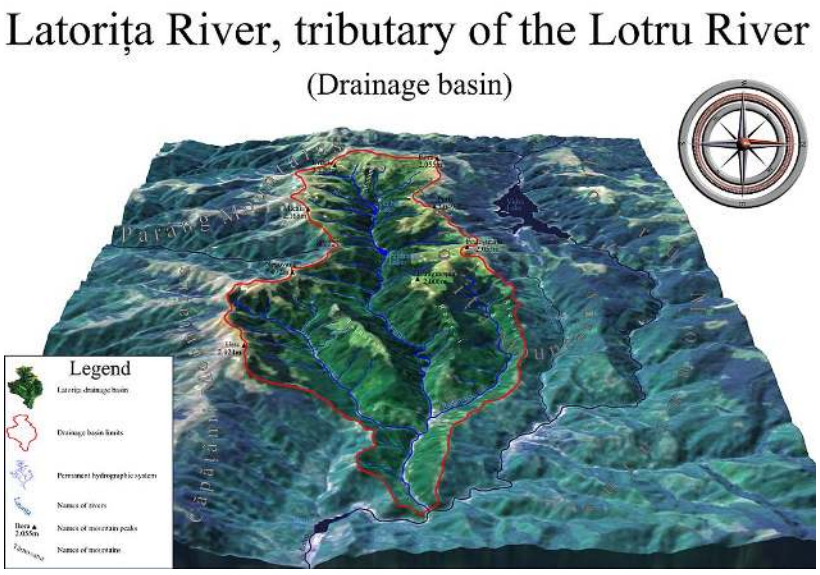


Figure 10.3.2: Oblique view of the drainage basin and divide of the Latorita River, Romania.

Streams only flow downhill and smaller **tributary** streams combine downhill to make the larger trunk of the stream. Where a stream begins is called the **headwaters** and where it finally reaches its end is called the **mouth**. Most streams have the mouth of the stream at the ocean. However, a rare number of streams do not flow to the ocean, but rather end in a **closed basin** where the water evaporates from a stream or lake before it can reach the ocean.

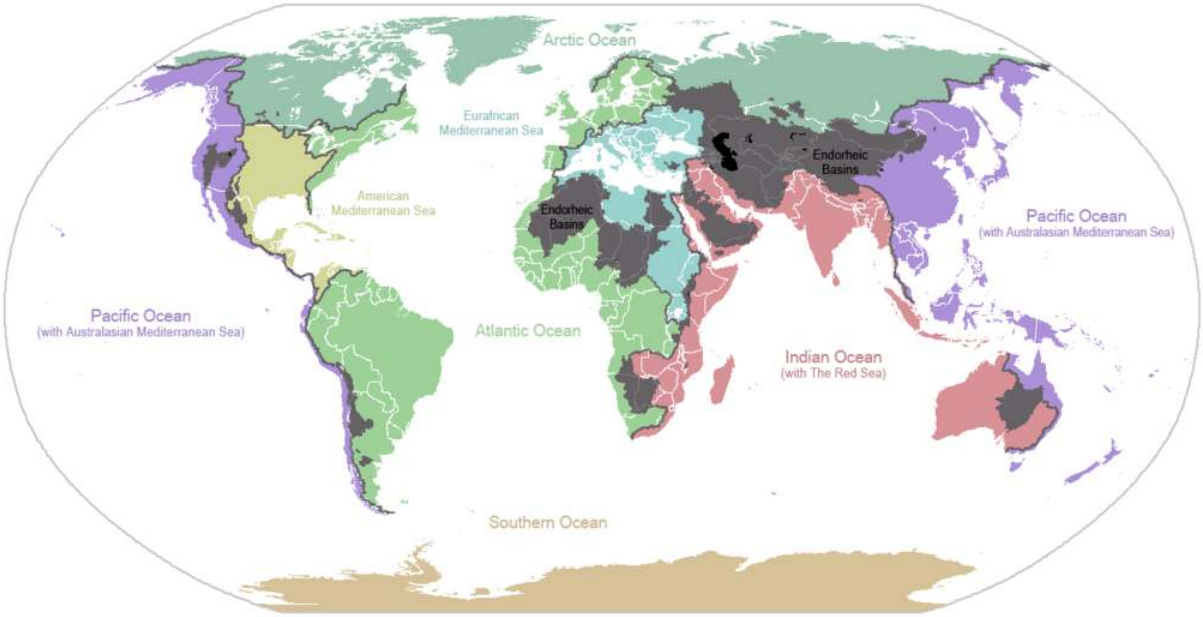


Figure 10.3.3: Major drainage basins color-coded to match the related ocean. Closed basins (or endorheic basins) are shown in gray.

In humid climates, many streams are perennial and in arid climates like Utah, many streams are ephemeral. **Perennial streams** flow all year round. They occur in humid or temperate climates where there are 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. These are often **arroyos** or dry washes for much of the year. They are above the water table and occur in dry climates with low amounts of rainfall and high evaporation rates. They flow mostly during flash floods.

Water Budgets

Scientists can make groundwater budgets within any designated boundary, but they are generally made for drainage basisn 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, where measured data are used with a conceptual workflow of the model to better understand the water system.

Like budgets for families and organizations, there is income from precipitation and expense from discharge and evaporation, with the ultimate expense for freshwater by returning to the salty ocean. For dry regions, the water budget is critical for sustaining human activities and understanding and managing it is an ongoing political challenge.

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10.4: Water Use and Distribution

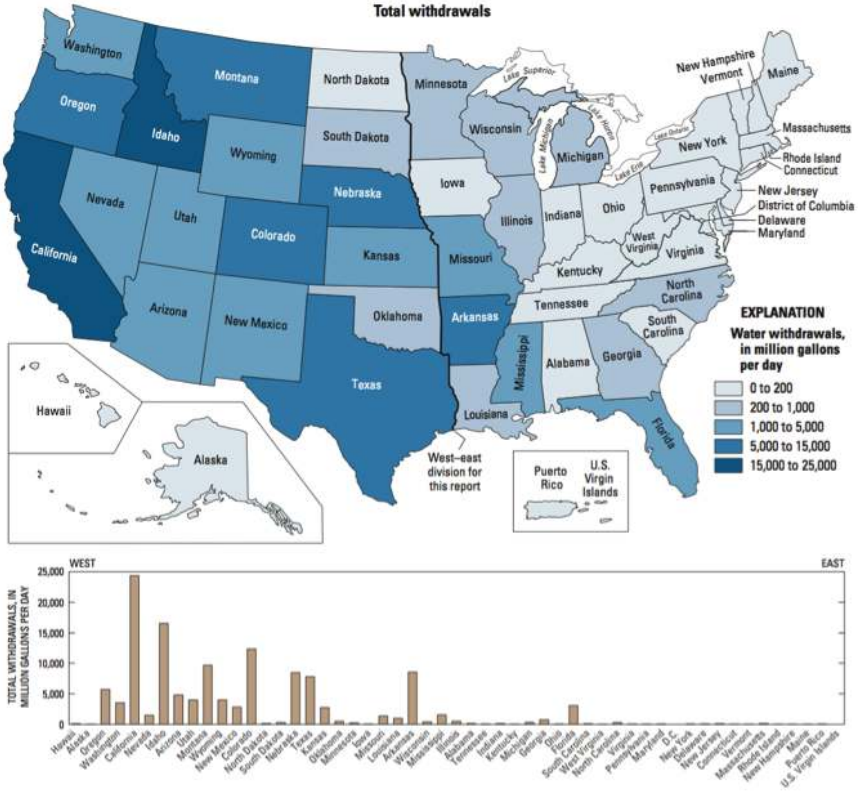


Figure 10.4.1: Agricultural water use in the United States by state.

In the United States, 355 billion gallons of ground and surface water are withdrawn for use each day, of which 76 billion gallons are fresh groundwater. The state of California accounts for 16% of national groundwater withdrawals [8].

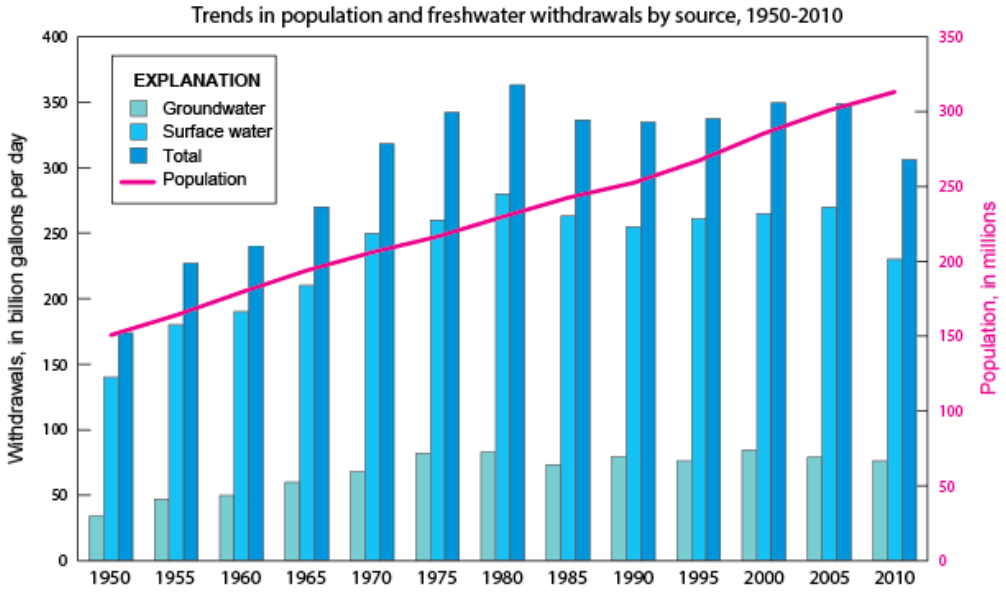


Figure 10.4.2: Trends in water use by source

Surface Water Distribution

Surface water makes up only 1.2% of the freshwater available on the planet, and 69% of that surface water is trapped in ground ice and permafrost. Water in rivers accounts for only 0.006% of all freshwater and lakes contain only 0.26% of the world's freshwater

[9]. Freshwater is a precious resource and should not be taken for granted, especially in dry climate areas.

Global circulation patterns are the most important factor in precipitation, and thus, the distribution of surface water. For reasons discussed in the chapter on Deserts, some areas receive a lot of precipitation, and others very little.

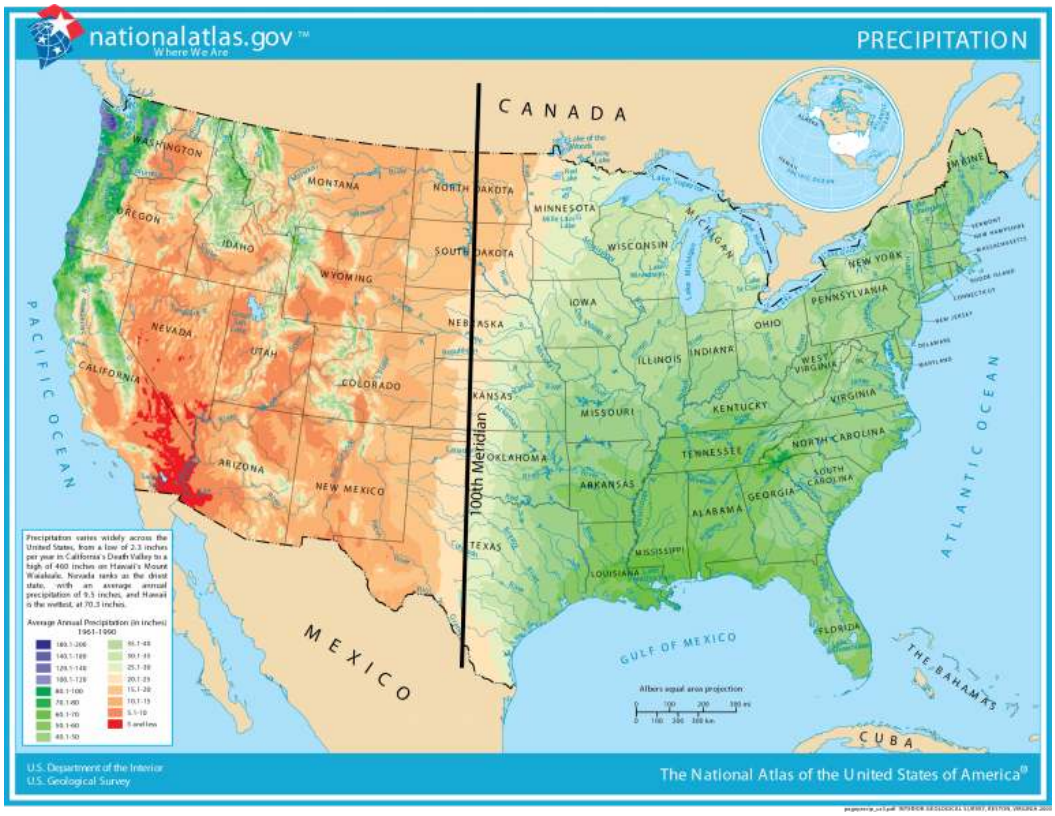


Figure 10.4.3: Distribution of precipitation in the United States. The 100th Meridian is approximately where the average precipitation transitions from relatively wet to dry. (Source: U.S. Geological Survey)

In the United States, the 100th Meridian roughly marks the boundary between the humid and arid parts of the country while west of the 100th Meridian, irrigation is required to grow crops [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.

Groundwater Distribution

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	5,614,000	—	1.69
— Fresh	2,526,000	30.1	0.76
— Saline	3,088,000	—	0.93
Soil Moisture	3,959	0.05	0.001
Ground Ice & Permafrost	71,970	0.86	0.022
Lakes	42,320	—	0.013
— Fresh	21,830	0.26	0.007
— 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)

Groundwater makes up 30.1% of the freshwater on the planet, making it the most abundant reservoir of freshwater accessible to most humans. The majority of freshwater, 68.7%, is stored in glaciers and ice caps as ice [9]. As the glaciers and ice caps melt due to global warming, this freshwater is lost as it flows into the oceans.

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10.5: Water Law

Federal and state governments have put laws in place to ensure the fair and equitable use of water. Based on the distribution of precipitation in the United States, the states are in a position that requires them to create a fair and legal system for sharing water.

Water Rights

Because of the limited supply of water, especially in the western United States, some states have adopted a system of legally dispersing ownership of natural waters. A claim to a portion or all of a water source, such as a spring, stream, well, or lake is known as a **water right**. 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 navigation 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 populated and while California was becoming a key agricultural producer, the states along the Colorado River realized that the river was important to sustaining population and agriculture 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 (formerly Boulder) and Glen Canyon Dams, and a treaty between the American and Mexican governments. Combined, the agreements became known as “The Law of the River.” Despite adjustments to the 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.

Quality 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 dumpage 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 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. It allows state governments and/or the U.S. Environmental Protection Agency power to remediate polluted sites through either actions or funds provided by the polluter that caused the contamination.

References

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10.6: Surface Water

A **stream** or river is a body of flowing surface water confined to a channel. Terms such as creeks and brooks are social terms not used in geology. Streams are the most important agents of erosion and transportation of sediments on the earth's surface. They create much of the surface topography and are an important water resource. Most of this section will focus on stream location, processes, landforms, and flood hazards. Water resources and groundwater processes will be discussed in later sections.

Discharge

Several factors cause streams to erode and transport sediment, but the two main factors are stream channel gradient and velocity. The stream **gradient** is the slope of the river channel. A steeper gradient promotes downward stream erosion. When tectonic forces lift up a mountain, the increased stream gradient causes the stream to erode downward and make a valley. Stream velocity is the speed of the flowing water in the channel. Velocity can increase by increasing the gradient, decreasing cross-sectional area (narrowing) of the channel (reducing friction), or by increasing the discharge.

Stream size is measured in terms of **discharge,** i.e. the volume of water flowing past a point in the stream over a defined time interval. Smaller streams have a smaller discharge, therefore generally stream discharge increase downstream**.** Volume is commonly measured in cubic feet (length x width x depth), shown as $feet^3$ or ft^3. Therefore, the units of discharge are cubic feet per second (ft^3/sec or cfs). For example, the Mississippi River is the largest river in North America, with an average flow of about 600,000 cfs [19]. For comparison, the average discharge for the Jordan River at Utah Lake is about 574 cfs [20] and for the Amazon River (the world's largest river), annual discharge is about 6,200,000 cfs [21].

When the 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 rainstorms or heavy snowmelt, runoffs will increase which increases stream discharge and thus velocity.

Velocity varies within the stream channel as well. Generally, when the channel is straight and uniform in-depth, 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 channel bottom and sides. When the channel curves, the highest velocity will be on the outside of the bend as it has to flow faster to keep up!

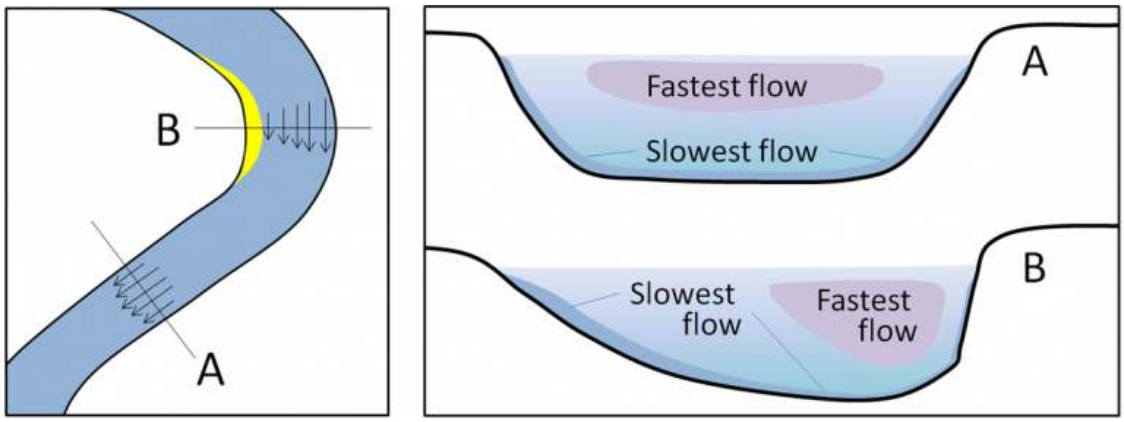


Figure 10.6.1: In a river bend, the fastest moving particles are on the outside of the bend, causing erosion. The slowest are on the inside bend, causing deposition. In section A, water flows fastest in the center (longer arrows indicate faster velocity). In section B, water flows fastest on the outside bend (Earle 2015).

Runoff vs. Infiltration

There are many factors dictating whether water will infiltrate into the ground or runoff over the land after precipitation. These include but are not limited to the amount, type, and intensity of precipitation, the type and amount of vegetative 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 precipitation as 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 groundwater via caves and subterranean drainage and this creates a **deranged pattern**.

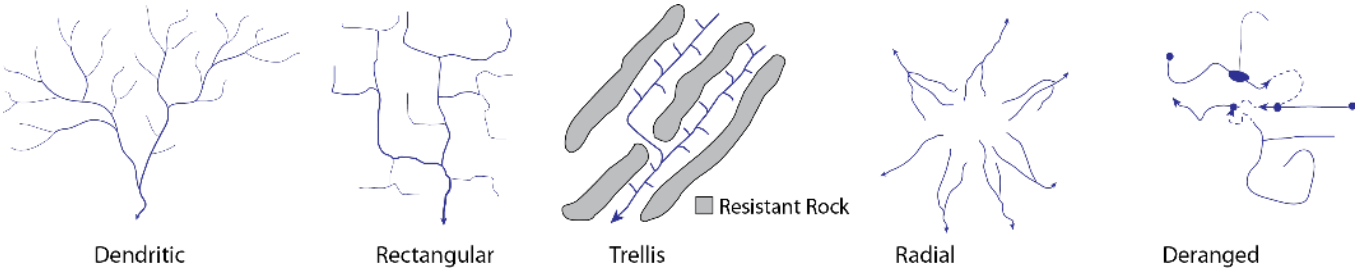


Figure 10.6.2: Various stream drainage patterns.

Fluvial Processes

Fluvial processes are the mechanisms that dictate how a stream functions and include factors controlling fluvial sediment production, transport, 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. These can be defined as zones of sediment production (erosion), transport, and deposition.

Zone of Sediment Production (Erosion)

The zone of sediment production is located in the **headwaters** (the start) of the stream were very small channels 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 most sediment production and erosion. Headwater streams tend to be narrow and straight. 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 contain much of the headwaters for the Colorado River which then flows from Colorado through Utah, Arizona, to Mexico.

Zone of Sediment Transfer (Transportation)

Downstream of the headwaters, the stream erodes less sediment but transports the sediment provided from the headwaters in the zone of sediment transfer.

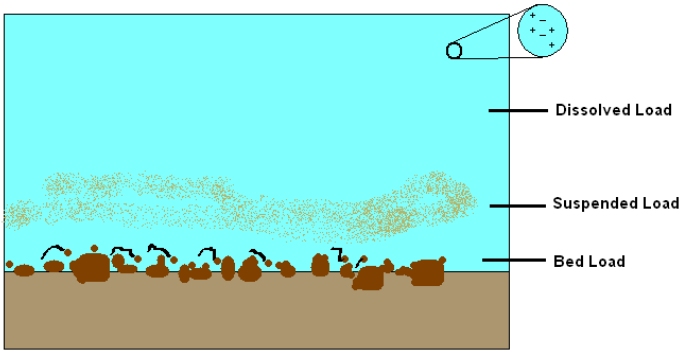


Figure 10.6.3: A stream carries dissolved load, suspended load, and bedload.

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.

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. Bedload is moved by saltation (bouncing) and traction (being pushed or rolled along by the force of the flow). When stream velocity increases, smaller bedload sediments can be picked up by flowing water and held in suspension as **suspended load**. The faster streams can carry larger grains as suspended load. **Dissolved load** in a stream is the sum 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^+).

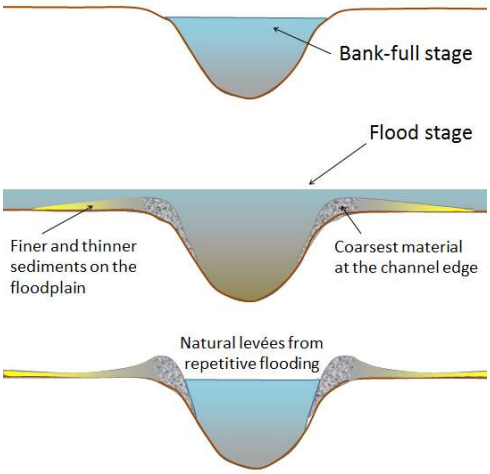


Figure 10.6.4: Profile of stream channel at bank-full stage, flood stage, and deposition of natural levee

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. As soon as the flooding stream overtops its banks and occupies the wide area of its floodplain, the velocity decreases. Sediments are added to the floodplain during this flooding process.

Zone of Disposition

The process of **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 (discussed later), however, ultimate deposition occurs at the mouth (the end) 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.

Landforms

Stream landforms are the land features formed on the surface by either erosion or deposition. The primarily stream-related landforms described here are related to channel types.

Channel Types



Figure 10.6.5: Braided stream pattern on the Waimakariri River in New Zealand.



Figure 10.6.6: Air photo of the meandering river, Río Cauto, Cuba.

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 channels have multiple smaller channels splitting and recombining downstream creating numerous mid-channel bars. These are found in broad terrain with low gradients near sediment source areas such as mountains or in front of glaciers, for example in Alaska.

Meandering channels are composed of a single channel that curves back and forth like a snake within its floodplain. Meandering channels tend to have a wide floodplain, high discharge, natural levees, and flood regularly. Meandering channels 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. In areas of uplift, like has occurred on the Colorado Plateau, meanders that formed on the upland can become entrenched or incised as the stream cuts its meandering pattern down into bedrock.

Entrenched channels occur 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 but an example of this process is where the Colorado River and other streams crossed the Colorado Plateau as meandering streams. As the Colorado Plateau has

uplifted over the past several million years, the Colorado River has incised into the flat-lying rocks of the plateau by hundreds of feet.



Figure 10.6.7: Entrenched meander of the Colorado River, downstream of Page, Arizona.



Figure 10.6.8: The San Juan River has incised meanders into the flat Colorado Plateau at Goosenecks State Park, southeastern Utah.

Alluvial Fans



Figure 10.6.9: Alluvial fan in Iraq seen by a NASA satellite. A stream emerges from the canyon and creates this cone-shaped deposit.

Alluvial fans are a depositional landform created where streams emerge from mountain canyons into a valley. The channel that had been confined by the canyon walls and suddenly is no longer confined slows down and spreads out, dropping its bedload of all sizes. It is like a delta above water. 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 (streams that run periodically) emerge from canyons in the ranges of the Basin and Range.

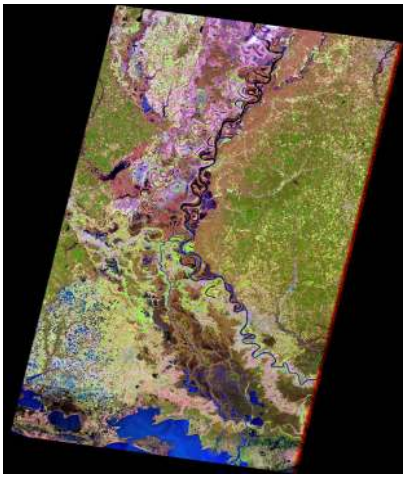


Figure 10.6.10: Landsat image of lower Mississippi River floodplain

Many fluvial landforms occur in a floodplain near a meandering stream. 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 location and width of floodplains naturally vary, however, humans build artificial levees on flood plains to limit flooding. Floodplains are nutrient-rich from the fine-grained deposits 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 10.6.11: Point bar and cut bank on the Cirque de la Madeleine in France.

Meandering rivers create additional landforms as the channel migrates within the floodplain. Meandering rivers 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 is called a **cut bank**. Opposite the cut bank on the inside bend of the channel is the lowest stream velocity and therefore becomes an area of deposition call a **point bar**.



Figure 10.6.12: Meander nearing cutoff on the Nowitna River in Alaska

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. The former channel becomes isolated from streamflow and forms an **oxbow lake**.

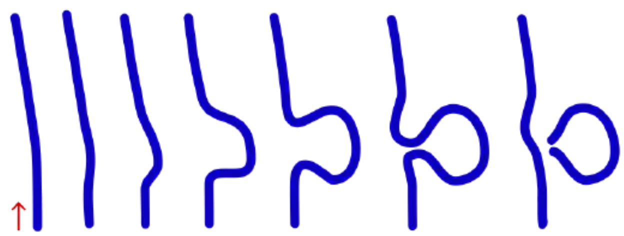


Figure 10.6.13: Meander nearing cutoff on the Nowitna River in Alaska

Deltas



Figure 10.6.14: Location of the Mississippi River drainage basin and Mississippi River delta.

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 come to rest, forming a **delta**. If wave erosion from the water body is greater than deposition from the river, the deposition will not occur and a 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 river carries a large amount of sediment that is supplied to the delta. The Mississippi River is a major shipping route and human engineering has ensured that the channel no longer meanders significantly within the floodplain. In addition, 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 (follow the link) 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].

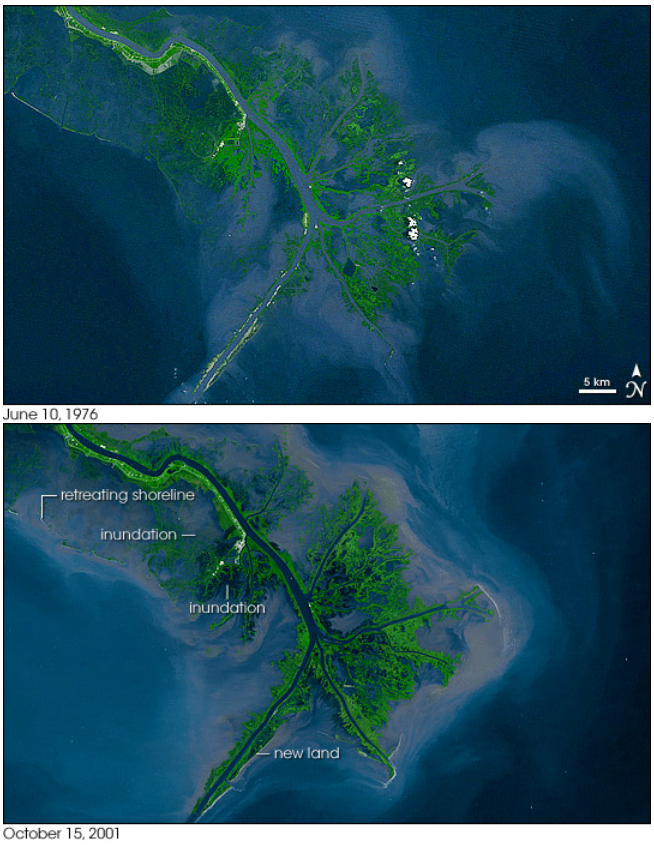


Figure 10.6.15: Before (left) vs. After (right)

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. A tide-dominated delta is 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 and mangrove swamp called the Sundarban.



Sundarban Delta in Bangladesh, a tide-dominated delta of the Ganges River Tidal forces creates linear segments in the delta shoreline by ocean intrusion into the delta deposits. This delta also holds the world's largest mangrove swamp, and incidentally is the only place where the Bengal tiger still actively hunts humans as prey.



Figure 10.6.16: Nile Delta showing its classic "delta" shape.

Terraces

Stream terraces are remnants of older floodplains located above the existing floodplain and river. 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.



Figure 10.6.17: Terraces along the Snake River, Wyoming.



Figure 10.6.18: Terraces in Glen Roy, Scotland

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10.7: Groundwater

Groundwater, freshwater stored under the surface, is an important source of freshwater. It can be found in all places under the ground but is limited by extractable quantity and quality.

Properties

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. Clays generally have very high porosity, but the pores are poorly connected, thereby causing low permeability, which makes it a poor aquifer.

Aquifers and Confining Layers

In order for a geologic material to be considered an aquifer, it must be porous and at least partially saturated, where its open spaces are filled with water, and be permeable, i.e. able to transmit water. Aquifers can vary dramatically in scale, from spanning several formations to being limited to a small area on the side of a hill.

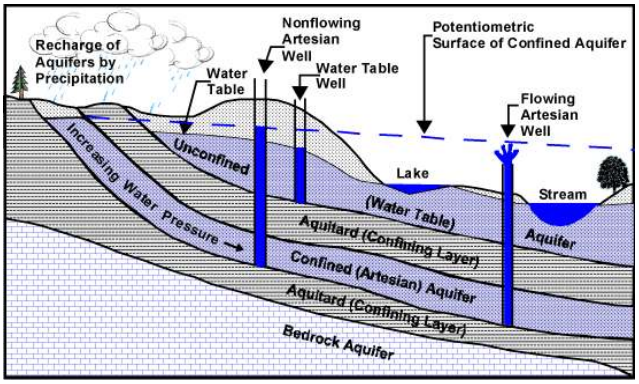


Figure 10.7.1: Potentiometric surface and water table in an aquifer system.

A good aquifer will provide a sufficient quantity of water to meet demand. The quantity of water that an aquifer can hold and transmit is governed by its physical properties. Most simply, the aquifer's porosity and permeability (defined above) are variables that govern its ability to serve as an aquifer.

A **confining layer** is a layer of low permeability geologic material that restricts the flow of water to or from 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 due to their low permeability.

Groundwater Flow

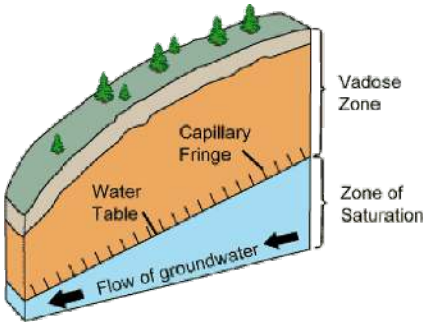


Figure 10.7.2: Zone of saturation.

When surface water infiltrates or seeps into the ground, it usually enters the unsaturated zone (a.k.a. vadose zone, a.k.a. zone of aeration). The **vadose zone** is the volume of material between the land surface and the zone of saturation, which consists of geologic materials in which the pore spaces are not completely filled with water [34]. Plants' roots inhabit the upper vadose zone. 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, described in the Properties of Water section above. Below the capillary fringe is the saturated 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.

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 porosity in sediments and bedrock. The **water table** is the area at which the pores are fully saturated with water. The most simple case of a water table is when the aquifer is unconfined, meaning it does not have a confining layer above it. 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**.

The water table will generally mirror surface topography, though more subdued (i.e. the highs are not quite as pronounced as the highs in the topography). 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.

Geologists measure the height of the water table and potentiometric surface using wells. Graphs of the depth to groundwater over time are known as hydrographs and can indicate changes in the water table over time. The water level in a well can change very frequently (every minute), seasonally, and over long periods of time, and is controlled by many forces.

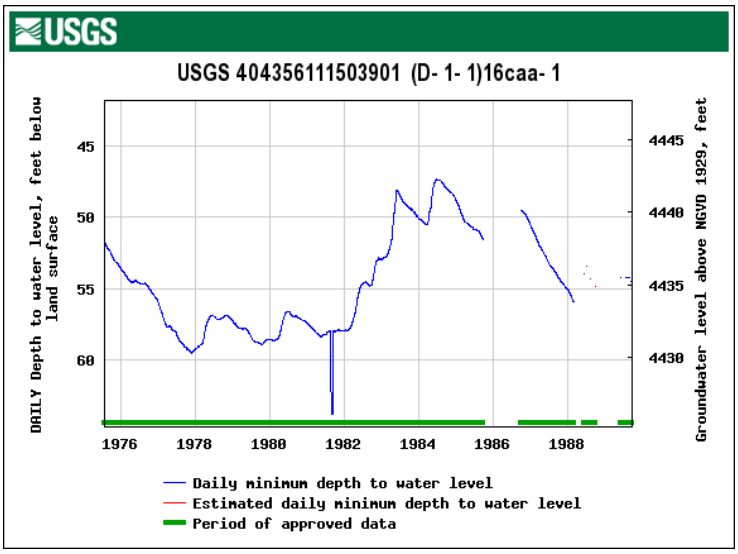


Figure 10.7.3: Example of a hydrograph.

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. In a confined aquifer, the pressure around the well is reduced. 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**.

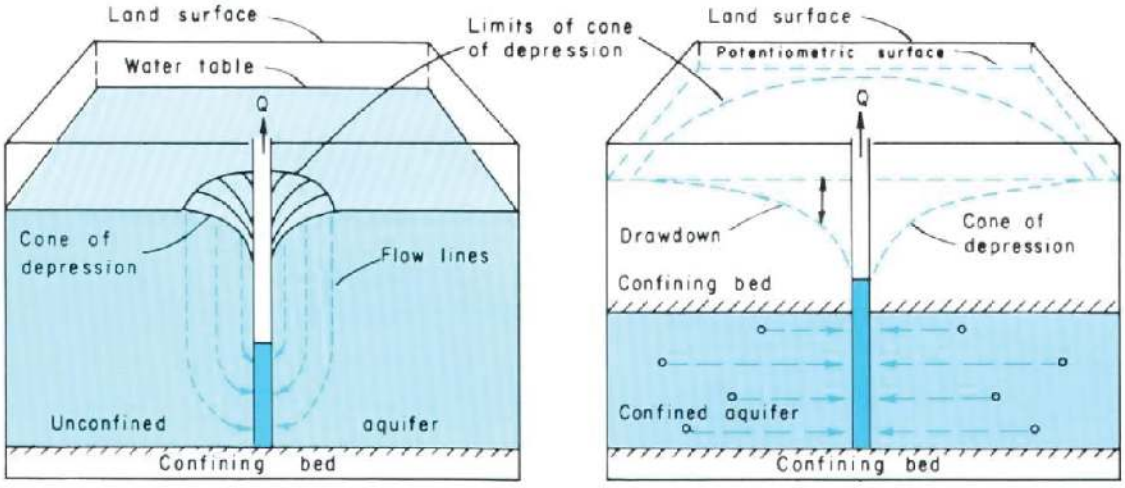


Figure 10.7.4: Cones of depression (Heath 1983).

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.

Recharge

The **recharge** area is where surface water enters an aquifer through the process of infiltration. Recharge areas are generally the topographically highest location of an aquifer. They are characterized by sediment or rock that allows infiltration into the subsurface. Recharge areas mark the beginning of groundwater flow paths.

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. 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.

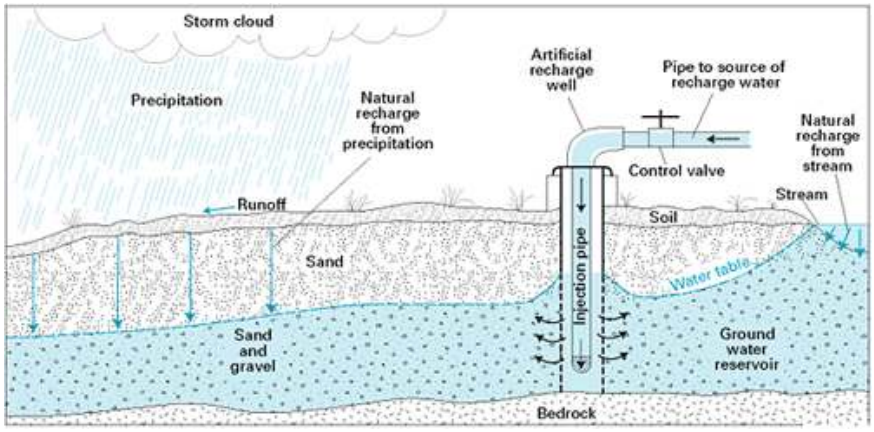


Figure 10.7.5: Different ways an aquifer can be recharged.

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. Discharge areas mark the end of groundwater flow paths.

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 recharge can renew it, 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. Groundwater mining can lead to dry wells, reduced spring and streamflow, and subsidence. Where there is a continual lowering of the water table in an area, e.g. extraction by pumping is greater than replenishment by precipitation, groundwater mining is happening.

Subsidence



Figure 10.7.6: Example of an earth fissure in Cedar City.

In many places, water actually helps hold up the skeleton of the aquifer by the water pressure exerted on the grains in an aquifer. If pore pressure decreases because of groundwater mining, the aquifer can compact, causing the surface of the ground to sink. 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].



Figure 10.7.7: The pole shows subsidence from groundwater pumping for irrigation in the San Joaquin Valley of California. The dates marked on the pole show land elevation in the past.

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.

The lowering of the landsurface due to groundwater removal is called **subsidence**.

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10.8: Karst



Figure 10.8.1: Steep karst towers in China left as remnants as limestone is dissolved away by acidic rain and groundwater.

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.

For example, the dissolution of underlying salt deposits has caused sinkholes to form in the Kaibab Limestone on the Colorado Plateau in Arizona.



Figure 10.8.2: 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 10.8.3: Sinkhole from the collapse of the surface into an underground cavern that appeared in the front yard of this home in Florida.

Karst forms when natural water, in combination with carbon dioxide, creates carbonic acid. This causes precipitation to be slightly acidic such that it can dissolve the calcite (calcium carbonate) that forms limestone.



Figure 10.8.4: Mammoth hot springs, Yellowstone National Park.

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. Speleotherms are secondary deposits, typically made of travertine, deposited in a cave. Travertine speleotherms form by water dripping through cracks and dissolution openings in caves and evaporating, leaving behind the travertine deposits. Speleotherms commonly occur in the form of stalactites, when extending from the ceiling, and stalagmites, when extending from the floor.

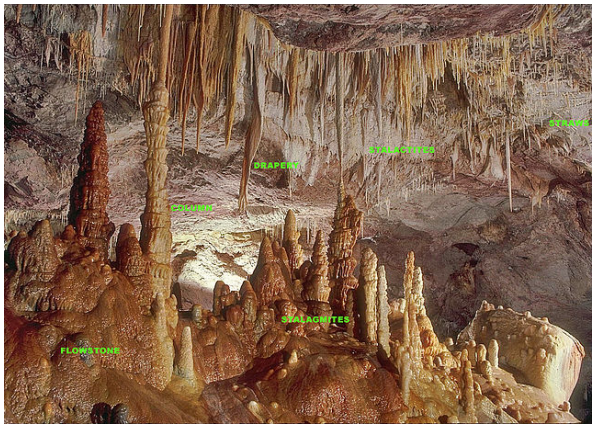


Figure 10.8.5: Varieties of speleotherms.



Figure 10.8.6: This stream disappears into a subterranean cavern system to re-emerge a few hundred yards downstream.

Meteoric (surface) water enters the karst system through sinkholes, or as rivers run over limestone and dissolve the limestone and sink into the ground. As the water dissolves its way through the limestone, it can leave behind intricate networks of caves and narrow passages, with caverns dropping thousands of feet. Often dissolution will follow and expand fractures in the limestone. Water exits the karst system as a spring.

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10.9: 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 the water, 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 pollution are 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 a fertilizer can lead to significant nonpoint 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. 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.

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 concentrate 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].

Metals are common inorganic contaminants. Lead, mercury, and arsenic are some of the more problematic inorganic groundwater contaminants. 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 intrusion, where cones of depression around fresh groundwater pumping near ocean coasts induce the encroachment of saltwater into the freshwater body.

Another common groundwater contaminant is biological, 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. Bacteria can also be used for remediation (see below).

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 reactive barriers, 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.

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CHAPTER OVERVIEW

11: Glaciers

The hydrosphere, liquid water, is the single most important agent of erosion and deposition. The cryosphere, the solid state of water in the form of ice also has its own unique erosional and depositional features. Large accumulations of year-round ice on the land surface are called **glaciers**. Masses of ice floating on the ocean as sea ice or icebergs are not glaciers, although they may have had their origin in glaciers.



Figure 11.1: Avalanche Lake in Glacier National Park, Montana is an example of a glacially-carved cirque basin.

Glaciers cover about 10% of the surface of the Earth. Glaciers form when more snow accumulates over a long span of time than melts and eventually turns into ice. This usually occurs in mountainous areas that have both cold temperatures and high precipitation. But snow can also accumulate and turn into ice in extremely cold low lying areas such as Greenland and Antarctica. This chapter focuses on types of glaciers, how glaciers function, erosional and depositional landforms created by glaciers, and how glaciers are connected to past climates and modern day climate change.

11.1: Types of Glaciers
11.2: Glacier Formation and Movement
11.3: Glacial Budget
11.4: Glacial Landforms
11.5: Ice Age Glaciations

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11.1: Types of Glaciers



Figure 11.1.1: Glacier in the Bernese Alps.

There are two general types of glaciers – alpine glaciers and ice sheets. **Alpine glaciers** form in mountainous areas either at high elevations or near cool and wet coastal areas like the Olympic Peninsula of Washington. Alpine glaciers are often confined to long, narrow valleys located in mountainous areas especially at higher latitudes (closer to either the north or south pole). Most alpine glaciers are located in the major mountain ranges of the world such as the Andes, Rockies, Alps, and Himalayas.



Figure 11.1.2: Greenland ice sheet.

The other major glacier type is a **continental glacier** (also called an ice sheet). These are thick accumulations of ice that occupy a large geographical area. The main ice sheets on the earth today are located on Greenland and Antarctica. The Greenland Ice Sheet has an extensive surface area and thickness up to two miles [1].

The Antarctic Ice Sheet is much larger and covers almost the entire continent. The thickest parts of this massive ice sheet are over 2.5 miles thick 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.

Former ice sheets, present during the last glacial maximum event (also known as the last ice age) in North America, are called the Laurentide Ice Sheet.

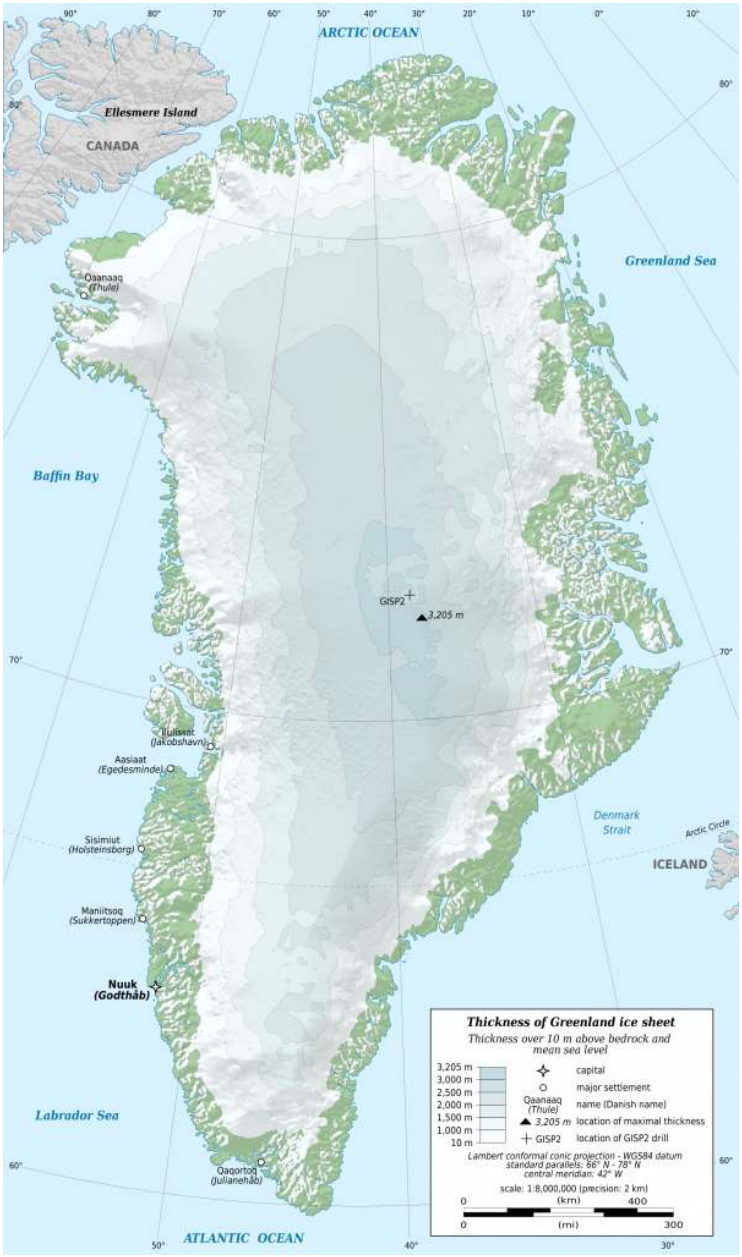


Figure 11.1.3: Thickness of Greenland ice sheet in meters (Source: Eric Gaba).

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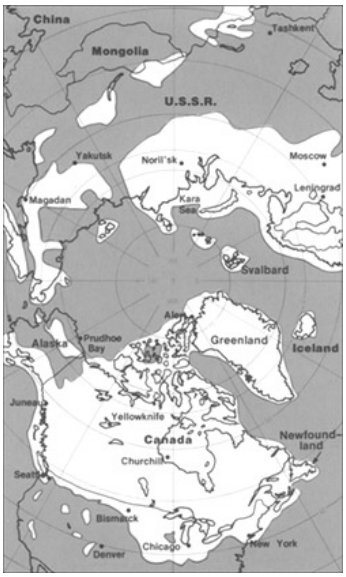


Figure 11.1.4: Maximum extent of Laurentide Ice Sheet

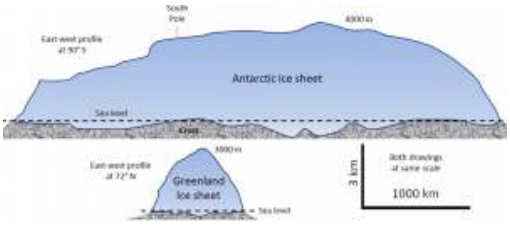


Figure 11.1.5: Cross-sectional view of both Greenland and Antarctic ice sheets drawn to scale for size comparison (Source: Steve Earle)

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11.2: Glacier Formation and Movement

Glaciers form when accumulating snow compresses and eventually turns into ice. In some cases, perennial snow accumulates on the ground and lasts all year. This makes a snowfield and not a glacier since it is a thin accumulation of snow. Snow and glacial ice actually have a fair amount of void space (porosity) that traps air. As the snow settles, compacts, and bonds with underlying snow, the amount of void space diminishes. When the snow gets buried by more snow, it compacts into granular **firn** with less air and it begins to resemble ice more than snow. Continual burial, compression, and recrystallization make the firn denser and more ice-like. Eventually, the accumulated snow turns fully to ice, however, small air pockets remain trapped in the ice and form a record of the past atmosphere.



Figure 11.2.1: Glacial crevasses.



Figure 11.2.2: Crevasse on the Easton Glacier in the North Cascades

As the ice accumulates, it begins to flow downward under its own weight. An early study of glacier movement conducted in 1948 on the Jungfraufirn Glacier in the Alps installed hollow vertical rods in the ice and measured the tilt over two years. The study found that the top part was fairly rigid and the bottom part flowed internally, more like a plastic. A P-T diagram of ice shows that 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 a 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 to form large cracks along the top of a glacier called **crevasses**. These crevasses can be covered and hidden by a snow bridge and thus are a hazard for glacier travelers.

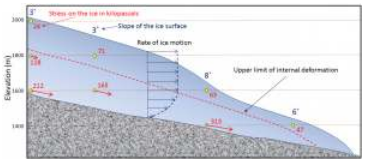


Figure 11.2.3: Cross-section of a valley glacier showing stress (red numbers) increases with depth under the ice. The ice will deform and flow where the stress is greater than 100 kilopascals, and the relative extent of that deformation is depicted by the red arrows. Downslope movement is shown with blue arrows. The upper ice above the red dashed line does not flow but is pushed along en masse. (Source: Steve Earle)

Below the brittle zone, there is so much weight of the overlying ice that it no longer breaks when force is applied to it but rather it bends or flows. This is the **plastic zone** and, within this zone, the ice flows. The plastic zone represents the great majority of the ice of a glacier and often contains a fair amount of sediment from as large as boulders and as small as silt and clay which act as grinding agents. The bottom of the plastic zone slides and grinds across the bedrock surface and represents the zone of erosion.

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11.3: Glacial Budget

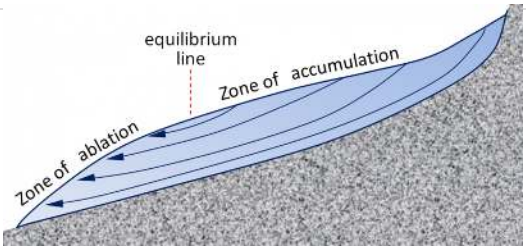


Figure 11.3.1: Cross-sectional view of an alpine glacier showing internal flow lines, zone of accumulation, snow line, and zone of melting.

Glaciers gain mass during the winter as snow accumulates. During summer the snow melts. The glacier is like a bank account, if more money is coming in (snow accumulating in winter) than going out (snow melting in summer), then the bank account grows. The glacial budget works in a similar way. The balance of accumulating ice (**zone of accumulation**) is weighed against melting ice (**zone of ablation** or zone of melting), and whichever is greater determines whether the glacier will advance or retreat. In the zone of accumulation, the rate of annual snowfall is greater than the rate of melting. In other words, not all of the snow that falls each winter melts during the following summer, and the ice surface is always covered with snow. In the zone of melting or ablation, more ice melts then accumulates as snow during the year. The **equilibrium line** marks the boundary between the zones of accumulation and ablation. Below the equilibrium line, in the zone of melting, bare ice is exposed because last winter's snow has all melted; above that line, the ice is still mostly covered with snow from last winter. The position of the equilibrium line changes from year to year as a function of the balance between snow accumulation in the winter and snowmelt during the summer. More winter snow and less summer melting obviously favors the advance of the equilibrium line (and of the glacier's leading-edge or **terminus**), but of these two variables, it is the summer melt that matters most to a glacier's budget. Cool summers promote glacial advance and warm summers promote glacial retreat [5].

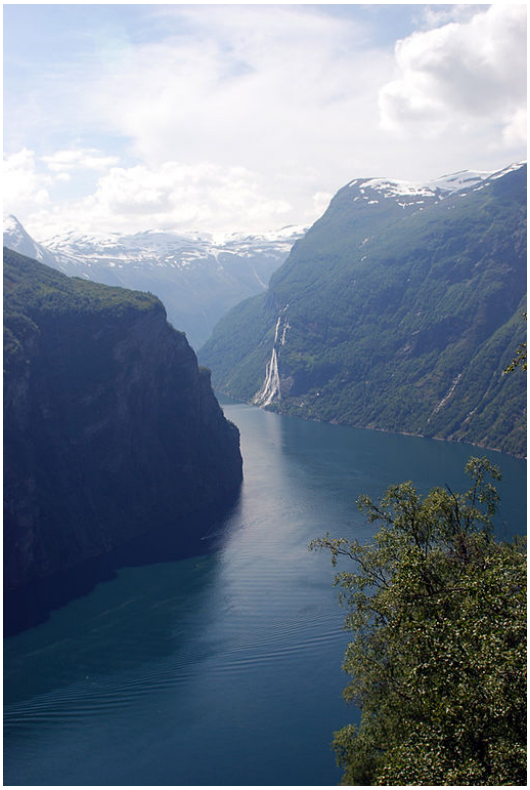


Figure 11.3.2: Fjord

If warmer summers promote glacial retreat, then overall climate warming over many decades and centuries causes the glacier to melt and retreat significantly. Since the global climate has been warming [6], this warming is likely causing the ice sheets to melt (or lose mass) at an increasing rate over years and decades rather than over centuries [7].

When ice sheets start to melt, such as those in Antarctica and Greenland, their flow into the ocean speeds up eventually creating floating ice sheets. The edges of the glacier or its extension as floating ice break off in a process called **calving**.

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11.4: Glacial Landforms

Glacial landforms are of two kinds, erosional and depositional landforms. Erosional landforms are formed by removing material and consist of the material that remains. Depositional landforms are formed by the deposition of sediment.

Erosional Glacial Landforms

Both alpine and continental glaciers erode bedrock and create erosional landforms. Glaciers are heavy masses of abrasive ice that grind over the surface. Elongated grooves are created by fragments of rock embedded in the ice at the base of a glacier scraping along the bedrock surface called **glacial striations**. In addition, rock flour as fine grit in the ice can polish a hard granite or quartzite bedrock to a smooth surface called **glacial polish**. This figure illustrates these glacial landforms.



Figure 11.4.1: A polished and striated rock in Yosemite.



Figure 11.4.2: Glacial striations in Mt. Rainier National Park.

Following is a description of erosional landforms produced by alpine glaciers. Since glaciers are typically much wider than rivers of similar length, and since they tend to erode more at their bases than their sides, they transform former V-shaped stream valleys into broad valleys that have relatively flat bottoms and steep sides with a distinctive “U” shape [5]. In contrast, river-carved canyons have a V-shaped profile when viewed in cross-section.



Figure 11.4.3: The U-shape of the Little Cottonwood Canyon, Utah, as it enters into the Salt Lake Valley.



Figure 11.4.4: V-shaped Big Cottonwood Canyon, carved by a stream.

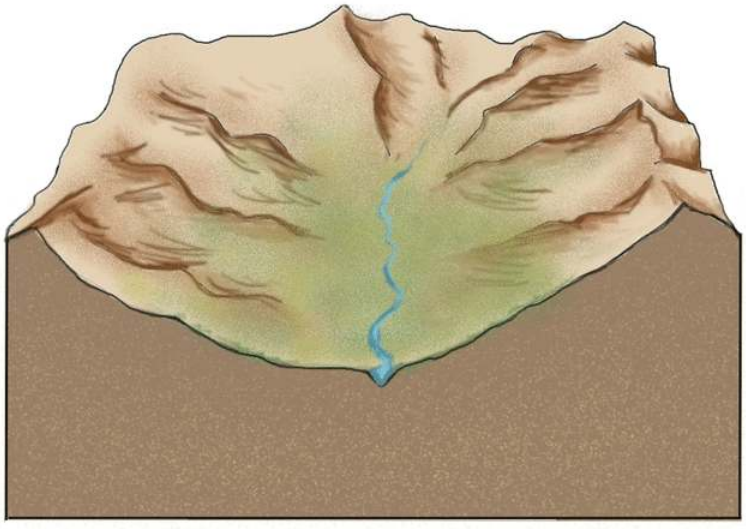


Figure 11.4.5: Formation of a glacial valley. Glaciers change the shape of the valley from a “V” shape to a “U”.

When two U-shaped valleys are adjacent to each other, the ridge between them can be carved into a sharp ridge called an **arête**. Since glaciers erode a broad valley, the arêtes are left behind with steep walls separating them. At the head of a glacially carved valley is a bowl-shaped feature called a **cirque** representing where the head of the glacier is eroding against the mountain by plucking rock away from it and the weight of the thick ice is eroding out a bowl. Headward cirque erosion by three or more mountain glaciers produces **horns,** which are steep-sided, spire-shaped mountains with pronounced cirques on three or more sides. When a smaller tributary glacier intersects a larger trunk glacier, the smaller glacier erodes down less. Therefore, once the ice has been removed, the tributary valley is left as a **hanging valley**, sometimes with a waterfall.



Figure 11.4.6: Cirque with Upper Thornton Lake in the North Cascades National Park, Washington. Photo by Walter Siegmund.



Figure 11.4.7: An example of a horn, Kinnerly Peak, Glacier National Park, Montana. Photo by USGS.

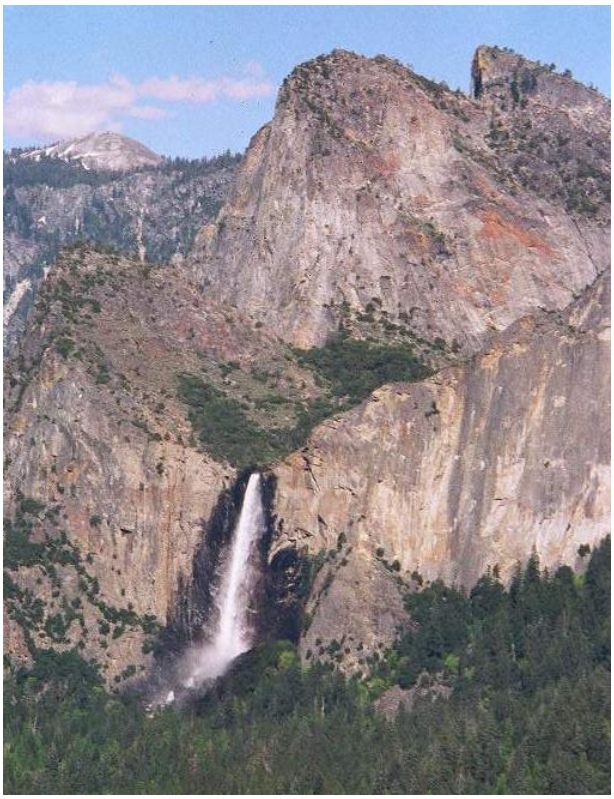


Figure 11.4.8: Bridalveil Falls in Yosemite National Park, California, is a good example of a hanging valley.

Depositional Glacial Landforms



Figure 11.4.9: Boulder of diamictite of the Mineral Fork Formation, Antelope Island, Utah, United States.

Sediment is deposited by glaciers in both alpine and continental environments. Since ice is responsible for a large amount of erosion, there is a lot of sediment in glacial ice. When sediment is left behind by a melting glacier, it is called **till** and is characteristically poorly sorted with grain sizes ranging from clay and silt to pebbles and boulders. Many of the depositional landforms described in this section are composed of till.

Material carried by the glacier is called **moraine**, which is an accumulation of glacial till produced by the grinding and erosive effects of a glacier. In valley glaciers, moraine also includes material falling on the sides of the glacier by mass wasting from the valley walls. The glacier acts like a conveyor belt, carrying sediment inside and on the ice and depositing it at the end and sides of the glacier. Because ice in the glacier is always flowing downslope, the deposited moraines build up even if the end of the glacier doesn't advance. The type of moraine depends on its location. A **terminal moraine** is a ridge of unsorted till at the maximum extent of a glacier [11]. When glaciers retreat in episodes, they may leave behind deposits called **recessional moraines**. The recessional moraines look similar to terminal moraines but are formed when the glacier retreat pauses. Moraines located along the side of a glacier are called **lateral moraines** and mostly represent material that fell on the sides of the glacier from the mass wasting of the valley walls. When two tributary glaciers join together, the two lateral moraines combine to form a **medial moraine**. Behind the terminal and recessional moraines is a veneer (or thin sheet) of till on top of bedrock called **ground moraine**.



Figure 11.4.10: Lateral moraines of Little Cottonwood and Bell canyons, Utah.



Figure 11.4.11: Medial moraines where tributary glaciers meet. At least seven tributary glaciers from upstream have joined to form the trunk glacier flowing on out of the upper left of the picture.

In addition to moraines, as glaciers melt they leave behind other depositional landforms. The intense grinding process creates a lot of silt. Streams of water melting from the glacier carry this silt (along with sand and gravel) and deposit it in front of the glacier in an area called an **outwash plain**. In addition, when glaciers retreat, they may leave behind large boulders of a type of rock that doesn't match the local bedrock. These are called **glacial erratics**. When continental glaciers melt, large blocks of ice can be left behind to melt within the impermeable till and can create a depression called a **kettle** that can be later filled with surface water like a **kettle lake**.



Figure 11.4.12: A small group of Ice Age drumlins in Germany.

Also common in continental glacial areas of New York state and Wisconsin are drumlins. A **drumlin** is an elongated asymmetrical drop-shaped hill with its steepest side pointing upstream to the flow of ice and streamlined side (low angle side) pointing in the direction the ice is flowing.

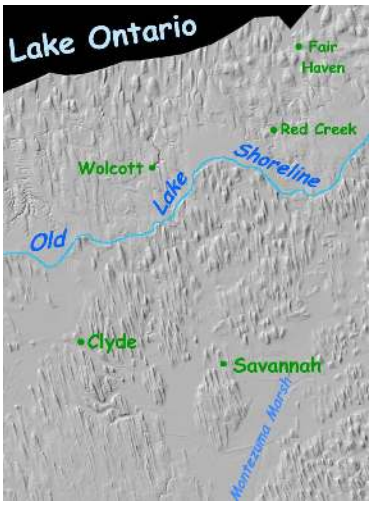
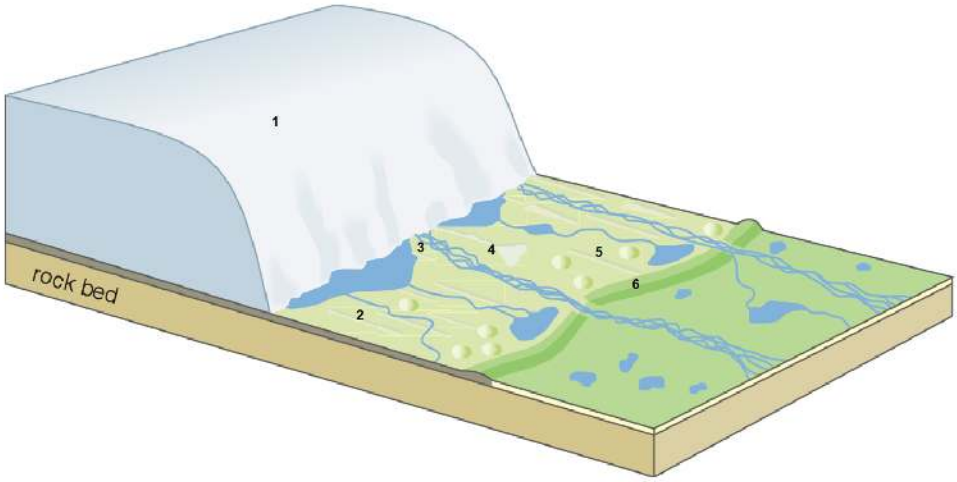


Figure 11.4.13: Drumlin field in Wayne County, New York. Their shape indicates the glacier moved in a NNW to SSE direction.

Drumlins can occur in great numbers in drumlin fields. The origin of drumlins is still debated and some leading ideas are the incremental accumulation of till under the glacier, large catastrophic meltwater floods located under the glacier, and surface deformation by the weight of the overlying glacial ice [12; 13].



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Number	Name	Note
1	Receding Glacier	Sheet of ice with a terminus that is moving back with time.
2	Ground Morraine	N/A
3	Esker	N/A
4	Esker	
5	Drumlin	N/A
6	Terminal Morraine	

Glacial Lakes



Figure 11.4.14: Tarn in a cirque.

Lakes are common features in alpine glacial environments. A lake that is confined to a glacial cirque is known as a **tarn** such as Avalanche Lake in Glacier National Park. Tarns are common in areas of alpine glaciation because the thick ice that carves out a cirque also typically hollows out a depression in bedrock that after the glacier is gone, fills with precipitated water.



Figure 11.4.15: Paternoster lakes

In some cases, recessional moraines may isolate a series of basins within a glaciated valley, and the resulting chain of lakes is called **paternoster lakes**.

During the last glaciation, most of the western United States had a cooler and wetter climate. Due to less evaporation and more precipitation, many large lakes formed in the basins of the Basin and Range Province called pluvial lakes. One of the largest pluvial lakes was Lake Bonneville. Lake Bonneville occupied much of western Utah and eastern Nevada (Figure from Miller et al 2013) and filled Salt Lake Valley, which is densely urbanized today, with water hundreds of feet deep. During the last glaciation, the level of the lake fluctuated many times leaving several pronounced old shorelines in the western portion of Utah including Salt Lake Valley identified by wave-cut terraces. Lake level peaked around 18,000 years ago [14] and spilled over Red Rock Pass in Idaho into the Snake River causing rapid erosion and a very large flood that rapidly lowered the lake level and scoured the land in Pocatello Valley, Snake River Plain, and Twin Falls Idaho. The flood ultimately flowed into the Columbia River across part of the scablands area and had an incredible discharge of about 4,750 km^3 [15] which is about the volume of Lake Michigan. This would be as if one of the Great Lakes completely emptied within days.

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11.5: Ice Age Glaciations

A **glaciation** (or ice age) occurs when the Earth's climate is cold enough that large ice sheets grow on continents. There have been four major, well-documented glaciations in Earth's history: one during the Archean-early Proterozoic (~2.5 billion years ago), another in late Proterozoic (~700 million years ago), another in the Pennsylvanian (323 to 300 million years ago), and the most recent Pliocene-Quaternary glaciation (2.5 million years ago to now). The best-studied glaciation is, of course, the most recent. The Pliocene-Quaternary glaciation is a series of many glacial cycles, possibly 18, during the last 2.5 million years. The last of these, known in popular media as "The Ice Age" but known by geologists as the last glacial maximum, reached its height between 26,500 and 19,000 years ago [10; 17]. The warm periods between these glacials are called interglacials. We are currently in an interglacial.

Causes of Glaciations

Why do glaciations occur? The causes include both long-term and short-term factors. In the geologic sense, long-term means a scale of 10's to 100's of millions of years and short-term means a 100 to 200,000-year scale. Ideas about long-term causes of glaciations over geologic time include the positioning of continents near poles by plate tectonics and changes in ocean circulation due to re-positioning of the continents such as the closing of the Panama Strait. Short-term factors are more recognizable for the most recent Pliocene-Quaternary Glaciation and are most relevant to today's anthropogenic climate change, but may have taken place in the earlier glaciations.

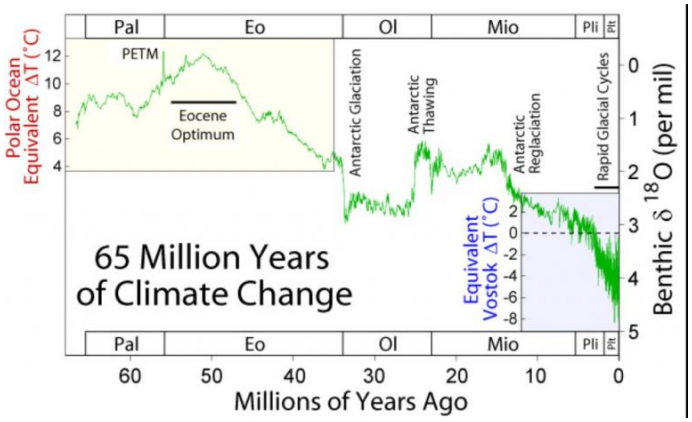
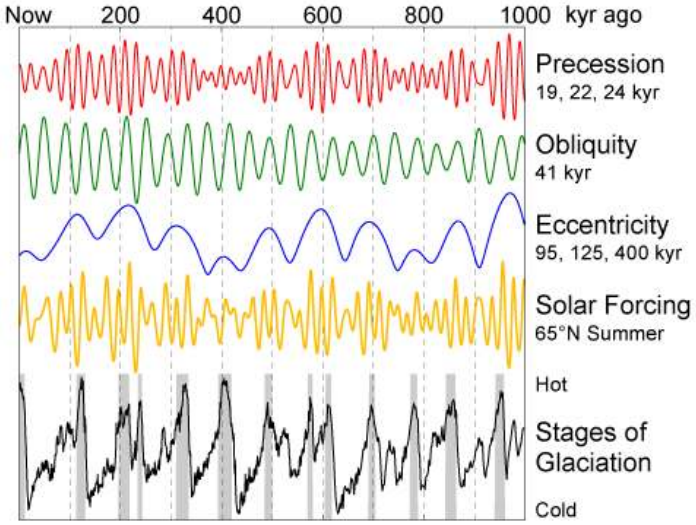
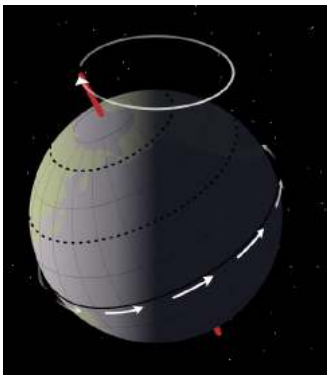
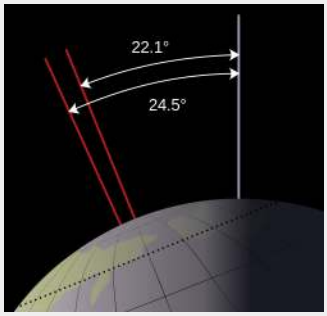
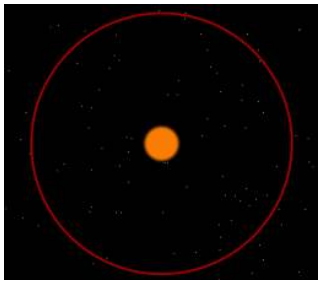


Figure 11.5.1: Atmospheric CO_2 has declined during the Cenozoic from a maximum in the Paleocene-Eocene to its dramatic rise since the Industrial Revolution.

Short-term causes of glacial fluctuations are attributed to cycles in the rotational axis of the Earth and in Earth-Sun relations due to variations in the earth's orbit called **Milankovitch Cycles**. These cycles affect the amount of incoming solar radiation, and changes in carbon dioxide in the atmosphere. During the Cenozoic, carbon dioxide levels steadily decreased from a maximum in the Paleocene causing a gradual climatic cooling [18]. As the climate cooled, the effects of the Milankovitch Cycles began to influence climate with regular cycles of warming and cooling. Milankovitch Cycles are three orbital changes named after the Serbian astronomer Milutin Milankovitch. The three orbital changes are the wobbling of Earth's axis called **precession** with a span of 21,000 years, the angle of Earth's axis called **obliquity** with a span of about 41,000 years, and variations of the distance from the sun in Earth's orbit around the sun referred to as **eccentricity** with a span of 93,000 years [19]. These orbital changes created a glacial-interglacial cycle of 41,000 years from 2.5 to 1.0 million years ago and a longer cycle of ~100,000 years from 1.0 million years ago to today. The combination of these three Milankovitch Cycles changes the angle at which the sun's energy strikes the surface of the earth near the poles and the amount of energy (insolation) received by Earth. As the climate cooled during the Cenozoic Era, the subtle changes in energy received by the planet were expressed as a warmer and cooler climate cycle, thus the glacial-interglacial cycles.

This chart illustrates the effect of the Milankovitch Cycles.



Name	Illustration
Precession	
Obliquity	
Eccentricity	

This video summarizes ice ages: their characteristics and causes.

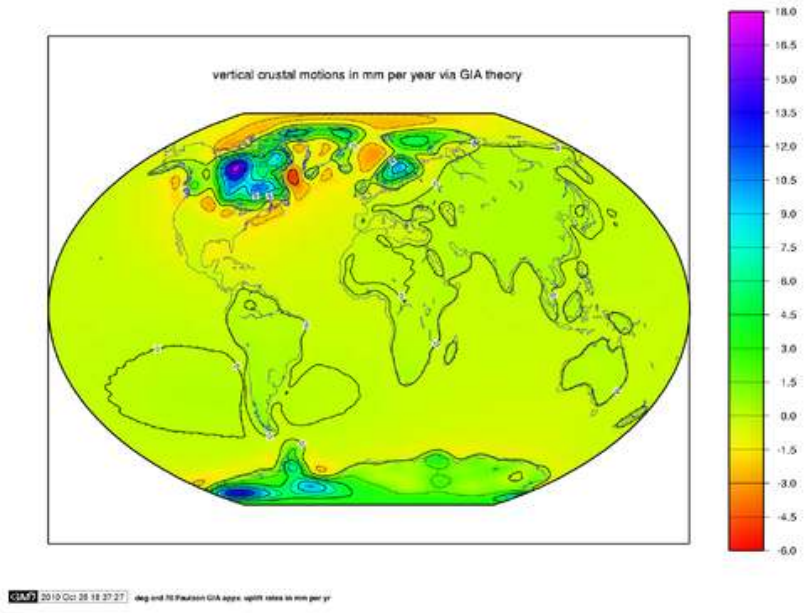
Sea-Level Change and Isostatic Rebound

Since glaciers are ice located on land (not floating in the ocean), when glaciers melt and retreat two things happen, sea-level rises globally and the land rises locally due to isostatic rebound. Melting glacial water runs off into the ocean and sea-level worldwide will rise. For example, since the last glacial maximum about 19,000 years ago [17] sea-level has risen about 400 feet (125 meters) [20]. Another important factor causing sea-level rise is thermal expansion. According to basic physics, thermal expansion occurs when a solid, liquid, or gas expands in volume when the temperature increases. About half of the sea-level rise during the last century has been the result of thermal expansion, the rest from the melting of glaciers [21; 22].

However, tectonics and isostatic rebound can move the land up and down. The change of sea level as it relates to a more local continental landscape is called relative sea-level change. The relative sea-level change includes both the vertical movement of the sea-level and the vertical movement of land, so that the sea-level change is measured *relative* to the land. Therefore, if the land rises a lot and sea-level rises only a little, then the sea-level would appear to go down.

The lithosphere can move vertically as a result of two main processes, tectonics, and isostatic rebound. Tectonic uplift occurs when tectonic plates collide as discussed in the Plate Tectonics chapter. **Isostasy** describes the equilibrium that exists for the earth's lithosphere, where denser lithosphere "sinks" lower on top of the asthenosphere and less dense lithosphere "floats" higher on the asthenosphere. **Isostatic rebound** is when some weight is removed from the continental lithosphere causing it to "float" higher on the asthenosphere. Erosion can remove this weight very slowly or the relatively rapid removal of glaciers can remove a lot of weight in a short amount of time. Melting glaciers removes weight from the continental lithosphere causing it to rise or "rebound" from being previously depressed. Most glacial isostatic rebound is occurring where glaciers recently melted (19,000 years ago) such as Canada and Scandinavia. Glacial isostatic rebound causes the relative sea-level to fall or rise less quickly as seen from the land. Isostatic rebound also occurred in Utah when the water from Lake Bonneville was removed [23]. Isostatic rebound is still taking place wherever Ice Age ice or water bodies were present on continental surfaces.

This map shows the rates of vertical crustal movement worldwide. Note that the greatest upward movement occurs in regions affected by recent glaciation, isostatic rebound. Also, note that crustal depression has also occurred in adjacent regions as subcrustal material displaced by isostatic lowering from the weight of the ice has flowed back in under the rebound.



Rates of isostatic rebound worldwide, greatest in regions of recent glaciation.

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CHAPTER OVERVIEW

12: Coastlines

Coastlines are the great interface between the 29% of earth's surface that is land and 71% of earth that is covered by the oceans. Therefore, it is the longest visible boundary on earth. To understand the processes that take place there, we must first consider waves, currents, and tides since they are the main agents that shape shorelines. Most coastal landforms can be attributed to moving sand via longshore drift, longterm rising sea levels, or longterm falling sea levels. Human intervention in beach processes, like jetties, have negative consequences that need to be mitigated.

12.1: Waves and Wave Processes

12.2: Shoreline Features

12.3: Current and Tides

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12.1: Waves and Wave Processes

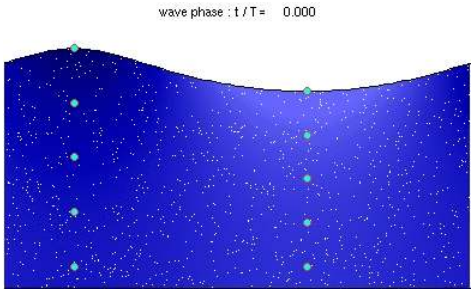


Figure 12.1.1: Particle motion within a wind-blown wave.

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, the water moving forward with the crest and moving backward in the trough. This can be demonstrated by watching the movement of a cork or some floating object as a wave passes.

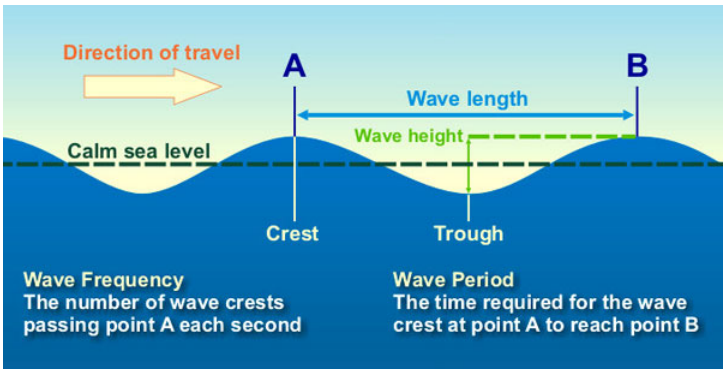


Figure 12.1.2: Aspects of water waves, labeled.

Important terms to understand in the operation of waves include: **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. **Wavelength** is the horizontal distance between adjacent wave crests or corresponding features of the wave.

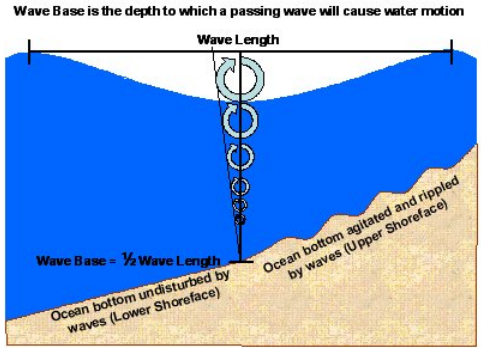


Figure 12.1.3: Diagram describing the wave base.

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. The vertical reach of waves in the water is called the **wave base**. Looking at incoming waves at a beach, one will appreciate that most ocean waves have a wavelength on the order of a few tens of feet. Thus, seawater is typically disturbed by a wave motion to a depth of a few tens of feet. This is known as a fair-weather wave base. In strong storms such as hurricanes, both wave length 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 300 feet [1].

Since 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 were typically generated hundreds of miles from the coast by storms and may have been traveling across the ocean for days.

Behavior of Waves Approaching Shore

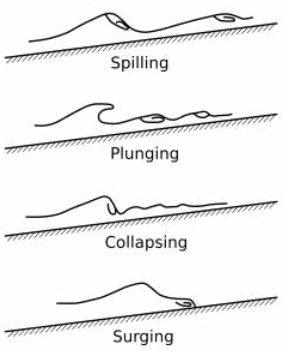


Figure 12.1.4: Types of breakers

On the open sea, waves generally appear choppy because wave trains from many directions are interacting with each other. Where crests converge with other crests (called constructive interference) they add together producing peaks, a process referred to as wave amplification. Constructive interference of troughs produces hollows. Where crests converge with troughs, they cancel each other out (called destructive interference). 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, but 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 front of the wave to supply a complete circle, and the crest pours over creating a **breaker**.

A special type of wave is generated by any energetic event affecting the seafloor, such as earthquakes, submarine landslides, and volcanic eruptions. Such waves are called **tsunamis** and, in the case of earthquakes, are created when a portion of the seafloor is suddenly elevated by movement in the crustal rocks below that are involved in the earthquake. The water is suddenly lifted and waves spread out in all directions from the mound carrying enormous energy and traveling very fast (hundreds of miles per hour). Tsunamis may pass unnoticed in the open ocean because the wavelength is very long and the wave height is very low. But as the waves approach the shore, each wave makes contact with the shallow seafloor, friction increases, and the wave slows down. Wave height builds up and the wave strikes the shore as a wall of water a hundred or more feet high. The massive wave may sweep inland well beyond the beach. Tsunamis deliver a catastrophic blow to observers at the beach as the water in the trough in front of it is drawn back toward the tsunami wave, exposing the seafloor. 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.

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12.2: Shoreline Features

Many different erosional and depositional features exist in the high energy of the coast. The **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 **shoreline** itself is the direct interface between water and land that migrates with the tides and with deposition and erosion of sediment. So the coastline is an area and the shoreline is a line that moves.

Shoreline Zones

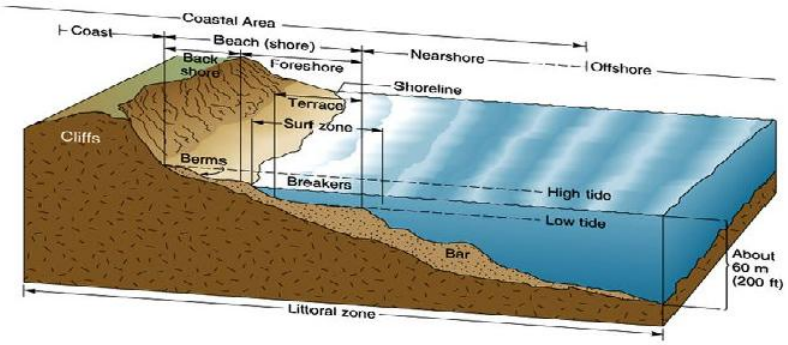


Figure 12.2.1: Diagram of zones of the shoreline.

Shoreline zones can be viewed by looking at the beach profile, which is divided into four primary zones – offshore, nearshore, foreshore, and backshore. The **offshore** is below any shoreline-derived process but is still geologically active due to underwater landslides and deeper currents. The **nearshore** is affected by the waves, i.e., that part of the shore where water depth is a half wavelength or less. The width of this zone thus depends on the maximum wavelength of the approaching waves and with the slope of the seafloor. The **surf zone** is where the waves break. The area (mostly overlapping the surf zone) that is periodically wet and dry, because of wave action and tides, is called the **foreshore**, which is made of planer-laminated, well-sorted sand.

The **beach face** is where the swash of the breaking wave runs up and the backwash flows back down. Above the beach face are low ridges called **berms** which is where beachgoers park their items. There may be a zone of dunes behind the berms representing sand blown behind the beach by onshore winds. This area behind the berms that is always above the ocean in normal conditions is known as the **backshore**.

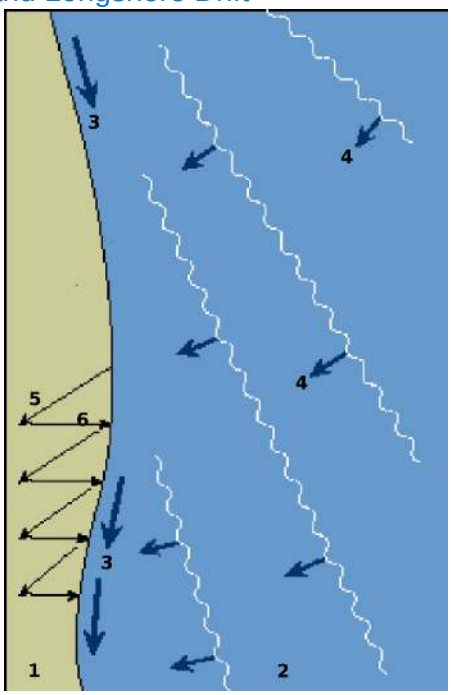


Figure 12.2.2: Longshore Drift. 1=beach, 2=sea, 3=longshore current direction, 4=incoming waves, 5=swash, 6=backwash

As waves enter shallower water, they slow down. Waves usually approach the shoreline at an angle, with one end of the waves slowing down first. This causes the waves to bend toward the beach. Such bending of the waves as they enter shallower water is called **wave refraction**, which produces the appearance from the beach that waves are approaching the beach face generally straight on, parallel to the beach. However, refracted waves on average approach the shoreline at somewhat of an angle creating a slight difference between the swash as it moves up the beach face and the backwash as it flows back down. This results in a net movement of the water along the beach creating a current called the **longshore current**. Sand stirred up by waves in the surf zone is thus moved along the shore by **longshore drift**. Longshore drift along both the west and east coasts of North America moves sand on average from north to south.



Figure 12.2.3: Farewell Spit, New Zealand

Longshore drift can be carried down the coast until it reaches a bay or inlet where it begins to deposit in the quieter water. Here, a **spit** begins to form. As the spit grows, it may extend across the mouth of the bay forming a **baymouth bar**. Where the bay or inlet serves as the anchorage for boats, such spit growth and baymouth bars are a severe inconvenience. Communities thus affected attempt measures to keep their harbor open.



Figure 12.2.4: Jetties near Carlsbad, California. Notice the left jetty is loaded with sand, while the right jetty is lacking sand. This is due to the longshore drift going left to right.

One means to do this is a **jetty**, often built of concrete or large stones, forming a long barrier to deflect the sand away from the harbor mouth or other ocean waterways in which transport is desired. If the jetty does not succeed in deflecting the sand far enough out, it may continue to flow along the shore, building a spit around the end of the jetty. A more expensive but effective method is then to dredge the sand from the growing spit, put it on barges, and deliver it back to the drift downstream of the harbor opening. An even more expensive (but effective) means 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.

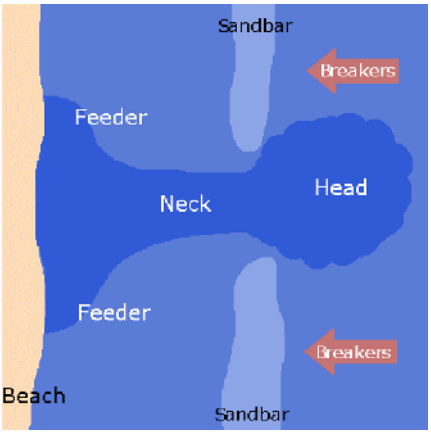


Figure 12.2.5: Animation of rip currents.

Another coastal phenomenon related to longshore currents is the presence of **rip currents**. These involve the nearshore configuration of the seafloor and/or the arrival of waves straight onto the shore**.** In areas where wave motion pushes water directly toward the beach, or the shape of the nearshore seafloor refracts and focuses the water movement toward a point on the beach, the water piling up there must find an outlet back to the sea. The outlet is provided by relatively narrow rip currents that carry the water directly away from the beach. Swimmers caught in such currents find themselves being carried out to sea. They may attempt to return to shore by swimming directly against the current. This is generally a fruitless effort because they tire against the strong current. A better solution is to ride it out to where it dissipates, then swim around it and return to the beach or swim laterally, parallel to the beach, until out of the current, then return to the beach.

Emergent and Submergent Coasts



Figure 12.2.6: Island Arch, a sea arch in Victoria, Australia.

Coastlines that have a relative fall in sea level, either caused by uplift of land due to tectonics or sea level change due to sea level falling, are called **emergent**. Where the shoreline is rocky, perhaps with a sea cliff, there are often headlands, or areas that more effectively resisted erosion and extend out into the ocean. Waves refracting around headlands attack the rocks behind the point of the headland.



Figure 12.2.7: This tombolo, called “Angel Road,” connects the stack of Shodo Island, Japan.

They may cut out the rock at the base forming a **sea arch** which may collapse to isolate the point as a **sea stack.** Rocks behind the stack may be eroded away and sand eroded from the point collects behind it forming a **tombolo**, a sand strip that connects the stack to the shoreline. Where sand supply is low, wave energy may erode a **wave-cut platform** across the surf zone, exposed as a bare rock with tidal pools at low tide.

Sea cliffs tend to be persistent features as the waves cut away at their base and higher rocks calve off by mass wasting. If the coast is emergent, these erosional features may be elevated relative to the wave zone. Wave-cut platforms become **marine terraces**, perhaps with remnant sea cliffs inland from them.



Figure 12.2.8: Landsat image of Chesapeake Bay, eastern United States. Note the barrier islands parallel to the coastline.

Tectonic subsidence or sea level rise produces a **submergent** coast. Features associated with the submergent coasts include estuaries, bays and river mouths flooded by the higher water. **Fjords** are former glacial valleys now flooded by sea level rise caused by the melting of glaciers. Elongated bodies of sand called **barrier islands** form parallel to the shoreline from the old beach sands, often isolated from the mainland by lagoons behind them [3].

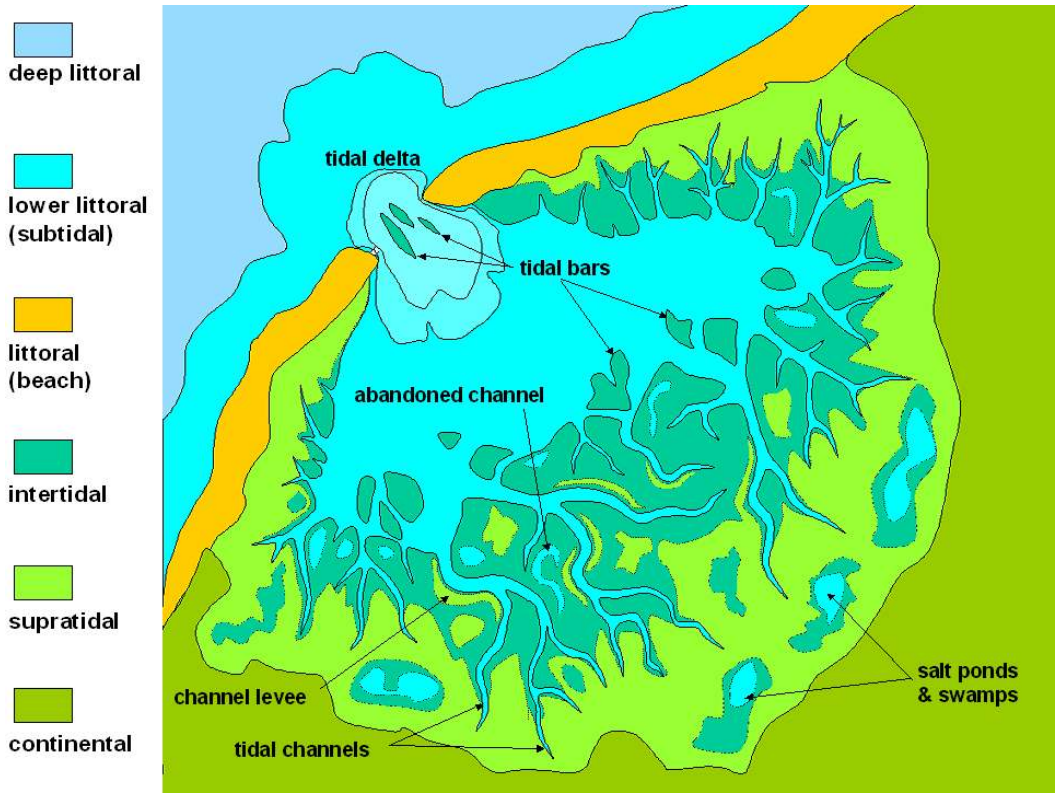


Figure 12.2.9: General diagram of a tidal flat and associated features.

Tidal flats or mudflats form where tides flood and expose low areas along the coast as the tide rises and falls. 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 ripples and cross-bedding common. Marshes are vegetated with common 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 12.2.10: Kara-Bogaz Gol lagoon, Turkmenistan.

Lagoons are locations where spits, barrier islands, or other features have partially cut off a body of water from the ocean. **Estuaries** are a (typically vegetated) type of lagoon where freshwater is flowing into the area as well, making the water a mix of 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 geologic environments, where beach, lake, and river processes can overlap.

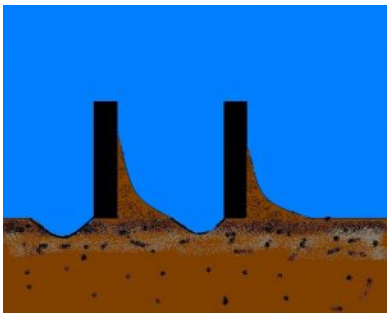


Figure 12.2.11: Groins gathering sediment from longshore drift.

Coasts are prime real estate land that attracts the development of beach houses, condominiums, and hotels. This kind of interest and investment leads to ongoing efforts to manage the natural processes in coastal areas. Humans who find longshore drift is removing sand from their beaches often use **groins** in an attempt to retain it.



Figure 12.2.12: Groin system on a coast in Virginia

Similar but smaller than jetties, groins are bits of wood or concrete built across the beach perpendicular to the shoreline at the downstream end of one's property. Unlike jetties, they are used to preserve sand on a beach, rather than to divert it from an area. Sand erodes on the downstream side of the groin and collects against the upstream side. Every groin thus creates the need for another one downstream. The series of groins along a beach develops a scalloped appearance for the shoreline.

Sand for longshore drift and beaches comes from rivers flowing to the oceans from inland areas. Beaches may become starved of sand if sediment carried by streams and rivers is trapped behind dams. To mitigate this, beach replenishment may be employed where sand is 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 sandy material and foreign microorganisms and can even bring in foreign objects that impact humans on replenished beaches. Visitors to one replenished east coast beach found munitions and metal shards in the sand which had been brought from abandoned test ranges from which the sand had been dredged [6].



Figure 12.2.13: A tombolo has formed behind the breakwater at Venice, CA

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.

Submarine Canyons

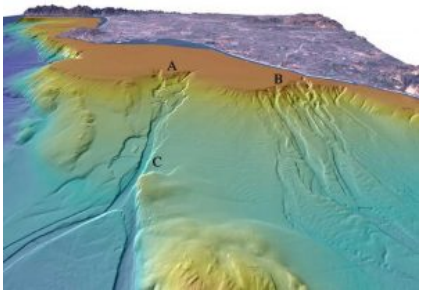


Figure 12.2.14: Submarine canyons off of Los Angeles. A=San Gabriel Canyon, B=Newport Canyon. At point C, the canyon is 815 m wide and 25 m deep.

Submarine canyons are narrow and deep canyons located in the marine environment on continental shelves. They typically form at the mouths of large landward river systems, both by cutting down into the continental shelf during times of low sea level and also by continual material slumping or flowing down from the mouth of the river or a delta. Underwater currents rich in sediment pass through the canyons, erode them and drain onto the ocean floor. Steep delta faces and underwater flows of sediments are released down the continental slope as underwater landslides.

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12.3: Current and Tides

Water in the ocean, when moving, can move via waves, currents, and tides. Waves have been discussed in chapter 12.1, and this section will focus on the other two. Currents in the ocean are driven by persistent global winds blowing over the surface of the water and water density. They are part of the Earth's heat engine in which solar energy is absorbed by the ocean water (remember the specific heat of water). The absorbed energy is distributed by ocean currents.

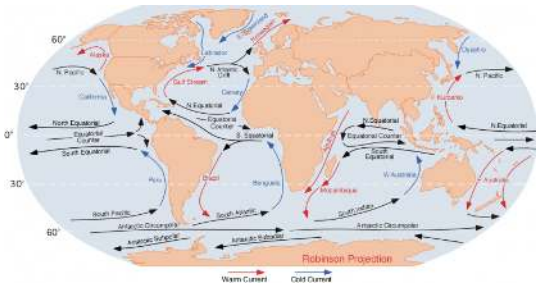


Figure 12.3.1: World ocean currents.

Surface Currents

In the above figure, notice the large sub-circular currents in the northern and southern hemispheres in the Atlantic, Pacific, and Indian Oceans. These are driven by prevailing atmospheric circulation and are called **gyres** [9] and rotate clockwise in the northern hemisphere and counterclockwise 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 and cold currents affect 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 Humboldt Current flowing north along the west coast of South America.

Deep Currents

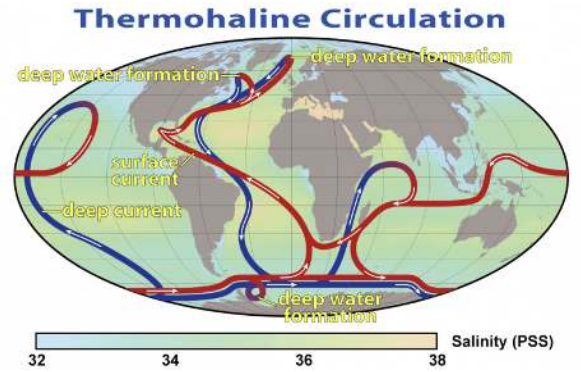


Figure 12.3.2: Global thermohaline circulation. PSS=practical salinity units.

Whether an ocean current moves horizontally or vertically depends on its density. The density of seawater is determined by factors such as temperature and salinity. Evaporation and the influx of freshwater from rivers also affect salinity and therefore the density of seawater. As the western boundary currents cool, some of the cool, denser water sinks to become the deep water of the oceans. Movement of this deep water is called the **thermohaline circulation** (*thermo* refers to temperature, *haline* refers to salinity) and connects the deep waters of all the world's oceans. This can be best illustrated by the Gulf Stream. After the warm water within the current promotes much evaporation and the heat dissipates, the resulting water is much colder and saltier. As the denser water reaches the North Atlantic and Greenland, it begins to descend and becomes a deepwater current.

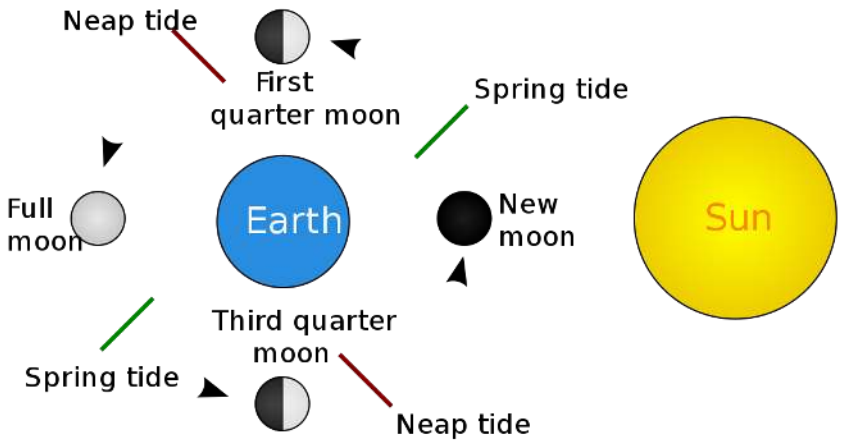


Figure 12.3.3: Diagram showing tides, in relation to the sun and moon.

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 exerts a more powerful gravity, 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, 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. 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 a **neap tide**.

Typical tidal ranges are on the order of 3 feet. Extreme tidal ranges occur where the tidal wave enters a restrictive zone. An example is the English Channel between Great Britain and the European continent where tidal ranges of 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 40 feet and extremes of around 60 feet have been observed. 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.

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CHAPTER OVERVIEW

13: Deserts

Approximately 30 percent 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. The lack of moisture, including the lack of humidity and cloud cover, allow temperature extremes to occur. The sun’s energy is more absorbed by the Earth’s surface without cloud cover, and nighttime cooling is more drastic without cloud cover and humidity to absorb the emitted heat, so temperature extremes are common in deserts.

13.1: Prelude to 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

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13.1: Prelude to Deserts



Figure 13.1.1: A playa filled with evaporite minerals (such as gypsum) erodes and forms ripple-covered dunes in White Sands National Monument, New Mexico.

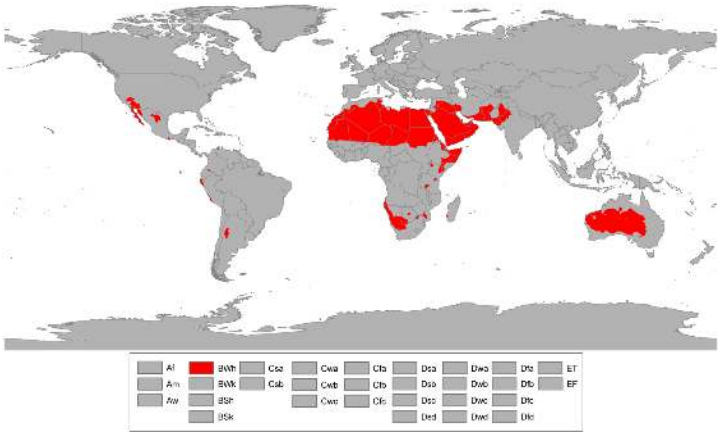


Figure 13.1.2: World hot deserts (Koppen BWh)

The location of climates on Earth's surface is not random. Jungles, tundras, and deserts have scientific explanations for their locations. Approximately 30 percent 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. The sun's energy is more absorbed by the Earth's surface without cloud cover, and nighttime cooling is more drastic without cloud cover and humidity to absorb the emitted heat, so temperature extremes are common.

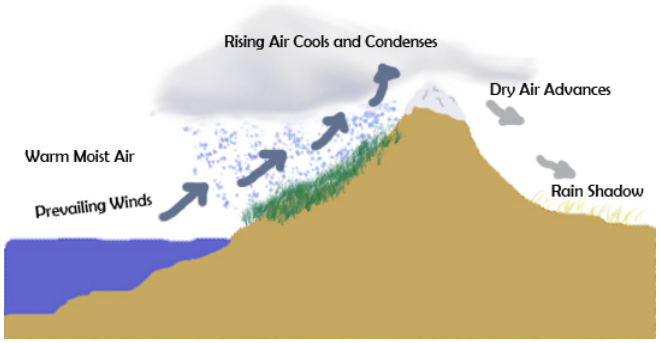


Figure 13.1.3: Diagram of rain shadow.

Deserts tend to occur at latitudes of around 30° and at the poles, both north and south, driven by circulation and prevailing wind patterns in the atmosphere. At approximately 30° north and south of the equator, sinking air produces **trade wind deserts** like the Sahara and the Outback of Australia [1]. **Rainshadow deserts** are produced on the backsides of mountains, where prevailing winds with moist air dries as it is forced to rise over the mountains. The Western Interior Desert of North America and the Atacama Desert of Chile (the driest warm desert on earth) are examples of rain shadow deserts.



Figure 13.1.4: 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 covered by sinking cold air which is usually too cold to hold much moisture. Though covered with ice and snow, the average annual precipitation is very low, with Antarctica being Earth's driest continent.

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13.2: The Origin of Deserts

Atmospheric Circulation

The engine that drives circulation in the atmosphere and oceans is solar energy which is determined by the average position of the sun over the earth's surface. Direct light provides uneven heating depending on latitude and angle of incidence, with high solar energy in the tropics, and little or no energy at the poles. Atmospheric circulation and geographic location are the primary causal agents of deserts.

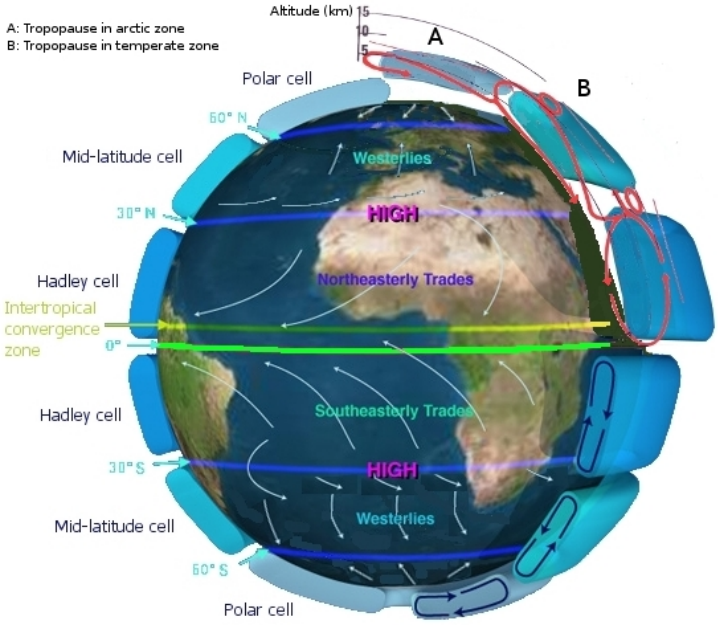


Figure 13.2.1: Generalized atmospheric circulation

The figure shows the generalized circulation of the atmosphere. There are three generalized circulating cells of rising and sinking air called the **Hadley Cell**, the **Ferrel** or **Mid-latitude Cell**, and the **Polar Cell**. Solar energy falling on the equatorial belt heats the air and causes it to rise. The rising air cools and its contained moisture falls back on the tropics as rain. The drier air then continues to spread toward the north and south where it collides with the Ferrel Cell and they sink back at about 30 degrees north and south latitudes. This sinking drier air creates belts of predominant high pressure along which desert conditions prevail in what are called the "horse latitudes." These belts of predominantly high pressure have air that descends along these belts and flows either north to become the westerlies or south to become the trade winds (which is why deserts created here are called trade-wind deserts). These circulation cells in the atmosphere rising in the tropics and polar regions and sinking in the horse latitudes produce the desert belts along the horse latitudes at approximately 30 degrees north and south of the equator [2]. Note the arrows indicating general directions of winds in the latitude zones. The trade winds are predominant in the tropics and the westerlies in the mid-latitudes.



Figure 13.2.2: USGS Map of the Great Basin Desert.

https://geo.libretexts.org/@go/page/28302

Other deserts have other atmospheric phenomena to owe (at least part of) their origin, like the desert of Utah, Nevada, and the surrounding areas called the Great Basin Desert [3]. This desert, while having some sinking air effects due to global circulation, is also a rain shadow desert produced as moist air from the Pacific rises by orthographic lifting over the Sierra Nevada (and other) Mountains and loses moisture from previous condensation and precipitation on the rainy side of the range(s).



Figure 13.2.3: Map of the Atacama desert (yellow) and surrounding related climate areas (orange).

One of the driest places on earth is the Atacama Desert of northern Chile [4]. This is a strip along the west coast of South America, west of the Andes, lying north of 30 degrees south latitude, at the southern edge of the trade wind belt. Warm moist air moves west across the Amazon basin and rises over the Andes where it loses moisture, its precipitation falling on the rain forest side of the mountains. Once over the mountains, it descends onto the Atacama where it meets air cooled by the cold Peru (Humboldt) ocean current flowing north along the coast. This is considered to be the driest (non-polar) place on earth with locations in the Atacama having not received any precipitation for periods of years [5].

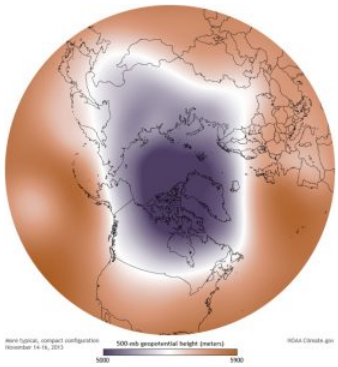


Figure 13.2.4: The polar vortex of mid-November, 2013. This cold, descending air (shown in purple) is characteristic of polar circulation.

Referring again to the figure above, note that the polar regions are also predominantly high-pressure areas of descending cold and dry air. Another circulation cell occurs there known as the **Polar Cell** [6]. Here, air not only descends convectively because it is cold, but cold air can hold much less moisture than warm air, and thus the driest and coldest places on Earth are the polar deserts. Antarctica is not only currently the driest land on Earth today, but any land that occupies the poles in Earth history should always be dry.

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13.3: Desert Weathering and Erosion



Figure 13.3.1: Weathering and erosion of Canyonlands National Park have created a unique landscape, including arches, cliffs, and spires.

Weathering takes place in desert climates by the same means as other climates, only at a slower rate. While the higher temperatures of some deserts would spur faster weathering, water is the main agent of weathering, so the lack of it slows weathering. Precipitation occurs in deserts, only less than in other climatic regions. When precipitation does occur, often in the form of flash floods, a large amount of mechanical weathering can happen quite quickly.



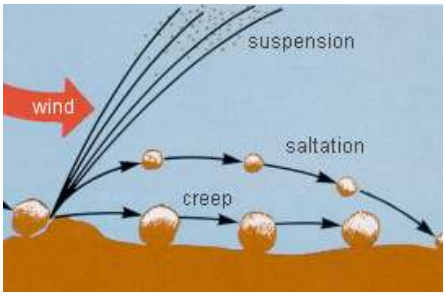
Figure 13.3.2: Newspaper rock, near Canyonlands National Park, has many petroglyphs carved into the desert varnish.

One unique weathering product of deserts is **desert varnish**. Also known as desert patina or rock rust, they are thin dark brown layers of clays and iron and manganese oxides that form on very stable surfaces within arid environments. The exact cause of the material is still unknown, though cosmogenic and biologic mechanisms have been proposed.



Figure 13.3.3: A dust storm (haboob) hits the Mongolian Gobi.

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 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 dust that melted out of the ice sheets during the last ice age [7]. 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 (like air) is distributed among bedload and suspended load. As with water, in wind, these components depend on wind velocity.



Sand size 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].

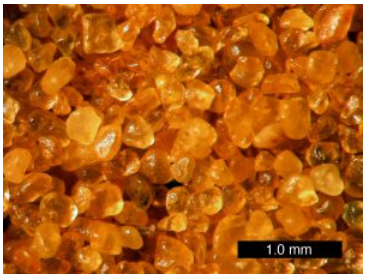


Figure 13.3.4: Enlarged image of frosted and rounded windblown sand grains

Since saltating sand grains are constantly impacting other sand grains, windblown sand grains are commonly pretty well rounded with frosted surfaces. In dust storms, the fine-grained suspended load is effectively removed from the sand and the surface as silt is carried away, leaving the sand behind. Wind is thus an effective sorting agent separating sand and dust-sized particles [9]. When wind velocity is high enough, materials may slide or roll along the surface.



One extreme version of sediment movement was shrouded in mystery for years: sliding stones. 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.

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13.4: Desert Landforms

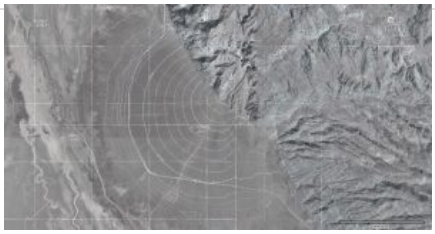


Figure 13.4.1: Aerial image of an alluvial fan in Death Valley

In deserts like those of the American Southwest, streams draining mountains flow through canyons and emerge into adjacent valleys. As the stream emerges from the narrow canyon and spreads out, and with a lower slope angle and slower speeds [17] and no longer constrained by the canyon walls [18], it drops its coarser load. As the channel fills with this larger material, the stream is deflected around it. This process causes the stream to be deflected back and forth, developing a system of radial distributaries and constructing a fan-shaped feature call an **alluvial fan**, similar to a delta made by a river entering a body of water [18].



Figure 13.4.2: Bajada along Frisco Peak in Utah

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 13.4.3: Inselbergs in Mojave Desert

As the mountains erode away and the debris accumulates first in alluvial fans, then bajadas, the mountains eventually are buried in their own erosional debris. Such residual buried mountain remnants are called **inselbergs** [20], “island mountains,” as first described by the German geologist Wilhelm Bornhardt.



Figure 13.4.4: Satellite image of desert playa surrounded by mountains

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 13.4.5: Dry wash (or ephemeral stream)

Playas are among the flattest of all landforms. Such a lake may cover a large area and be only a few inches deep, and that only after a heavy thunderstorm. Playa lakes and desert streams that fill or flow only after rainstorms are called intermittent or **ephemeral**. Drainage basins of ephemeral streams gather water from large areas and ephemeral channels may suddenly fill with water from storms many miles away and not even visible at that location.



Figure 13.4.6: Flash flood in a dry wash

Floods are a major factor in desert deposition and a serious concern for desert travelers who need to pay attention to regional weather. Water is less able to infiltrate because the flow compacts the surface, plants are less common to slow flows, and soils in deserts can become more hydrophobic. So water typically runs off as **sheet wash** (similar to how water runs over concrete) to steep-walled stream channels covered in gravel called **arroyos** that may be dry part or most of the year. Dry ephemeral channels can fill quickly, creating a mass of water and debris that charges down the channel, possibly even overflowing the banks of the arroyo. These events are called flash floods and people entering such channels or camping by them have been swept away.

Sand



Figure 13.4.7: Sahara Desert erg

While deserts are defined by dryness, not sand, the popular conception of a typical desert is a sand sea called an **erg**. An erg is a broad area of desert covered by a sheet of fine-grained sand often blown by wind into dunes [21]. Probably the best-known erg is the Empty Quarter (Rub' al Khali) of Saudi Arabia, but other ergs exist 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). It is not only deserts that form dunes; the high supply of sand can form ergs anywhere.



The way dunes form creates an internal feature called **cross-bedding**. As the 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 the fossil ergs represented by the Navajo Sandstone and Zion National Park of Utah.



Figure 13.4.8: Cross beds in the Navajo Sandstone at Zion National Park

Dune Types

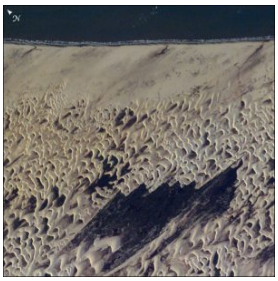


Figure 13.4.9: NASA image of a barchan dune field in coastal Brazil

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: Satellite image of longitudinal dunes in Egypt

Longitudinal dunes or **linear dunes** form where sand supply is greater and the wind is variable 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: Parabolic dunes, Cape Cod

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.

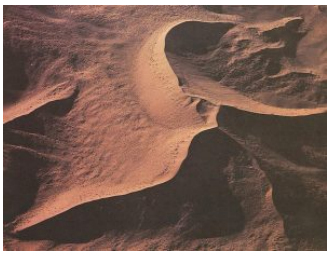


Figure 13.4.12: Star dune in Sahara

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.

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13.5: The Great Basin and the Basin and Range



Figure 13.5.1: Map of the Great Basin

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 previous chapters. 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. This has created the bowl-like shape of the region, which creates an overall internal drainage. Each of these is lined by alluvial sediments leading into playa or lake depositional environments. Even without the arid conditions, there would be these types of deposits, with lake becoming more common in place of playa. This most recently occurred with pluvial lakes that formed during the last glacial maximum.



Figure 13.5.2: Typical Basin and Range scene. Ridgecrest, CA sits just east of the southern Sierra Nevada Mountains.

The desert of the Basin and Range extends in the western US from about 35° to near 40° and has a rain shadow effect created by westerly winds from the Pacific rising and cooling over the Sierras, 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 young fault blocks of Utah with sharp peaks and alluvial fans at the mouths of canyons, through older landscapes in Southern Nevada with bajadas along the mountain fronts, to the oldest landscapes

in the Mojave Desert of California with subdued inselbergs sticking up through a sea of old bajadas. These landscapes illustrate the evolutionary stages of desert landscape development.

Desertification

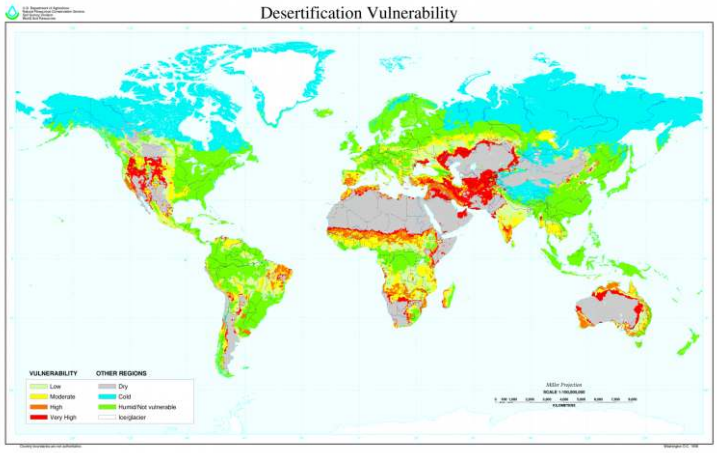


Figure 13.5.3: World map showing desertification vulnerability

Previously arable and usable land may be turned into desert by climate change and the activities of humans, such as poor farming practices, livestock overgrazing, and overuse of available water. This is a process called **desertification** and it is a serious problem worldwide [22]. Plants and soil types that are non-arid specifically help groundwater infiltration and water retention. Adding aridity to an area converts these soils and plants to be less effective in retaining water, and via a positive feedback loop (meaning that the processes feed on themselves promoting an increasing spiral). This only increases the aridity and spreads the desert further. The figure shows areas of the world and their vulnerability to desertification. Note the red and orange areas in the western and midwestern United States. The Dust Bowl of the 1930s is a classic example of human-caused desertification.

Reference

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CHAPTER OVERVIEW

14: Energy and Mineral Resources

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. without mining, modern civilization would not exist. Geologists are essential in the process of mining.

14.1: Prelude to Energy and Mineral Resources

14.2: Mining

14.3: Fossil Fuels

14.4: Mineral Resources

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14.1: Prelude to Energy and Mineral Resources



Figure 14.1.1: The Latrobe Gold Nugget, as seen on display in the London Natural History Museum, is 717 grams and displays the rare cubic form of native gold. Most gold, even larger nuggets, grow in confined spaces where the euhedral nature of the mineral is not seen.

This text 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 14.1.1: A Mode 1 Oldowan tool used for chopping

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, 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, modern civilization would not exist. Geologists are essential in the process of mining.

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14.2: Mining

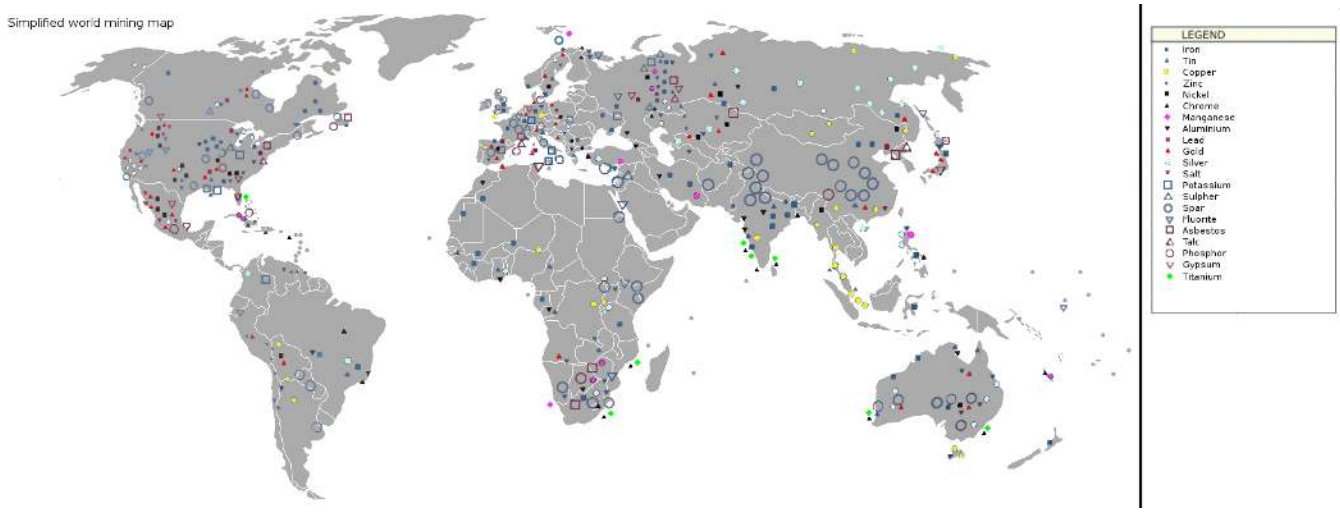


Figure 14.2.1: Map of world mining areas.

Mining is defined as the extraction, from the Earth, of valuable material for societal use. Usually, this includes solid materials (e.g. 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 evidence of mining, with a concentrated area of digging into the Earth for materials, has a history that may go back 40,000 years to the hematite (used as red dye) of the Lion Cave in Swaziland [4]. Resources extracted by mining are generally considered to be nonrenewable.

Renewable vs. Nonrenewable Resources

Resources generally come in two major categories: renewable, which can be reused over and over, or replicate over the course of a short (less than a human life span) time, and nonrenewable, which cannot.



Figure 14.2.2: Hoover Dam provides hydroelectric energy and stores water for southern Nevada.

Renewable resources are items that are present in our environment which can be exploited and replenished. Some of the more common energy sources in this category are linked with green energy resources because they are associated with environmental impacts that are relatively small or easily remediated. Solar energy is the energy that comes from fusion within the Sun, which radiates electromagnetic energy. This energy reaches the Earth both constantly and consistently and should continue to do so for about 5 billion more years [5]. Wind energy is maybe the oldest form of renewable energy, used in sailing ships and windmills. Both solar and wind-generated energy are variable on the Earth's surface. These limitations may be offset through the use of energy storing devices such as batteries or electricity exchanges between producing sites. The heat of the Earth, known as geothermal, can be viable anywhere if drilling goes deep enough. In practice, it is more useful where heat flow is great, such as volcanic zones or

regions with thinner crust [6]. Hydroelectric dams provide energy by allowing water to fall through the dam activating turbines that produce energy. Ocean tides can also be a reliable source of energy. All of these types of renewable resources can provide the energy that powers society. Other renewable resources that are not directly energy-related are plant and animal matter, which are used for food, clothing, and other necessities.

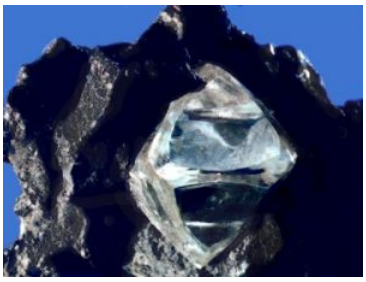


Figure 14.2.3: Natural, octahedral shape of diamond.

Nonrenewable resources cannot be replenished at a sustainable rate. They are finite within a human lifetime. Many nonrenewable resources are chiefly a result of planetary, tectonic, or long-term biologic processes, and include items such as gold, lead, copper, diamonds, marble, sand, natural gas, oil, and coal. Most nonrenewable resources are utilized for their concentration of specific elements on the periodic table. For example, if society needs sources of iron (Fe), then it is the exploration geologist who will search for iron-rich deposits that can be economically extracted. Non-renewable resources may be abandoned when other materials become cheaper or serve their purpose better. For example, abundant coal is available in England, but the availability of North Sea oil and natural gas (at a lower cost and lower environmental impact) led to the decrease in coal usage.

Ore

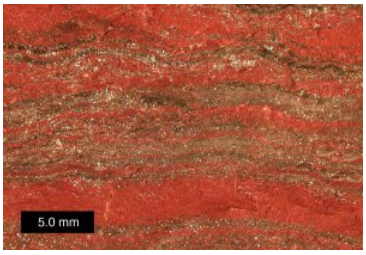


Figure 14.2.4: Banded-iron formations are an important ore of iron (Fe).

The elements of the periodic table are found within the materials that make up the Earth. However, it is rare for the amount of the element to be concentrated to the point where the extraction and processing of the material into usable product becomes profitable. Any place where the amount of valuable material is concentrated is a geologic and geochemical anomaly. If the material can be mined at a profit, the body constitutes an **ore** deposit. Typically, the term ore is used for only metal-bearing minerals, though the concept of ore as a non-renewable resource can be applied to valuable concentrations of fossil fuels, building stones, and other non-metal deposits, even groundwater. The term “natural resource” is more common than ore for these types of materials.

Mining Techniques



Figure 14.2.5: Bingham Canyon Mine, Utah. This open-pit mine is the largest man-made removal of rock in the world.

The style of mining is a function of 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. 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 14.2.6: A surface coal mine in Wyoming.

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. Typically, the pit progressively deepens through additional mining cuts to extract the ore, and the walls of the pit are as steep as can safely be managed. A steep wall means there is less waste (non-valuable) rock or overburden to remove and is an engineering balance between efficient mining and mass wasting. Occasionally landslides do occur, including a very large landslide that occurred in the Bingham Canyon mine in 2013. These events are costly and dangerous, though careful monitoring gave the Bingham Canyon mine ample warning time. **Strip mining** and **mountaintop mining** are surface mining techniques also used for resources that cover large areas, especially layered resources like coal. In this case, an entire mountaintop or rock layer is removed to gain access to the ore below. The environmental impacts of surface mining are usually greater due to the larger surface disturbance footprint [7].



Figure 14.2.7: Underground mining in Estonia of Oil Shale.

Underground mining is often used for higher-grade, more localized, or very concentrated resources. Some ore minerals are mined underground by introducing chemical agents that dissolve the target mineral followed by solution extraction and subsequent precipitation in a surface operation, but more often a mining shaft/tunnel (or a large network of these shafts and tunnels) is dug to access the material. Whether mining occurs underground or from Earth's surface is dictated by ore deposit depth, geometry, land-use policies, economics, strength of the surrounding rock, and physical access to the ore to be mined. For example, deeper deposits might require removal of too much material, it may be too dangerous or impractical to remove, or it may be too expensive to remove the entire overburden. These factors may prevent materials from being mined from the surface, and cause a project to be mined underground. Also, places where the mining footprint can not be large may force underground mining to occur. The method of mining and whether mining is feasible depends on the price of the commodity and the cost of available technology to remove it and deliver it to market. Thus mines and the towns that support them come and go as the price of the commodity varies. Technological advances and market demands may reopen mines and revive ghost towns.

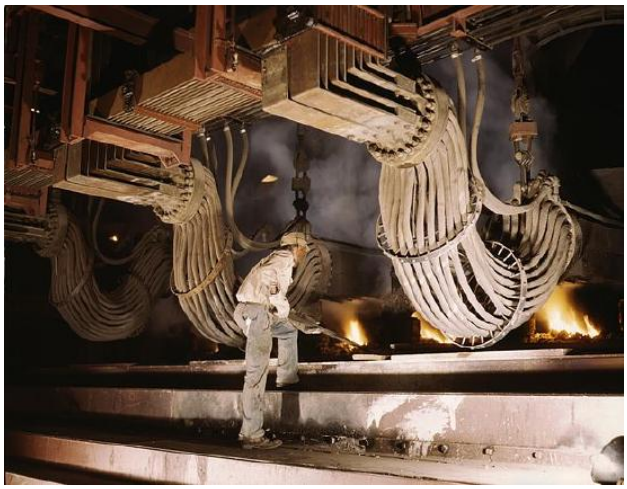


Figure 14.2.8: A phosphate smelting operation in Alabama, 1942.

All ore minerals are 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 in the presence of other minerals) 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, this is done one of three ways:

1. Items 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. Items can also be heated to chemically separate desired components, like refining crude oil into gasoline
3. Items 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 amounts of energy. Continual advances in metallurgy and mining practices aim to develop ever more energy-efficient and environmentally benign processes and practices.

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14.3: Fossil Fuels



Figure 14.3.1: Coal power plant in Helper, Utah.

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. This energy was originally formed via photosynthesis by living organisms such as plants, phytoplankton, algae, and cyanobacteria. Sometimes this is known as fossil solar energy since the energy of the sun in the past has been converted into the chemical energy within a fossil fuel. Of course, as the energy is used, just like respiration from photosynthesis that occurs today, carbon can enter the atmosphere, causing climate consequences. Fossil fuels account for a large portion of the energy used in the world.



Figure 14.3.2: Modern coral reefs and other highly-productive shallow marine environments are thought to be the sources of most petroleum resources.

Oil and Gas

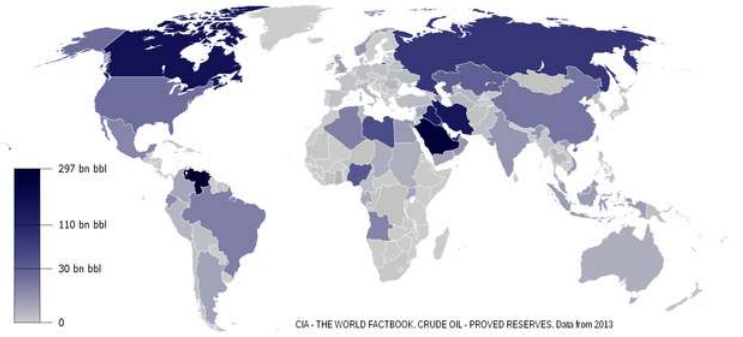


Figure 14.3.3: World Oil Reserves in 2013. Scale in billions of barrels.

Petroleum, with the liquid component commonly called **oil** and gas component called **natural gas** (mostly made up of methane), is principally derived from organic-rich shallow marine sedimentary deposits [12]. As the rock (which is typically shale, mudstone, or limestone) lithifies, the oil and gas leak out of the **source rock** due to the increased pressure and temperature, and migrate to a different rock unit higher in the rock column. Similar to the discussion of good aquifers in the chapter covering groundwater, if the rock is sandstone, limestone, or other porous and permeable rock, then that rock can act as a **reservoir** for the oil and gas.



Figure 14.3.4: A structural or anticline trap. The red on the image represents pooling petroleum. The green layer would be a non-permeable rock, and the yellow would be a reservoir rock.

A **trap** is a combination of a subsurface geologic structure 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 (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 extraction is less viable.

Oil Shale

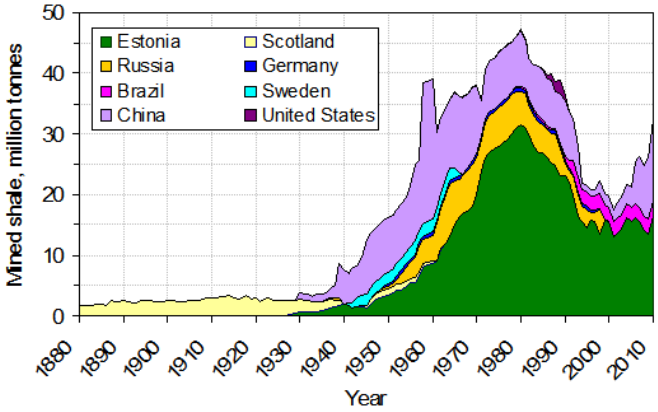


Figure 14.3.5: Global production of Oil Shale, 1880-2010.

Oil shale (or tight oil) is a fine-grained sedimentary rock that has a significant quantity of petroleum or natural gas. Shale is a common source of fossil fuels with high porosity but it has very low permeability. In order to get the oil out, the material has to be mined and heated which is expensive and typically has a negative impact on the environment [17].

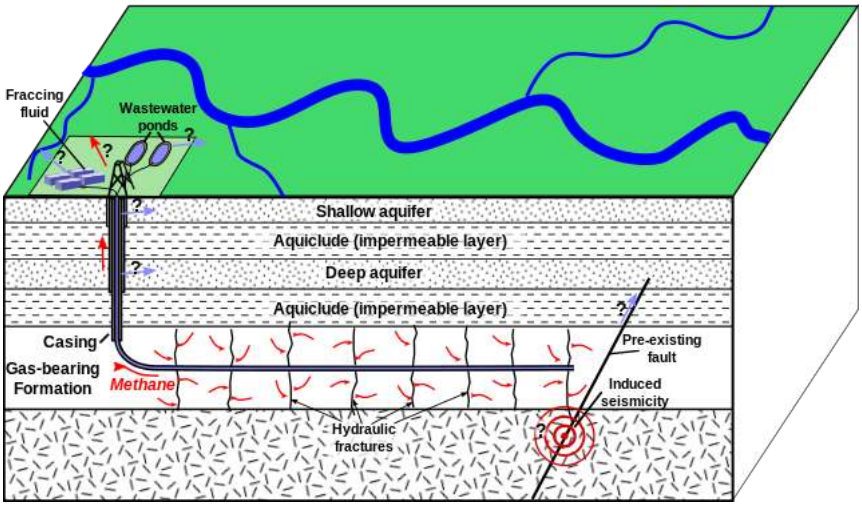


Figure 14.3.6: Schematic diagram of fracking.

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, creating and holding open fractures in the rocks, which aids in the release of the hard-to-access fluids, mostly natural gas. This 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 due to the potential for groundwater contamination [19] and induced seismicity [20] and represents a balance between public concerns and energy value.

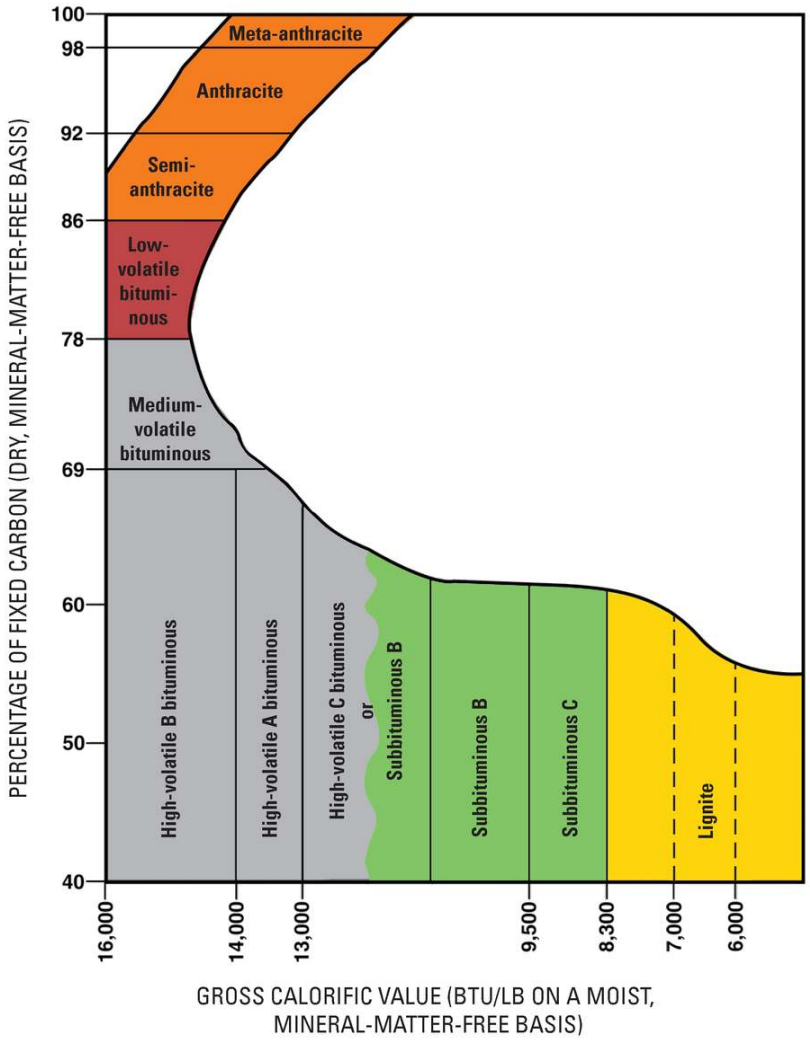


Figure 14.3.7: USGS diagram of different coal rankings.

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]. It is chiefly carbon, hydrogen, nitrogen, sulfur, and oxygen, with minor amounts of other elements [23]. As this plant material is incorporated into sediments, it undergoes a series of changes due to heat and pressure which concentrates fixed carbon, the combustible portion of the coal. In this sense, the more heat and pressure that coal undergoes, the greater is its fuel value and the more desirable is the coal. The general sequence of a swamp turning into the various stages of coal are:

Swamp => Peat => Lignite => Sub-bituminous => Bituminous => Anthracite => Graphite.

As swamp materials collect on the floor of the swamp, they turn to peat. 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 14.3.8: Anthracite coal, the highest grade of coal.

Coal has been used by humans 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) in 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 for energy sources and recognition of its negative environmental impacts, including increased very fine-grained particulate matter, greenhouse gases [27], acid rain [28], and heavy metal pollution [29]. Seen from this point of view, the coal industry is unlikely to revive.

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14.4: Mineral Resources

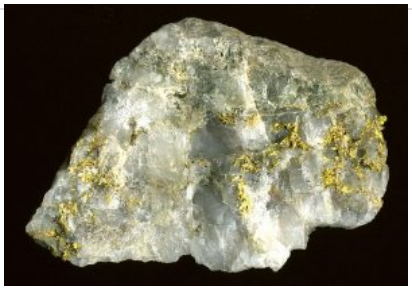


Figure 14.4.1: Gold-bearing quartz vein from California.

Mineral resources, while principally nonrenewable, are generally placed in two main categories: **metallic** (containing metals) or **nonmetallic** (containing other useful materials). Most mining is focused on metallic minerals. A significant part of the advancement of human society has been developing the knowledge and technologies that yielded metal from the Earth and allowed the machines, buildings, and monetary systems that dominate our world today. The location and recovery of these metals have 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.

Types of Metallic Mineral Deposits

The number of ways that 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



Figure 14.4.2: Layered intrusion of dark chromium-bearing minerals, Bushveld Complex, South Africa

Crystallization and differentiation of a magmatic body can cause the concentration of certain minerals and elements. **Layered intrusion** (typically ultramafic to mafic) can be host to deposits that contain copper, nickel, platinum-palladium-rhodium, and chromium. The Stillwater Complex in Montana is an example of an economic 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]. Rocks of the Bushveld Igneous Complex have an areal extent larger than the state of Utah. The chromite occurs in layers, which resemble sedimentary layers, except this occurred within a crystallizing magma chamber.

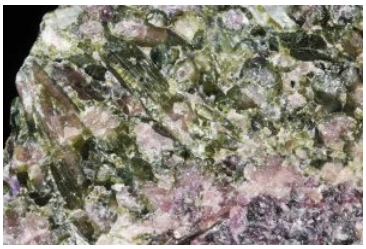


Figure 14.4.3: This pegmatite from Brazil contains lithium-rich green elbaite (a tourmaline) and purple lepidolite (a mica).

Water and other volatiles that are not incorporated into mineral crystals while a magma crystallizes 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, masses of these large crystals are called **pegmatites** that form from the concentration of magma fluids near the end of crystallization 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.

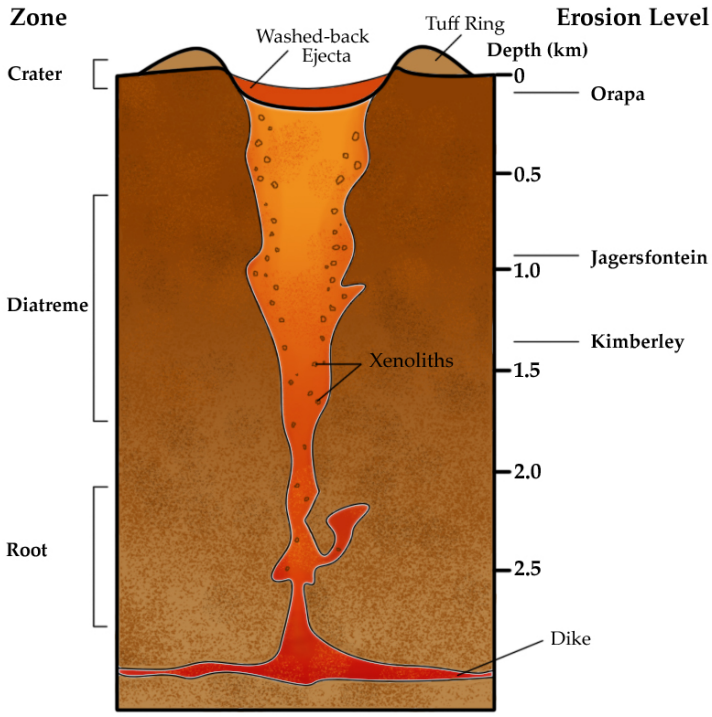


Figure 14.4.4: Schematic diagram of a kimberlite pipe.

An unusual magmatic process 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 temperature and depth, are transported this way to locations where they can be mined. The process that emplaced these kimberlite (ultramafic) rocks is no longer common on Earth, and most of the known deposits are Archean [33].

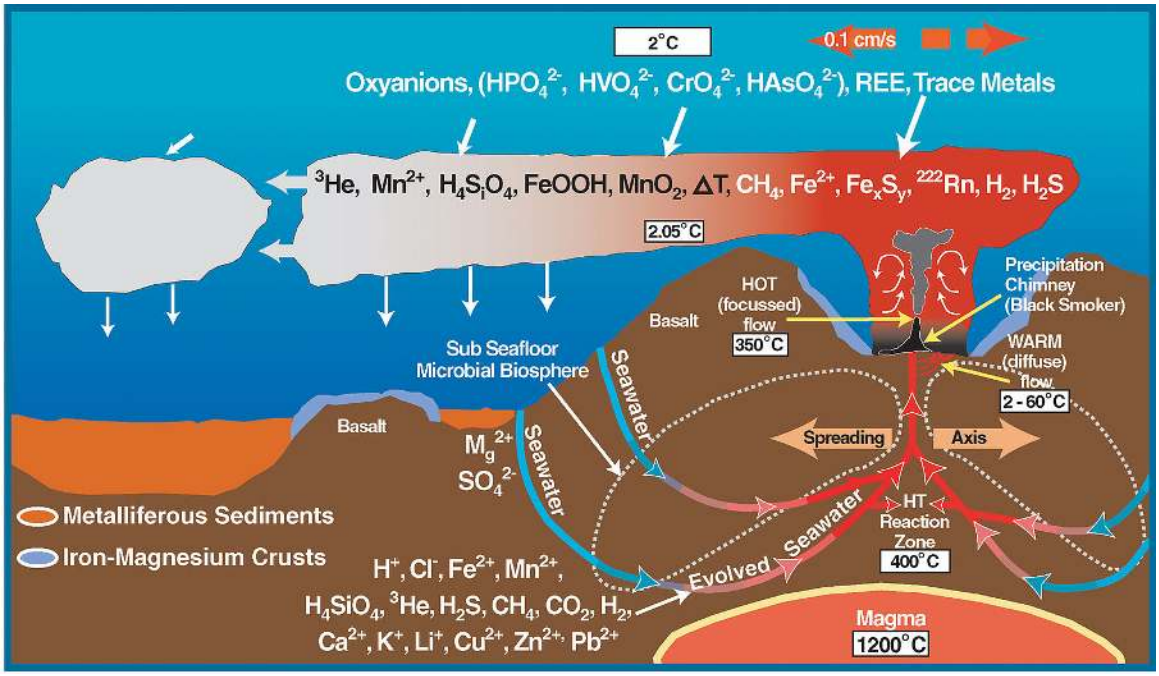


Figure 14.4.5: The complex chemistry around mid-ocean ridges.

Fluids rising from crystallizing magmatic bodies or heated by the geothermal gradient cause a wide range of geochemical reactions that can form a variety of mineral deposits. The most active hydrothermal process today produces **volcanogenic massive sulfide** (VMS) deposits, which form from black smoker activity near mid-ocean ridges all over the world, and commonly contain copper, zinc, lead, gold, and silver when found on the surface [34]. The largest of these deposits occur in Precambrian age rocks. The Jerome deposit in central Arizona is a good example.

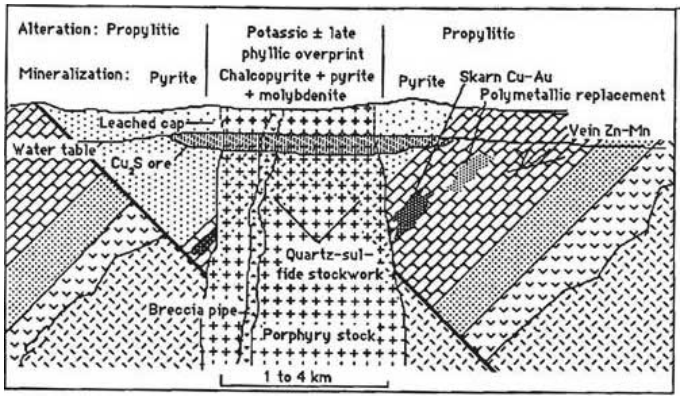


Figure 14.4.6: USGS schematic of a Porphyry copper deposit.

Another type of deposit which draws on heated water from magma 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 in a porphyry deposit. Several types of porphyry deposits exist: porphyry copper, porphyry molybdenum, and porphyry tin. They are characterized by the presence of low-grade disseminated ore minerals closely associated with intermediate and felsic intrusive rocks over a very large area [35]. Porphyry deposits are typically the largest mines on Earth. One of the largest, richest, and possibly best-studied mines in the world is Utah's Bingham Canyon open-pit mine, which has had over 100 years of high production of several elements including copper, gold, molybdenum, and silver. Associated underground carbonate replacement deposits have produced lead, zinc, gold, silver, and copper [36]. Past open pit production at this mine was dominated by copper and gold from chalcopyrite and bornite. Gold occurs in minor quantities in the copper-bearing mineral, but the large scale of production makes Bingham Canyon one of the largest gold mines in the U.S. Future production may be more copper and molybdenum (molybdenite) from deeper underground mines.

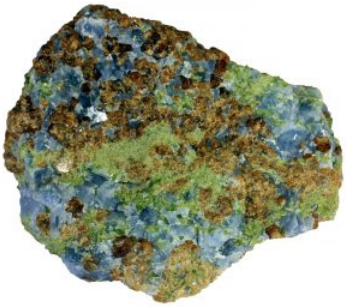


Figure 14.4.7: Garnet-augite skarn from Italy.

If limestone or other calcareous sedimentary rocks are present adjacent to the magmatic body, then another type of ore deposit called a **skarn** deposit can form. 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 zones of iron, copper, and zinc minerals and gold [38]. Intrusions that are genetically related to the intrusion that made the Bingham Canyon deposit have also produced copper-gold skarns that were mined by the early European settlers in Utah [39; 40]. Metamorphism of iron and/or sulfide deposits commonly results in an increase in grain size that makes separation of gangue from the desired sulfide or oxide minerals much easier.

Non-Magmatic Geochemical Processes



Figure 14.4.8: Underground uranium mine near Moab, Utah.

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 environments are considered reducing. Uranium deposition is an example of redox mobilization. Uranium is soluble in oxidizing groundwater environments and precipitates as uraninite when reducing conditions are encountered. Many of the deposits across the Colorado Plateau (e.g. Moab, Utah) were formed by this method [43].

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. Cyclic oxygenation of iron-rich waters initiated the precipitation of the iron beds. Because BIFs are generally Precambrian in age, they are only found in some of the older exposed rocks in the United States, in the upper peninsula of Michigan and northeastern Minnesota [44].

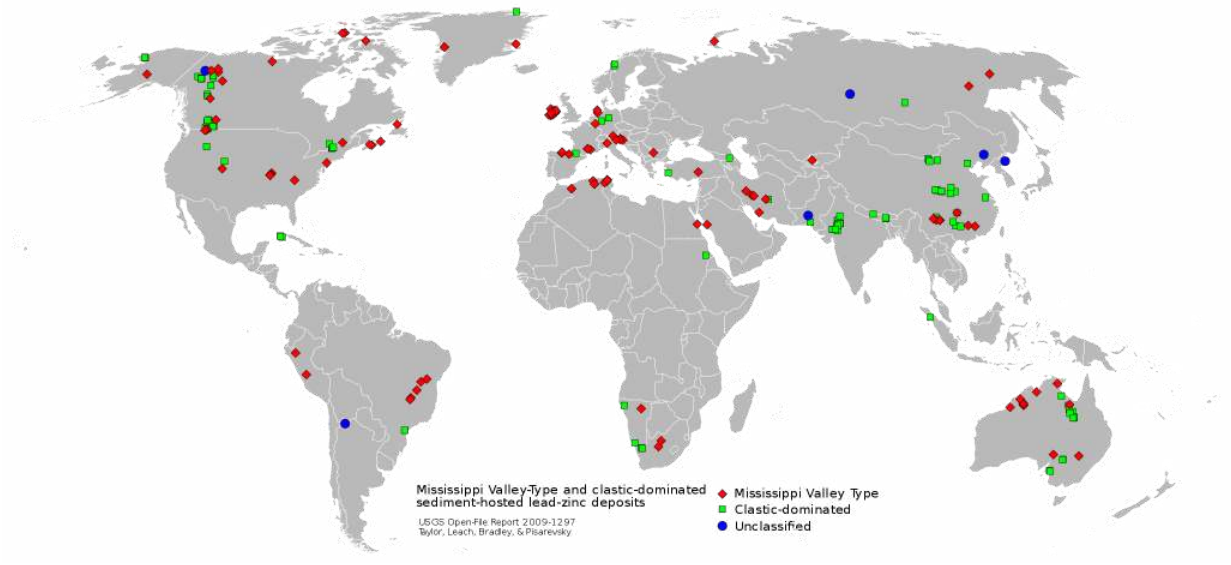


Figure 14.4.9: Map of Mississippi-Valley type ore deposits.

Deep, saline, connate fluids (trapped in the pore spaces), within sedimentary basins may be highly metalliferous. When expelled outward and upward during basin compaction, these fluids may form lead and zinc deposits in limestone by replacement or by filling open spaces (caves, faults) and in sandstone by filling pore spaces. The most famous of these 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) which form from fluids in the temperature range of 100 to 200°C. Although they are named for occurrences along the Mississippi River Valley in the United States, they are found worldwide.

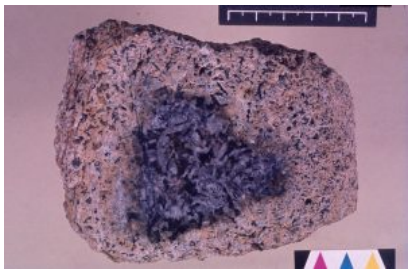


Figure 14.4.10: A sample of bauxite. Note the unweathered igneous rock in the center.

Deep and intense weathering of soils and mineral deposits exposed at the surface can result in the formation of surficial deposits. **Bauxite**, an ore of aluminum, is preserved in karst topography and laterites (soils formed in wet tropical environments) [46]. Aluminum concentrates in soils as feldspar and ferromagnesian minerals in igneous and metamorphic rocks undergo chemical weathering processes. Weathering of ultramafic rocks results in the formation of nickel-rich soils and weathering of magnetite and hematite in banded iron formation results in the formation of goethite, a friable mineral that is easily mined for its iron content.

Surficial Physical Processes



Figure 14.4.11: Lithified heavy mineral sand (dark layers) from a beach deposit in India.

At the earth's surface, the physical process of mass wasting or fluid movement concentrates high-density minerals by hydraulic sorting. When these minerals are concentrated in streams, rivers, and beaches, they are called **placer** deposits, whether in modern

sands or 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.

Environmental Impacts of Metallic Mineral Mining



Figure 14.4.12: Acid mine drainage in the Rio Tinto, Spain.

The primary impact of metallic mineral mining comes from the mining itself, including disturbance of the land surface, covering of landscapes by tailings impoundments, and increased mass wasting by accelerated erosion [48]. In addition, many metal deposits contain pyrite, an uneconomic sulfide mineral placed on waste dumps, which may generate **acid rock drainage** (ARD) during weathering. In the presence of oxygenated water, sulfides such as pyrite react undergo complex reactions to release metal ions and hydrogen ions, lowering pH to highly acidic levels. Mining and processing of mined materials typically increase the surface area to volume ratio in the material, causing reactions to occur even faster than what would occur naturally. If not managed properly, these reactions may lead to acidification of streams and groundwater plumes that can carry dissolved toxic metals. In mines where limestone is a waste rock of carbonate minerals like calcite or dolomite are present, their acid-neutralizing potential helps reduce the likelihood of generating ARD. Although this is a natural process too, it is very important to isolate mine dumps and tailings from oxygenated water, both to prevent the dissolution of sulfides and subsequent percolation of the sulfate-rich water into waterways. Industry has taken great strides in preventing contamination in recent decades, but earlier mining projects are still causing problems with local ecosystems.

Nonmetallic Mineral Deposits



Figure 14.4.13: Carrara marble quarry in Italy, source to famous sculptures like Michelangelo's David.

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.

Evaporites



Figure 14.4.14: Salt-covered plain known as the Bonneville Salt Flats, Utah.

Evaporite deposits form in restricted basins, such as the Great Salt Lake or the Dead Sea, where evaporation of water exceeds the recharge of water into the basin [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 and rock gypsum and other minerals can accumulate (see chapter 5).



Figure 14.4.15: Hanksite, $Na_{22}K(SO_4)_9(CO_3)_2Cl$, one of the few minerals that is considered a carbonate and a sulfate

Evaporite minerals like halite are used in our food as common table salt. Salt was a vitally important economic resource prior to refrigeration as a food preservative. While still used in food, now it is 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, being 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 hyperarid locations, even rarer and more complex evaporites, like borax, trona, ulexite, and hanksite, are found and mined. They can be found in such localities as Searles Dry Lake and Death Valley, California, and in ancient evaporite deposits of the Green River Formation of Utah and Wyoming.

Phosphorous



Figure 14.4.16: Apatite from Mexico.

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. Phosphorous is a major component of bone and a key component of DNA. Bone ash and guano are natural sources of phosphorus.

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