INTRODUCTION TO HISTORICAL GEOLOGY



Chris Johnson, Callan Bentley, Karla Panchuk, et al.

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Introduction to Historical Geology

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Russ Kohrs, Paul Inkenbrandt, Cam Mosher,
Brian Ricketts, and Charlene Estrada

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Licensing

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1.1: Stories from Stone

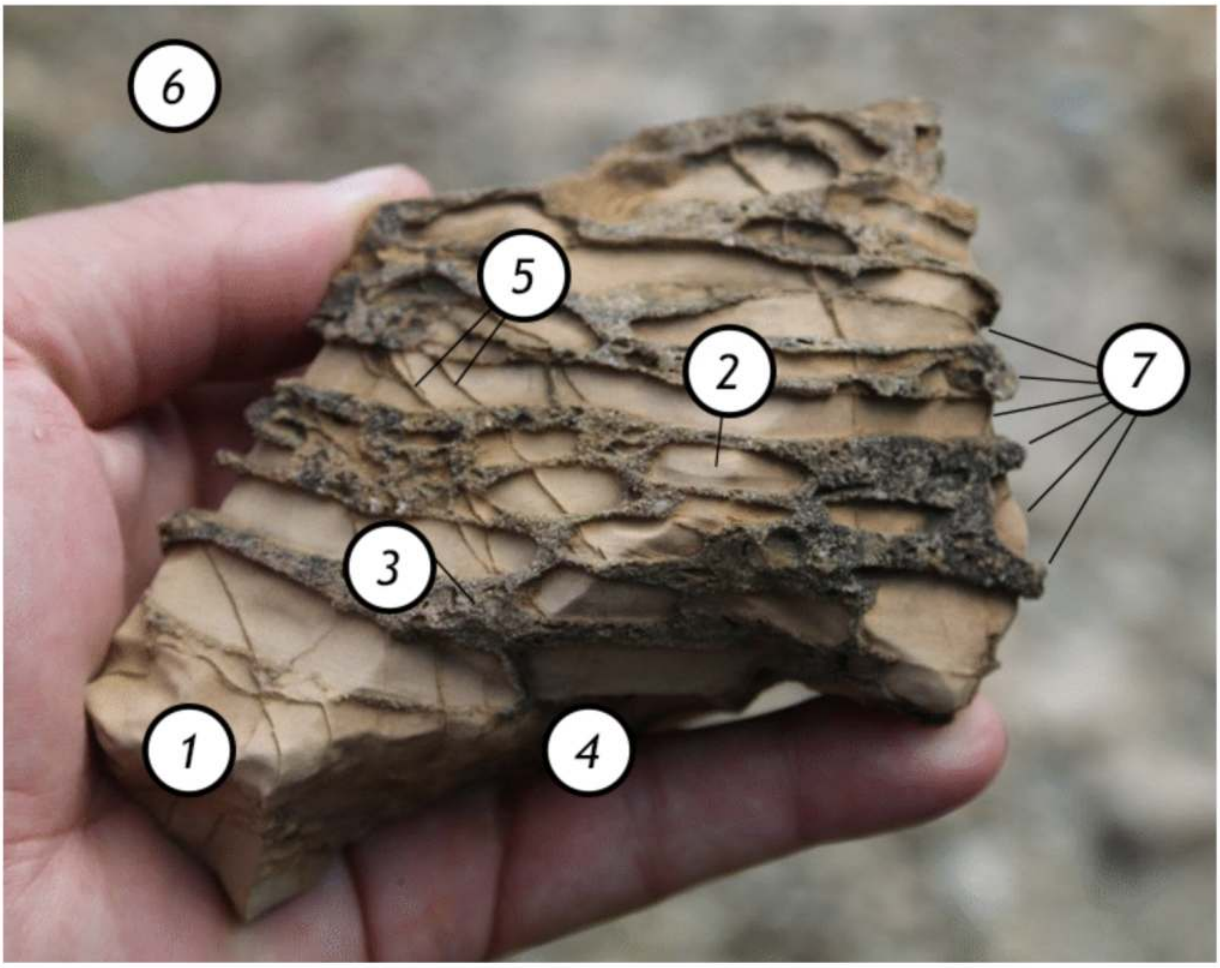
The most common question I get as a geologist is, “What is this?” People not only want to know the name of a rock but also what it *MEANS* to be that rock.



A photograph of a sample of rock, held in a person’s hand. The rock overall is a mix of light-colored, fine-grained material (in chunks) and darker-colored coarser-grained material. The fine-grained stuff has been etched away, leaving the coarse-grained stuff standing out in high relief.

This chunk of rock contains features that we can translate into information. The variety of materials it is made from, how those materials are arranged, other features that cut across them, and the shape of the sample all convey information about past processes and circumstances. What would it take to make a rock just like this one? Why does it exist in this particular way?

Let’s translate:



Points of annotation showing the 7-chapter history of this one rock sample.
Callan Bentley photo.

Step one is identifying the materials within the rock.

1. Smooth, tan limestone
2. Fragments of the same limestone
3. Quartz sand surrounding the fragments
4. …..
5. Quartz veins: typically occurs underground when heat is present
6. …..
7. …..

Step two is determining what those materials *imply about how they formed*.

1. Limestone: deposited in a calm, water environment – typically an ocean
2. Fragments: pieces were ripped up an energetic event
3. Quartz sand: the same event that ripped up (2) also moved the quartz sand
4. …..
5. Quartz veins: typically occurs underground when heat is present
6. …..
7. …..

Step three is looking at what mountain building/rock-forming events occurred.

1. …..

2. …..
3. …..
4. buried and turned to rock
5. …..
6. uplifted to be exposed at the Earth's surface
7. piece is eroded away to give it the distinctive texture (different materials weather differently)

Then we put it all together:

A smooth, tan limestone (1) was deposited in a calm, water environment. A storm event, perhaps a hurricane, ripped up fragments (2) and quartz sand (3) from the ocean floor. These sediments were buried over time and eventually turned into rock, a process called lithification (4). They were buried deeply enough with mountain building forces that the rocks were fractured, and quartz filled the small cracks (5). The rock was eventually uplifted to the surface (6), and a piece broke off. That piece then weathered differently (7) based on the materials it was made of to look the way it does today.

Whew. That was a lot. To understand all of this, we must build the skills to identify the rock and tell the story.

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1.2: Sequences of Strata

Decoding the handful of past processes encoded into a single cobble is powerful, but more power still comes from looking at sedimentary rock in outcrop, and comparing the conditions of one layer with those below it and those above it. The *lingua franca* of historical geology is sedimentary rock, and vast epic tales have been inscribed in the sequence of accumulated strata, or rock layers.

If a shale means former mud, and that mud's deposition implies calm, offshore conditions, and sandstone means former sand, and that sand's deposition implies more energetic nearshore conditions, then what does this photo show?



Figure 1.2.1: A sequence of sedimentary layers can tell a story of changing conditions over time. Photograph by Zoltán Sylvester; reproduced with permission.

A sedimentary sequence of mud-rich strata overlain by sand-rich strata is the record of a shift in depositional conditions at a site over time. In the image above, from Chilean Patagonia, the vertical sequence of layers shows that as time went by, there was less and less mud being deposited at this location and more and more sand. The progressive stack of layers is a record of time, with older, earlier chapters at the bottom and more recent chapters at the top.

So we've made an observation (the shift from muddy to sandy), and that observation begs the question: Why was there less mud and more sand? (In other words, what's the story?)

Maybe it shows a shift from deeper water to shallower water conditions. Maybe it shows the tectonic drift of a landmass into a more energetic, storm-dominated location. Maybe it shows the advance of a submarine fan system spreading onto the abyssal plains of the deep ocean. Maybe it shows the development of a nearby mountain belt, shedding the source sediment. Each of these "maybes" is a hypothesis, a potential explanation for the observation of the sandier layers overlying the muddier layers. The goal of historical geologists is to probe each of these possible explanations and test the ideas against other local details of the geology. Storms rip up mud chips, for instance. Are there mud chips in these sands? Submarine fans often show graded beds. Are there any? A change in the fossil fauna often marks shallower conditions. Tectonic drift can be marked by changes in the paleomagnetic inclination. There is a profusion of additional data we can collect to match up with the different potential interpretations. Generating and considering multiple working hypotheses allows historical geologists a "toolkit of the imagination" when they head into the field.



Figure 1.2.2: Outcrops in Inuyama, Japan, show a Triassic sequence of sedimentary strata that record changes in deep ocean oxygen levels after the end-Permian mass extinction event. The strata have been tectonically rotated due to accretion with the Japanese mainland. Photo by Yukio Isozaki; reproduced with permission.

Here is another example from Inuyama, Japan:

Here, the strata are not in their original horizontal orientation. They have been rotated during tectonic accretion of these deep ocean strata to the Japanese mainland. The layers get younger to the right (rather than “up”). Don’t let that throw you off – You can twist your head to the right if you need to.

The important thing is to note that a distinctive color change occurs across the outcrop: the layers shift from black to red. This is a signature of changes in the amount of oxygen in the deep ocean: there was almost no oxygen (black), and then there was plenty (red). This is interesting because a major mass extinction occurs just before this sequence: the end-**Permian mass extinction** (also called “The Great Dying”). These strata are 247 to 242 Ma, early middle Triassic, and show the recovery after the mass extinction.

Evidence such as this sequence of strata suggests that ocean anoxia (low oxygen levels) were associated with this major die-off in animals, either as a cause or as a consequence. And most everything on the planet died: 95% of species. It was the worst thing that ever happened to life on our planet, and this sequence of jet black layers transitioning to red layers is trying to tell us why it happened. Can we speak Rock well enough to hear the story? Our own species’ future might depend on our ability to translate. This work is important.

Again, no *single* layer tells the whole story – we must look at the succession of layers over time.

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1.3: Fossil Vocabulary

Within those layers, we find clues of various kinds. One of those varieties of information is fossils. Earth is a planet that has hosted life for most of its 4½ billion years, and especially for the last ½ billion years, there have been plenty of animal skeletons and plant tissues to be included in the strata as clues to when they were deposited. For instance, with the Japanese cherts we just examined; we know how old they are based on the fossil radiolaria that they contain.



Stacks of old newspapers, ready for reading or recycling. The words in these periodicals are clues to when they were published, like fossils in strata are clues to when those strata were deposited.

Consider an analogy: these piles of old newspapers. Imagine reaching into these stacks and extracting three pages at random. You read them, and find many familiar words and phrases: each is written in English. But also on each of them, you see distinctive words and phrases:

- On one: *Hepcat. Daddy-O. Supermurgitroid. Peepers. Knuckle sandwich. Made in the shade.*
- On another: *Moxie. Bee's knees. Gams. Flapper. Sockdollager. Zozzled.*
- On a third: *Fain. Forsooth. Betwixt. Thou. Verily. Swain. Hugger-mugger.*

These words are all archaic in some sense (*i.e.*, they are no longer in widespread use), but they are also specific to certain periods of time (*as well as to English-speaking areas of the world*). One set of words derives from the 1950s, another from the roaring 1920s, and the last from Elizabethan England (~1600, as recorded by William Shakespeare). The words alone evoke the time in which the newspaper was written.



Trilobite fossil.

Similarly, in historical geology, fossils are the rich vocabulary that serves to indelibly mark their strata with the time of their formation. Some sediments speak a strain of Rock with a pachycephalosaur patois or a trilobite tone. Others communicate with an ichthyosaur inflection or a brachiopod brogue. As with language, these distinctive fossil "words" are emblematic of a specific time or a specific place, or both. They tell us not only *when* but also *what* the environment was like at the time.

While sandstones and shales are enduring elements in the language of Rock, fossils are the slang that changes with time and space. Fossils mark the passage of time and tectonics with their unique lexicon.



You'll never have a pet Anomalocaris because you live at the wrong time in Earth history.
Photo illustration mashup orchestrated by Callan Bentley.

It would be anachronistic to find an Ordovician fossil in Cretaceous strata like a digital watch on a medieval knight. Shakespeare's characters never said "groovy" or "coronavirus." And you can't buy an *Anomalocaris* at the pet store today. Certain fossils — the ones most useful for constraining the age and place of formation of their host strata — are distinctive enough that we can rely on them as signals of the origins of the sedimentary strata that host them, markers of particular times or places.

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1.4: Magma, Metamorphism and Mountains

Traditionally, Historical Geology has focused its attention on sedimentary rocks, built up in layers and host to distinctive fossils. But this is only one dialect that Earth is capable of speaking. She is equally eloquent when she writes igneous and metamorphic stories or deforms her earlier works through later episodes of mountain building. In this text" book," we embrace the full gamut of Earth's works, giving equal value to the Earth stories we can decode from metamorphic and igneous rocks and tectonic structures too.

For example, the outcrop below shows a pattern that's quite striking.



The pattern you observe above is a particularly regular system of fractures that developed in a cooling lava flow. The iron-rich lava erupted 50 million years ago in Northern Ireland, congealed, and cooled down, shrinking as it did so. (Cold rock takes up less space than warm rock.) The contraction induced tensional stresses across the surface of the lava flow. These stresses were greater than the strength of the freshly-solidified rock, and it cracked. The fractures met each other in a "honeycomb" like pattern on the surface of the flow and then propagated down into the flow's interior, dividing the warm young rock into a series of cold polygonal pillars.

The locals were so impressed with the precise regularity of this pattern they were convinced a conscious mind must have constructed it. They spun a legend to explain it – a tempestuous giant named Finn MacCool (*Fionn mac Cumhaill*), trying to pick a fight with a Scottish giant named Benandonner. Benandonner had been so bold as to claim Ireland as his own, and Finn wouldn't stand for it! Finn started pitching rocks toward his rival and realized he could build out a causeway to connect the two landmasses. He journeyed to Scotland (specifically, to the island of Staffa) to confront the usurper but got scared by Benandonner's size and rushed back across the causeway to Ireland. The Scot pursued but was tricked into a hasty retreat thanks to some clever subterfuge by Finn's wife Sadhbh (*pronounced "Siive"*). She dressed Finn as an infant and told Benandonner that he was, in fact, merely

Finn's baby and that big daddy Finn would be home soon! Benandonner beat his hasty retreat, knocking the causeway to pieces as he fled. The site is therefore known as the Giant's Causeway.

Geologists have a different way of explaining the Giant's Causeway, of course. Because they speak Rock rather than Giant, they note that on a landscape scale, this lava flow is one of many that erupted in a vast region between 60 and 50 million years ago, smothering the local landscape as they flowed and pooled. The volcanic activity was not limited to Ireland and Scotland. It also produced basalt in the Faroe Islands, Norway, and Greenland. All these sites are in the North Atlantic Igneous Province (NAIP). The Paleogene-aged lava flows link them all to be in the same spot around 60 million years ago, marking them all with a common stamp before the various landmasses separated from each other, with the Atlantic Ocean opening up in between.

The cause of this eruption may have been the Icelandic hotspot, which continues erupting similar lava today at the Mid-Atlantic Ridge. The now-widely-dispersed basalts of the NAIP help link together landmasses that experienced divergent tectonics during the Paleogene and have been moving apart ever since.

Furthermore, the timing of the NAIP's eruption overlaps with a time of extreme global warming, the Paleocene-Eocene Thermal Maximum (PETM), and some scientists have suggested that CO_2 emissions from the lava may have been the cause of this superlative episode of global warming. As with the end-Permian extinction, that is potentially really important: a matter of some importance to Rock-fluent humans who are living in their own time of global warming.

Bottom line: Igneous rocks are just as vital as sedimentary rocks in interpreting Earth's history.

On the metamorphic front, consider this blueschist:



Blueschist is a metamorphic rock that is blue and flaky. But like all metamorphic rocks, it didn't start that way. This blueschist started as basalt – the same sort of mafic volcanic rock we just saw at the Giant's Causeway! The fact that it doesn't look like that anymore is evidence of a profound journey it took – from Earth's surface down deep into its interior, down 40 or maybe 50 kilometers into the mantle. Down there, temperatures were hotter, but more importantly, the pressure was insanely high. The minerals making up the basalt couldn't take it: they reacted and fell apart, and their atoms recombined to make new minerals. The new minerals were chemically stable under the crushing pressure.

This distinctive blend of minerals is, therefore, evidence of a *journey*. Like an astronaut returning from the Moon, the blueschist has returned from a place none of us will ever get to go. Its minerals tell a story, like stamps in a passport. The best travel stories are written in Metamorphic.

Scars tell stories too, and Earth has picked up her share of scars through the years. Mostly, these "scars" are tectonic *structures* like folds, faults, and shear zones. Consider this photograph of the Andes Mountains in southern Argentina:



Folded turbidite strata in southern Argentina. Photograph by Zoltán Sylvester; reproduced with permission.

Rock layers are laid down in layers that are pretty close to horizontal, perhaps as steep as 30° (the angle of repose). But when you see them as in this Patagonian mountainside, with some of the layers upside-down, you know that they must have been seriously contorted sometime *after* their original deposition. After all, you cannot deform a rock layer without first forming it.

Stories can be extracted not only from the sequence of sedimentary layers in this case but also from their contortion, orientation, their elevation, and their shape. The structure is the most passionate strain of Rock vernacular. If we ignore the structure of the mountainside, we potentially miss out on the most dramatic stories these rock layers have to tell.

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1.5: Earth's Story is Your Story!

Why should you care about any of this?

You are an Earthling. You are were born on this planet, and you will live out ***your entire life on its surface***. The atoms from which your body is constructed are atoms that have been on this planet for more than four billion years, repurposed endlessly before becoming part of you, and destined for continued eternal recycling after you die. The carbon that makes up your eyeballs, your neurons, and your fingertips was derived from food you ate, and the plants who captured that carbon pulled it from the atmosphere, and it got *there* perhaps from volcanic emissions, or exsolution from the ocean, the rotting of a dinosaur, or the combustion of ancient coal. You are literally made of Earth bits. When you eat or breathe or pee, you are connecting with the Earth, as surely as a child poking around in the soil.



We instinctively connect with the Earth from a very young age.
Callan Bentley photo.

You and I are fortunate to live on a very interesting planet. Earth is big enough to have differentiated. This allowed the development of a metallic core, whose circulation powers a magnetic field that protects our atmosphere from erosion by the solar wind. Differentiation also provides for mantle convection, the power source driving plate tectonics. The eruption of volcanoes and gravitational acquisition of comets both yield water, and this water is critical for life. Every critter and microbe needs liquid water, and for 4.5 billion years, our planet's surface has had the right blend of solar heating and greenhouse gas insulation to keep the water flowing. *What fortune!* Every move we make, every breathe we take, we rely on geology.

How has this charmed situation come to pass? If you speak Rock, you can learn some of the backstory. If you speak Rock, a drive through the mountains or the patterns in a fancy countertop become interesting tales, potentially shocking revelations. If you speak Rock, you can answer questions about the circumstances that led to your very existence.

Understanding Historical Geology is essential work for the sake of our species' future, and —more personally — for the sake of feeling at home here on the most fascinating planet in the neighborhood.

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CHAPTER OVERVIEW

2: Planet Earth - An Introduction

2.1: Origin of the Universe- From Big Bang to Planet Earth

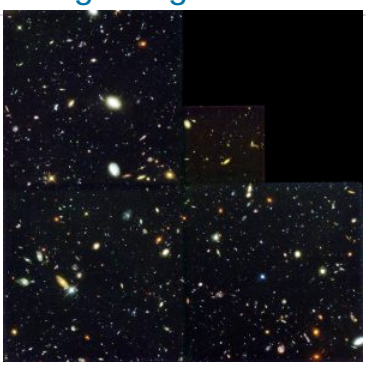
2.2: Geologic Time

2.3: Key Events in Earth History

2.4: Earth Systems

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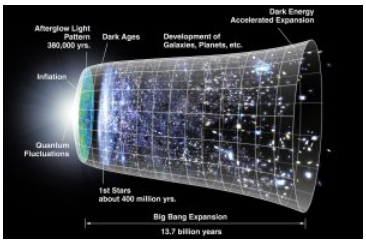
2.1: Origin of the Universe- From Big Bang to Planet Earth



The Hubble Deep Field. This image, released in 1996, is a composite long-exposure picture of one of the darkest parts of the night sky. Every light on this image that does not have diffraction spikes is believed to be an entire galaxy, with hundreds of billions of stars, demonstrating the immense size and scope of the universe.

The universe appears to have an infinite number of galaxies and solar systems and our solar system occupies a small section of this vast entirety. The origins of the universe and solar system set the context for conceptualizing the Earth's origin and early history.

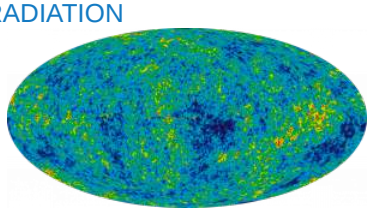
Big-Bang Theory



Timeline of the expansion of the Universe

The mysterious details of events before and during the universe's origin are subject to great scientific debate. The prevailing idea about how the universe was created is called the **big-bang theory**. Although the ideas behind the big-bang theory feel almost mystical, they are supported by Einstein's theory of general relativity. Other scientific evidence, grounded in empirical observations, supports the big-bang theory. The big-bang theory proposes that the universe was formed from an infinitely dense and hot material core. The bang in the title suggests there was an explosive, outward expansion of all matter and space that created atoms. Spectroscopy confirms that hydrogen makes up about 74% of all matter in the universe. Since its creation, the universe has been expanding for 13.8 billion years, and recent observations suggest that this expansion is increasing.

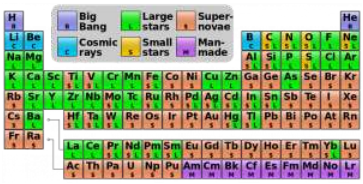
COSMIC MICROWAVE BACKGROUND RADIATION



Heat map showing slight variations in background heat, which is related to cosmic background radiation.

A strong indication of the big bang is **cosmic microwave background radiation**. Cosmic radiation was accidentally discovered by Arno Penzias (1933–) and Robert Woodrow Wilson (1936–) when they were trying to eliminate background noise from a communication satellite. They discovered very faint traces of energy or heat that are omnipresent across the universe. This energy was left behind from the big bang, like an echo.

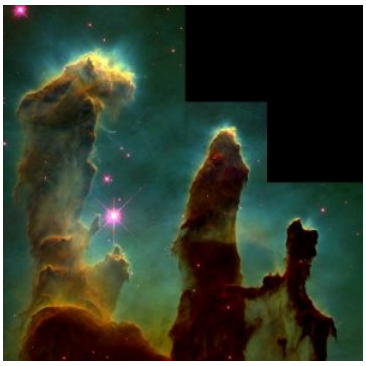
Stellar Evolution



The origin of the elements on the periodic table shows the star life cycle's important role.

Astronomers think the big bang created lighter elements, mostly hydrogen, and smaller amounts of elements helium, lithium, and beryllium. Another process must be responsible for creating the other 90 heavier elements. The current model of stellar evolution explains the origins of these heavier elements.

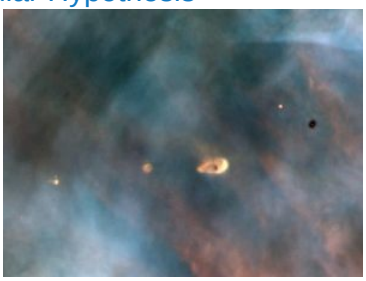
BIRTH OF A STAR



Section of the Eagle Nebula known as "The Pillars of Creation."

Stars start their lives as elements floating in cold, spinning clouds of gas and dust known as **nebulas**. Gravitational attraction or perhaps a nearby stellar explosion causes the elements to condense and spin into a disk shape. In the center of this disk shape, a new star is born under the force of gravity. The spinning whirlpool concentrates material in the center, and the increasing gravitational forces collect even more mass. Eventually, the immensely concentrated mass of material reaches a critical point of such intense heat and pressure it initiates fusion.

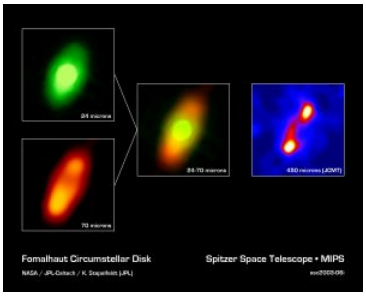
Origin of the Solar System: The Nebular Hypothesis



Small protoplanetary discs in the Orion Nebula

Our solar system formed at the same time as our Sun, as described in the nebular hypothesis. The **nebular hypothesis** is the idea that a spinning cloud of dust made of mostly light elements, called a nebula, flattened into a protoplanetary disk and became a solar system consisting of a star with orbiting planets. The spinning nebula collected the vast majority of material in its center, which is why the sun Accounts for over 99% of the mass in our solar system.

Planet Arrangement and Segregation



This disk is asymmetric, possibly because of a large gas giant planet orbiting relatively far from the star.

As our solar system formed, the nebular cloud of dispersed particles developed distinct temperature zones. Temperatures were very high close to the center, only allowing condensation of metals and silicate minerals with high melting points. Farther from the Sun, the temperatures were lower, allowing lighter gaseous molecules such as methane, ammonia, carbon dioxide, and water condensation. This temperature differentiation resulted in the inner four planets of the solar system becoming rocky, and the outer four planets becoming gas giants.

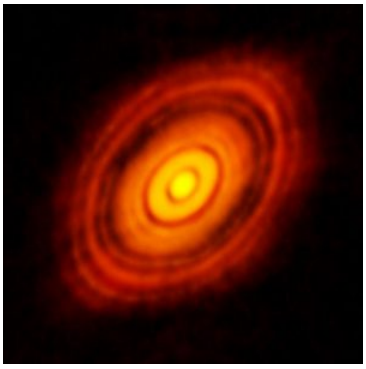
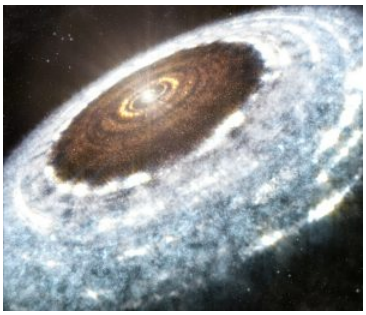


Image by the ALMA telescope of HL Tauri and its protoplanetary disk, showing grooves formed as planets absorb material in the disk.

Image by the ALMA telescope of HL Tauri and its protoplanetary disk, showing grooves formed as planets absorb material in the disk.

Both rocky and gaseous planets have a similar growth model. Particles of dust, floating in the disc were attracted to each other by static charges and eventually, gravity. As the clumps of dust became bigger, they interacted with each other—colliding, sticking, and forming proto-planets. The planets continued to grow over the course of many thousands or millions of years, as material from the protoplanetary disc was added. Both rocky and gaseous planets started with a solid core. Rocky planets built more rock on that core, while gas planets added gas and ice. Ice giants formed later and on the furthest edges of the disc, accumulating less gas and more ice. That is why the gas-giant planets Jupiter and Saturn are composed of mostly hydrogen and helium gas, more than 90%. The ice giants Uranus and Neptune are composed of mostly methane ices and only about 20% hydrogen and helium gases.



This artist’s impression of the water snowline around the young star V883 Orionis, as detected with ALMA.

This artist’s impression of the water snowline around the young star V883 Orionis, as detected with ALMA.

The planetary composition of the gas giants is clearly different from the rocky planets. Their size is also dramatically different for two reasons: First, the original planetary nebula contained more gases and ices than metals and rocks. There was abundant hydrogen, carbon, oxygen, nitrogen, and less silicon and iron, giving the outer planets more building material. Second, the stronger gravitational pull of these giant planets allowed them to collect large quantities of hydrogen and helium, which could not be collected by the weaker gravity of the smaller planets.



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

Jupiter's massive gravity further shaped the solar system and growth of the inner rocky planets, including our planet Earth. As the nebula started to coalesce into planets, Jupiter's gravity accelerated the movement of nearby materials, generating destructive collisions rather than constructively gluing material together. These collisions created the asteroid belt, an unfinished planet located between Mars and Jupiter. This asteroid belt is the source of most **meteorites** currently impacting the Earth. The study of asteroids and meteorites help geologist determine the age of Earth and the composition of its core, mantle, and crust. Jupiter's gravity may also explain Mars' smaller mass, with the larger planet consuming material as it migrated from the inner to the outer edge of the solar system.

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2.2: Geologic Time

Since 4.54 byr is a large chunk of time, geologists have divided it into more manageable chunks by creating a time scale. The commonly accepted time scale comes from the International Commission on Stratigraphy (Figure 3.1). It is continually revised as new research fine-tunes numbers between time scale divisions. The one in Figure 3.1 is the most up-to-date at the time of this writing and will be referenced throughout this manual. The divisions on the time scale are often based on significant events that have taken place tectonically, biologically, or climatically. The numerical ages are derived from the radiometric dating of rocks, minerals, and fossils.

Geologic time is first divided into eons; these are the Hadean, Archean, Proterozoic, and Phanerozoic. The first three eons are often called the Precambrian, which we'll call a "super" eon. The eons are subdivided into eras, and eras are subdivided into periods, periods into epochs, and epochs into ages.

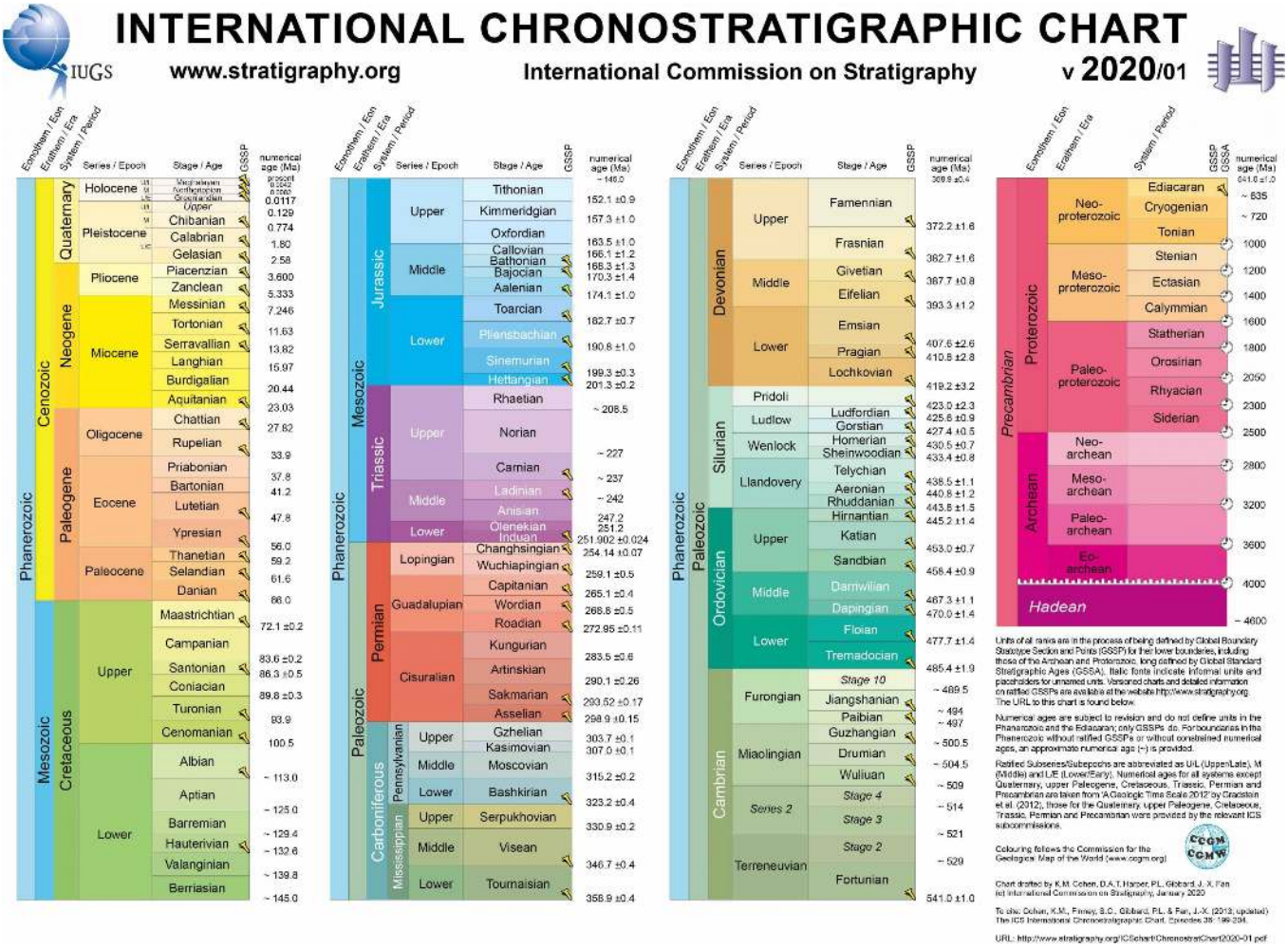


Figure 3.1 – The geologic time scale subdivided by eon, era, system, series, epoch, and stage. What do you observe about each period of time? It is simple; they are color-coded. We will use these standard colors throughout this lab manual. The golden spikes at age boundaries indicate a specific place used to define the age boundary. Image credit: International Commission on Stratigraphy.Geologists use abbreviations to refer to the different portions of geologic time (Table 3.2), particularly on geologic maps. A few of these abbreviations may seem puzzling because the abbreviation isn't always the first letter of the name. Carboniferous got the "C", so Cambrian got a C with a slash through it "€". This left Cretaceous with a "K." Tertiary had originally received the "T", which left Triassic to be symbolized with a T that had an R subscript. In 2003, "Tertiary" was stricken from the geologic timescale and replaced by the Paleogene and Neogene. You will still see "T" on older maps for Tertiary, and some geologists still use the term today. So, be careful! There were already some M's and P's; so, Proterozoic eras were assigned as X, Y, and Z.Geologists use two methods for dating events in Earth's history. The first is called relative dating, meaning how events relate to each other in time, or more plainly, they figure out the sequence of events (what came first, second, third, etc.). Relative dating

has no regard for numerical ages. The second method is absolute dating, where geologists use radioactive isotopes to figure out the numerical age of a rock or mineral.

Geologic Time	Abbreviation
Quaternary	Q
Neogene	Ng
Paleogene	Pg
Cretaceous	K
Jurassic	J
Triassic	T_R
Permian	P
Carboniferous	C
Devonian	D
Silurian	S
Ordovician	O
Cambrian	€
Neoproterozoic	Z
Mesoproterozoic	Y
Paleoproterozoic	X
Archean	A

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2.3: Key Events in Earth History

At this point, you do not need to worry about all the small details of the Geologic Timescale. What is important is remembering how it is divided up into several very different eras that reflect completely different conditions on Earth. For instance, the main Eons in the Geologic Timescale are the Hadean, the Archean, the Proterozoic, and the Phanerozoic. How are these four Eons in Earth history different?

The **Hadean** is the earliest Eon in Earth's history. It is derived from the word "Hades" which reflects the hellish conditions that were thought to persist on the planet. These conditions are thought to include a global magma ocean as Earth was just forming and differentiating or ordering and separating out its internal layers, the crust, mantle, and core. At the beginning of this Eon, Earth did not have a moon, water, continents, or plate tectonics – it was completely uninhabitable to life as we know it today. Much of that changed throughout the Hadean as Earth developed a liquid ocean, atmosphere, and the moon formed – more on how all of this happened later in the course.



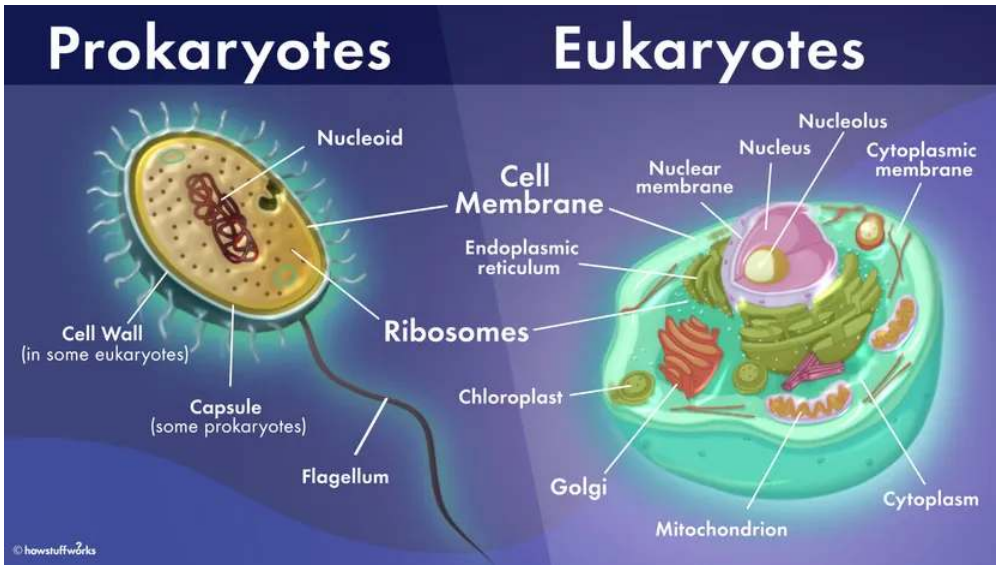
The Hadean Era by Lori Mayden

The **Archean** (Greek for "ancient") is the next Eon that follows in Earth's history. For some geologists, this is the most exciting of the Eons. Earth cooled down enough to form the continental crust, differentiated enough to begin an early process of plate tectonics. Most notably, the first signs of life have been found in the geologic record as far back as 3.5 billion years ago. Early life was only single-celled bacteria, and the atmosphere of Earth lacked any free oxygen compared to what the modern atmosphere is like today.

The Archean Eon (4,600 – 2,500 Million Years Ago) | Smithsonian Ocean

The **Proterozoic** (Greek for "early life") Eon marked a transition to a more oxygenated atmosphere, courtesy of photosynthesizing bacteria. Eukaryotes began to evolve on the planet during this Eon, and these single-celled organisms were the first to develop a

central nucleus and organelles capable of performing tasks for the whole organism. The development of eukaryotes was important because all multicelled life, including animals and humans, evolved from this domain of life.



Geologically, the Earth experienced several episodes of what is called a “Snowball Earth” event; in which most or all of the Earth was thought to freeze over during a glacial age. The end of these glaciations is thought to be marked by a large release of nutrients and a movement of carbon reservoirs throughout the planet that might have affected the evolution of the earliest multicellular organisms. These organisms are best preserved in the Ediacaran Era (635-541 Ma).

The story of Snowball Earth | Astronomy.com

The **Phanerozoic** (Greek for visible life) Eon is the time in which we are currently living. This Eon is characterized by a huge diversification in animals and plants. First many hard-shelled organisms appeared in oceans in what is called the Cambrian Explosion beginning in 541 Ma, although vertebrate fishes also began to evolve. Around the same time, land plants colonized the continents.



The Phanerozoic Eon is divided into three distinct Eras – the Paleozoic (Ancient Life), Mesozoic (Middle Life), and Cenozoic (Recent Life). The Paleozoic is marked by diverse ocean life, early land reptiles and amphibians, large insects, and enormous plants. The Mesozoic is well-known as the age of the Dinosaurs, although it is also marked by the evolution of mammals and flowering plants. The Cenozoic saw the diversification of mammals on land and sea and the evolution of humans.

Mass Extinctions

The Phanerozoic is also marked by Mass Extinction events, which mark a significant die-off of species in a short time. There have been five (perhaps even six!) extinctions during the past 541 million years, each with a different cause. Let's learn about these extinctions below!



A YouTube element has been excluded from this version of the text. You can view it online here: https://pb.libretexts.org/historicalgeology/?p=40

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2.4: Earth Systems

Throughout its entire 4.54 Ga history, every event in the history of our planet happened because something forced it to happen. As we've reviewed in the previous section, an asteroid caused the extinction of the dinosaurs at the end of the Cretaceous period. Extinctions led to new evolutionary pathways for the origination of species. Plate tectonics caused continents to move around. In each of these realities, we might be tempted to simply describe them as they have been here, in a linear fashion. A caused B. We could leave it at that. However, is it that simple?

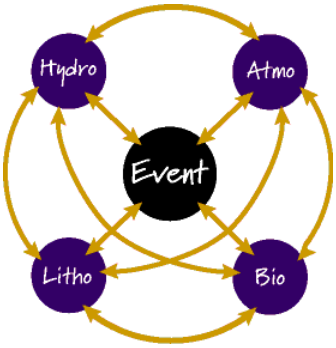
Within the Earth system, there are subsystems. These subsystems encompass the space environment (exosphere), gaseous environment (atmosphere), liquid environment (hydrosphere), solid environment (lithosphere and geosphere, hereafter lithosphere), and living environment (biosphere). In some cases, the icy environment (cryosphere) is broken away from the hydrosphere and the anthroposphere (human environment) from the biosphere for emphasis.

Energy, coming from the Sun on the one hand and the Earth's interior on the other, powers these systems. As energy flows through these systems, so do nutrients and elements through biogeochemical cycles. Examples of these include the carbon and nitrogen cycles.



Earth systems science seeks to understand the interconnectedness of our planet's air, water, land, and life (M. Ruzek 1999).

The Earth system is made up of four fundamental subsystems, or spheres. The atmosphere, hydrosphere, biosphere, and lithosphere/geosphere. In the diagram below and to the left, these are abbreviated. In a moment, we will explore an expanded view of these spheres that includes six of them.



Earth system interactions due to an "Event". Events can be a wide variety of things that trigger interactions, or forcings, within the Earth system.
Graphic: Wheeling Jesuit University/NASA Classrooms of the Future; reproduced with permission.

Events affect the Earth system as a whole, though they usually begin within a particular sphere. Let us explore a volcanic eruption as an example. When a volcanic eruption event occurs, what happens?

- Ash and rock are blasted into the atmosphere;
- Gases are emitted into the atmosphere;
- Lava may run downslope;
- Pyroclastic flows, superheated ash, may rush downslope at great velocity;
- Lahars, or mudflows, may result from the rapid melting of snow and ice, bringing large amounts of debris downslope;
- Flora and fauna near the volcano may be killed.

There are other effects associated with volcanic eruptions beyond these. If you were to place the eruption in the very center of this diagram and write some of these interactions along the golden arrows toward a sphere, where would you place them? These are direct effects on each sphere from the event.

These direct effects are not the end of the story. You can see that there are also arrows connecting each sphere. What are some of the indirect interactions (forcings) that result from the eruption? How is the biosphere affected, for example, by how the eruption directly affects the lithosphere (geosphere)? Ultimately, understanding the interconnectedness of the Earth system is critical for gaining a holistic view of how our planet is affected by individual events. In the history of our planet, there has been no shortage of such events!

The table below shows some of the relationships between these events that cause a relationship to form between the different spheres, called forcing events. They are called forcing events because they cause a shift in energy flow from one system to another.

Forcing Event	Lithosphere/Geosphere	Hydrosphere	Atmosphere	Biosphere
Melting of Polar Ice	Direct – Isostatic adjustment as ice located on land melts	Direct – influx of freshwater into the marine realm affects thermohaline circulation. Change in planetary albedo leads to warmer seawater due to the loss of light-colored ice	Direct – Loss of sea ice leads to warming of oceans and changes in evaporation and weather patterns	Habitat fragmentation of organism ranges in polar regions
Stratovolcanic Eruption	Direct – Release of eruptive materials in the vicinity of the mountain and the creation of volcanic strata	Indirect via Atmosphere – the influx of sulfur and carbon dioxide may lead to lower water pH	Direct – influx of ash, sulfur, and carbon dioxide	Direct – Major effects on life in the vicinity of the volcano; Temporary regional or even global cooling affects some groups of organisms.
Supervolcano Eruption	Direct – Major new volcanic deposits form over a large area. Vicinity of eruption subject to collapse.	Indirect via Atmosphere – a surge of carbon and sulfur dioxide in the atmosphere leads to lower pH in marine and freshwater environments.	Direct – Major influxes of volcanic gases and ash for extended periods.	Direct – Local, regional, and perhaps global extinctions due to deposits of massive amounts of volcanic material and extended global cooling
Deforestation	Indirect via Biosphere – Loss of vegetation leads to more mass wasting events	Indirect via Biosphere – Loss of vegetation affects regional transpiration, water storage, and drainage patterns	Direct – Loss of plant life leads to higher concentrations of carbon dioxide in the atmosphere	Direct – Habitat fragmentation and biodiversity loss. Possible localized or regional extinctions
Asteroid Impact	Direct – Major crater formed, ejecta deposits created over a large area; high-pressure metamorphism in the zone of impact; Seismic energy release	Direct – Vaporization of large amounts of surface and groundwater from a local to a global level, depending upon the size and composition of the impactor	Direct – Frictional heating as impactor enters atmosphere; Heating as hot ejecta is thrown into atmosphere and exosphere that rains back down on Earth; Introduction of impact-related particulate matter	Direct – local to a global extinction event
Earthquake	Direct – Local to regional changes in the landscape near the fault	Indirect via Geosphere – Possible tsunamis or coastal landslides and turbidity currents	Indirect via Geosphere – Release of dust, debris, and toxic gases in some instances	Direct and Indirect – Depending on the species, building collapse and liquefaction can lead to death; Ecosystem changes are possible
Burning of Fossil Fuels	Indirect via Atmosphere – Addition of particulate matter and black carbon to sedimentary deposits;	Direct – Changes in regional evaporation rates due to warming; Changes in the hydrologic cycle	Direct – addition of carbon dioxide warms the troposphere; Changes weather patterns.	Direct – Climate warming leads to biodiversity loss, extinction, etc.

Forcing Event	Lithosphere/Geosphere	Hydrosphere	Atmosphere	Biosphere
	Climate change leads to other downstream effects			
Viral Pandemic Among Human Population	Indirect – Social isolation and pause in economic activity lead to lower ambient seismic activity.	Indirect – Reduction in pollution; Reduction in habitat disturbance from recreational boating	Indirect – Reduction in air pollution due to changes in economic activity and energy production	Direct – Widespread sickness and mortality among the human species

Consider how the different systems are connected with these event scenarios below:

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CHAPTER OVERVIEW

3: Minerals

3.1: Atoms
3.2: Bonding and Lattices
3.3: Mineral Groups
3.4: Silicate Minerals
3.5: Mineral Properties
3.6: Summary Key Term Check

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3.1: Atoms

Protons Are What Make Elements Distinct

All matter, including mineral crystals, is made up of atoms. All atoms comprise three main particles: **protons**, **neutrons,** and **electrons**. Protons have a positive charge, neutrons have no charge, and electrons have a negative charge. Protons and neutrons have approximately the same mass, but electrons have a mass that is 10,000 times smaller.

The element hydrogen (H) has the simplest atoms. Most hydrogen atoms have just one proton and one electron. The proton forms the **nucleus** (the center of the atom), while the electron orbits around it (Figure below, on the left). All other elements have more than one proton in their nucleus.

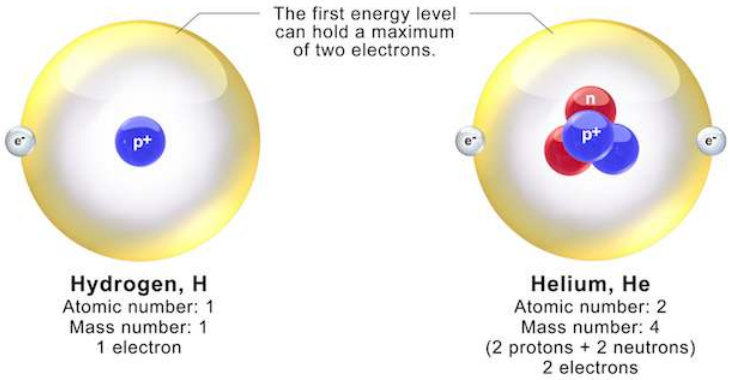


Figure Atomic structure of hydrogen and helium showing protons (p+), neutrons (n), and electrons (e-). Source: Bruce Blaus (2014), CC BY 3.0. View source.

Protons repel each other because they are positively charged, but they can have more than one proton in a nucleus because neutrons hold them together. The next most complex atom, helium (He) has two protons and two neutrons in its nucleus, in its most common form. Some atoms of the same element can have different numbers of neutrons. For example, forms of hydrogen exist with one and two neutrons, and a tiny fraction of He atoms have only one neutron. Forms of an element with different numbers of neutrons are called **isotopes**.

The number of protons in an atom determines what element it will be, so the number of protons is called the **atomic number** of that element. The total number of protons and neutrons in the nucleus is the **mass number**. The mass number distinguishes between isotopes of an element. Isotopes of an element are denoted by putting the mass number as a subscript in front of the symbol for that element. For example, the isotopes of hydrogen are ^{1}H (1 proton), ^{2}H (1 proton + 1 neutron), and ^{3}H (1 proton + 2 neutrons).

For most of the 16 lightest elements (up to oxygen), the number of neutrons equals the number of protons. For most of the remaining elements, there are more neutrons than protons. This is because the more protons are concentrated in a small space, the more neutrons are needed to keep the nucleus together. For example, the most common isotope of uranium (U) is ^{238}U. It has 92 protons but requires 146 neutrons to keep them together. The neutrons are only partly successful: uranium is radioactive, meaning its nucleus will eventually split apart and release energy. What remains of the nucleus has fewer protons, so after decay, the atom is a different element.

Did You Get All of That? Let's See!

Electrons Are What Control How Atoms Interact

Electrons orbiting around the nucleus of an atom are arranged in **shells** (also called **energy levels**). The first shell can hold only two electrons, but the next shell holds up to eight electrons. An atom can have many shells of electrons, but there are never more than 8 outermost electrons interacting with surrounding atoms.

The outermost electrons determine how atoms can be bonded together. Elements with a full outer shell (e.g., neon, Figure below, to the right) are **inert** because they do not react with other elements to form compounds. These are the **noble gases** (including helium, argon, krypton, and radon, in addition to neon) in the far-right column of the periodic table. For elements that do not have a full

outer shell (e.g., lithium, Figure below, on the left), the outermost electrons can interact with the outermost electrons of nearby atoms to create chemical bonds.

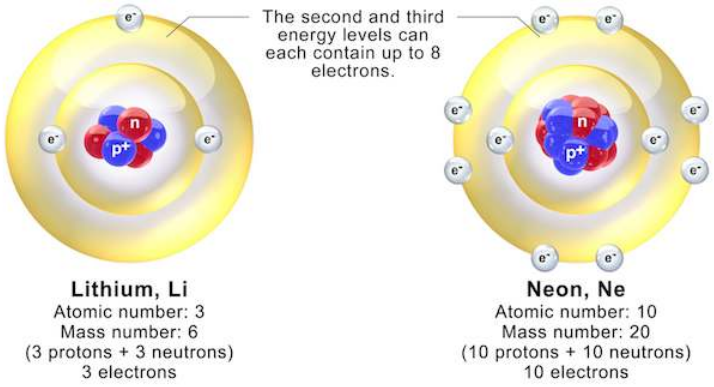


Figure The number of electrons in an atom's outermost shell (or energy level) determines whether it will bond to other atoms and how it will bond. Right- Neon has a filled outer shell with 8 electrons. It does not bond with other atoms. Left- Lithium has only one electron in its outer shell. It bonds with other atoms. Source: Bruce Blaus (2014), CC BY 3.0. View source.

The electron shell configurations for 29 of the first 36 elements are listed in the Table below. Note that some of the shells in the table below have more than 8 electrons. This is because they contain subshells. For example, the third shell can hold up to 18

electrons because it contains one subshell that can hold 2 electrons and two subshells that can hold 8 electrons each.

Table Electron shell configurations of some of the elements up to krypton. Inert elements (those with filled outer shells) are shaded.

			Number of Electrons in Each Shell			
Element	**Symbol**	**Atomic Number**	**First**	**Second**	**Third**	**Fourth**
Hydrogen	H	1	1			
Helium	**He**	**2**	**2**			
Lithium	Li	3	2	1		
Beryllium	Be	4	2	2		
Boron	B	5	2	3		
Carbon	C	6	2	4		
Nitrogen	N	7	2	5		
Oxygen	O	8	2	6		
Fluorine	F	9	2	7		
Neon	**Ne**	**10**	**2**	**8**		
Sodium	Na	11	2	8	1	
Magnesium	Mg	12	2	8	2	
Aluminum	Al	13	2	8	3	
Silicon	Si	14	2	8	4	
Phosphorus	P	15	2	8	5	
Sulphur	S	16	2	8	6	
Chlorine	Cl	17	2	8	7	
Argon	**Ar**	**18**	**2**	**8**	**8**	
Potassium	K	19	2	8	8	1
Calcium	Ca	20	2	8	8	2
Scandium	Sc	21	2	8	9	2
Titanium	Ti	22	2	8	10	2
Vanadium	V	23	2	8	11	2
Chromium	Cr	24	2	8	13	1
Manganese	Mn	25	2	8	13	2
Iron	Fe	26	2	8	14	2
.	.	.	.	.	.	.
Selenium	Se	34	2	8	18	6
Bromine	Br	35	2	8	18	7

Krypton	**Kr**	**36**	**2**	**8**	**18**	**8**

Do You Understand Why Elements React?

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3.2: Bonding and Lattices

Atoms seek to have a full outer shell. For hydrogen and helium, a full outer shell means two electrons. For other elements, it means 8 electrons. The outer shell is filled by transferring or sharing electrons with other atoms in chemical bonds. The type of chemical bond is important for the study of minerals because the type of bond will determine many of a mineral's physical and chemical properties.

Ionic Bonds

Consider the example of halite again, which is made up of sodium (Na) and chlorine (Cl). Na has 11 electrons: two in the first shell, eight in the second, and one in the third (see the top half of the figure below). Na readily gives up the third shell electron so it can have the second shell with 8 electrons as its outermost shell. When it loses the electron, the electron's total charge is -10, but the proton's total charge is +11, so it is left with a +1 charge overall.

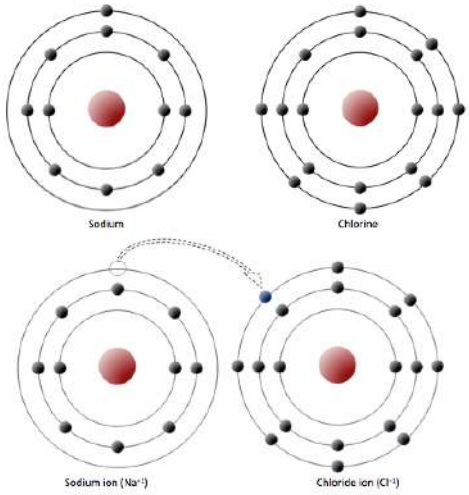


Figure Electron configuration of sodium and chlorine atoms (top). Sodium gives up an electron to become a cation (bottom left) and chlorine accepts an electron to become an anion (bottom right). Source: Steven Earle (2015), CC BY 4.0. View source.

Chlorine has 17 electrons: two in the first shell, eight in the second, and seven in the third. Cl readily accepts an eighth electron to fill its third shell, and therefore becomes negatively charged because it has a total charge of -18 from electrons, and a total charge of +17 from protons.

In changing their number of electrons, these atoms become **ions**—the sodium loses an electron to become a positive ion or **cation**, [1] and the chlorine gains an electron to become a negative ion or **anion** (look at the bottom half of the figure above). Because negative and positive charges attract, sodium and chlorine ions stick together, creating an **ionic bond**. In an ionic bond, electrons can be thought of as having transferred from one atom to another.

Cation or Anion?

Covalent Bonds

Some elements can also form bonds by sharing electrons rather than giving away or receiving them. This will happen when electrons are held especially tightly by their atoms. Chlorine gas (Cl_2, Figure below) is formed when chlorine atoms share two outer-shell electrons so that each has a complete outer shell. This is called a **covalent bond**.

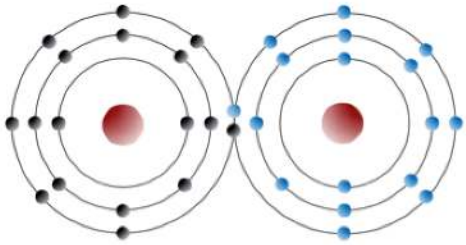


Figure A covalent bond between two chlorine atoms. The electrons are black in the left atom and blue in the right atom. Two electrons are shared (one black and one blue) so that each atom appears to have a full outer shell. Source: Steven Earle (2015), CC BY 4.0. View source.

Carbon is another atom that participates in covalent bonding. An uncharged carbon atom has six protons and six electrons. Two of the electrons are in the inner shell and four are in the outer shell (left half of figure below). Carbon would need to gain or lose four electrons to have a filled outer shell, creating too great a charge imbalance. Instead, carbon atoms share electrons to create covalent bonds (see right half of the figure below).

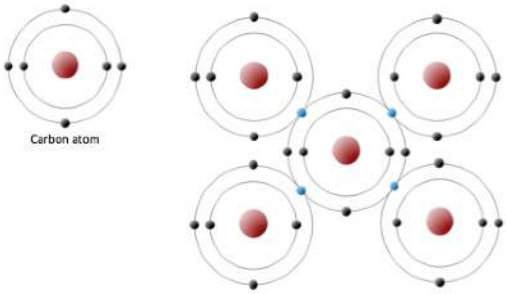


Figure The electron configuration of carbon (left) and the sharing of electrons in covalent C bonding (right). The electrons shown in blue are shared between adjacent C atoms. Source: Steven Earle (2015), CC BY 4.0. View source.

Carbon atoms are linked in a three-dimensional framework in the mineral diamond (shown on the left-hand side of the figure below). Each carbon atom is bonded to four other carbon atoms. Every bond is a very strong covalent bond.

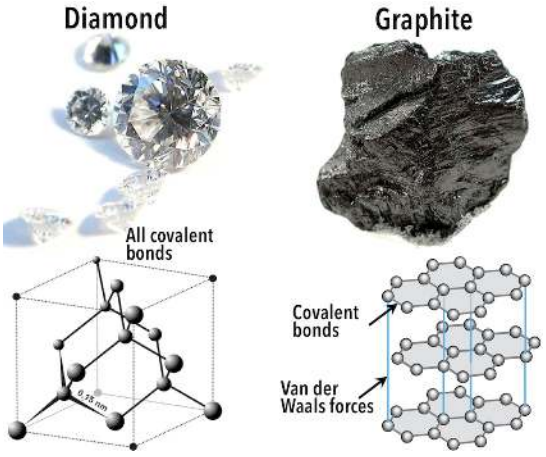


Figure Covalently bonded structures. Left: Diamond with a three-dimensional structure of covalently bonded carbon. Right: Graphite with covalently bonded sheets of carbon. Sheets are held together by weaker van der Waals forces. Source: Karla Panchuk (2018), CC BY 4.0. Modified after Materialscientist (2009), CC BY-SA 3.0. View source.

Other Types of Bonds

Most minerals are characterized by ionic bonds, covalent bonds, or a combination of the two, but other types of bonds are important in minerals.

Van der Waals Forces and Hydrogen Bonds

Consider the mineral graphite (Figure above, right-hand side): carbon atoms are linked in sheets or layers in which each carbon atom is covalently bonded to three others. Graphite-based compounds are strong because of the covalent bonding between carbon atoms within each layer, which is why they're used in high-end sports equipment such as ultralight racing bicycles. However, Graphite is soft because the layers themselves are held together by relatively weak **Van der Waals forces**.

Van der Waals forces, like **hydrogen bonds**, work because molecules can be electrostatically neutral but still have an end that is slightly more positive and an end that is slightly more negative. In water molecules (Figure below, left-hand side), the bent shape puts the hydrogen atoms on one side of the molecule, and the oxygen atom, with more electrons, on the other. The charge is distributed asymmetrically across the water molecule. Contrast this with the straight carbon dioxide molecule (Figure below, right-hand side). The slightly more negative oxygen atoms on the ends are distributed symmetrically on either side of the carbon atom.

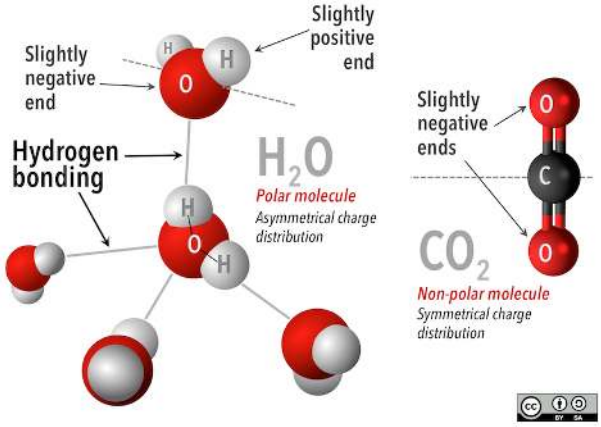


Figure Hydrogen bonding. Water molecules (left) are polar molecules (their charge is distributed asymmetrically). Slightly negative parts of the molecule attract slightly positive parts of other water molecules. Carbon dioxide (right) is a non-polar molecule. The slightly negative oxygen atoms are distributed symmetrically on either side of the carbon atom. Source: Karla Panchuk (2018), CC BY-SA 4.0. Modified after Querter (2011, CC BY-SA 3.0; view source) and Jynto (2011, CC0 1.0; view source).

Metallic Bonds

Metallic bonding occurs in metallic elements because they have outer electrons that are relatively loosely held. (The metals are highlighted on the periodic table in Appendix 1.) When bonds between such atoms are formed, the **dissociated electrons** can move freely from one atom to another. This feature accounts for two very important properties of metals: their electrical conductivity and their malleability (they can be deformed and shaped).

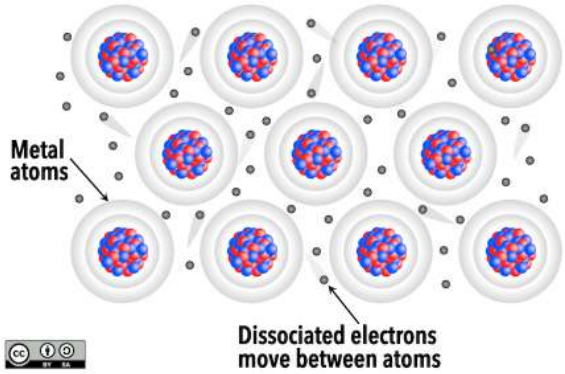


Figure Metallic bonding. Dissociated electrons (grey dots) move between metal atoms. Source: Karla Panchuk (2018), CC BY-SA 4.0. Nucleus by Fornax (2010), CC BY-SA 3.0. View source.

Chemical Bond Types

1. You can remember that a cation is positive by remembering that a cat has paws (paws sounds like "pos" in "positive"). You could also think of the "t" in "cation" as a plus sign. ↵

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3.3: Mineral Groups

Review: Definition of a Mineral

Minerals are organized according to the anion or anion group (a group of atoms with a net negative charge, e.g., SO_4^{2-}) they contain, because the anion or anion group has the biggest effect on the properties of the mineral. Silicates, with the anion group SiO_4^{4-}, are by far the most abundant group in the crust and mantle. The different mineral groups and some examples of minerals in each group are summarized below.

Oxide Minerals: O^{2-} Anion

Oxide minerals have oxygen (O^{2-}) as their anion (see figure below). They don't include anion groups with other elements, such as the carbonate (CO_3^{2-}), sulfate (SO_4^{2-}), and silicate (SiO_4^{4-}) anion groups. The iron oxides hematite and magnetite are two examples of iron's important ores. Corundum is an abrasive but can also be a gemstone in its ruby and sapphire varieties. The mineral is known as a hydroxide if the oxygen is combined with hydrogen to form the hydroxyl anion (OH^-). Some important hydroxides are limonite, bauxite, and iron and aluminum ores.

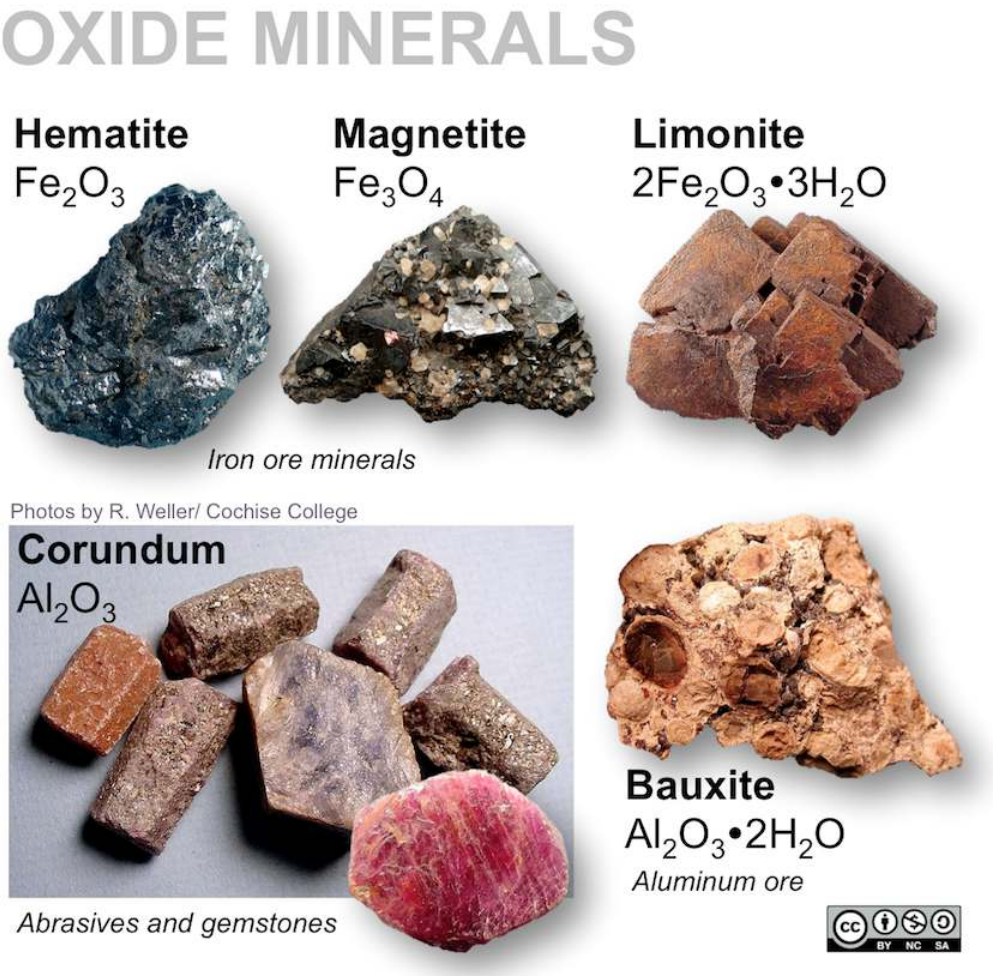


Figure: Oxide minerals include metal ore minerals, industrial minerals, and gemstones. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Photos by R. Weller/ Cochise College.

Sulfide Minerals: S^{2-} Anion

Sulfide minerals (Figure below) include galena, sphalerite, chalcopyrite, and molybdenite, which are the most important ores of lead, zinc, copper, and molybdenum, respectively. Some other sulfide minerals are pyrite, bornite, stibnite, and arsenopyrite. Sulfide minerals tend to have a metallic sheen.



Figure Sulphide minerals often have a metallic luster and include metal ores. Source: Karla Panchuk (2018) CC BY-NC-SA 4.0. Photos by R. Weller/ Cochise College. Click for more attributions.

Sulfate Minerals: SO_4^{2-} Anion Group

Many **sulfate** minerals form when sulfate-bearing water evaporates. A deposit of sulfate minerals may indicate that a lake or sea has dried up at that location. Sulfates with calcium include anhydrite and gypsum (Figure below). Sulfates with barium and strontium are barite and celestite, respectively. The cation has a +2 charge in all these minerals, which balances the –2 charge on the sulfate ion.



Figure Sulphate minerals. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Click for more attributions.

Halide Minerals: Anions from the Halogen Group

The anions in **halides** are the **halogen** elements, including chlorine, fluorine, and bromine. Examples of halide minerals are cryolite, fluorite, and halite (Figure below). Halide minerals are made of ionic bonds. Like the sulfates, some halides also form when mineral-rich water evaporates.



Figure Halide minerals. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Click for more attributions.

Carbonate Minerals: CO_3^{2-} Anion Group

The **carbonate** anion group combines with +2 cations to form minerals such as calcite, magnesite, dolomite, and siderite (Figure below). The copper minerals malachite and azurite are also carbonates. The carbonate mineral calcite is the main component of rocks formed in ancient seas by organisms such as corals and algae.

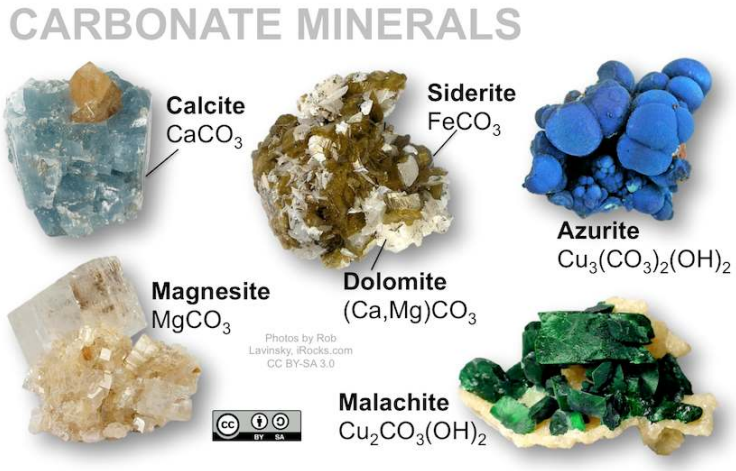


Figure Carbonate minerals. Source: Karla Panchuk (2018), CC BY-SA 4.0. Photos by Rob Lavinsky, iRocks.com, CC BY-SA 3.0. Click for more attributions.

Phosphate Minerals: PO_4^{3-} Anion

The apatite group of **phosphate** minerals (see Figure below, to the left) includes hydroxyapatite, which makes up the enamel of your teeth. Turquoise is also a phosphate mineral (see Figure below, to the right).



Figure Phosphate minerals. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Photos by R. Weller/ Cochise College. Click for more attributions.

Silicates (SiO_4^{4-})

The **silicate** minerals include the elements silicon and oxygen in varying proportions. These are discussed at length in the next section.

Native Element Minerals

These are minerals made of a single element, such as gold, copper, silver, or sulfur/sulphur (Figure below).



Figure Native element minerals are made up of a single element. Source: Karla Panchuk (2018), CC BY-SA 4.0. Click for more attributions.

Beginner Practice with Anionic Groups

Minerals are grouped according to the anion part of the mineral formula, and mineral formulas are always written with the anion part last. For example, in pyrite (FeS_2), Fe^{2+} is the cation, and S^- is the anion. This helps us to know that it's a sulfide, but it isn't always that obvious…

Hematite (Fe_2O_3) is an oxide; that's easy, but anhydrite ($CaSO_4$) is a sulfate because SO_4^{2-} is the anion, not O. Similarly, calcite ($CaCO_3$) is a carbonate, and olivine (Mg_2SiO_4) is a silicate.

Minerals with only one element (such as S) are native minerals, while those with an anion from the halogen column of the periodic table (Cl, F, Br, etc.) are halides.

IMPORTANT: To get all the points for the activity below, scroll through the THREE questions to the FINISH button to report your score! Otherwise, you will lose out on points!

Anionic Groups: Level Up!

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3.4: Silicate Minerals

Silicon and oxygen bond covalently to create a **silicate tetrahedron** (SiO_4^{4-}), which is a four-sided pyramid shape with oxygen at each corner and silicon in the middle (Figure below). This structure is the building block of many important minerals in the crust and mantle. Silicon has a charge of +4, and oxygen has a charge of -2, so the total charge of the silicate anion is -4.

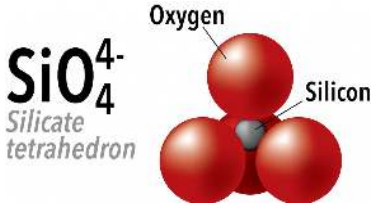


Figure The silica tetrahedron is the building block of all silicate minerals. Source: Karla Panchuk (2018), CC BY-SA 4.0. Modified after Helgi (2013), CC BY-SA 3.0. View source.

In silicate minerals, these tetrahedra are arranged and linked together in various ways, from single units to chains, rings, and more complex frameworks. In the rest of this section, we will look at the structures of the most common silicate minerals in Earth's crust and mantle.

Isolated Tetrahedra

The simplest silicate structure—that of the mineral **olivine** (Figure below)—is composed of isolated tetrahedra bonded to iron and/or magnesium ions. In olivine, the –4 charge of each silica tetrahedron is balanced by two iron or magnesium cations, each with a charge of +2.

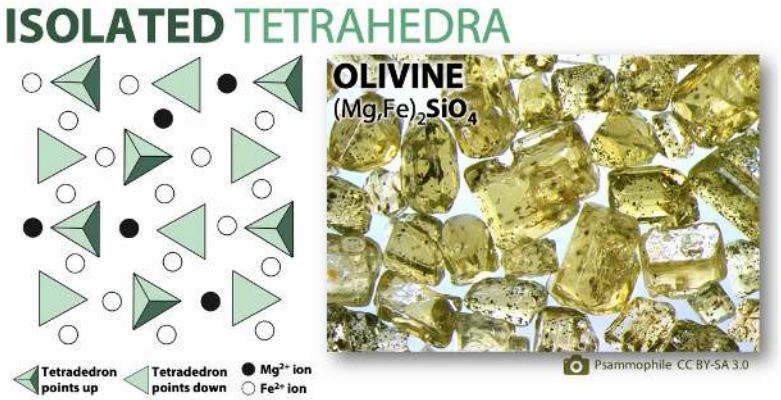


Figure Olivine is a silicate mineral made of isolated silica tetrahedra bonded to Fe and Mg ions (left). Source: Karla Panchuk (2021), CC BY-SA 4.0. Click for more attributions.

Olivine can be pure Mg_2SiO_4 or pure Fe_2SiO_4, or a combination of the two, written as $(Mg,Fe)_2SiO_4$.

Chain Silicates

The mineral **pyroxene** is an example of a **single-chain silicate** (Figure 5.25), where one oxygen from each tetrahedron is shared with the next tetrahedron. Sharing means that fewer oxygens are needed to make the tetrahedra, so there's less oxygen in this structure compared to olivine. This can be expressed as a silicon-to-oxygen ratio (Si:O).

In olivine, tetrahedra aren't connected directly to each other, so each silicon atom must have four oxygen atoms all to itself to complete a tetrahedron. That's a ratio of 1:4. In pyroxene, each silicon atom only needs three unique oxygen atoms because it can borrow one from a neighbor to have a tetrahedron. For pyroxene, the Si:O is 1:3. With one less oxygen in the mix per tetrahedron, the net charge per silicon atom that cations must balance is lower (-2 instead of -4).

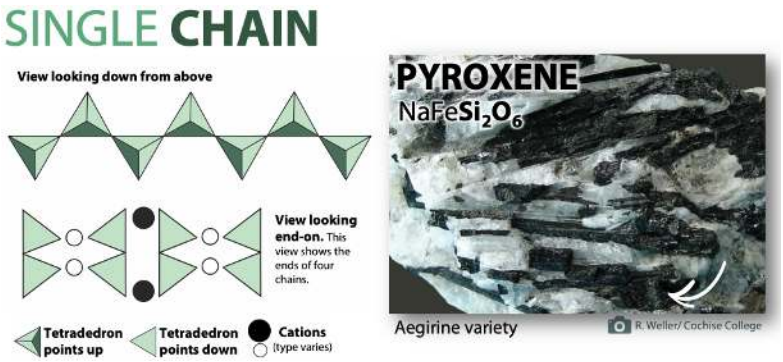


Figure Pyroxene (dark mineral in the photo) is a silicate mineral in which tetrahedra are linked in strings, with adjacent tetrahedra sharing an oxygen atom. Source: Karla Panchuk (2021), CC BY-SA 4.0. End-on view modified after Klein & Hurlbut (1993).Click for more attributions.

The way tetrahedra share oxygens in single-chain silicates is why, even though pyroxene is built out of silicate tetrahedra (with the silicate anion being SiO_4^{4-}), its formula has the tetrahedra represented as SiO_3 (e.g., $MgSiO_3$, $FeSiO_3$, and $CaSiO_3$.

The mineral **amphibole** is also a chain silicate, but the silica tetrahedra are linked in **double chains** (Figure below). Amphibole has a silicon-to-oxygen ratio higher than pyroxene; hence, fewer cations are necessary to balance the charge. Amphibole is even more permissive than pyroxene, and its compositions can be very complex.

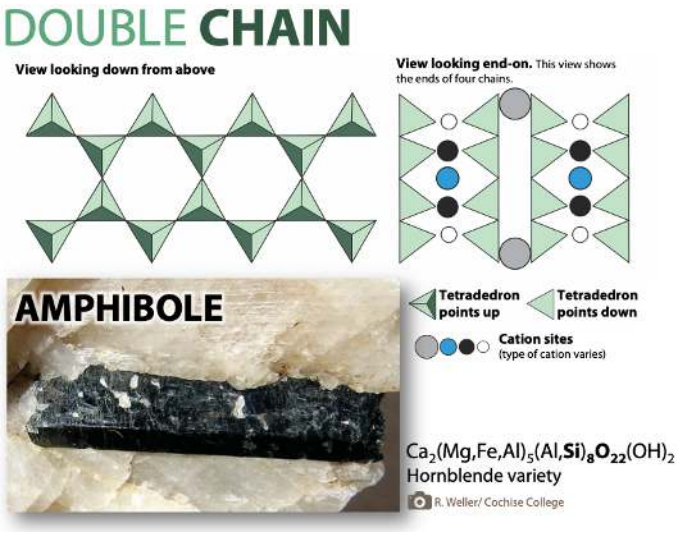


Figure Amphibole (dark mineral in the photo) is a silicate mineral in which tetrahedra are linked in doubled-up strings. Tetrahedra share two oxygen atoms with adjacent tetrahedra. End-on view modified after Klein & Hurlbut (1993). Source: Karla Panchuk (2021), CC BY-SA 4.0. Click for more attributions.

Sheet Silicates

In **mica** structures, the silica tetrahedra are arranged in continuous sheets (Figure below), where each tetrahedron shares three oxygen anions with adjacent tetrahedra. Because even more oxygens are shared between adjacent tetrahedra, fewer charge-balancing cations are needed for sheet silicate minerals.

Bonding between sheets is relatively weak, which accounts for the tendency of mica minerals to split apart in sheets (Figure below, to the bottom right). Two common micas in silicate rocks are **biotite** (Figure below, to the bottom left), which contains iron and/or magnesium, making it a dark mineral, and muscovite (Figure below on the right), which contains aluminum and potassium and is light in color. All of the sheet silicate minerals have water in their structure, in the form of the hydroxyl (OH-) anion.

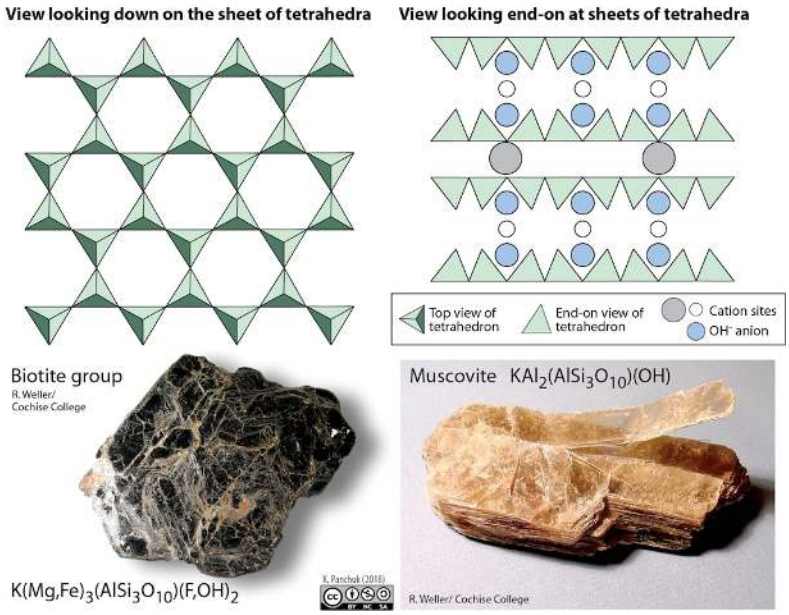


Figure Micas are sheet silicates and split easily into thin layers along planes parallel to the sheets. Biotite mica (lower left) has Fe and Mg cations. Muscovite mica (lower right) has Al and K instead. The muscovite mica shows how thin layers can split away in a sheet silicate. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Top left- Modified after Steven Earle (2015), CC BY 4.0. Top right- Modified after Klein & Hurlbut (1993). Click for more attributions.

Some sheet silicates typically occur in clay-sized fragments (i.e., less than 0.004 mm). These include the clay minerals **kaolinite**, **illite,** and **smectite**, which are important components of rocks and especially of soils.

Framework Silicates

In framework silicates, tetrahedra are connected in three-dimensional structures rather than in two-dimensional chains and sheets.

Feldspar

Feldspars are a group of very abundant framework silicates in Earth's crust. They include alumina tetrahedra as well as silicate tetrahedra. In alumina tetrahedra, an aluminum cation is at the center instead of a silicon cation.

Feldspars are classified using a ternary (3-fold) system with three end-members ("pure" feldspars). This system is illustrated with a triangular diagram with each end member at one corner (Figure below). The distance along a side of the diagram represents the relative abundance of the composition of each end member.

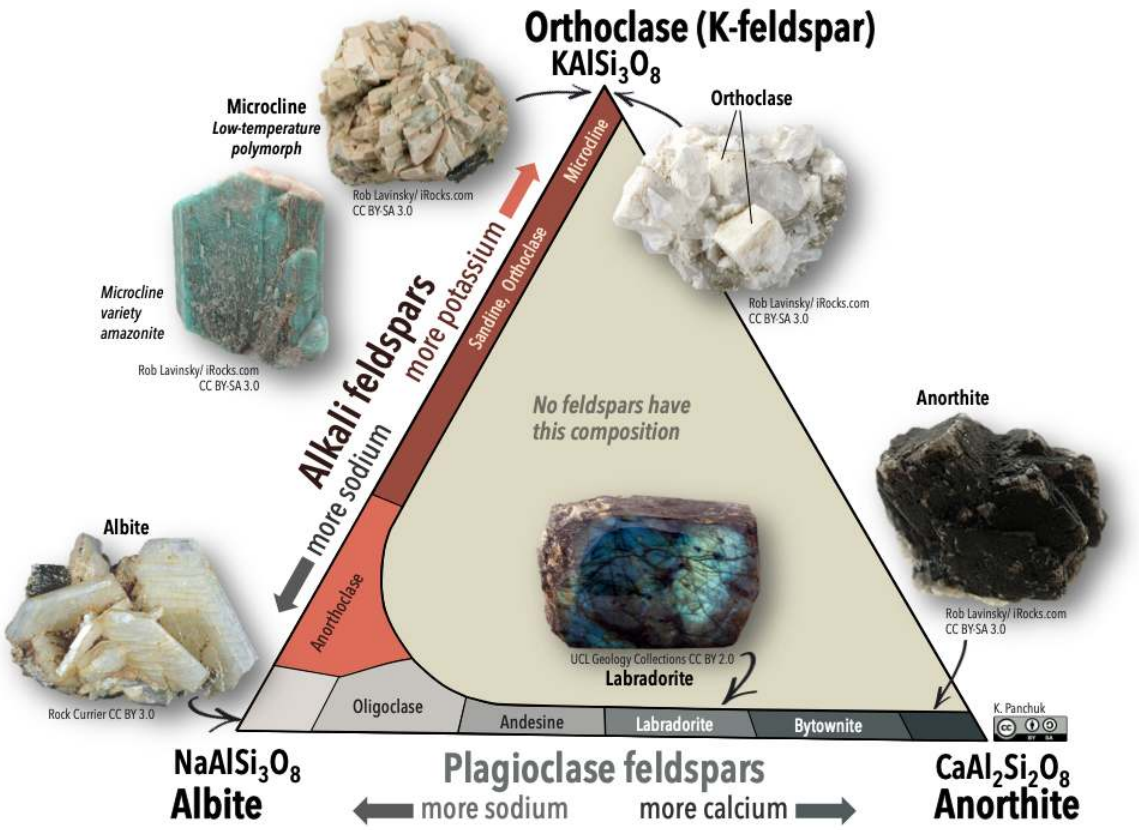


Figure Ternary diagram showing the feldspar group of framework silicate minerals. Alkali feldspars are those with compositions ranging between albite (with a Na cation) and orthoclase and its polymorphs (with a K cation. Plagioclase feldspars are those with compositions ranging between albite and anorthite (with a Ca cation). Source: Karla Panchuk (2018) CC BY-SA 4.0. Ternary diagram modified after Klein & Hurlbut (1993). Click for more attributions and a ternary diagram without mineral images.

One end-member is **potassium feldspar** (also referred to as **K-feldspar**), which has the composition $KAlSi_3O_8$. Another end member is **albite**, which has sodium instead of potassium (formula $NaAlSi_3O_8$). As is the case for iron and magnesium in olivine, there is a continuous range of compositions between albite and orthoclase. Feldspars in this series are referred to as **alkali feldspars**.

The third end-member is **anorthite,** which has calcium instead of potassium or sodium (formula $CaAl_2Si_3O_8$). Feldspars with a mixture of sodium and calcium in their composition are called **plagioclase feldspars**.

Quartz

Quartz (SiO_2; Figure below) contains only silica tetrahedra. In quartz**,** each silica tetrahedron is bonded to four other tetrahedra (with oxygen shared at every corner of each tetrahedron), making a three-dimensional framework. As a result, the ratio of silicon to oxygen is 1:2. The hardness of quartz and the fact that it breaks irregularly (notice the bottom of the crystal in the right of the figure below) and not along smooth planes result from the strong covalent/ionic bonds characteristic of the silica tetrahedron.

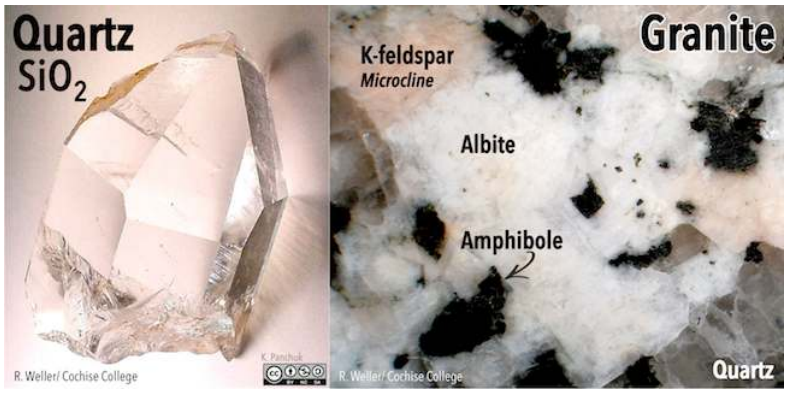


Figure Quartz is another silicate mineral with a three-dimensional framework of silica tetrahedra. Sometimes quartz occurs as well-developed crystals (left), but it also occurs in common rocks such as granite (right). In addition to quartz, the granite contains potassium feldspar, albite, and amphibole. Source: Karla Panchuk (2018) CC BY-NC-SA 4.0. Click for more attributions.

References

Klein, C. & Hurlbut, C. S., Jr. (1993). *Manual of Mineralogy (after J. D. Dana)*. New York, NY: John Wiley & Sons, Inc.

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3.5: Mineral Properties

Minerals are universal. A crystal of hematite on Mars will have the same properties as one on Earth and the same as one on a planet orbiting another star. That's good news for geology students who are planning interplanetary travel because they can use the same properties to identify minerals anywhere. That doesn't mean that it's easy, however. Identification of minerals takes practice. Some of the mineral properties that are useful for identification are color, streak, luster, hardness, habit, cleavage or fracture, and density.

Color

Some minerals have distinctive colors that are useful as diagnostic criteria. The mineral sulfur (Figure below, to the left) is always a characteristic bright yellow. For other minerals, the color might vary. Hematite is an example of a mineral for which color is not necessarily diagnostic. In some forms, hematite is a deep dull red (a fairly unique color), but in others, it is a metallic silvery black (Figure below, to the right).

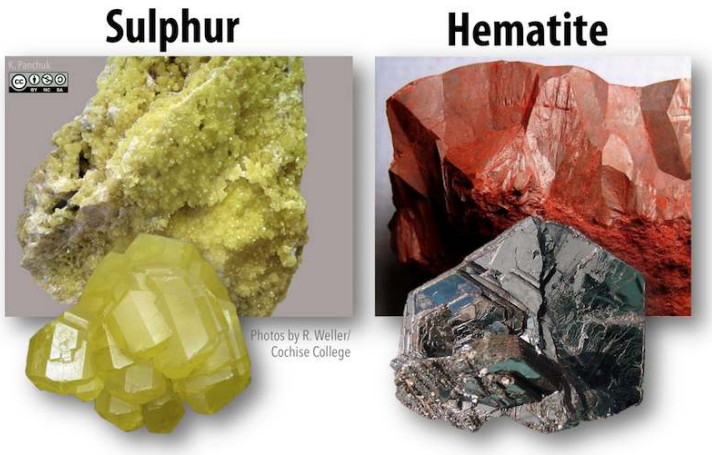


Figure Colour is a useful diagnostic property for sulfur (left) and some types of hematite (right) because the yellow and dark red colors are unique to those minerals. In contrast, silvery metallic forms of hematite appear similar to many other minerals. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Click for more attributions.

For other minerals, the problem is that a single mineral can have a wide range of colors. The color variations can result from varying proportions of trace elements within the mineral or structural defects within the crystal lattice. In the case of quartz (Figure below), milky quartz gets its white color from millions of tiny fluid-filled cavities. Smoky quartz gets its grey color from structural damage caused by natural radiation. Amethyst and citrine get their colors from trace amounts of iron, and rose quartz gets its pink hue from manganese.



Figure The many colors of quartz. Quartz can be colorless, milky, greyish smoky color, purple, yellow, and pink. Source: Karla Panchuk (2018), CC BY-NC-SA 4.0. Photos by R. Weller/ Cochise College. Click for more attributions.

Streak

The color of a mineral is what you see when light reflects off the sample's surface. One reason that color can be so variable is that the surface texture is variable. A way to get around this problem is to grind a small amount of the sample to a powder and observe the color of the powder. This color is the mineral's **streak**. The mineral can be powdered by scraping the sample across a piece of unglazed porcelain called a **streak plate.** In the figure below, two hematite samples have been scraped across the streak plate. Even though one sample is metallic and the other is deep red, both have a similar reddish-brown streak.



Figure Hematite leaves a distinctive reddish-brown streak whether the sample is metallic or deep red. Source: Karla Panchuk (2015) CC BY 4.0

Streak is an especially helpful property when minerals look similar. The next figure below shows four different minerals, but all are silvery-black in color, with varying degrees of metallic sheen. The streaks of these minerals are much more distinctive, ranging from dark grey to yellowish brown.



Figure Similar dark-grey minerals with varying degrees of metallic sheen leave different colors of streaks. The minerals are from the upper left clockwise: hematite (reddish brown streak), magnetite (grey streak), sphalerite (yellowish brown streak), and galena (darker grey streak). Source: Karla Panchuk (2015), CC BY 4.0.

Luster

Luster is how light reflects off a mineral's surface and the degree to which it penetrates the interior. The main distinction is between **metallic luster** and **non-metallic luster**. Light doesn't pass through metals, which is why they look metallic (e.g., look at hematite in the figure above). Even a thin sheet of metal—aluminum foil, for example—will not permit light to pass through it. Many non-metallic minerals may look as if the light will not pass through them, but if you look closely at a thin slice of the mineral, you'll see that it is translucent or transparent.

If a non-metallic mineral has a shiny, reflective surface, it is said to have a **glassy** luster. The quartz crystals in the previous section are examples of minerals with glassy luster. If the mineral surface is dull and non-reflective, it has an **earthy** luster (like the reddish

brown hematite in the previous section on Streak). Other types of non-metallic lusters are silky, pearly, and resinous (like amber or candied pineapple). Luster is a good diagnostic property because most minerals will always appear either metallic or non-metallic, although, as the two hematite samples show, there are exceptions.

Practice with Luster

Hardness

One of the most important diagnostic properties of a mineral is its hardness. In practical terms, hardness determines whether or not a particular material can scratch a mineral.

In 1812 German mineralogist Friedrich Mohs came up with a list of 10 minerals representing a wide range of hardness and numbered them 1 through 10 in order of increasing hardness (see graph below, on the horizontal axis). While each mineral on the list is harder than before, the measured hardness (vertical axis) is not linear. Notice the difference in hardness between talc and gypsum, then compare that to the difference between corundum and diamond.

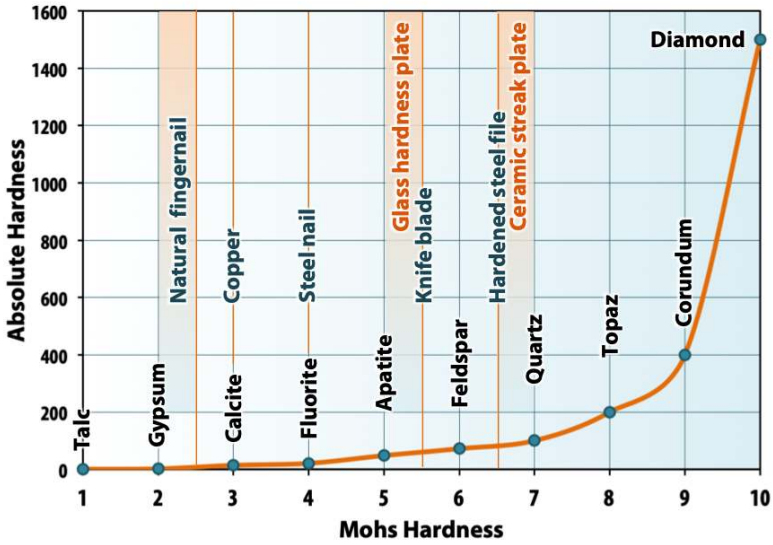


Figure Minerals and reference materials in the Mohs scale of hardness. Absolute hardness values are measured values. Mohs hardness values do not increase proportionately with absolute hardness. Source: Karla Panchuk (2021), CC BY 4.0.

Some commonly available reference materials are also shown on this diagram and summarized in the table below. Note that fingernails are often included as reference material for testing minerals, but this only applies to natural fingernails. Artificial fingernails may be much harder than natural fingernails. Some materials used for artificial nails are harder than quartz.

Table Mohs Hardness of Scale		
Mohs Hardness	**Mineral**	**Common Reference Materials**
1	Talc (softest)	
2	Gypsum	2-2.5 – Natural fingernail
3	Calcite	3 – Copper
4	Fluorite	4 – Nail (steel)
5	Apatite	5-5.5 – Glass plate, knife blade
6	Orthoclase Feldspar	6.5-7 – Streak plates, hardened steel file
7	Quartz	
8	Topaz	
9	Corundum	
10	Diamond (hardest)	

Using these materials to determine hardness involves testing to see which material will or won't scratch the mineral so you can find upper and lower limits on hardness. For example, if you have a mineral that you can't scratch with your fingernail but can scratch with a copper wire, its hardness is between 2.5 and 3. A mineral with known hardness can be used to test other minerals.

Concept Check: Testing for Hardness

Crystal Habit

When minerals form within rocks, there is a possibility that they will form in distinctive crystal shapes if other pre-existing minerals do not crowd them out. Every mineral has one or more distinctive **crystalhabits** determined by its atomic structure, although it is not that common in ordinary rocks for the shapes to be obvious.

Quartz, for instance, will form six-sided prisms with pointed ends (see the left side of the figure below), but this typically happens only when it crystallizes from a hot water solution within a cavity in an existing rock. Pyrite can form cubic crystals (center of the figure below), but can also form crystals with 12 faces, known as dodecahedra. The mineral garnet also forms many-sided crystals with an overall rounded shape (right on the figure below).



Figure Hexagonal prisms of quartz (left), intergrown cubic crystals of pyrite (center), and 24-sided crystals of garnet (right). Source: Karla Panchuk (2018) CC BY-NC-SA 4.0. Photos by R. Weller/ Cochise College. Click for more attributions.

Some of the terms that are used to describe habit include bladed, botryoidal (grape-like), dendritic (branched), drusy (an encrustation of crystals), and equant (similar size in all dimensions), fibrous, platy, prismatic (long and thin), and stubby.

Cleavage and Fracture

Cleavage and fracture describe how a mineral breaks. These characteristics are the most important diagnostic features of many minerals and often the most difficult to understand and identify. **We see cleavage** when a mineral breaks along a plane or planes,

while the **fracture** is an irregular break. Some minerals tend to cleave along planes at various fixed orientations. Some, like quartz, do not cleave at all, only fracture. Minerals with cleavage can also fracture along surfaces that are not parallel to their cleavage planes.

The way minerals break is determined by the atoms' arrangement and, more specifically, by the orientation of weaknesses within their crystal lattice. Graphite and mica break off in parallel sheets (Figure below).

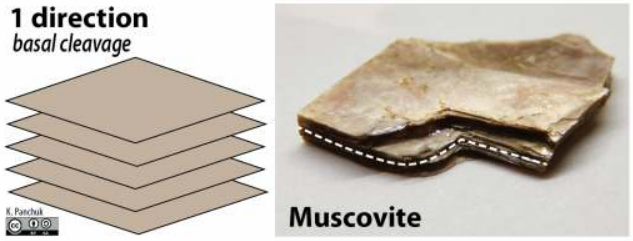


Figure One direction of cleavage (basal cleavage). Left: Schematic of basal cleavage. Right: Muscovite showing basal cleavage. The white dashed line marks the edge of the cleavage plane. *Source: Karla Panchuk (2018) CC BY-SA 4.0. Cleavage diagram modified after M.C. Rygel (2010) CC BY-SA 3.0* view source

Other minerals have two directions of cleavage, classified as two directions at 90° and two directions not at 90°. While the diagrams of planes on the left of Figure below make this difference clear, it may be less obvious in practice. The minerals in Figure both have two planes of cleavage that are very close to 90°. The white dashed lines mark the edges of the planes, as in Figure shown above. See if you can find the planes repeated in the images. The images are close-up views of the minerals, only a few cm across. Sometimes you must look very carefully to find cleavage planes.

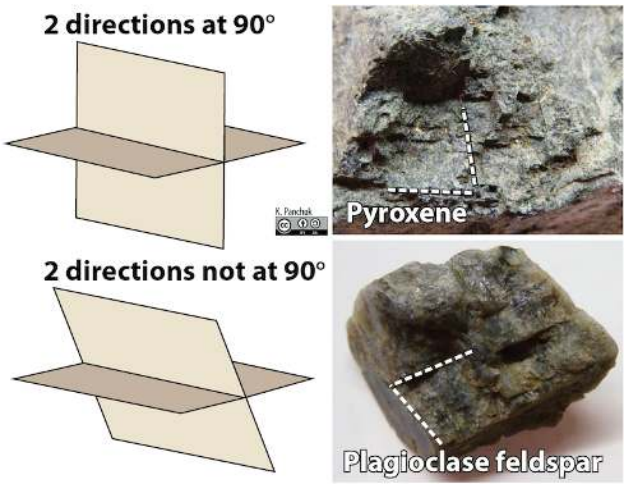


Figure Two directions of cleavage. Top: Two directions at 90° in pyroxene. Bottom: two directions not at 90° in plagioclase feldspar. Edges of cleavage planes are marked with dashed lines. *Source: Karla Panchuk (2018) CC BY-SA 4.0. Cleavage diagrams modified after M.C. Rygel (2010) CC BY-SA 3.0* view source

Some minerals have many directions of cleavage. Figure 5.38 shows minerals with three directions of cleavage. Halite (shown at the top of the figure below) has three directions at 90°, and calcite (as shown by the bottom half of the figure below) has three directions, not at 90°.

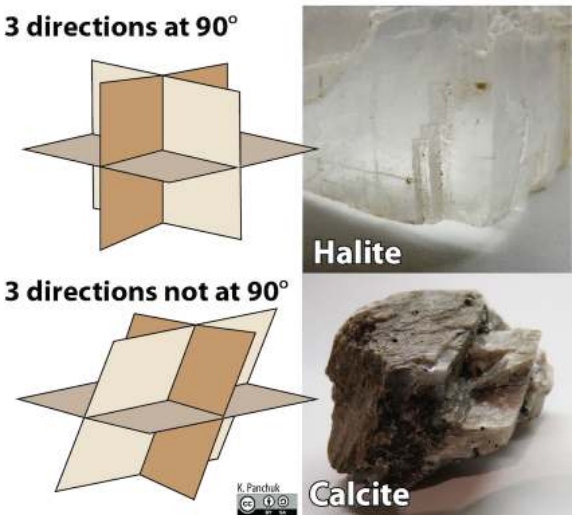


Figure Three directions of cleavage. Top: Three directions at 90° in halite. Bottom: Three directions not at 90° in calcite. *Source: Karla Panchuk (2018) CC BY-SA 4.0. Cleavage diagrams modified after M.C. Rygel (2010) CC BY-SA 3.0* view source

Students encounter a few common difficulties when learning to recognize and describe cleavage. One is that it might be necessary to look very closely at a sample to see mineral cleavage. The key features of cleavage are only cm or mm in scale. If crystals are very small, it may not be possible to see cleavage at all. Another issue is that sometimes cleavage is present, but it is poor, meaning the cleavage surface isn't perfectly flat. Finally, it can be difficult to know whether a flat surface on a crystal is a cleavage plane, a crystal face, or simply a surface that happens to be flat.

Cleavage planes tend to repeat themselves at different depths throughout the mineral, so if you are unsure whether the surface you are looking at is a cleavage plane, try rotating the mineral in bright light. If cleavage is present, you will generally find that all of the cleavage surfaces will glint in the light for a given cleavage direction. Crystal faces will also glint in the light, but they do not repeat themselves at depth throughout the mineral. The best way to overcome these problems is to look at many examples. It's worth it to be able to identify cleavage and fracture because cleavage is a reliable diagnostic property for most minerals.

Can You Pick Out the Broken Piece?

IMPORTANT: To Receive points for this activity, make sure you advance the slides to the end!

Density

Density is a measure of the mass of a mineral per unit volume, and it is a useful diagnostic tool in some cases. Most common minerals, such as quartz, feldspar, calcite, amphibole, and mica, are of average density (2.6 to 3.0 g/cm^3), and it would be difficult to tell them apart based on their density. On the other hand, many of metallic minerals, such as pyrite, hematite, and magnetite, have densities over 5 g/cm^3. If you picked up a sample of one of these minerals, it would feel much heavier compared to a similarly sized sample of a mineral with average density. A limitation of using density as a diagnostic tool is that one cannot assess it in minerals that are a small part of rock with other minerals in it.

Other properties

Several other properties are useful for the identification of some minerals. Some of these are:

- Calcite reacts with dilute acid and will give off bubbles of carbon dioxide.
- Magnetite is strongly magnetic, and some other minerals are weakly magnetic.
- Sphalerite ((Zn,Fe)S) gives off a smell of sulfur when drawn across a streak plate.
- Halite tastes salty.
- Talc feels soapy to the touch.
- Plagioclase feldspar has striations (parallel razor-thin lines etched on the surface) and some varieties show a play of colors when light hits them at the right angle (Google the mineral labradorite!)

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3.6: Summary Key Term Check

Chapter Main Ideas

Atoms

An atom comprises protons and neutrons in the nucleus and electrons arranged in energy shells around the nucleus. The first shell holds two electrons, and the outer shells hold more. Atoms strive to have eight electrons in their outermost shell (or two for H and He). Atoms gain, lose or share electrons to achieve this. In so doing, they become either positively charged cations (if they lose electrons) or negatively charged anions (if they gain them).

Bonding and Lattices

The main types of bonding in minerals are ionic bonding (electrons transferred) and covalent bonding (electrons shared). Some minerals have metallic bonding or weak Van der Waals forces. Minerals form in three-dimensional lattices. The configuration of the lattices and the type of bonding within help determine mineral properties.

Mineral Groups

Minerals are grouped according to the anion part of their formula. Some common types are oxides, sulfides, sulfates, halides, carbonates, phosphates, silicates, and native minerals.

Silicate Minerals

Silicate minerals are the most common minerals in Earth's crust and mantle. They all have silica tetrahedra (four oxygens surrounding a single silicon atom) arranged in different structures (chains, sheets, etc.).

Do You Know the Common Silicate Minerals?

The most common rock-building minerals come up frequently in this textbook. Now is a good time to make sure you know what they are.

Mineral Properties

Some important mineral identification properties include hardness, cleavage/fracture, density, luster, color, and streak color.

Key Term Check

What key term from Chapter 5 is each card describing? Turn the card to check your answer.

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CHAPTER OVERVIEW

4: The Rock Cycle

4.1: What Is A Rock?

4.2: The Rock Cycle

4.3: Igneous Rocks

4.4: Sedimentary Rocks

4.5: Metamorphic Rocks

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4.1: What Is A Rock?

A **rock** is a solid mass of geological materials. Geological materials include individual mineral crystals, inorganic non-mineral solids like glass, pieces broken from other rocks, and even fossils. The geological materials in rocks may be inorganic, but they can also include organic materials such as the partially decomposed plant matter preserved in coal. A rock can be composed of only one type of geological material or mineral, but many are composed of several types (Figure below).

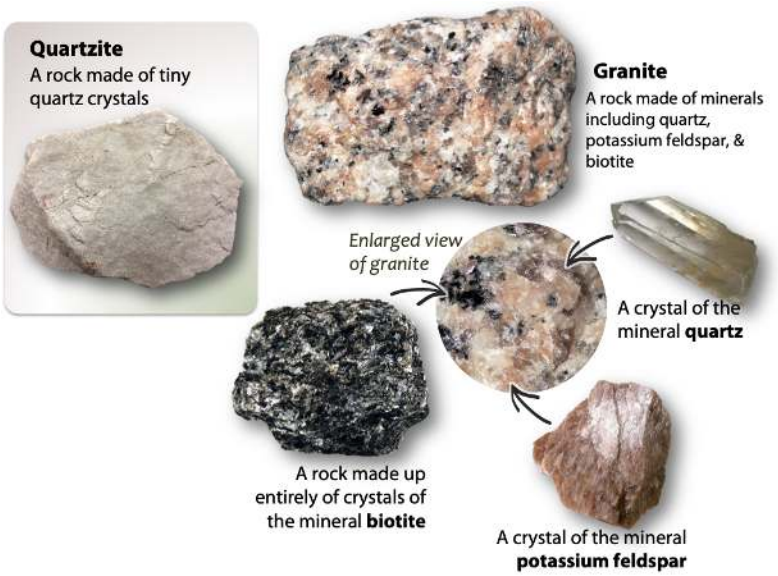


Figure Rocks versus minerals. Rocks in the image are made up of crystals of one or more minerals. Source: Karla Panchuk (2021), CC BY-NC-SA. Photographs by R. Weller/ Cochise College and James St. John. Click for attributions.

Concept Check: Rock or Mineral?

Three Main Types of Rock

Rocks are grouped into three main categories based on how they form. **Igneous rocks** form when melted rock cools and solidifies. **Sedimentary rocks** form when fragments of other rocks are buried, compressed, and cemented together; or when minerals precipitate from solution, either directly or with the help of an organism. **Metamorphic rocks** form when heat and pressure alter a pre-existing rock. Although temperatures can be very high, metamorphism does not involve the melting of the rock.

Do You Know Your Rock Types?

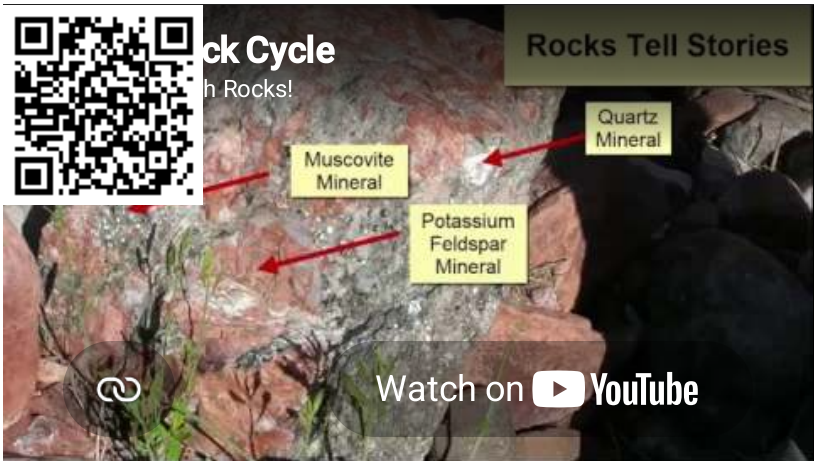
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4.2: The Rock Cycle

The Rock Cycle: Recycle and Reuse

Earth is an efficient recycler of its solid materials through the processes of plate tectonics, in which the rigid oceanic lithosphere will eventually descend into the asthenosphere (mantle), melt, and form again at spreading centers. Another way in which the Earth can rework and recycle its crust is through the Rock Cycle.

Video The rock cycle (3:22).



The **Rock Cycle** describes how the three main rock types that compose the Earth's lithosphere are constantly transformed into one another through geologic processes. These three rock types are those we learned about in the previous chapter: igneous, sedimentary, and metamorphic rocks. For now, let's focus on the process that creates these different rock types.

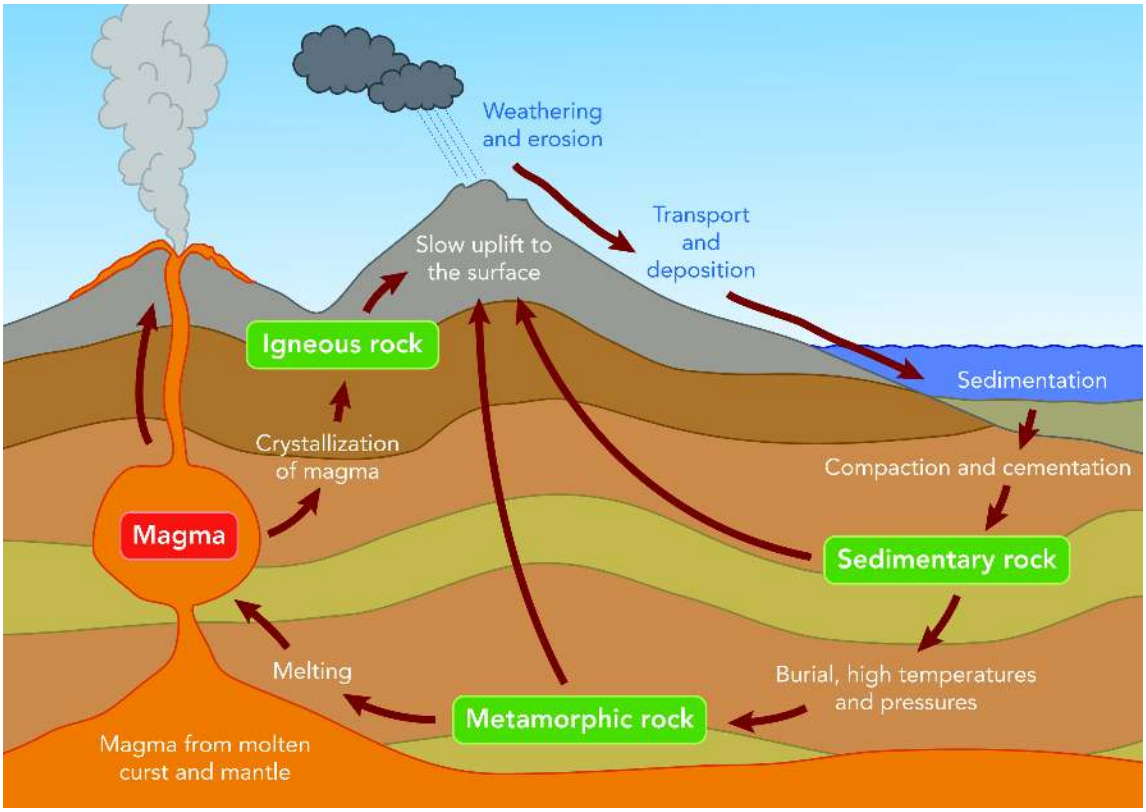


Figure "Rock Cycle" by Siyavula Education is licensed under Creative Commons Attribution 2.0 Generic.

Igneous rocks: melting, cooling, and crystallizing

Extremely hot rock from the mantle that has melted is called molten rock or **magma**. It is called **lava** when magma rises to the crust and erupts from volcanoes at the earth's surface. Eventually, these molten materials cool and solidify and might grow minerals through **crystallization**. The solid products from magma, lava, or volcanic activity are called **igneous rocks.**



Figure. "Sarychev Peak Eruption, Kuril Islands" by NASA's Earth Observatory is licensed under Public Domain.

Igneous rocks are not only formed from mantle material. Sometimes a sedimentary rock can be buried deep within the crust and come in contact with magma, which causes the molten rock's composition to change. The same can happen with metamorphic rocks or even preexisting igneous rocks. When ANY type of rock melts, it becomes molten, and it thus has the potential to become an igneous rock.

Sedimentary rocks: Weathering, deposition, and lithification

When any rock is exposed to the Earth's surface, it will undergo weathering and erosion, which produces sediment. **Weathering** is the physical and chemical breakdown of rocks into smaller fragments by the atmosphere, hydrosphere, or biosphere. **Erosion** is the removal of those fragments from their original location. Streams, glaciers, or wind will transport the sediment from the original rock, but it will ultimately accumulate on the earth's surface due to gravity. This accumulation is called **deposition.**



Figure. Sedimentary layers of cross-bedded sandstone at Zion National Park, Utah.

The deposited sediments will eventually build in layers and will become buried on the Earth's surface. As the sediments reach deeper, they eventually become a solid rock through a process called **lithification,** which requires both compaction and cementation of the loose solids. The weight of the overlying layers will **compact** the sediment closer together, and as groundwater leaks between the individual grains, it will glue or **cement** the sediment as solid rock.

In other circumstances, weathering strips rocks of their very elements at an atomic level; these elements later precipitate as new solids in oceans or lakes. Both lithification and precipitation will produce **Sedimentary Rocks.**

Metamorphic rocks: Burial, deformation, and exhumation

When any type of rock – igneous, sedimentary, or metamorphic – is buried within the earth's lithosphere, it will reach zones of higher temperatures and/or pressures at depth. The alteration of rock by heat and pressure is called **metamorphism.**



Figure. Folded metamorphic rock gneiss found in Soldier Canyon, Arizona

Rocks will change during metamorphism because the minerals that compose them are only stable under a specific range of temperatures and pressures. Therefore, greater heat and pressure will cause new minerals to grow, and in some rocks, the pressure will squeeze and stretch minerals in patterns of lines or waves called foliation. These transformed and deformed products of heat and pressure are called **metamorphic rocks.**

Metamorphic rocks do not form at the Earth's surface as the heat and/or pressures required for metamorphism are found kilometers deep within the lithosphere. But we see metamorphic rock formations on the surface of the Earth. Why is this? Sometimes the deeply buried layers of metamorphic rock are forced toward the light of day by mountain building processes or the sudden weathering and erosion of overlying rocks. This process is called **exhumation**, which is why we can see a variety of rocks from different periods in Earth's history!

Written in stone

The Rock Cycle is truly a cycle. There is no single point at which it "begins" or "ends," and it has been operating for billions of years. There is a natural tendency to think that the rocks on Earth's surface progress as igneous -> sedimentary -> metamorphic -> igneous, but that is not the case. ***Any type of rock on Earth's surface has the potential to become any other type of rock through geologic processes!***

For example, a sedimentary rock might be buried deep within the Earth's surface and melt in contact with a magma plume. The magma and molten sedimentary rock mix and produce a new igneous rock. That igneous rock is buried deep in the crust and becomes warped and deformed; it transforms into a metamorphic rock. Finally, after hundreds of millions of years, the metamorphic rock formation is exhumed and weathered away. The fragments of that metamorphic rock lithify to form a brand new sedimentary rock. What a cycle!

One last thing! The processes involved in the rock cycle, and the rocks themselves, tell a story of the events that happened in Earth's 4.54 billion-year history. While even the best geologic cannot reconstruct every page of Earth's story from a single rock formation, they can get a glimpse of what might have happened in a region to form a certain type of rock.

Igneous rock can tell us a story of magma chambers or volcanic activity. Sedimentary rocks tell us where rivers, deserts, beaches, and oceans once resided, and metamorphic rocks help us reconstruct the times when tectonic plates collided or spread apart from one another.

> *In his poem, Auguries of Innocence, Walt Whitman once wrote:* ***"To see a World in a Grain of Sand"*** *What will you see in an entire rock?*

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4.3: Igneous Rocks

Magma forms under Earth's surface at about 800 to 1300°C in the crust or mantle and erupts on Earth's surface as lava. When magma or lava cools, it solidifies by crystallization in which minerals grow within the magma or lava. The rock that results from this is an **igneous rock** from the Latin word *ignis*, meaning "fire." [2] Igneous rocks are traditionally defined as the solid products from the cooling and hardening of molten magma in many different environments. We identify these rocks by their composition and texture.

Igneous Rock Composition

Composition refers to a rock's chemical and mineralogical make-up. For igneous rock, the composition is generally divided into four groups: ***ultramafic, mafic, intermediate, and felsic***. These groups refer to differing amounts of *silica (SiO_2), iron (Fe), and magnesium (Mg)* found in the minerals that make up the rocks.

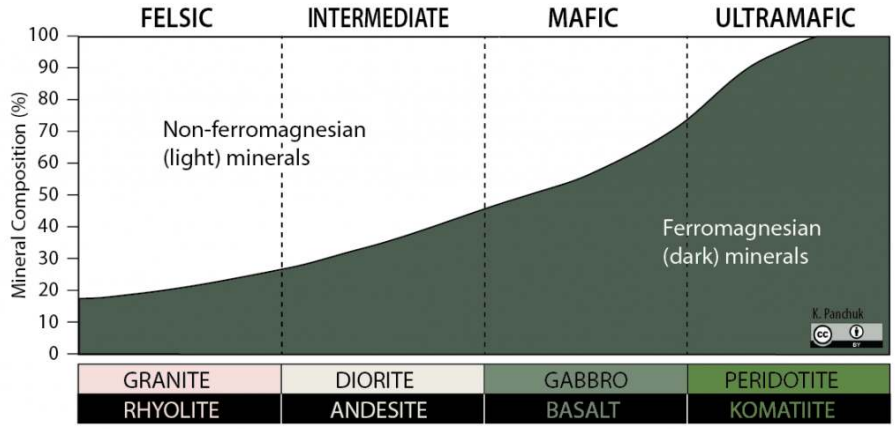


Figure. Classification of igneous rocks by composition. Darker rocks are ultramafic or mafic, whereas lighter rocks are intermediate/felsic. Y-axis refers to the abundance of Fe and Mg-containing minerals in each rock type.

Ultramafic refers to rocks composed of mostly olivine and some pyroxene. These rocks have even more magnesium and iron and even less silica than 'ordinary' mafic rocks. Ultramafic rocks are rare on the surface, but they make up the primary components of the upper mantle. Ultramafic rocks are very poor in silica, in the 40% or less range (this means that the rock would be less than 40 weight percent silica).

Mafic refers to igneous rocks with abundant ferromagnesian minerals (those with the elements Mg and Fe in their chemical formulae) plus plagioclase feldspar. Such minerals are dark-colored and include pyroxene and olivine. Mafic rocks are low in silica (in the 45-50% range), but they make up most of the oceanic crust and lithosphere.

Intermediate describes the igneous rock composition between mafic and felsic. It contains roughly equal amounts of light and dark minerals, including light grains of plagioclase feldspar and dark grains of amphibole It is intermediate in silica (in the 55-60% range).

Felsic refers to a predominance of light-colored minerals, including **Fel**dspar and **si**li**c**a quartz. These minerals have more silica as a proportion of their overall chemical formulae. Minor amounts of dark-colored minerals, such as biotite mica, may sometimes be present. Felsic igneous rocks are rich in silica (in the 65-75% range) and tend to represent the continental crust or lithosphere composition.



Figure. Phaneritic/Coarse-grained texture, where the naked eye easily spots individual crystals and minerals within the rock.

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



Figure. Aphanitic/Fine-grained texture, where the naked eye within the rock cannot spot individual crystals and minerals.

When lava erupts onto the surface or rises into shallow in a mountain and cools, the rock that will cool from it is called an **extrusive** igneous rock. Extrusive igneous rocks have a fine-grained or “aphanitic” texture, in which the grains are too small to see with the unaided eye. This fine-grained texture tells us that the quickly-cooling lava did not have time to grow large crystals.



Figure Porphyritic texture illustrated by large, visible crystals found within a fine-grained matrix in igneous rock.

Some igneous rocks have a mixture of large crystals within a fine-grained matrix. Such a texture is called **porphyritic**. A porphyritic texture tells us that the magma underwent multiple cooling stages; it first cooled slowly when it was deep under the surface, then rose to a shallow depth where it cooled quickly.



Figure. Vesicular texture, in which the rock contains many pores where gases escaped during formation.

All magmas contain dissolved gases called volatiles. When magma quickly rises to the surface as lava, these volatiles sometimes become trapped in the cooling molten rock and form a bubbling texture that appears sponge-like. Such a texture is called **vesicular** because the holes in the rock are called vesicles by scientists.



Figure. Glassy texture, which is very smooth without the presence of crystals or minerals

Lava will sometimes cool so quickly that not even microscopic crystals will form in it. As a result, volcanic glass will form with a shiny, smooth appearance that reflects light. This texture is called **glassy**. Just like the mineral quartz, a glassy rock will have a conchoidal fracture with distinctive, rounded fracture edges. This is because, like quartz, most glassy rocks are made of the compound SiO_2 in the form of the mineraloid amorphous silica.

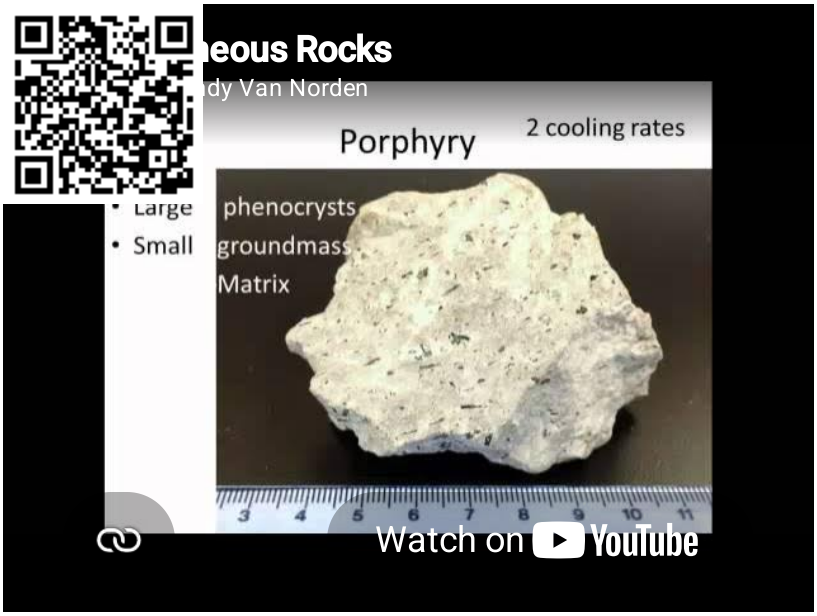


Figure Pyroclastic texture shown by angular rock fragments mixed with fine ash and volcanic glass in igneous rock.

The final volcanic texture is a result of explosive, violent eruptions. These eruptions produce lava and clouds of ash, rock, gases, and glass. The eruption's solid material, called tephra, eventually falls back onto the earth and consolidates into a solid mass. The rock that will form from this process has a **pyroclastic** texture. This texture consists of volcanic ash, glass shards, and small rock fragments.

Igneous Rock Field Guide

Video. Classifying igneous rocks by texture and composition explained (7:37).



Komatiite



Figure. Komatiite, South Africa. Click on this image to see a 3D interactive model by Sara Carena CC BY-NC.

"KO-MAT-EE-ITE"

Most commonly confused with: basalt

An ultramafic, fine-grained extrusive igneous rock. This rock will form from rapidly cooling lava, but it is very rare. Molten ultramafic rock was more prevalent on early Earth, and it can be found in the mantle. Komatiite is composed primarily of olivine and pyroxene minerals, which causes the rock to take on a dark, greenish color.

Chances are that if you are holding a fine-grained, dark igneous rock, it will be basalt since komatiites are not very common but look twice if it has a strong green tint. Olivine and pyroxene are more susceptible to weathering than minerals found in felsic rocks; therefore, this igneous rock may erode easier than a felsic counterpart.

Peridotite



Figure. Peridotite. Click on this image to go to a 3D interactive model by Sara Carena CC BY-NC.

"PUR-ID-DOH-TITE"

Most commonly confused with: olivine (mineral), gabbro

An ultramafic, coarse-grained (intrusive) igneous rock. Peridotite will form under Earth's surface from slowly-cooling magma. It is less rare than Komatiite, but still not very common. Ultramafic magma composes the Earth's upper and lower mantle; therefore, when a plume of magma rises to the lithosphere and cools as a "Xenolith," peridotite will form.

Peridotite is composed of visible, and sometimes large, crystals of olivine and pyroxene. It is often dark and with distinctively green crystals. When distinguishing this rock from gabbro, consider the percentage of olivine in the rock; peridotite has at least more than 25%. As with komatiite, peridotite will also erode more easily than other igneous rocks because it contains minerals that are more susceptible to weathering.

Basalt



Figure. Basalt. Click on this image to go to a 3D interactive model by Sara Carena CC BY-NC.

"BAH-SALT"

Most commonly confused with: komatiite, shale (sedimentary), limestone (sedimentary)

A mafic, fine-grained (extrusive igneous rock. About 90% of all volcanic rocks that form are on the Earth's surface are basalt as this rock represents a common magma composition of the upper mantle mixed with the crust. Basalt is the final product when this magma erupts as lava and cools.

Basalts are characterized by low (~50%) silica content and minerals with iron (Fe) and magnesium (Mg). Such minerals typically include amphibole and pyroxene, and sometimes small amounts of olivine. Additionally, basalts include a significant amount of calcium-plagioclase feldspar in their matrix. Basalt often is dark gray, and unlike komatiite, does not have a green tint. When distinguishing this rock from shale and limestone, use a hand lens to identify the notable minerals in its matrix. Additionally, basalt is more resistant to scratching than shale and limestone; try scratching it with a penny!

Gabbro



Figure. Gabbro. **Click on this image** to see a 3D interactive model by Sara Carena CC BY-NC.

"GAB-BRO"

Most commonly confused with: peridotite

A mafic, coarse-grained (intrusive) igneous rock. Gabbro makes up the majority of oceanic lithosphere. It often forms at divergent boundaries when magma with a composition similar to the upper mantle rises and slowly cools beneath the surface. This rock has the same general composition as basalt, but its minerals are easily visible to the naked eye due to its slow cooling history.

Gabbro is dark gray or black, sometimes with noticeable flecks of white calcium plagioclase or green olivine. Gabbro can be distinguished from peridotite by its limited olivine content. If the number of olivine crystals in an unknown, coarse-grained rock is small or nonexistent, then you are looking at gabbro.

Like basalt, gabbro will erode at Earth's surface faster than other felsic, intrusive rocks. The effects of weathering by the atmosphere may be more noticeable on gabbro because it contains larger crystals of iron-rich minerals.

Andesite



Figure. Andesite. **Click on this image** to go to a 3D interactive model by Sara Carena CC BY-NC.

"AN-DEH-SITE"

Most commonly confused with: rhyolite

An intermediate, extrusive igneous rock. Andesite cools from lavas that are between mafic and felsic in composition. As such, andesite typically has a silica content of 55-60%. Andesite can often be found near volcanoes along the Pacific Ring of Fire or at stratovolcanoes, which are sometimes called Andesite volcanoes. Andesitic lavas are typical of subduction zones, such Andes Mountains subduction zone for which it received its name.

Andesite is commonly light gray. Unlike other extrusive igneous rocks that cool from lava, andesite is also porphyritic, which means that medium and small dark crystals can be seen in its fine-grained matrix by the naked eye. These crystals are usually amphibole and pyroxene minerals, whereas the light matrix is mostly composed of calcium plagioclase feldspar. The porphyritic texture of andesite clearly distinguishes it from rhyolite, which is sometimes a light beige-grey color.

Diorite



Figure. Diorite. **Click on this image** to go to a 3D interactive model by Sara Carena CC BY-NC.

"DYE-O-RITE"

Most commonly confused with: granite

An intermediate, coarse-grained (intrusive) igneous rock. Diorite cools slowly from molten rock that forms beneath the Earth's surface along subduction zones, such as the Andes Mountains and other convergent margins at the Ring of Fire.

Diorite has a "cookies and cream" or "Dalmation"-like appearance, which is caused by black amphibole and white plagioclase crystals that are easily identifiable by the naked eye. Although this rock is much lighter in color than gabbro, it should be easily distinguished from granite due to the abundance of dark crystals in its matrix.

Rhyolite



Figure. Rhyolite. Click on this image to go to a 3D interactive model by Dr. Parvinder Sethi CC BY

"RYE-O-LITE"

Most commonly confused with: andesite, tuff

A felsic, fine-grained (extrusive) igneous rock. Rhyolite rapidly cools from a high-silica (65-75%) lava. Rhyolite is not as common as its coarse-grained counterpart, granite, because felsic lavas cannot move very far once they erupt.

Rhyolite is typically light tan to pinkish tan in color, and individual crystals are usually difficult to see with the naked eye. Therefore, this rock can be distinguished from andesite, which often has medium and small crystals within its matrix. There is also no pyroclastic debris within rhyolite, which can be confirmed by examining the rock under light; unlike tuff, rhyolite does not have visible flecks of volcanic glass in its matrix. Rhyolite can be found at explosive volcanoes, such as at Yellowstone National Park.

Granite

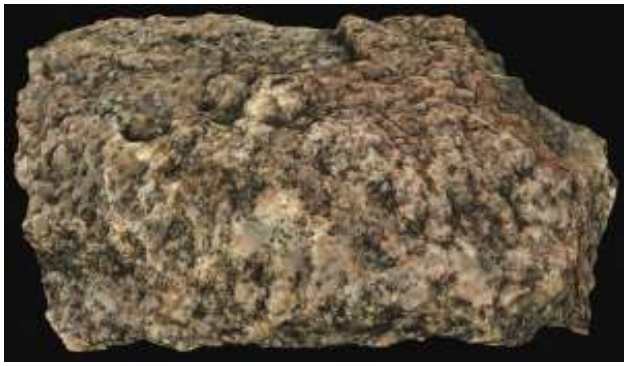


Figure. Granite. Click on this image to go to a 3D interactive model by Sara Carena CC BY-NC.

"GRAN-IT"

Most commonly confused with: diorite

A felsic, coarse-grained (intrusive) igneous rock. Granite is often used to approximate the composition of the continental crust in both composition and density. This estimation is typical because granite usually forms within the cores of mountains and thick lithosphere when magma with high silica content slowly cools and crystallizes.

Granite can be identified by its light to pinkish color and the presence of abundant quartz. Granite commonly has large amounts of salmon-pink potassium feldspar and white sodium-feldspar (plagioclase), which can even have visible cleavage planes on the crystals. Some varieties of granite have black flecks in the matrix, which are typically biotite mica. To distinguish granite from the intermediate coarse-grained rock, diorite carefully examines the minerals in its matrix. Granite, unlike diorite, has few dark minerals and a tendency to have more pink K-Feldspar.



Figure. Obsidian. **Click on this image** to go to a 3D interactive model by jonathan.davidson CC BY.

"OB-CID-DEE-AN"

Most commonly confused with: chert (sedimentary)

A felsic, glassy (extrusive) igneous rock. Obsidian is commonly found along cooled lava fields with rhyolite. Obsidian forms when lava cools so quickly that it does not have time to even develop microscopic crystals. When cooled, obsidian is very smooth, brittle, and shiny. Although obsidian is felsic and has a high silica content, is often black, brown, or red.

Obsidian is usually identifiable by its glassy texture alone. Like the mineral quartz, it displays conchoidal fracture, which can appear as curved or rounded fracture edges. The sedimentary rock chert, which is mostly composed of silica, also has conchoidal fracture; however, this rock is often much duller in both luster and color.

Because obsidian is both brittle and hard, it has been traditionally used as tools and weapons by indigenous people for thousands of years. Many of these weapons hold an edge after centuries, which demonstrates this rock's resistance to weathering.

BACKYARD GEOLOGY: APACHE TEARS



Figure An obsidian Apache Tear.

Volcanic deposits are common throughout the western United States and Mexico, and these rocks have played an important role Native American culture and history. Obsidian can be reshaped and carved to form surgical instruments or weapons such as knives. There are also naturally occurring varieties of rounded obsidian that are said to absorb grief and depression. Today these stones are called "Apache Tears". There are variations in the folklore behind this stone, the tragic theme of loss unites each story.

The most popular legend behind the Apache Tears focuses upon a Pinal Apache Tribe that was vastly outnumbered by the U.S. military during the nation's westward expansion in the 1800s. The majority of the warriors were killed in a surprise raid near Picacho Peak, AZ. The remaining Native Americans preferred suicide over imprisonment or execution.

The family of the warriors gathered not far from the peaks, where the remains of the brave warriors remained at the base. The women grieved deeply, day and night, for their lost men. The Great Father placed their tears in the obsidian. Should a person find

one of these stones, they would not need to grieve since the women of the Pinal Apache Tribe have already given their tears inside the stone [3].

Scoria



Figure. Scoria. **Click on this image** to go to a 3D interactive model by Sara Carena CC BY-NC.

"SCOR-EE-AH"

Most commonly confused with: pumice

A mafic, vesicular (extrusive) igneous rock. Scoria forms when lava containing volatile gases erupts. The gases trapped within the lava form large cavities, or vesicles, as it rapidly cools at the surface. Scoria cools from mafic lava, and contains microscopic crystals of silica-poor, ferromagnesian minerals such as amphibole, pyroxene, and calcium plagioclase.

Scoria is usually dark gray and is easily distinguished from other mafic rocks by its vesicular texture. Although pumice is also vesicular, scoria is much darker and denser. Despite having large cavities, scoria will never float above water.

Some varieties of scoria are rust-red, which reflects the tendency of the minerals within the mafic rock to oxidize, or rust, under Earth's atmosphere. Furthermore, the abundant vesicles make a greater surface area available for weathering, which causes this rock to weather faster than other mafic rocks.

Pumice



Figure. Pumice. **Click on this image** to see a 3D interactive model by Sara Carena CC BY-NC.

"PUHM-IS"

Most commonly confused with: tuff, scoria

A felsic, vesicular (extrusive) igneous rock. Pumice forms when felsic lava, which can contain significantly higher amounts of volatile gases, very rapidly cools. As with scoria, the cooling rock traps many vesicles of gas, and it takes on a "frothy" appearance as it solidifies. Pumice often cools too quickly to form minerals, and it is sometimes referred to as volcanic glass. We can find this rock along the slopes of explosive volcanoes with pyroclastic deposits.

Pumice is usually a light tan, pink, or gray. It is a very low density, and most varieties contain enough vesicles that when placed in a bowl or cup of water, the rock will float. This unique characteristic will often distinguish pumice from similar-looking rocks such as tuff or scoria.

Video. How to tell if a rock is a pumice? (0:32).



A YouTube element has been excluded from this version of the text. You can view it online here: https://pb.libretexts.org/historicalgeology/?p=122

Tuff



Figure. Rhyolite. **Click on this image** to go to a 3D interactive model by rocksandminerals CC BY.

"TOUGH"

Most commonly confused with: pumice, rhyolite

A felsic to intermediate, pyroclastic igneous rock. Tuff forms from the solid debris or tephra of an explosive volcanic eruption. When ash, volcanic glass, and rock fragments are ejected by the volcano or related pyroclastic flow eventually accumulate together, they form tuff. This type of rock usually indicates a violent eruption, and it is commonly found along the Ring of Fire and stratovolcanoes.

Tuff often does not have any single distinguishing mineralogical composition, although it often contains angular rock fragments and shiny flecks of glass within a fine-grained matrix of ash. Tuff is typically pale tan or gray. Its pyroclastic texture sometimes makes it less dense than other volcanic rocks; however, unlike pumice, it does not float in water.

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4.4: Sedimentary Rocks

When geologists encounter a sedimentary outcrop, they can reconstruct ancient landscapes – rapidly flowing mountain streams, frightening avalanches, or vast deserts – from these rocks. As the name suggests, **Sedimentary Rocks** are composed of sediments that have been cemented and compacted together, or lithified, over long periods.

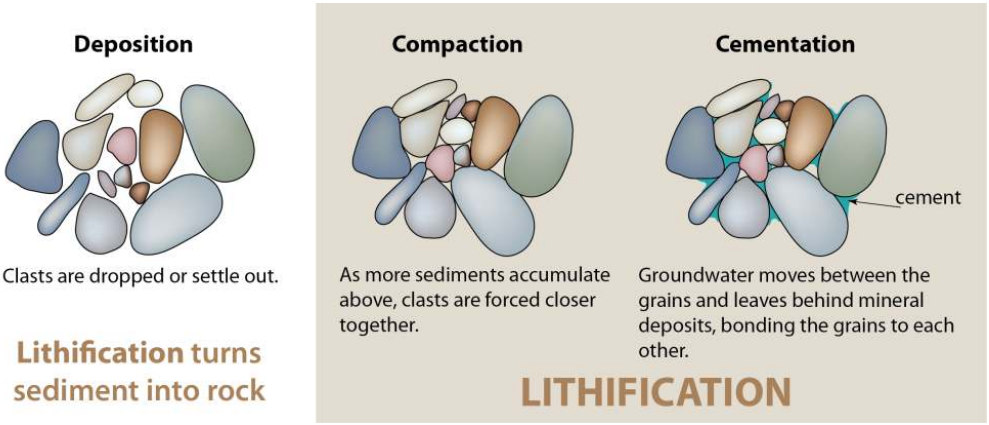


Figure. The process of lithification to form sedimentary rocks: Deposition is the settling of clasts, compaction is the movement of clasts closer together, and cementation is the bonding of the clasts together.

Sediments may include:

- Fragments of other rocks have often been worn down into small pieces, such as sand, silt, or clay.
- Organic materials, or the remains of once-living organisms.
- Chemical precipitates, which are materials that get left behind after the water evaporates from a solution.

Both physical and chemical weathering processes that take place at the Earth's surface are responsible for accumulating this source of sediment from preexisting igneous rocks, metamorphic rocks, and even other sedimentary rocks. **Physical weathering** breaks the rocks apart, while **chemical weathering** dissolves the less stable minerals. These original elements of the minerals end up in solution, and new minerals may form. Water, wind, ice, or gravity remove and transport sediments in a process called **erosion**.



Figure. Sandstone cliff formation near a beach in San Diego shows evidence of physical weathering.

We can classify sedimentary rocks in two broad categories: **clastic** and **chemical**. **Clastic sedimentary rocks** are made from fragments of eroded bedrock and sediment, which is usually derived from physical weathering. We classify clastic rocks by their grain size, shape, and sorting. **Chemical Sedimentary rocks** are precipitated from water that contains a very high concentration of dissolved elements, or ions. We usually classify these chemical rocks based on the precipitated minerals or whether biological processes play a role in the rock's formation.

Clastic Sedimentary Rocks

Clastic sedimentary rocks consist of preexisting rock fragments, or clasts, that come from weathered bedrock. The majority of this sediment has been physically weathered, but some of it may also be chemically weathered. We classify this type of sedimentary rock with three main identifiers: grain size, shape, and sorting.

Grain size

The size of the individual clasts making up a sedimentary rock strongly indicates the amount of energy a landscape had available for transporting the sediment. For example, extremely large fragments, such as boulders, might only be deposited by high energy processes, such as avalanches, whereas wind environments may be able to deposit smaller grains such as silt, mud and clay. Below is a diagram that gives an approximate size range for the different types of clasts that geologists encounter in clastic sedimentary rocks.

	Clast name	Diameter Range
Coarse-grained	Boulder	Larger than 256 mm
	Cobble	64 mm - 256 mm
	Pebble	2 mm - 64 mm
Medium-grained	Sand	63 µm - 2 mm
	coarse	500 µm - 2 mm
	medium	250 µm - 500 µm
	fine	63 µm - 250 µm
Fine-grained	Silt	2 µm - 63 µm
	Clay	Smaller than 2 µm

Figure. Classification of grain sizes.

Grain shape

The general shape of the individual clasts within a sedimentary rock is also an excellent method to determine the type of geologic environment that may have deposited the sediment long ago. Clastic rocks with **rounded** grains were likely transported by water over long distances and with more energy, in which the individual sediments were abraded and physically rounded as they moved along the riverbed (Fig. 3.6.5). By contrast, **angular** grains indicate the sediments were transported over a relatively short distance by a landslide or mass wasting event, which did not involve water.

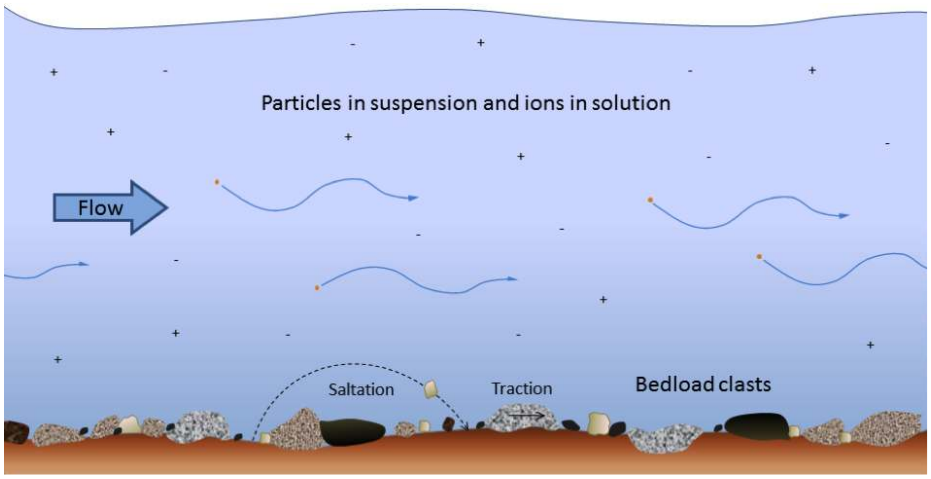


Figure Transportation of clasts by stream flow. Water can move clasts over long distances and cause the abrasion of clasts which results in rounding. This abrasion happens by the movement of clasts on the stream bed.

Grain sorting

The third identifying property of clastic rocks is the sorting of the individual clasts. A **well-sorted** clastic rock contains **clasts with the same general size** in its matrix (Fig. 3.6.5). Such an arrangement means that the clasts were transported over longer distances and/or with greater energy. A **poorly sorted** clastic rock contains clasts that are a **mixture of different sizes** in its matrix. This type of clastic rock typically tells a story of sediment being transported over very short distances.

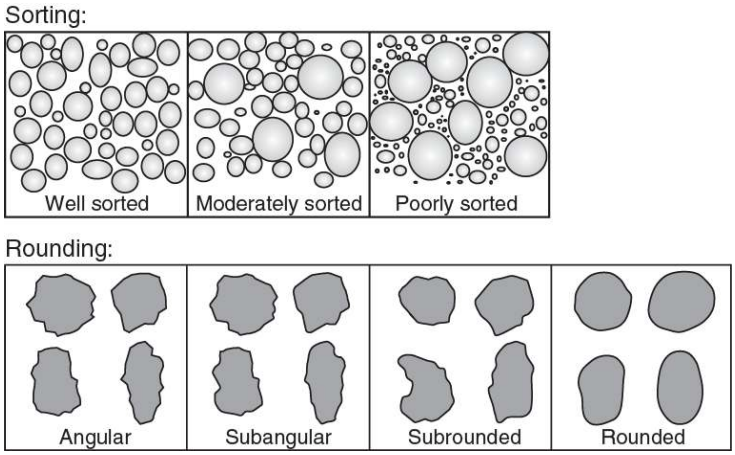


Figure 3.6.5. Sorting of clasts/grains from well-sorted to poorly sorted and the rounding of grains from angular to rounded.

Chemical Sedimentary Rocks

Chemical sedimentary rocks are formed by processes that do not directly involve physical weathering. When preexisting bedrock is weathered by chemical reactions that take place in water, the atmosphere, or the biosphere, that rock is broken down into its constituent elements or ion[/pb_glossary]s that are dissolved and transported in water.

The dissolved ions in the planet's water supply will eventually precipitate as solid, chemical sedimentary rocks. The type of chemical sedimentary rocks that form broadly varies based on the type of elements that precipitate the rock, as well as whether organic materials contribute to the process. Chemical sedimentary rocks may contain silica, evaporite ions (sulfate and chloride), and carbonate.

We can divide chemical sedimentary rocks into two general categories: inorganic and biochemical/organic. **Inorganic chemical sedimentary rocks** are precipitated from only dissolved ions in the water. However, if a chemical rock requires microscopic fossils, shells, or organic material to precipitate, it is called a **Biochemical or Organic sedimentary rock**. As you will discover with some examples in the field guide below, the sedimentary rock can sometimes be both inorganic and organic!

Conglomerate



Figure. Conglomerate. **Click this image** to go to an interactive model by Sara Carena CC-BY.

"CON-GLOM-ER-AT"

Most commonly confused with: breccia

A clastic sedimentary rock. Conglomerate is poorly sorted with well-rounded clasts that are larger than sand (cobble or pebble-sized) within a cementing fine-grained matrix. In a conglomerate, the rock fragments within are sub-rounded to rounded, implying that they traveled along a high-energy water source such as a steep stream or river. To tell conglomerate apart from breccia, take a careful look at the individual clasts that compose it. If most of the clasts are rounded, then your rock sample is a conglomerate.

Breccia



Figure. Breccia. **Click this image** to go to an interactive model of breccia by Sara Carena CC-BY.

"BRECH-AH"

Most commonly confused with: conglomerate

A clastic sedimentary rock. Breccia is a poorly sorted rock with angular clasts that are larger than sand. These clasts are cemented within a much finer-grained matrix. Breccia usually indicated that the clasts were transported over a short distance by a high-energy event, such as a landslide or the movement of a glacier across the Earth's surface. To tell breccia apart from conglomerate, look at the individual clasts within the rock. If the edges of the clasts are angular or sharp, then the rock sample is breccia.

Sandstone



Figure . Sandstone. **Click this image** to go to an interactive model by Sara Carena CC-BY.

"SAND-STONE"

Most commonly confused with: quartzite (metamorphic)

A clastic sedimentary rock. Sandstone contains sand-sized clasts, is most easily identified by its "sandpaper" feel. Sandstone usually appears as a uniform accumulation of cemented sand, which can vary in color as pink, gray, or beige. Desert environments, as well as beaches, reliably deposit sandstone.

Sandstone can sometimes be confused with the metamorphic rock quartzite, and for good reason; sandstone is quartzite's *parent rock* before it is subjected to high temperatures and pressures! Sandstone is typically rougher to the touch and contains smaller individual grains of quartz, whereas quartzite has larger, recrystallized grains of quartz throughout its matrix.

Light tan varieties of sandstone with well-sorted grains of quartz are among the most resistant sedimentary rocks to physical and chemical weathering. Sandstone is often used as bricks and other construction materials for its durable properties.

Shale



Figure. Shale. **Click this image** to go to an interactive model by Dr. Parvinder Sethi CC-BY.

"SHAIL"

Most commonly confused with: slate, limestone, basalt

A clastic sedimentary rock. Shale is well-sorted with silt, mud, or clay-sized clasts that are tightly packed into a solid matrix. It is most easily identified by its tendency to split into thin planes, which is a property called fissility. Shale derives from very low-energy environments, where fine-grained sediment can slowly settle from still water and accumulate over time. Examples of these environments are lagoons, lakes, and continental shelves.

"Shale" is sometimes an umbrella term for fine-grained clastic rocks, and the term "mudstone", which refers to rocks made of clasts smaller than sand, also applies to shale. For this course, "shale" and "mudstone" are interchangeable. Shale is often composed of quartz, feldspar, and clay minerals, although its color can significantly vary depending on the presence of other minerals. Shale can be pigmented red, black, green, gray, brown, etc., by different minerals.

Shale can easily be confused with metamorphic counterpart, slate, which is also fine-grained and separates into thin planes. However, with slate, the breakage is much more pronounced and foliated lines can be observed. Light and dark gray varieties of shale are similar in appearance to limestone, but shale does not contain significant quantities of calcium carbonate, and it will not react to dilute hydrochloric acid.

Shale is very susceptible to physical weathering due to the ease with which it breaks into planes. However, when buried under the Earth's surface, shale's fine-grained matrix can prevent liquids from seeping past, and consequently, shale has held important roles in natural gas and oil exploration.

Rock salt



Figure. Rock Salt/Halite. **Click on this image** to go to a 3D interactive model by Dexter Perkins (CC BY-NC)

Most commonly confused with: rock gypsum and calcite.

A chemical sedimentary rock. Almost every variety of rock salt precipitates inorganically from excess sodium (Na^+) and chloride (Cl^-) ions in water. Rock salt is one of the few rocks that is composed of a single mineral, halite (NaCl), and as such, it has a predictable chemical formula and structure. As with halite, rock salt is typically white or colorless with a cubic shape or clusters of cubic crystals with a distinctive, salty taste.

The presence of rock salt usually indicates that a wet environment has or is undergoing significant drought. Designated as an evaporite, rock salt will commonly precipitate along the edges of an evaporating lake; as water evaporates into the atmosphere, the sodium and chloride ions remaining in the dwindling water become more concentrated until rock salt forms. Rock salt may also be common to dry plains that receive periodic rainfall or regions with briny water.

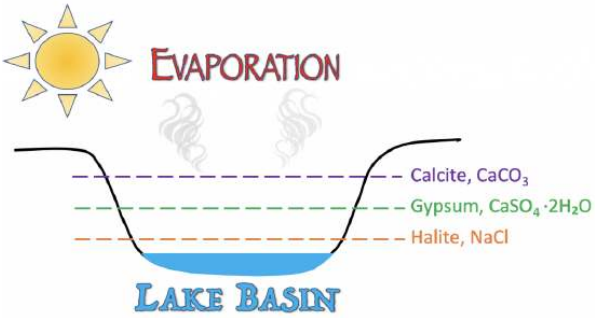


Figure The formation of evaporite sedimentary rocks. As a closed-off body of water, such as a lake, evaporates over time, minerals will precipitate in the following order: calcite, gypsum, and halite.

Rock salt is sometimes confused with another clear, single-mineral sedimentary rock called rock gypsum. The primary difference between these two rocks can be found in the shape of the crystals. Rock salt has cubic-shaped crystals, whereas rock gypsum can have rhomb-shaped crystals. And, you cannot scratch rock salt with a fingernail.

Rock Salt has been popular in the food industry, city planning (de-icing frozen streets), and agriculture. This rock can completely dissolve in freshwater, and it has a Mohs hardness of 3. Therefore, it is very susceptible to both physical and chemical weathering and often causes sinkholes if dissolved underground.

Rock gypsum



Figure. Rock gypsum. **Click on this image** to go to a 3D interactive model by EDUROCK – Educational virtual rock collection (CC BY)

"ROCK JIP-SOME"

Most commonly confused with: rock salt, calcite

A chemical sedimentary rock. Almost every variety of rock gypsum precipitates inorganically from excess calcium (Ca^{2+}) and sulfate (SO_4^{2-}) ions in water. Rock gypsum is an evaporite rock that is composed of a single mineral, gypsum ($CaSO_4 \cdot 2H_2O$). Therefore, this rock has a predictable chemical formula and atomic structure. Rock gypsum is usually white or colorless with rhomb-shaped crystals or sometimes prismatic crystals. Rock gypsum is also very soft, and a fingernail can scratch its surface.

Similar to rock salt, the presence of rock gypsum in an environment or outcrop tells you that it was deposited in an arid region with either a briny or evaporating water source. These environments are often evaporating lakes or salt flats that accumulate a high concentration of calcium carbonate. It is not unusual to find both rock gypsum, rock salt, and calcite in a sequence in the same outcrop. An evaporating body of water will usually precipitate calcite first, followed by rock gypsum, and then rock salt.

Rock gypsum does not react to dilute hydrochloric acid, unlike the mineral calcite, and its crystals are usually prismatic or rhomb-shaped, unlike rock salt. Rock gypsum is softer than both calcite and halite and, therefore, more susceptible to physical weathering. Rock gypsum has been traditionally used in the construction industry as a building material, such as drywall.

Chert



Figure. Chert. Click on this image to go to a 3D interactive model by rocksandminerals (CC BY).

" CHURT"

Most commonly confused with: obsidian, quartz

A chemical sedimentary rock. Chert can be both inorganically and biochemically precipitated in groundwater or the ocean, and it is usually composed of silica. Inorganic varieties of chert form exclusively from dissolved silica in high-temperature, high-pressure groundwater. When this geothermal water source rises to the surface, it can no longer keep the silica in a dissolved state, and it precipitates as solid chert.

Biochemical varieties are formed when silica-based skeletons such as radiolarians and diatoms accumulate on the seafloor after they die. That sediment cements together to form biogenic chert.

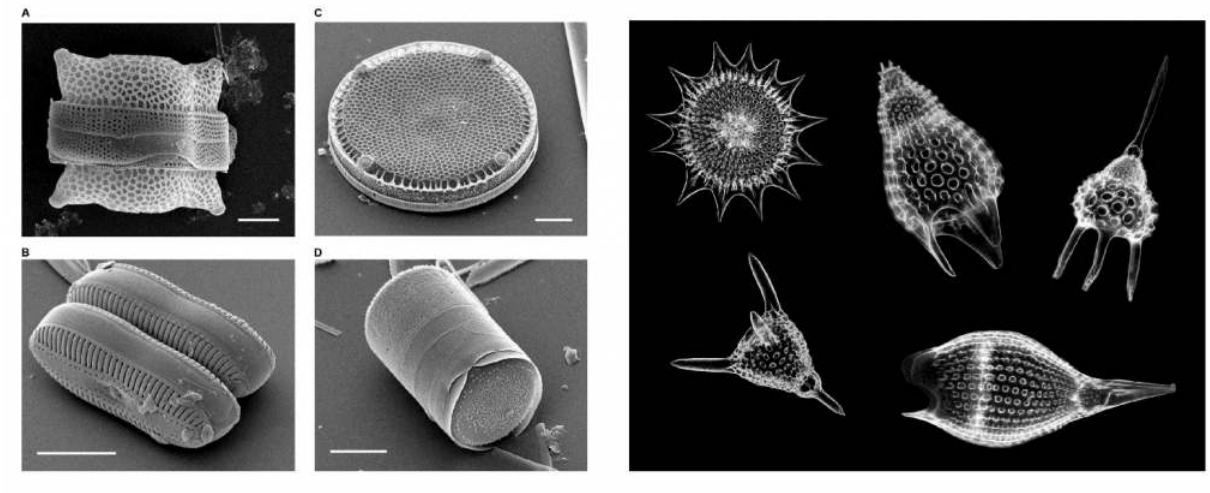


Figure Various diatoms (left) and radiolarian (right) phytoplankton.

We call chert “jasper,” “flint,” “onyx,” and “agate,” which reflect the wide variety of hues and colors that this rock can have. However, chert is most easily identified by the curved conchoidal fracture pattern it displays, which is also present in obsidian and quartz. But, chert is not as glassy or reflective as these rocks and minerals.

Because of its high silica content, chert is very durable and resistant to weathering. This property has made chert desirable as a road material and gravel in the construction trade. Like onyx, it has been traditionally used as a weapon in spearheads and arrowheads for thousands of years.

Limestone



Figure Outcrop of gray limestone that shows weathering. **Click on this image** to go to a 3D interactive model by Théobald GUFFON (CC BY-NC-SA)

Most commonly confused with: shale, basalt

A chemical sedimentary rock. Limestone can be both inorganically and biochemically precipitated in seawater, and it is primarily composed of calcium carbonate. Inorganic limestone forms in the deep ocean when seawater reaches colder depths that cause the dissolved ions calcium (Ca^{2+}) and carbonate (CO_3^{2-}) to precipitate as solid rock. There are several varieties of limestone that incorporate both the ions and organic material from marine organisms when forming a solid matrix. These types of limestone, described in more detail below, are biochemical sedimentary rocks.

Limestone is often light to dark gray, or tan, and it can be scratched by a penny. Limestone is composed of calcium carbonate (calcite, ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$). This rock will strongly fizz when exposed to dilute hydrochloric acid. This reaction

clearly distinguishes limestone from other dark gray rocks such as shale and basalt.

Limestone is very susceptible to chemical weathering by dissolution as rainwater and groundwater are both slightly acidic. Consequently, regions with underlying limestone often have caves and experience sinkholes. Despite its tendency to weather over long periods, limestone has been (and remains) a popular construction material for thousands of years, dating back to the Great Pyramid of Giza in Ancient Egypt.

Fossiliferous Limestone



Figure. Fossiliferous limestone. **Click on this image** to go to a 3D interactive model by rocksandminerals (CC BY)

"FOSSIL-LIFF-ER-RUS LIME-STONE"

Most commonly confused with: limestone

A biochemical sedimentary rock. Fossiliferous limestone specifies a type of limestone that includes visible fossils within the rock's matrix, and therefore, it is biochemically precipitated. Fossiliferous limestone may contain the remains of shells, such as brachiopods, trilobites, plants, or many other types of other animals.

Fossiliferous limestone can form in the deep ocean or near coral reefs, which are composed primarily of calcium carbonate. The surrounding matrix has the same properties as limestone (see above), and this rock will react when exposed to dilute hydrochloric acid. However, macrofossils within the rock distinguish fossiliferous limestone from inorganically precipitated limestone.

Coquina



Figure. Coquina. Click on the image to go to an interactive model by the Digital Atlas of Ancient Life (CC BY-SA)

"CO-KEEN-AH"

Most commonly confused with: n/a

A biochemical sedimentary rock. Coquina is a biochemical variety of limestone that is composed of shells, fossils, and sand that have been poorly cemented together. Coquina is often tan in color with shells that are easily visible to the naked eye. Most of the material within coquina is composed of calcium carbonate, and it fizzes in contact with dilute hydrochloric acid.

Coquina is often found along beaches or tidal pools with abundant shelly creatures. The leftover shells, sand, and fossils are eventually clustered together by wave action and cemented together upon burial.

Chalk



Figure. Chalk. **Click the image** to open an interactive model by PalomarESES CC BY-ND.

Most commonly confused with: n/a

A chemical sedimentary rock, sometimes also called marlstone or marl. Chalk is a biochemical variety of limestone that is composed of microscopic shells from an oceanic organism called a coccolithophore or coccolith. When these organisms die, their calcium carbonate-based shells accumulate along the bottom of the ocean as ooze-like sediment, which eventually cements as chalk.

Chalk is distinctively white, powdery, and extremely soft to the point of crumbling when touched. It is easily weathered on the Earth's surface and typically is collected by mining. Chalk has traditionally been used on school boards, but its use has since been widely discontinued in favor of gypsum.

Coal



Figure. Coal. **Click on the image** to go to an interactive model by the Byrd Polar & Climate Research Center (CC BY-ND)

Most commonly confused with: obsidian

An organic sedimentary rock. Coal is one of the few organic sedimentary rocks that is composed of dead plant matter. When enormous amounts of plant matter decay and accumulate from an environment, such as swamplands or dense forests, it eventually becomes buried within the Earth's surface. There, over millions of years and at high temperatures and pressures, this matter will transform into solid, black rock.

Coal is easily identified by its low density, dark black color, and soft, crumbling surface. Because organic matter has formed it, it is mostly made of carbon (C). Although obsidian is also black and shines like coal sometimes does, obsidian is much harder and more resistant to breaking.

Coal is combustible, and it has reshaped how societies are powered by electricity. Byproducts of burning this rock are commonly called fossil fuels, and they have contributed to significant climate problems that the world is currently facing.

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4.5: Metamorphic Rocks

The word “metamorphic” is Greek: *meta* means *change*; *morphos* means *form*. When rock units are buried very deeply within Earth’s crust, they are subjected to high temperatures and pressures. These rocks are squeezed and warped like putty in a process called *metamorphism;* some of these rocks grow new minerals and textures, and others lose minerals. The result of this high temperature and pressure transformation is a *metamorphic* rock.



Figure. Due to metamorphism, limestone and chert layers of rock have been folded at high temperatures and pressures.

Metamorphic rocks can involve high temperatures, but unlike igneous rocks, they do not melt into magma. They also involve transforming one rock type into another, but unlike sedimentary rocks, they do not reduce the original rock into fragments before reassembling. Metamorphism is a unique process that takes any type of preexisting rock (even old metamorphic units) and subjects it to heat and pressure over long periods until it has changed.

Metamorphism usually happens at least a couple kilometers beneath the Earth’s surface. Tectonic boundaries, in particular, cause different types of metamorphism at mountain-building cores, subduction zones, and spreading centers. The metamorphic rocks that form at elevated temperatures and pressures are categorized by both *metamorphic texture* and the amount of change that the original rock appears to have undergone: *metamorphic grade*.

Metamorphic Texture

Metamorphic texture describes the shape and orientation of mineral grains within a metamorphic rock. As the original rock is subjected to higher temperatures and pressures, some of its minerals might stretch out in a single direction, recrystallize, or enlarge. Therefore, the new metamorphic rock will have a different texture than the parent rock.

Metamorphic texture is broadly categorized as either *foliated* or *non-foliated*.

Foliated texture

Foliation is a term that describes how minerals line up along a preferred direction. Some minerals, particularly micas, are usually thin and planar by default. Rocks with foliated texture look like they have their minerals stacked together as though they were

pages in a book; hence the term “folia,” or leaflike.



Figure. Foliated texture can look like very thin horizontal layers in rock that appear like they can peel apart like leaves, like in this mica-schist formation.

Why do foliated rocks form during metamorphism? At least two conditions need to be satisfied. First, the original rock must contain minerals that will easily deform or align with applied pressure into a flat plane. We know that mica minerals such as muscovite and biotite will do this as well as amphibole; however, stronger minerals like quartz and feldspar will often resist most pressure in the Earth’s crust.



Figure. Mineral banding in the high-grade metamorphic rock gneiss. This segregation of minerals together is a type of foliation.

The second requirement needed for a foliated texture is directed pressure called differential stress. If pressure is applied unevenly, the weak minerals in the original rock will easily deform into long planes. This type of stress could be squeezing (compression),

stretching (tensional), or sliding (shear) the rock unit.

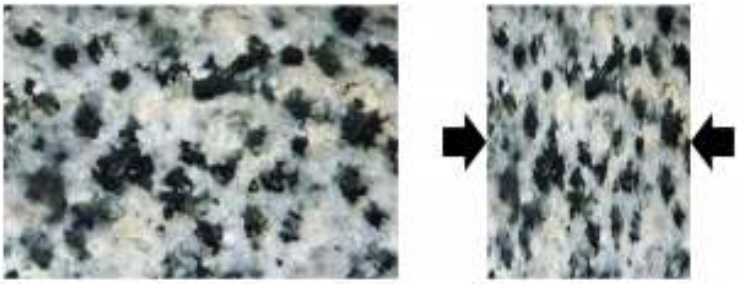


Figure. Foliation that develops when minerals are squeezed and deform by lengthening in the direction perpendicular to the greatest stress (indicated by black arrows). Left- before squeezing. Right- after squeezing.

Non-Foliated texture

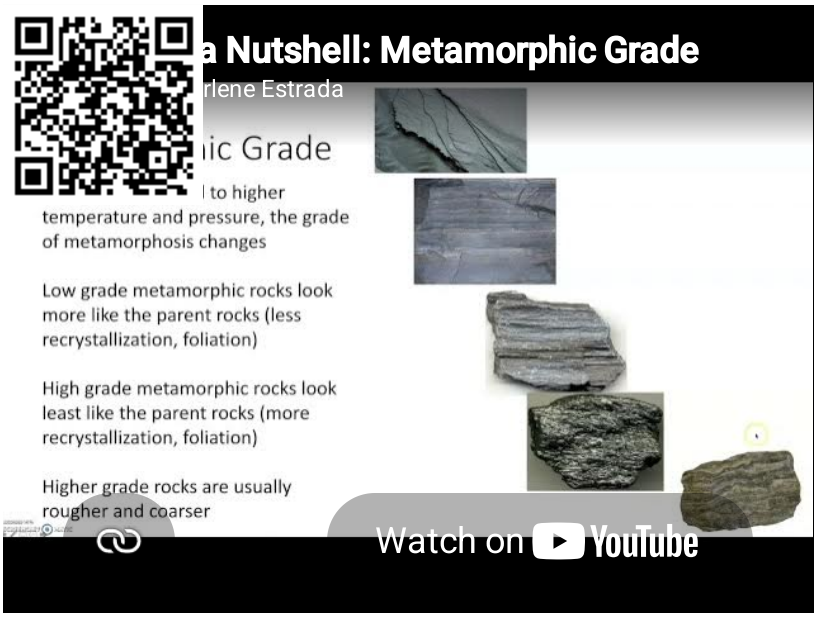


Figure. Non-foliated texture does not have any distinct layering or banding. Instead, the rock appears granular or crystalline, as seen in the rock quartzite.

Non-foliated metamorphic rocks do not have any preferential alignments of mineral grains. These rocks are also called "Granoblastic", which references the tendency of the individual grains to have somewhat equal shapes and dimensions. The majority of non-foliated metamorphic rock contains mostly one mineral. Nonetheless, metamorphism still occurs within these rocks; the mineral grains have recrystallized, interlocked, and grown larger. As a consequence, non-foliated rocks are much more durable and resistant to weathering than their parent rocks.

Metamorphic Grade

Video. What does a rock look like as it undergoes increasing grade in metamorphism? (1:22).



Metamorphic Grade refers to the extent to which metamorphism can transform the preexisting rock. This original rock is called the **parent rock**, and it can undergo **low-grade metamorphism** (little metamorphic change) to **high-grade metamorphism** (significant metamorphic change).

Low-grade metamorphism begins at temperatures and pressures that are not much higher than those that form sedimentary rocks. This type of metamorphism often results from rocks being buried at depths of at least 2 km. The conditions in such an environment are typically low temperature and pressure. The parent rock still transforms into a new metamorphic rock; however, it is often easy to identify visual similarities between the resulting metamorphic rock and original rock.

High-grade metamorphism requires both high temperatures and pressures. Burial depths for high-grade metamorphic rocks can be up to 35 km! Mountain-building centers and subduction zones are prime examples of regions where high-grade metamorphism might take place. When a parent rock undergoes metamorphism under these conditions, the resulting metamorphic rock bears very few similarities to the original specimen; it often has strong foliation or banding of minerals.

Metamorphic rocks are often not restricted to either low or high grades. Geologists typically rank these rocks based on how many (or few) similarities they bear to the parent rock. In a sequence or outcrop filled with metamorphic rocks, you might observe some units that appear similar to the parent rock and others that look nothing like it. In general, the following rocks have been ranked in order of increasing metamorphic grade:

Metamorphic Rock Field Guide

Slate



Figure. Slate. **Click on this image** to go to a 3D interactive model by rocksandminerals CC BY.

"SLAYTE"

Most commonly confused with: shale, phyllite

A foliated, low-grade metamorphic rock. Slate is fine-grained and composed of clays and mica minerals that are usually too small to see with the naked eye. Slate displays strong foliation in thin sheets or layers that sometimes resemble sedimentary bedding or the linae of shale. It is usually dark gray, but it can also be red, green, brown, and even blue.

Although slate often forms from shale and resembles this parent rock, its foliation pattern is typically more pronounced. Slate is also sometimes confused with a higher grade metamorphic rock, phyllite, which contains visible grains of mica minerals.

Slate will easily break into sheets along its foliation planes, but the silicate minerals that compose its fine-grained structure make it durable against physical weathering processes directed perpendicular to the foliation line. Slate is therefore used as a building material for roofs and tiles in construction.

Phyllite



Figure. Phyllite. **Click on this image** to go to a 3D interactive model by Dr. Parvinder Sethi CC BY.

"FY-LITE"

Most commonly confused with: slate, schist

A foliated, low to medium-grade metamorphic rock. Phyllite is fine-grained and composed mostly of quartz, feldspar, and visible flakes of mica minerals. These enlarged mica crystals give phyllite a shimmering or silky appearance under light. Phyllite, like slate, is foliated with thin sheets that sometime separate into uneven layers that give an individual rock a wavy appearance.

The color of phyllite is often gray, black, tan, or green, and it is sometimes confused for either slate or schist. Phyllite can generally be thought of as a version of slate that has undergone more metamorphism; it can be distinguished from slate by its visible, glittering mica minerals. By contrast, phyllite can very broadly be thought of as the lower-grade version of schist. Schist has much larger, foliated plates of mica.

Schist

Video. Let's observe a garnet schist. The sample is foliated. The shiny minerals are mica while the dark red grains are garnet (0:31).



"SHIST"

Most commonly confused with: phyllite

A foliated, medium-grade metamorphic rock. Schist contains large, leaflike grains of mica such as muscovite and biotite, that are strongly oriented in a single direction. Some varieties of schist have garnet, which only forms at elevated temperatures and pressures, although most schists also contain quartz and feldspar. Due to the presence of large mica plates and their flattened orientation, schist is usually shiny or vitreous (glassy) under light.

Schist is often gray or brown in color, and it is sometimes confused with phyllite, which contains smaller grains of mica. Schist can also be identified by its strong foliation pattern, which is called "schistosity". This pattern is much more pronounced than lower grade metamorphic rocks such as slate and phyllite.

Gneiss



Figure. Gneiss. **Click on this image** to go to a 3D interactive model by Sara Carena CC BY-NC.

"NICE"

Most commonly confused with: granite, diorite

A foliated, high-grade metamorphic rock. Gneiss has a visible separation of light and dark bands, which is called lineation. Gneiss is coarse-grained and mostly contains silicate minerals that are resistant to high temperatures such as quartz, feldspar, biotite, and garnet. The banding pattern on gneiss is usually wavy or folded, which reflects how the parent rock deformed like putty at extremely high temperatures and pressures; indeed, sometimes the rock partially melts.

Gneiss is occasionally mislabeled as the igneous rock granite or diorite. Although gneiss may contain similar silicate minerals as these rocks, it is distinctively banded, whereas the igneous rocks are granular and have no preferred orientation. Because gneiss is not easily broken into sheets, it is a useful construction material in landscaping and architecture.

Marble



Figure. Marble. **Click on this image** to go to a 3D interactive model by EDUROCK – EDUCATIONAL VIRTUAL ROCK COLLECTION CC BY.

Most commonly confused with: quartzite

A non-foliated metamorphic rock. Marble forms under low grade or high-grade metamorphism, although at the latter it will grow larger, more interlocked crystals that reflect its higher temperature and pressure origins. The parent rock of marble is limestone, although marble typically appears to have more identifiable grains. Marble contains the mineral calcite and/or dolomite, which may fizz with dilute hydrochloric acid.

Marble is light in color, and it can be white, pink, tan, or gray. Some varieties of marble look similar to another non-foliated metamorphic rock, quartzite. However, the minerals calcite and dolomite that primarily compose marble are much softer than those found in quartz, and marble is easier to scratch with an iron nail.

Although marble that contains calcite will react to acidic rainwater and groundwater over long periods, marble has traditionally been a sculpting material in art and architecture for thousands of years. Polished material remains a popular building and decorative material to this day.

Quartzite



Figure. Quartzite. **Click on this image** to go to a 3D interactive model by rocksandminerals CC BY.

Most commonly confused with: marble, sandstone

A non-foliated, high-grade metamorphic rock. Quartzite only forms from the sedimentary rock sandstone, and it almost exclusively contains quartz. The high temperatures and pressures in the metamorphic environment have caused the individual quartz grains to increase in size and interlock.

Quartzite is usually light tan or pink in color and coarse-grained. Like other rocks that predominately contain silica, it can display conchoidal fracture, although it is very resistant to mechanical breakage. Although this rock can look similar to marble, it will not fizz in contact with dilute hydrochloric acid, and it will not easily scratch.

Quartzite is also more durable than its parent rock, sandstone, which weakly cements sand grains together. For example, if sandstone is hit with a rock hammer, the individual grains will remain intact and the cement would crumble. If quartzite is hit with a rock hammer, it would probably spark from the friction and cause the hammer to bounce back, and, thus, is NOT recommended!

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CHAPTER OVERVIEW

5: Plate Tectonics

5.1: Alfred Wegener's Continental Drift Hypothesis

5.2: Development of Plate Tectonic Theory

5.3: Layers of Earth

5.4: Plate Boundaries

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5.1: Alfred Wegener's Continental Drift Hypothesis

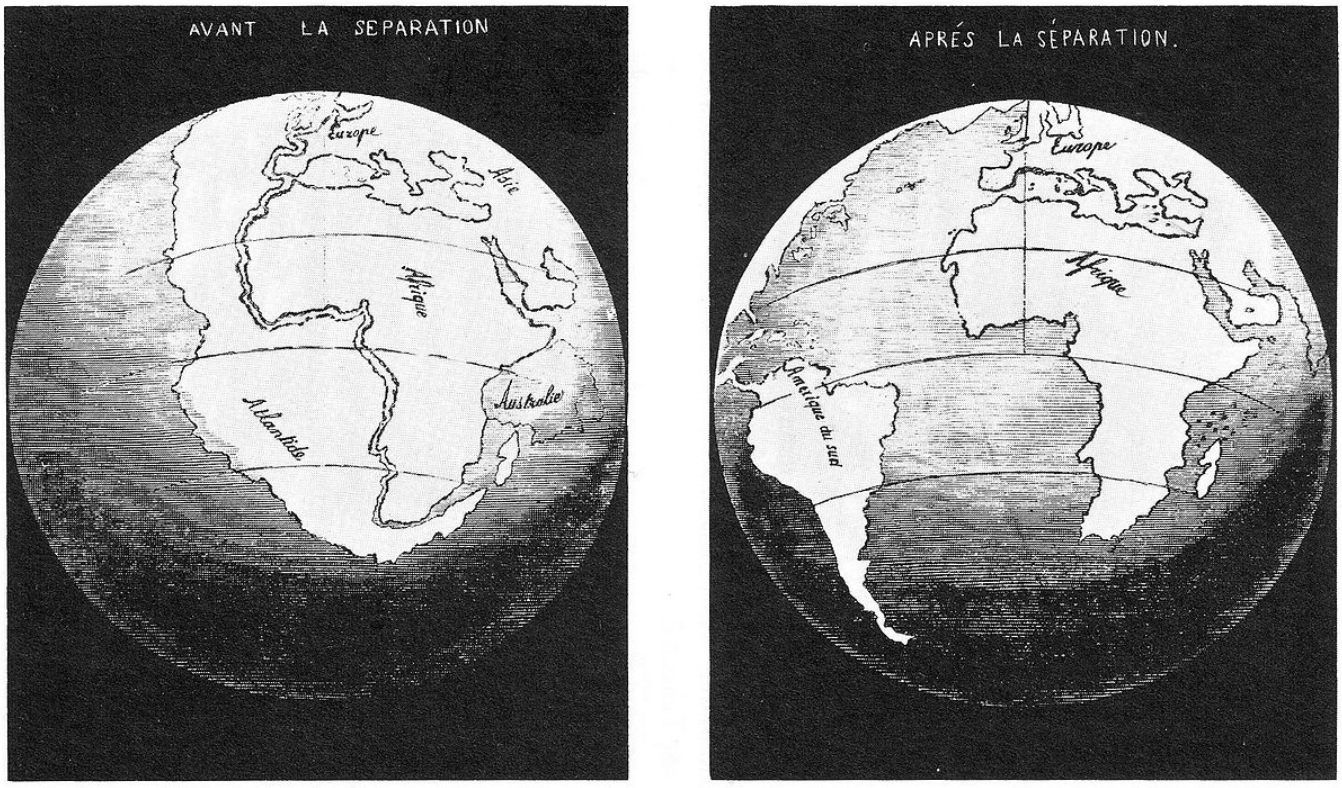
Alfred Wegener's Continental Drift Hypothesis



Alfred Wegener (1880-1930) was a German scientist specializing in meteorology and climatology. His knack for questioning accepted ideas started in 1910 when he disagreed with the explanation that the Bering Land Bridge was formed by isostasy and that similar land bridges once connected the continents. 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, his hypothesis was backed up by a long list of evidence.

Early Evidence for Continental Drift Hypothesis

Wegener's first evidence was that some continents' coastlines 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. Antonio Snider-Pellegrini did preliminary work on continental separation and matching fossils in 1858.



Public Domain, Link

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

What Wegener did differently was synthesizing a large amount of data in one place. He used true continents' edges based on the continental shelves' shapes. This resulted in a better fit than previous efforts that traced the existing coastlinesWegener 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 that these land-dwelling creatures could not have swum across an ocean.

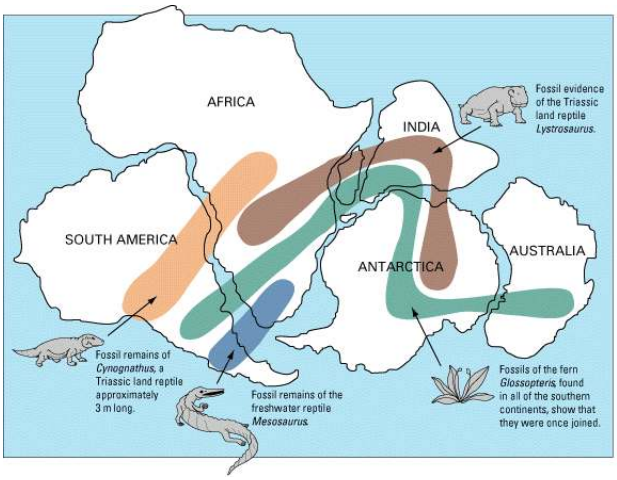


Image showing fossils that connect the continents of Gondwana (the southern continents of Pangea).

Opponents of continental drift insisted trans-oceanic land bridges allowed animals and plants to move between continents. The land bridges eventually eroded, 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. Land bridges could not explain these climate anomalies. 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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5.2: Development of Plate Tectonic Theory

Development of Plate Tectonic Theory

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

Mapping of the ocean floor

In 1947 researchers started using an adaptation of SONAR to map a region in the middle of the Atlantic Ocean with poorly-understood topographic and thermal properties. Using this information, Bruce Heezen and Marie Tharp created the first detailed map of the ocean floor to reveal the Mid-Atlantic Ridge. This basaltic mountain range spanned the length of the Atlantic Ocean, with rock chemistry and dimensions unlike the mountains found on the continents. Initially, scientists thought the ridge was part of a mechanism that explained the expanding Earth or ocean-basin growth hypotheses. In 1959, Harry Hess proposed the seafloor spreading hypothesis that the mid-ocean ridges represented tectonic plate factories, where a new oceanic plate was issuing from these long volcanic ridges. Scientists later included transform faults perpendicular to the ridges to better account for varying movement rates between the newly formed plates. When earthquake epicenters were discovered along the ridges, the idea that earthquakes were linked to plate movement took hold.



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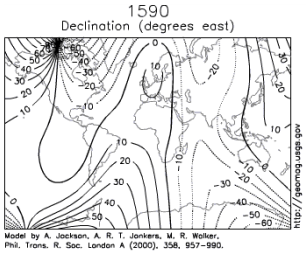
Marie Tharp

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

Paleomagnetism

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

Paleomagnetism studies magnetic fields frozen within rocks, basically a fossilized compass. The first hard evidence to support plate motion came from paleomagnetism.

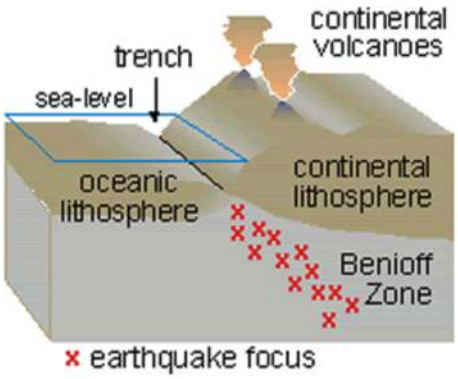


This animation shows how the magnetic poles have moved over 400 years.

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

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

Wadati-Benioff Zones



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

John Tuzo Wilson

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



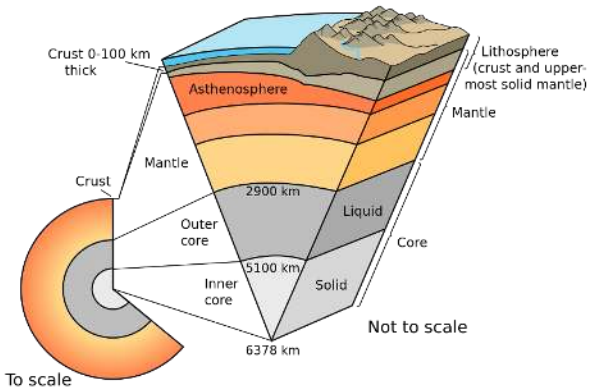
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5.3: Layers of Earth

Layers of the Earth

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



The core can be broken into two: the inner and outer core. The **inner core** is a solid iron-nickel alloy with a radius of about 760 mi. The inner core's temperature is believed to be the same as the sun's surface (approximately 5,430 °C). The transition between the inner and outer core is approximately 3,200 mi beneath the Earth's surface.

The **outer core** is also iron and nickel, but it's liquid. The liquid outer core flows, producing the Earth's magnetic field.

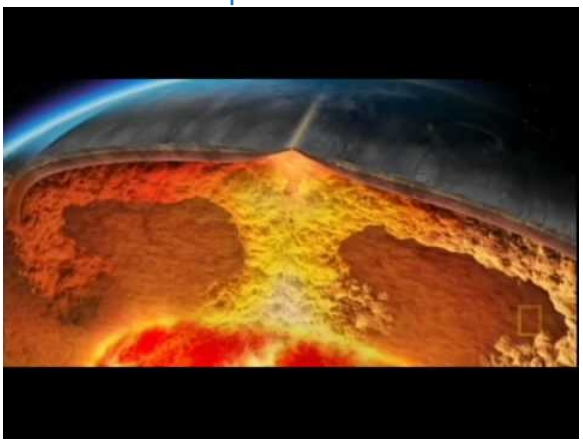
The next layer is called the **mantle**, the thickest layer on the Earth. The mantle is a plastic solid, which can flow very slowly. The heat from the Earth's core causes the mantle to convect, like water over a stove but much slower, and it is the mantle's convection that is the driving force of plate tectonics.

The Earth's surface layer is called the **crust,** and it makes up only 1% of Earth's mass. The crust is subdivided into two components: **oceanic** and **continental** crust. Referring to the image below, note that the **oceanic crust** is only about 3 miles thick but is slightly denser than the continental crust. Most of this oceanic rock is called basalt, a dark, dense rock.

Continental crust is much thicker than oceanic crust (averages between 20 to 25 miles thick) but is slightly less dense than oceanic crust. The main type of rock on continents is called granite. So if these two types of crust collide, what do you think would happen to the oceanic crust? As a whole, notice that the crust is lighter than the mantle. It is sometimes said that the crust "floats" on the mantle like an iceberg in water, which is not too far from the truth and is called **isostasy**. Finally, the crust is the coldest, most rigid, and brittle layer with many folds and fractures.

Two additional layers in the Earth are very important to plate tectonics. The **asthenosphere** is located in the upper mantle at a depth between ~ 50 and 124 miles below the surface. It is a mostly solid but weak layer that flows like silly putty. Above the asthenosphere is the **lithosphere.** The lithosphere consists of the uppermost mantle and crust (both oceanic and continental). The lithosphere makes up the tectonic plates.

What moves the plates?



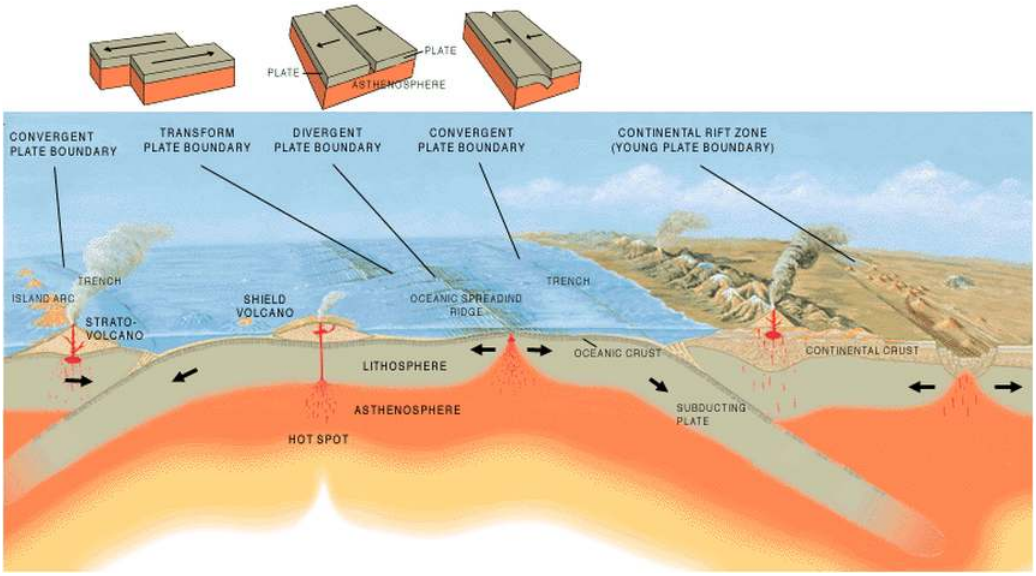
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5.4: Plate Boundaries

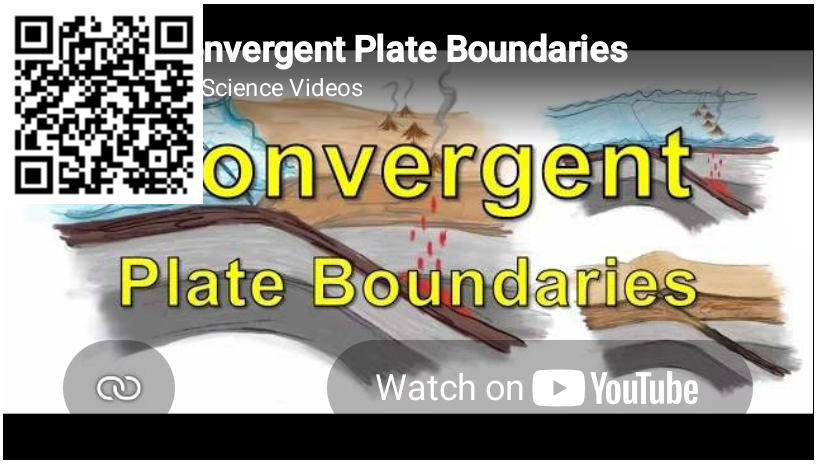
Plate Boundaries

In a simplified model, there are three major types of tectonic plate boundaries: ***convergent***, ***divergent***, and ***transform***.



Convergent plate boundaries

Let's first look at convergent plate boundaries, which can be broken down into three subcategories, depending on the type of crusts involved. Recall that oceanic crust (basaltic) is denser than continental rock (granitic). Thus when two tectonic plates collide, **the denser** oceanic crust **will subduct underneath the lighter** continental crust. If the subducting rock becomes stuck, vast amounts of energy build up. Once the pressure and energy accumulated exceed the resistance of the rock slab, the rock will rupture, creating powerful earthquakes. As the subducted material sinks further, it will begin to melt under great heat and pressure, becoming less dense as it melts and rises up as magma to form dangerous **composite volcanoes**. Mountain ranges created by **oceanic-to-continental convergence** are the Andes mountains in South America, the Cascades in the western United States, and the Ring of Fire in the Pacific Ocean.



Ocean to Continent Convergent Plate Boundary

When an oceanic plate boundary collides with a continental plate boundary, the oceanic plate will always subduct under the continent. Again you will get a trench marking the boundary between the two plates. As the oceanic plate subducts, it generates

magma that leads to a row of volcanoes embedded in a coastal mountain range. Earthquakes are generated along the subduction zone and within the mountain range.

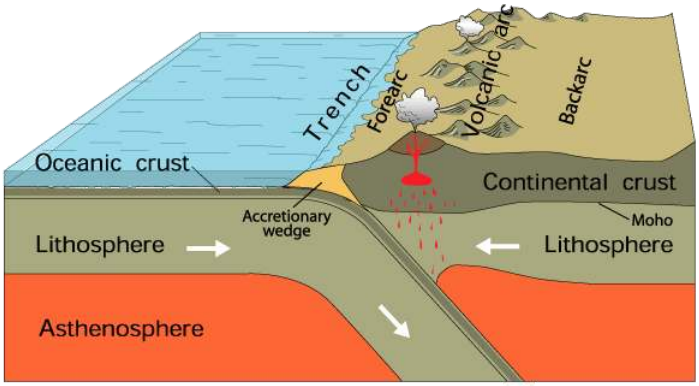
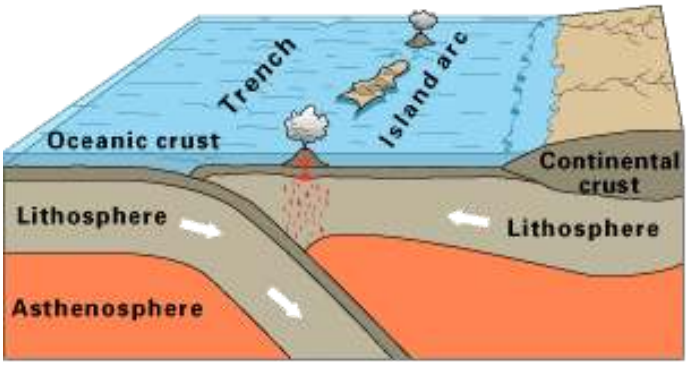


Diagram of Oceanic-continent subduction

Ocean to Ocean Convergent Plate Boundary

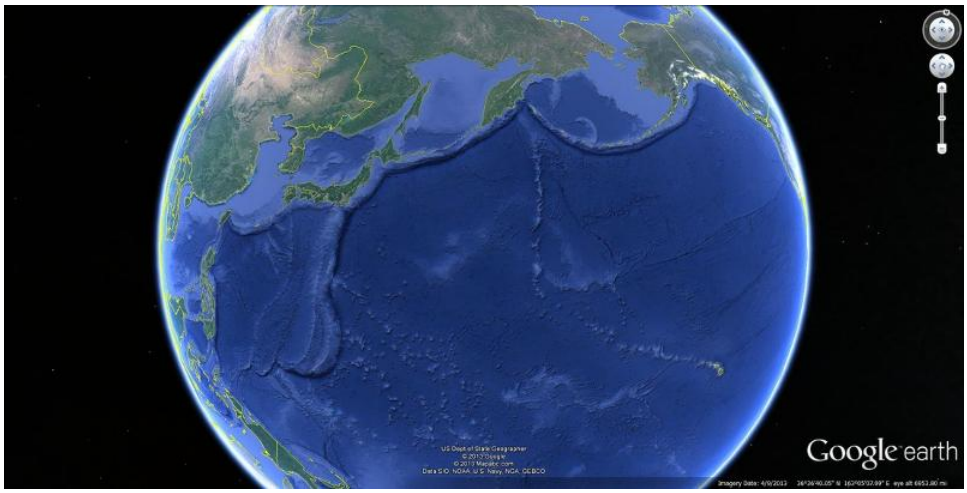
A trench marks the boundary of the two plates. When two oceanic plates collide, you'll notice that one plate slides under the other. We call that process subduction. As the subducting plate descends into the asthenosphere, part of it begins to melt, generating magma. The magma is less dense than the surrounding rock and begins to rise toward the surface. The rising magma creates a line of volcanic islands on the non-subducted plate. The colliding frictional forces between the two plates create earthquakes starting around the trench and progressing inward along the subduction plate. Just like continental-to-oceanic convergence, this plate boundary can generate powerful earthquakes and volcanoes.



Oceanic-oceanic convergence

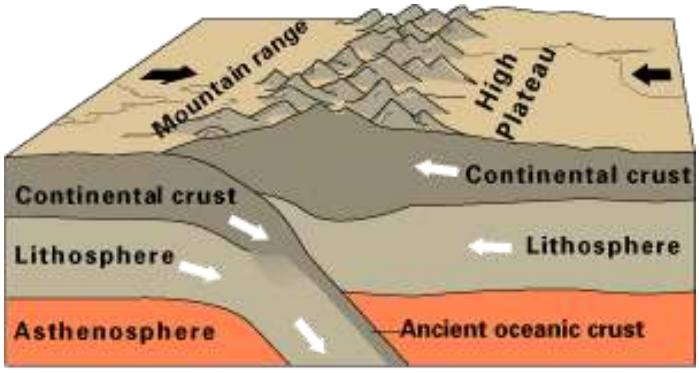
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Still, instead of volcanoes on land, volcanic islands form, such as Japan, the Aleutian Islands of Alaska, and Indonesia. Large shallow earthquakes can generate tsunamis in these regions. The great earthquake in Indonesia in 2004, which produced the devastating tsunami, was created by this process along with the 2011 earthquake and tsunami in Japan. Below is a Google Earth image showing a series of oceanic-to-oceanic subduction zones within the Pacific Ring of Fire. You can see the subduction zones that create the volcanic and powerful Aleutian Islands and the converging subduction plates that make the volcanic islands of Japan.

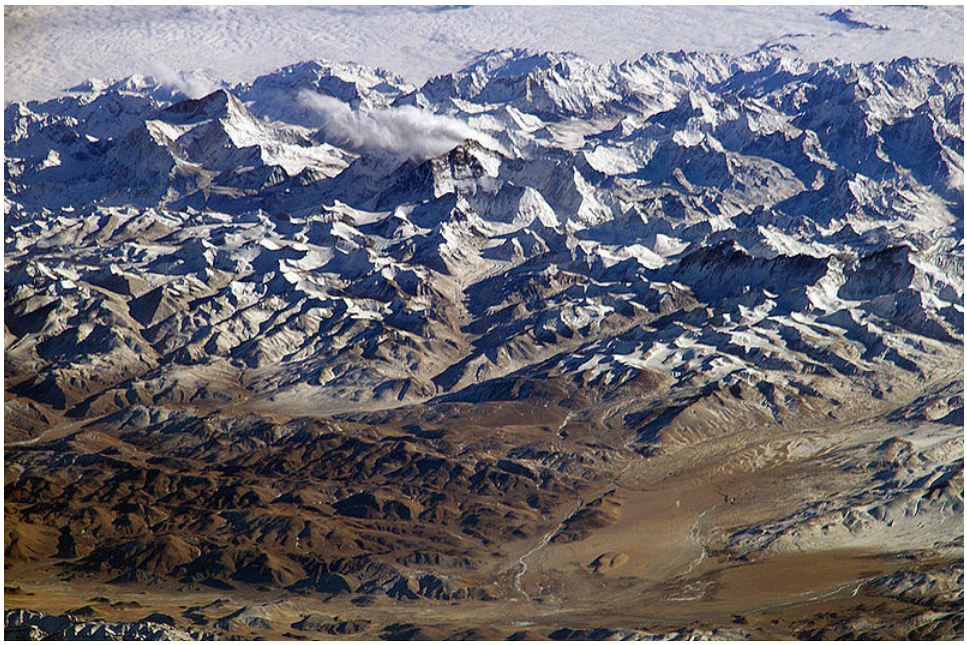


Continent to Continent Convergent Plate Boundary

When two continental plates converge, instead of subduction, the two similar tectonic plates will buckle up to create large mountain ranges like a massive car pile-up. This is called **continental-to-continental convergence** and geologically creates intense folding and faulting rather than volcanic activity. Examples of mountain ranges created by this process are the Himalayan mountains (taken from the International Space Station) as India collided with Asia, the Alps in Europe, and the Appalachian mountains in the United States as the North American plate collided with the African plate when Pangea was forming. The Kashmir India earthquake of 2005 killed over 80,000 people because of this process. And most recently, the 2008 earthquake in China killed nearly 85,000 people before the Summer Olympics because of this tectonic force. The Himalayan mountains are a result of India crashing into Eurasia. Because there is no subduction, there are few volcanoes, if any. However, the region is experiencing tremendous compressional stress over a large area, leading to the loss of earthquakes.

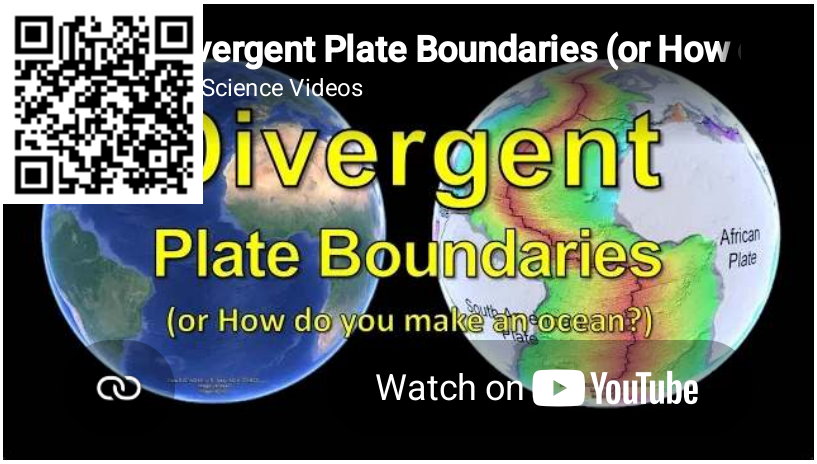


Continental-continental convergence



Divergent Plate Boundaries

A divergent plate boundary is when two plates move away from each other under tensional stress. The oceanic crust becomes thin, and cracks open up fissures. Magma moves up to the surface through these fissures and solidifies, creating new oceanic crust. You do not get the traditional cone-shaped volcanoes with a central vent forming at these locations, but magma moving up to the surface at the fissure cracks forming new crust. Shallow, low to medium magnitude earthquakes occur at divergent plate boundaries. As divergence occurs, shallow earthquakes and volcanoes along the rift areas can occur. When the process begins on land, it is called **continental rifting,** and a valley will develop, such as the Great Rift Valley in Africa. Over time that valley can fill up with water creating linear lakes. If divergence continues, a sea can form like the Red Sea and finally an ocean like the Atlantic Ocean.

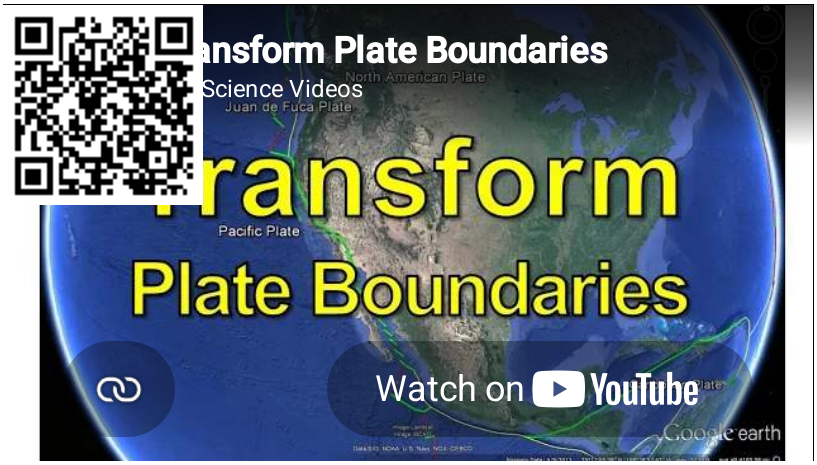


A divergent plate boundary under water is called a **mid-ocean ridge,** such as the mid-Atlantic ridge. Check out the eastern half of Africa and notice the lakes that look linear. Eastern Africa is separated from these linear lakes by the Great Rift Valley and the Red Sea. Notice how the Red Sea looks like it could be put back together again. The ultimate divergent boundary is the Atlantic Ocean, which began when Pangea broke apart. Below are two satellite images using Google Earth, both focusing on parts of Africa. On the left, you can see rift valleys filled with water to form linear lakes. On the right, in northern Africa, you can see the Red Sea with a rift valley in the center, which used to be a linear lake that grew into a sea. If the Red Sea continues to grow, it could form an ocean similar to the Atlantic Ocean with the mid-Atlantic Ridge.

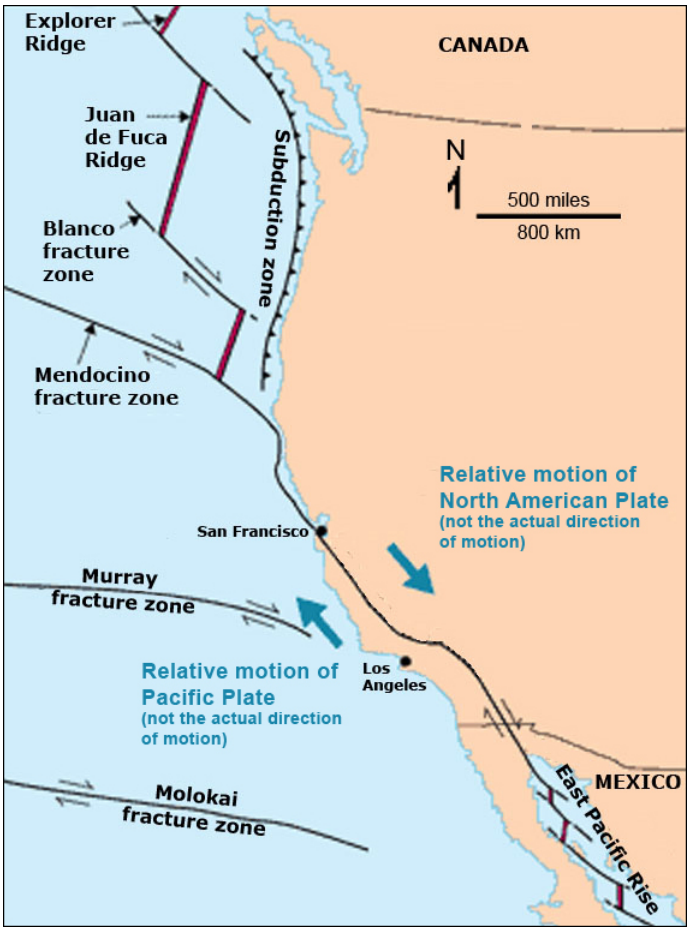
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Transform Plate Boundaries

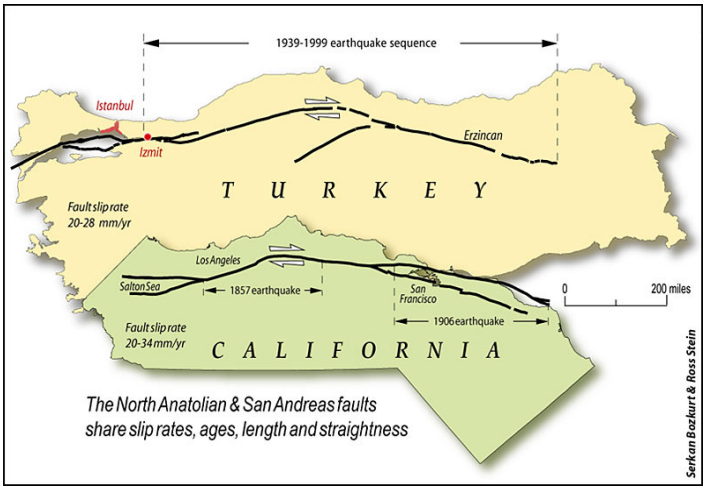


Transform boundaries occur when two tectonic plates slide (or grind) past parallel to each other. Transform plate boundaries form when two plates slide past each other under shear stress. There is no subduction, so volcanoes do not form. However, tremendous frictional stress can create large, destructive earthquakes. The most famous example of a transform plate boundary is the San Andreas Fault running through California. This fault marks the boundary between the North American and Pacific plates. The most famous transform boundary is the **San Andreas Fault,** where the Pacific plate (Los Angeles and Hawaii) is grinding past the North American plate (that San Francisco and the rest of the United States are on) 3 inches a year. Recently, geologists have stated that San Francisco should expect another disastrous earthquake in the next 30 years.



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Another important transform boundary is the **North Anatolian Fault** in Turkey. This powerful fault last ruptured in 1999 in Izmit, Turkey, killing 17,000 people in 48 seconds.



Comparison of the San Andreas Fault, CA, and the North Anatolian Fault in Turkey. Both are located along transform boundaries. Image source: United States Geologic Survey (USGS).

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CHAPTER OVERVIEW

6: Geologic Environments and Structures

6.1: Sedimentary Rocks, Weathering and Erosion

6.2: Sedimentary Environments

6.3: Plastic Deformation

6.4: Deformation- Stress and Strain

6.5: Brittle Deformation

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6.1: Sedimentary Rocks, Weathering and Erosion

Sedimentary rocks are types of rock that are formed by the deposition of material at the Earth's surface and within bodies of water. Sedimentation is the collective name for processes that cause mineral and/or organic particles (detritus) to settle and accumulate or minerals that precipitate from a solution. Particles that form a sedimentary rock by accumulating are called sediment. Before being deposited, sediment was formed by weathering and erosion in a source area and then transported to the place of deposition by water, wind, ice, mass movement, or glaciers.



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Weathering is the breaking down of rocks, soil, and minerals as well as artificial materials through contact with the Earth's atmosphere, biota, and waters. Weathering occurs in situ, or "with no movement", and thus should not be confused with erosion, which involves the movement of rocks and minerals by agents such as water, ice, snow, wind, waves, and gravity.

Two important classifications of weathering processes exist – **physical and chemical weathering;** each sometimes involves a biological component. Mechanical or physical weathering involves the breakdown of rocks and soils through direct contact with atmospheric conditions, such as heat, water, ice, and pressure. The second classification, chemical weathering, involves the direct effect of atmospheric chemicals or biologically produced chemicals, also known as biological weathering, in the breakdown of rocks, soils, and minerals.

The materials left over after the rock breaks down, combined with organic material, create soil. The parent material determines the mineral content of the soil thus a soil derived from a single rock type can often be deficient in one or more minerals for good fertility, while a soil weathered from a mix of rock types (as in glacial, aeolian, or alluvial sediments) often makes more fertile soil.

- Erosion: The process of picking up material.
- Transportation: The movement of material by wind, water, and gravity.
- Deposition: The laying down of material in different environments. Control on erosion, transportation, and deposition is the amount of Energy. High energy = carry bigger sizes.

Physical Weathering

Physical weathering, also known as mechanical weathering, is the class of processes that causes the disintegration of rocks without chemical change. The primary process in physical weathering is abrasion (the process by which clasts and other particles are reduced in size). Physical weathering can occur due to temperature, pressure, frost, etc.

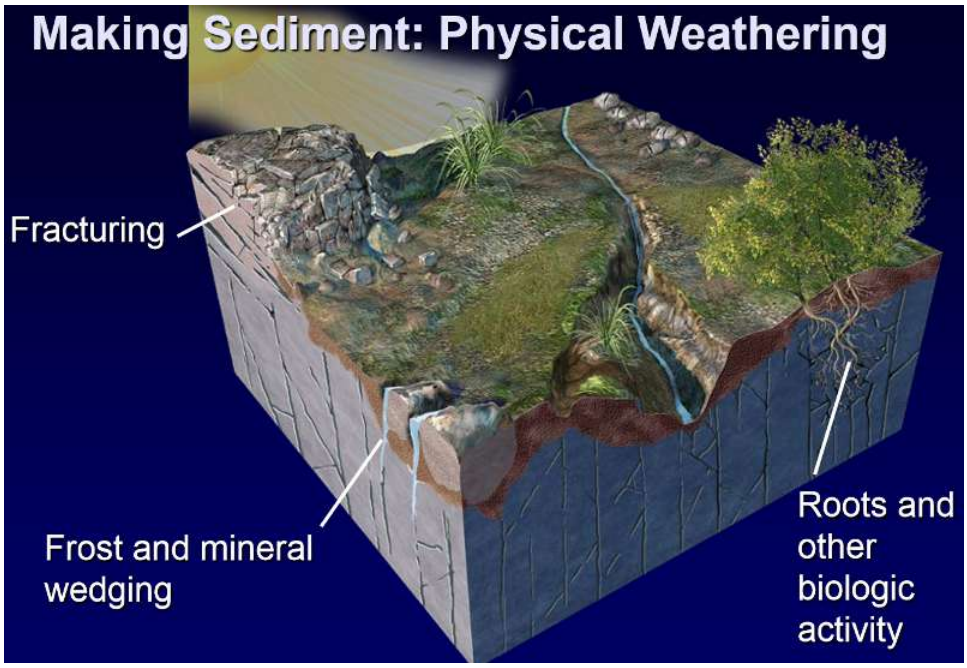


Figure Above: Different processes of physical weathering are shown. These include wedging by frost and minerals, fracturing, and wedging by roots and other biological activity.

Thermal stress

Thermal stress weathering results from the expansion and contraction of rock caused by temperature changes. For example, the heating of rocks by sunlight or fires can cause the expansion of their constituent minerals. As some minerals expand more than others, temperature changes set up differential stresses that eventually cause the rock to crack apart. Because the outer surface of a rock is often warmer or colder than the more protected inner portions, some rocks may weather by exfoliation – the peeling away of outer layers. This process may be sharply accelerated if ice forms in the surface cracks. When water freezes, it expands with a force of about 1465 Mg/m^2, disintegrating huge rock masses and dislodging mineral grains from smaller fragments.

Frost weathering

Frost weathering, frost wedging, ice wedging, or cryofracturing is the collective name for several processes where ice is present. Certain frost-susceptible soils expand or heave upon freezing due to water migrating via capillary action to grow ice lenses near the freezing front. This same phenomenon occurs within pore spaces of rocks. The ice accumulations grow larger as they attract liquid water from the surrounding pores. The ice crystal growth weakens the rocks, which, in time, break up. It is caused by the approximately 10% expansion of ice when water freezes, which can place considerable stress on anything containing the water as it freezes.

Pressure release

In pressure release, also known as unloading, overlying materials (not necessarily rocks) are removed (by erosion or other processes), which causes underlying rocks to expand and fracture parallel to the surface. Intrusive igneous rocks (e.g., granite) are formed deep beneath the Earth's surface. They are under tremendous pressure because of the overlying rock material. When erosion removes the overlying rock material, these intrusive rocks are exposed, and the pressure on them is released. The outer parts of the rocks then tend to expand. The expansion sets up stresses which cause fractures parallel to the rock surface to form. Over time, sheets of rock break away from the exposed rocks along the fractures, a process known as exfoliation. Exfoliation due to pressure release is also known as "sheeting."

Salt-crystal growth

Salt crystallization causes the disintegration of rocks when saline solutions seep into cracks and joints in the rocks and evaporate, leaving salt crystals behind. These salt crystals expand as they are heated up, exerting pressure on the confining rock. The salts which have proved most effective in disintegrating rocks are sodium sulfate, magnesium sulfate, and calcium chloride. Some of

these salts can expand up to three times or even more. It is normally associated with arid climates where strong heating causes strong evaporation and, therefore salt crystallization.

Chemical Weathering

Chemical weathering changes the composition of rocks, often transforming them when water interacts with minerals to create various chemical reactions. Chemical weathering is a gradual and ongoing process as the mineralogy of the rock adjusts to the near-surface environment. New or secondary minerals develop from the original minerals of the rock. More chemical weathering occurs in environments that are wetter and warmer.

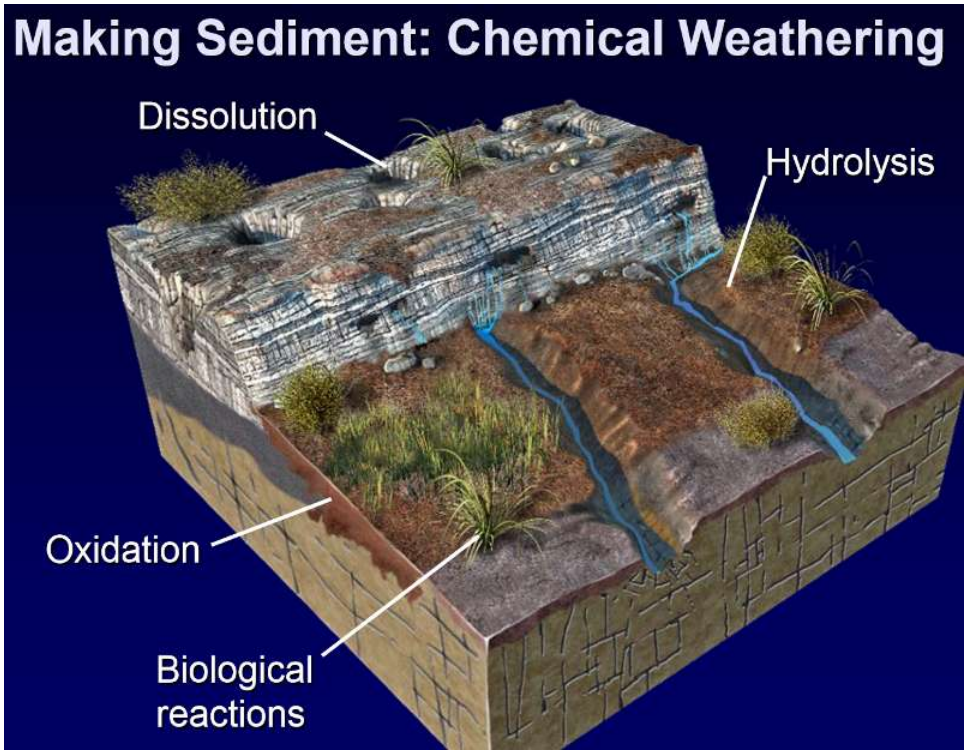


Figure Above: The different processes of chemical weathering are shown. These include oxidation, biological reactions, hydrolysis, and dissolution.

Dissolution/Carbonation

Rainfall is acidic because atmospheric carbon dioxide dissolves in the rainwater producing weak carbonic acid. In unpolluted environments, the rainfall pH is around 5.6. Acid rain occurs when gases such as sulfur dioxide and nitrogen oxides are present in the atmosphere. These oxides react in the rain water to produce stronger acids and can lower the pH to 4.5 or even 3.0.

One of the most well-known solution weathering processes is carbonation, the process in which atmospheric carbon dioxide leads to solution weathering. Carbonation occurs on rocks that contain calcium carbonates, such as limestone and chalk. This occurs when rain combines with carbon dioxide or an organic acid to form a weak carbonic acid that reacts with calcium carbonate (the limestone) and forms calcium bicarbonate.

$CO_2 + H_2O \rightarrow H_2CO_3$

carbon dioxide + water → carbonic acid

$H_2CO_3 + CaCO_3 \rightarrow Ca(HCO_3)_2$

carbonic acid + calcium carbonate → calcium bicarbonate

Carbonation on the surface of well-jointed limestone produces a dissected limestone pavement. This process is most effective along the joints, widening and deepening them.

Hydrolysis (Hydration)

Mineral hydration is a form of chemical weathering that involves the rigid attachment of H+ and OH- ions to the atoms and molecules of a mineral. When rock minerals take up water, the increased volume creates physical stresses within the rock. For example, iron oxides are converted to iron hydroxides, and the hydration of anhydrite forms gypsum.

Oxidation

Within the weathering environment, the chemical oxidation of a variety of metals occurs. The most commonly observed is the oxidation of Fe2+ (iron) and combination with oxygen and water to form Fe3+ hydroxides and oxides such as goethite, limonite, and hematite. This gives the affected rocks a reddish-brown coloration on the surface, which crumbles easily and weakens the rock. This process is better known as 'rusting', though it is distinct from the rusting of metallic iron. This process can occur through the atmosphere or under water.

Biological weathering

Plants and animals may create chemical weathering by releasing acidic compounds, i.e., the effect of moss growing on roofs is classed as weathering. Mineral weathering can also be initiated and/or accelerated by soil microorganisms. Lichens on rocks are thought to increase chemical weathering rates.

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6.2: Sedimentary Environments

The setting in which a sedimentary rock forms is called the sedimentary environment. Every environment has a characteristic combination of geologic processes and circumstances. The type of sediment that is deposited is not only dependent on the sediment that is transported to a place, but also on the environment itself.

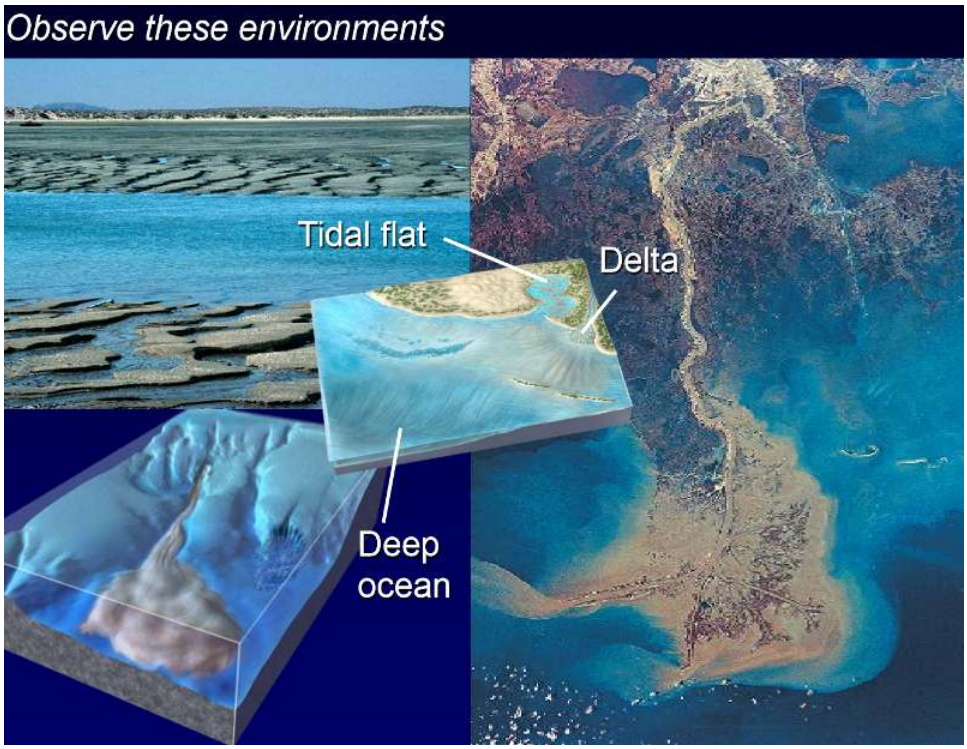


Figure Above: Different sedimentary environments are associated with a stream moving in the ocean. They include a delta where the river fans out (shallow marine), a tidal flat (shallow marine), and the deep ocean.

Marine environment means the rock was formed in a sea or ocean. Often, a distinction is made between deep and shallow marine environments.

- Deep marine usually refers to environments more than 200 m below the water surface. The water current over the sea bottom is small in deep marine environments. Only fine particles can be transported to such places. Typically sediments depositing on the ocean floor are fine clay or small skeletons of micro-organisms.
- Shallow marine environments exist adjacent to coastlines and can extend out to the boundaries of the continental shelf. The water in such environments has generally higher energy than that in deep environments because of wave activity. This means coarser sediment particles can be transported, and the deposited sediment can be coarser than in deep environments. When the available sediment is transported from the continent, an alternation of sand, clay, and silt is deposited. When the continent is far away, the amount of such sediment may be small, and biochemical processes dominate the type of rock that forms. Especially in warm climates, shallow marine environments far offshore mainly see the deposition of carbonate rocks. The shallow, warm water is an ideal habitat for many small organisms that build carbonate skeletons. When these organisms die, their skeletons sink to the bottom, forming a thick layer of calcareous mud that may lithify into limestone. Warm shallow marine environments are also ideal for coral reefs, where the sediment consists mainly of the calcareous skeletons of larger organisms.



Figure Above: More sedimentary environments are depicted. First, there is a gentle, meandering river. Second a river delta, third is a lake, and fourth is wetlands.

Meandering River

- Rocks created: mudstone (shale)

Delta

- Where along a coast (either the coast of a sea or a lake) rivers enter the body of water, deltas can form. Large accumulations of sediment transported from the continent to places in front of the mouth of the river. Deltas are dominantly composed of clastic sediment.
- Rocks created: sandstone, mudstone (shale)

Wetlands, Lake, & Lagoons

- Fine sediment is deposited in the quiet water of swamps, lakes, and lagoons, mingled with organic material from dead plants and animals.
- Rocks created: mudstone (shale), coal.



Figure Above: More shallow marine environments are shown. They include a sandy beach, sandy coastal dunes, a calm lagoon, and a coral reef.

Beach

- The coast is an environment dominated by wave action. At the beach, predominantly coarse sediment like sand or gravel is deposited, often mingled with shell fragments.
- Rocks created: Sandstone, conglomerate.

Coral Reef

- Rocks created: Limestone

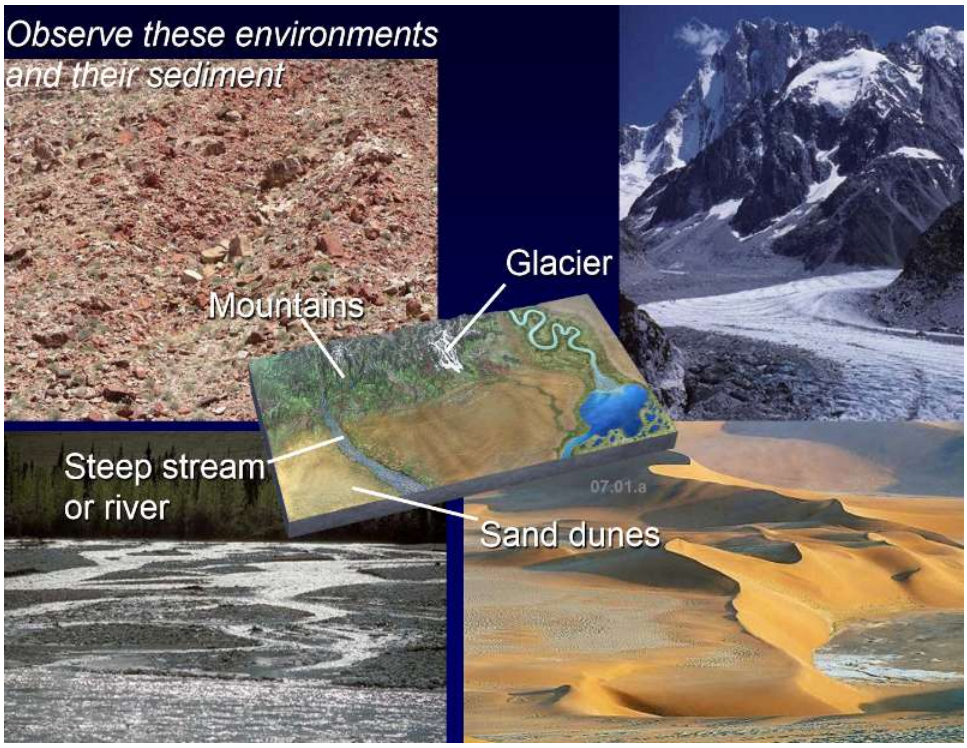


Figure Above: More high-energy sedimentary environments are depicted. They include mountain slopes with rocky, angular sediment, glaciers with angular sediment, steep streams or rivers with large stones, and dunes.

Mountains & Glaciers

- The sediment tends to be angular because it has not traveled far. Sediment transported by a glacier is called glacial and is characterized by very poor sorting.
- Rocks created: Breccia

Rivers

- In rivers, the energy of the water is much higher, and the transported material consists of clastic sediment that is rounded in shape with a large range of clast sizes.
- Rocks created: Conglomerate

Coastal Dunes & Dessert Dunes

- Sediment transported by wind is called Aeolian and is sand-sized and always very well sorted.
- Rocks created: Sandstone.

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6.3: Plastic Deformation

Solid rock exposed to stress can undergo folding, stretching, compression, and bending when there is sufficiently high temperature and pressure with depth. This is called plastic deformation.

First, we will explore the two main types of stress. **Confining pressure** is stress that is applied equally in all directions. **Differential stress** is stress that is applied unevenly.



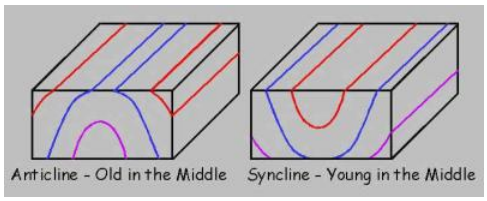
Photo: G. Shuttleworth

In plastic deformation, stretching and bending do not just occur on the scale of individual rocks! You can see folds in the rock layers themselves! This is something that you might have even seen driving around on road trips across America! There are two main types of folds that you learned about already in the video several slides ago: **Anticlines and Synclines.**

Anticlines tend to arch upward, and Synclines arch downward. This isn't all there is to them, however!

The **OLDEST** beds are found at the core of an anticline. The **YOUNGEST** beds are found at the core of a syncline.

Another way to remember: ANTICLINE has an "A" or arch shape. Syncline has a Smile shape.



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6.4: Deformation- Stress and Strain

Tectonic activity (i.e., the movement of the Earth's tectonic plates) has an important effect – it applies **stress** on the surrounding rock of the Earth's crust.

Rocks change as they undergo stress. **Stress** is a force applied to a given area. Since stress is a function of area, changing the area to which stress is applied will change the resulting stress. For example, imagine the stress created at the tip of the heel of a high-heeled shoe and compare it to the bottom of an athletic shoe. In the high-heeled shoe heel, the area is very small, so much stress is concentrated at that point. The stress is more spread out in an athletic shoe. If stress is not concentrated at one point in a rock, the rock is less likely to change (break or bend).

There are three main types of stress: **compression, tension, and shear.** When **compressional** forces are at work, rocks are pushed together. **Tensional** forces operate when rocks pull away from each other. Simple **shear** force is created when rocks move horizontally past each other in opposite directions. Rocks can withstand much more compressional stress than tensional stress.



The Roman Forum. Why did the Romans use so many vertical columns to hold up the one horizontal beam? If the horizontal beam spanned long distances without support, it would buckle under its weight. This beam is experiencing tensional stress, and rocks have very little strength when exposed to such stress.

Applying stress creates a deformation in the rock, known as **strain**. Initially, as rocks are subjected to increased stress which begins the process of strain, they behave **elastically**, meaning they return to their original shape after deformation ceases. This elastic behavior continues until the rocks reach their elastic limit, which will begin to deform plastically.

Plastic deformation may lead to the rocks bending into folds, or if too much strain accumulates, the rocks may behave in a brittle manner and fracture. An example of brittle behavior is a hammer hitting glass, which shatters the glass. With plastic deformation, the rocks do not return to their original shape when the stress is removed. The deformation that results from applied stress depends on many factors, including the type of stress, the type of rock, pressure and temperature conditions (e.g., rocks deeper in the crust will be subject to higher pressures and temperatures), and the length of time the rock is subjected to the stress. Rocks behave very differently at depth than at the surface. Rocks tend to deform in a more plastic manner at depth and in a more brittle manner near Earth's surface. The faster the event occurs (like an earthquake), the more likely rocks will experience brittle deformation.

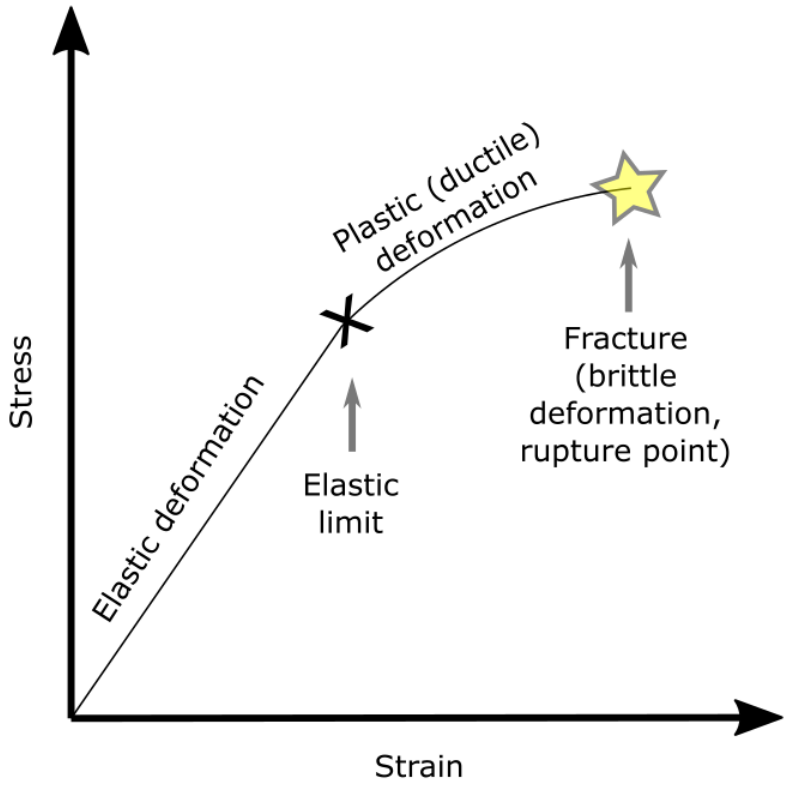


Figure Above: A stress and strain diagram. As stress and strain increase, rocks first experience elastic deformation and will return to their original shape if the stress is released. When the elastic limit is reached (point X), if stress accumulates as strain, the rocks will deform plastically and will not return to their original shape if the stress is released. If rocks are subjected to stress greater than they can accommodate with strain, they will fracture (brittle deformation, yellow star)

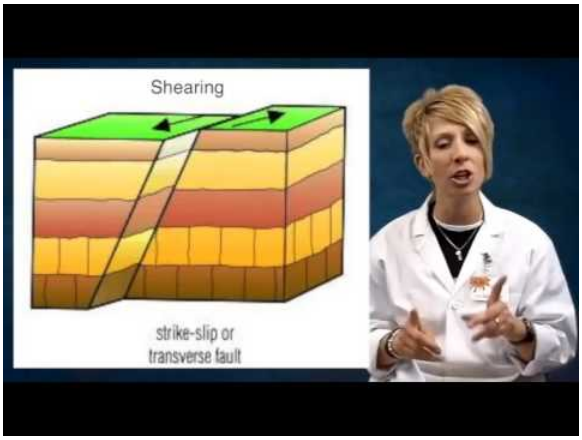
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6.5: Brittle Deformation

As we learned from our lesson on plate tectonics, the Earth's crust is constantly in motion. This motion can lead to the collision of mountains, the rifting of valleys, and the shearing of the surface. These processes cause **deformation** on our surface.

There are two main types of deformation that we can observe on Earth's surface: **brittle deformation** and **plastic deformation.** Although they are both consequences of similar processes, they occur at different temperatures and pressures, and as a result, these two types of deformation result in very different-looking features!

Please watch the video below for a summary of all these features: folds and faults!



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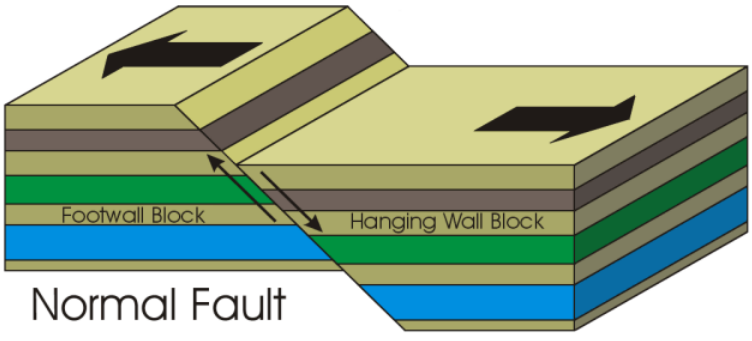
Brittle Deformation

When enough stress is applied to a rock or layer **not** subjected to significantly high temperatures and pressures, the rock can fracture along a point of weakness. This fracture is called brittle deformation. This type of deformation occurs commonly in the upper crust in the form of fractures, joints, and faulting. Faults involve the physical dislocation of rock layers along a shear plane or plane of weakness. Below are the classifications of faults:

Dip-Slip Faults

These faults move vertically. Two types of dip-slip faults include:

Normal Faults: These faults occur when blocks of rock are pulled apart.



Normal Fault

Reverse or Thrust Faults: These occur when blocks of rock are pushed together.

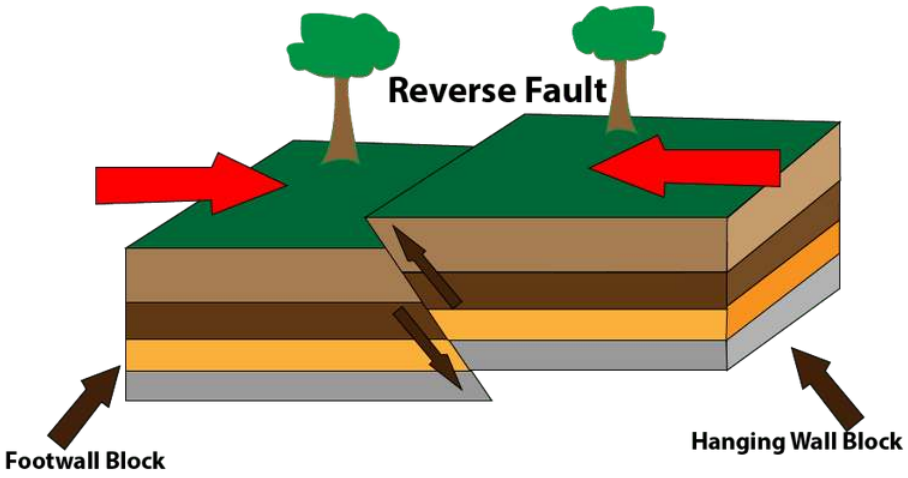


Image: InventionSky.com

Another terminology related to dip-slip faults is the **footwall**, which is the underlying block on the fault. The **hanging wall** is the overlying block on the fault.

Strike-Slip Faults

These faults move horizontally. Two types of strike-slip faults include:

Left-lateral Faults: If you stand on the fault line of this fault, the left block would move toward you, and the right would move away.

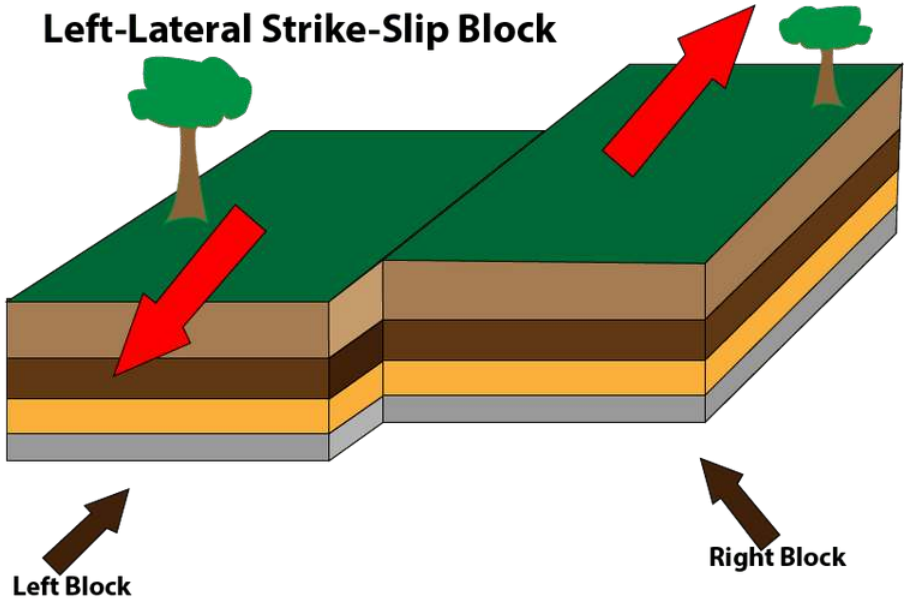


Image: InventionSky.com

Right-Lateral Faults: If you stand on the fault line of this fault, the right block will move toward you, and the left will move away.

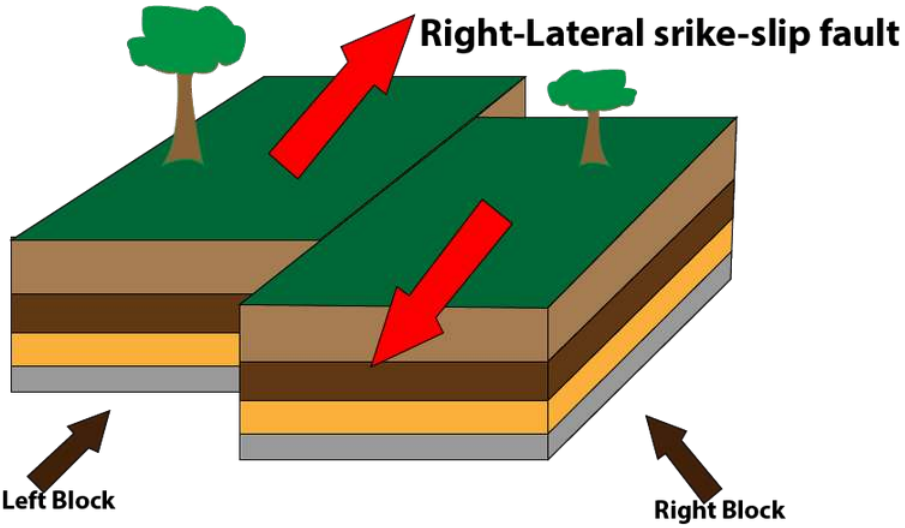


Image: InventionSky.com

Questions: Brittle Deformation



Image Source: SEG Wiki

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CHAPTER OVERVIEW

7: Geologic and Topographic Mapping

7.1: Basic Map Features

7.2: Latitude and Longitude

7.3: Contour Lines

7.4: Geologic Map Units

7.5: Mapping Geologic Structures

Thumbnail: An example of use of layers in a GIS application. In this example, the forest-cover layer (light green) forms the bottom layer, with the topographic layer (contour lines) over it. Next up is a standing water layer (pond, lake) and then a flowing water layer (stream, river), followed by the boundary layer and finally the road layer on top. The order is very important in order to properly display the final result. Note that the ponds are layered under the streams, so that a stream line can be seen overlying one of the ponds. (CC BY 3.0; Jaknouse via Wikipedia)

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7.1: Basic Map Features

Often topographic maps are superimposed on a geologic map, so the features you need to look for in both are often the same. The basic information available at the top, bottom, and sides of a map can help you interpret these maps easily.

Title and Legend

The title of a topographic map is usually named after a major geographic feature within the map area, such as the Bright Angel Quadrangle, Arizona. On most U.S. Geological Survey (U.S.G.S.) maps, the title is found along both the upper and lower margins of the right-hand side of the map. The legend of a map is the key that explains the symbols used on the map. All U.S.G.S. topographic maps use the same symbols. Legends on topographic maps are usually limited to various symbols used for road classification.

Author(s)

All maps will at least have these listed, usually with the title or at the bottom.

Spatial coordinate markings or grids

Many maps will include latitude/longitude (Figure 10.1) or UTM information for precise locations on the map. Latitude and longitude coordinates on topographic maps and global positioning system (GPS) devices have different formats but are most commonly given in degrees, minutes, and seconds (DMS).

1 degree (°) = 60 minutes (‘) 1 minute (‘) = 60 seconds (“)

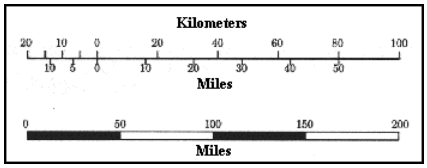
Topographic Map Base

We will dive more into this in another section, but this tells us what elevation is represented by each contour line in a topographic map so we can interpret elevation changes and easily find steep areas, hills, and valleys. The amount of elevation between contour lines is referred to as the **contour interval** and is usually printed on the map.

Scale

The scale of a map gives us the relationship between the true ground distance and the corresponding distance on the map. Topographic maps use three types of scale: ratio, verbal, and graphic. As part of this exercise, you will learn to convert from one scale to another.

- Verbal Scale—A verbal map scale gives the scale in words. A verbal scale in the metric system might be 1 cm = 4 km, meaning one centimeter on the map represents 4 kilometers on the ground. A verbal scale in the English system might be something like 1 in = 5 mi., meaning one inch on the map represents 5 miles on the ground. You will not find a verbal scale on most topographic maps. Therefore, you must be able to calculate the verbal scale from the fractional scale.
- Graphic Scale —A graphic, or bar, scale is shown as a line or bar divided into units representing distances on the ground. The distance between two points on a map can easily be measured with a ruler, then you use the graphic scale to determine the true ground distance this map distance represents. Graphic scales are always found on topographic maps along the lower margin. Maps usually have several graphic scales, showing several different units in the English and metric systems. If a map is enlarged or reduced, the bar scale will also be enlarged or reduced by the same amount.



- Ratio Scale —The fractional scale of a map is the ratio between one unit distance on the map and the corresponding distance on the ground. For example, if the fractional scale of a map is 1:50,000, it means that one unit on the map corresponds to 50,000 *of the same units* on the ground. The units can be either in the English or metric system. In the metric system, a scale of 1:50,000 would mean one centimeter on the map corresponds to 50,000 centimeters on the ground. Using the English system, one inch on the map would correspond to 50,000 inches on the ground. Any unit of measurement may be used, but the units *must* be the same. Fractional scales are found at the lower margin of the map, usually near the graphic scales.

- **Scale Conversion**Working with topographic maps requires you to convert from one type of scale to another.Common conversion units:

1 meter (m) =	100 centimeters (m)
1 kilometer(km) =	1000 meters (m)
1 inch (in) =	2.54 centimeters (cm)
1 foot (ft)=	12 inches (in)
1 mile (mi) =	5280 feet (ft)

To convert from a ratio to a verbal scale, simply divide by the conversion factors, and use these steps:

1. List the ratio scale. For example 1:48000
2. Convert to a verbal scale by turning the larger number into meters/kilometers/or miles.

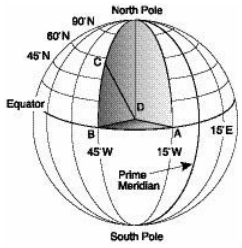
On the map	On the ground
1 unit =	48000 units
1 cm =	48000 cm ÷ 100 = 480 m 48000 cm ÷ 100000 = 0.48 km
1 in =	48000 in ÷ 12 = 4000 ft
1 in =	48000 in ÷ 12 ÷ 5280 = 0.76 mi

If you need more practice or help, go to The Math You Need When You Need It.

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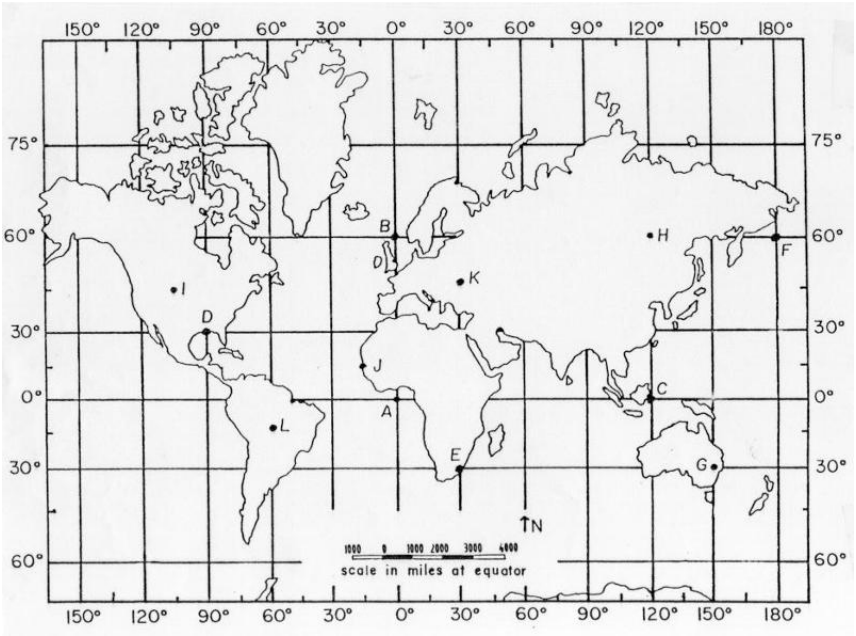
7.2: Latitude and Longitude

The surface of the Earth is divided into a grid system of two sets of reference lines. Lines running east-west are called lines of latitude or parallels; those running north-south are called lines of longitude, or meridians. The intersection of latitude and longitude lines describes any point on the Earth's surface. Latitude and longitude are obtained from the angle of lines passing through poles and the Earth's center. Latitude and longitude are measured in degrees (°), minutes (‘), and seconds (“). One degree (1°) = 60 minutes. One minute = 60 seconds. Therefore a point on Earth would be as 98° 45′ 30″ West Longitude, 42° 30’ 15” South Latitude.

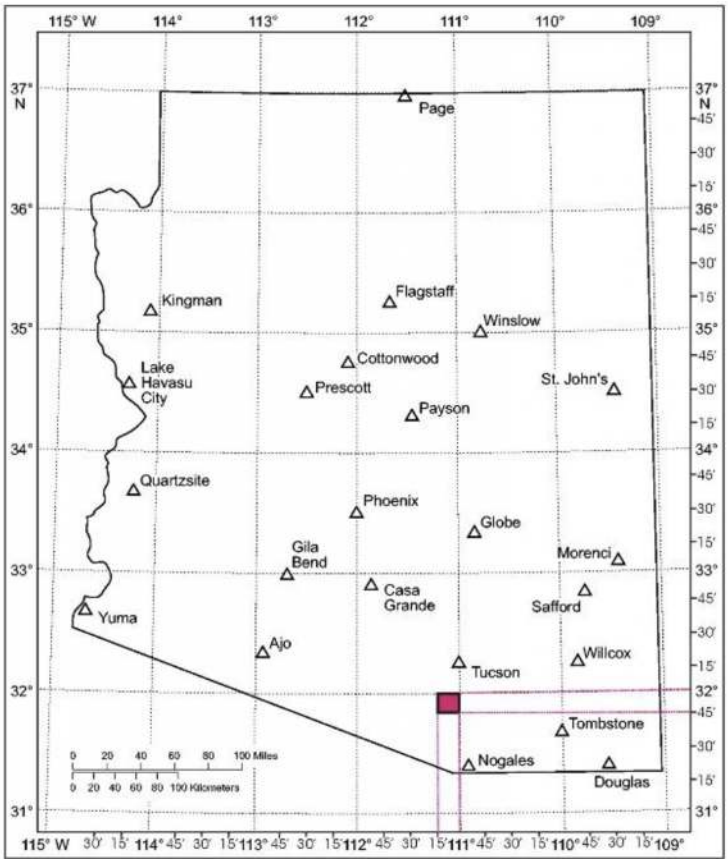


Latitude is measured from a 0° reference line that divides the Earth into a northern and a southern hemisphere. The 0° line of latitude is called the equator. Since lines of latitude run east-west, they will be measured from 0° (the equator) to 90° (the poles) in degrees either north or south of the equator.

Longitude is measured from a 0° reference line that passes through Greenwich, England. The 0° line of longitude is called the Prime Meridian. On the opposite side of the Earth, from 0° longitude, will be the 180° line of longitude, called the International Date Line. The Prime Meridian and the International Date Line divide the Earth into the western and eastern hemispheres. Because longitude lines run north-south, they will be measured in degrees either east or west of the Prime Meridian, from 0° to 180°. Since North America is north of the equator and west of the Prime Meridian, all lines of latitude in the continental U.S. are north, and all lines of longitude are west.



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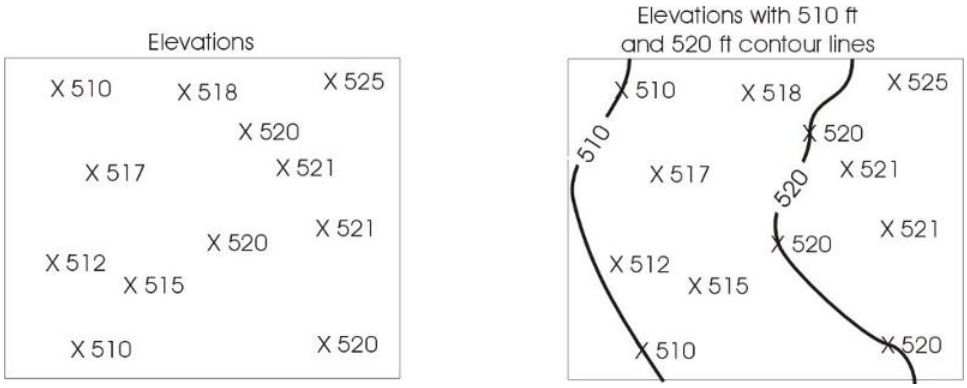
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7.3: Contour Lines

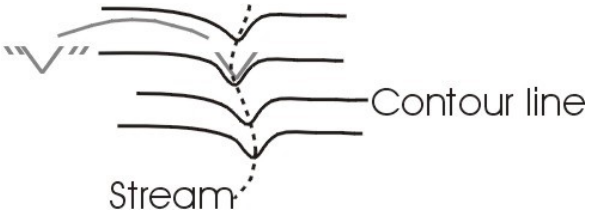
Contour lines are what allow topographic maps to show the three-dimensional configuration (i.e., topography) of the Earth's surface. Sea level is accepted worldwide as the zero contour line, so that elevation measurements are given in terms of either feet or meters above or below sea level.

Rules for Contour lines:

1. Every point on a contour line is the same elevation; **contour lines** connect points of equal elevation.

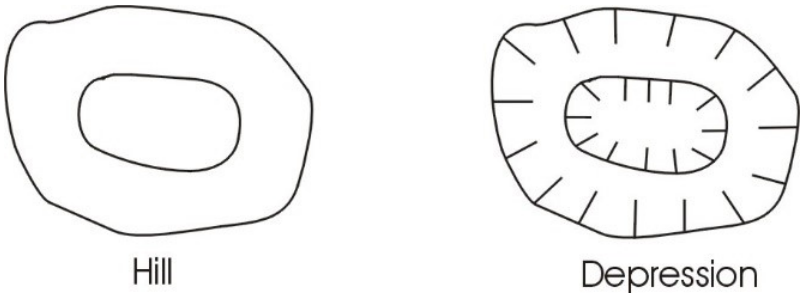


2. Contour lines are always a uniform vertical distance apart, called a **contour interval**, which is stated on the map (bottom middle).
3. Contour lines never branch, fork, or cross one another.
4. Contour lines form a "V" when crossing a stream valley; the bottom of the "V" always points upstream, so the water flows in the opposite direction.



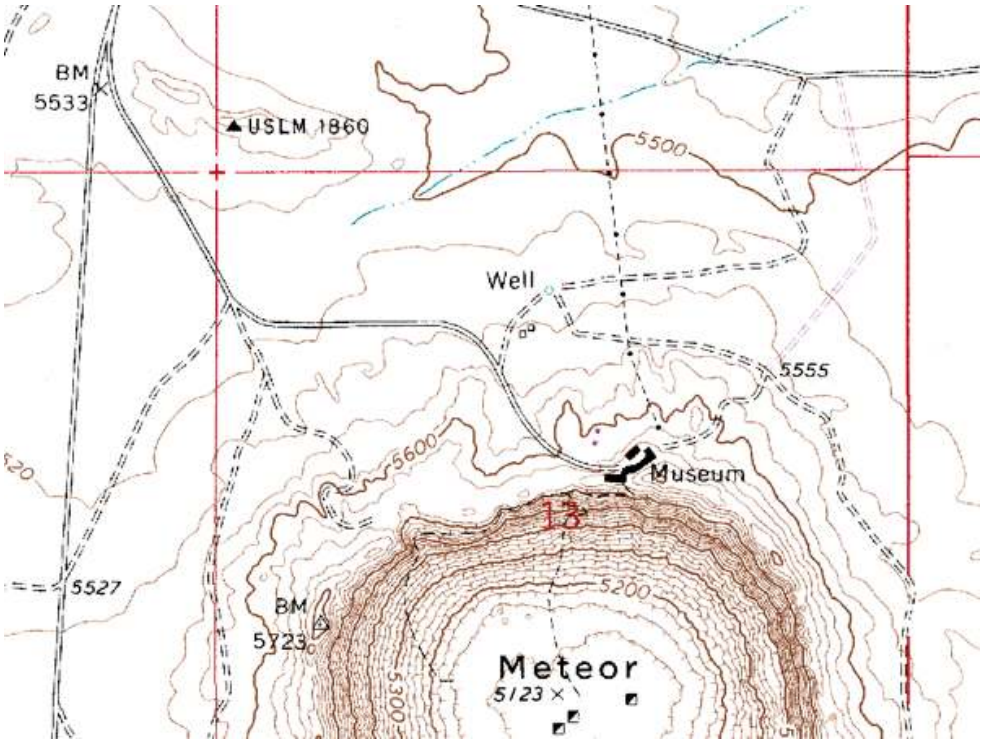
The V points upstream, south in this case. So the stream is flowing north.

5. Every 5th contour line, called an **index contour**, is drawn darker to aid in counting the lines.
6. Dotted contour lines, called **supplementary contours**, are used to show more detail in an area of little relief (areas that are flatter).
7. **Hachured** contour lines indicate a closed depression. The hachures are located on the downhill side.

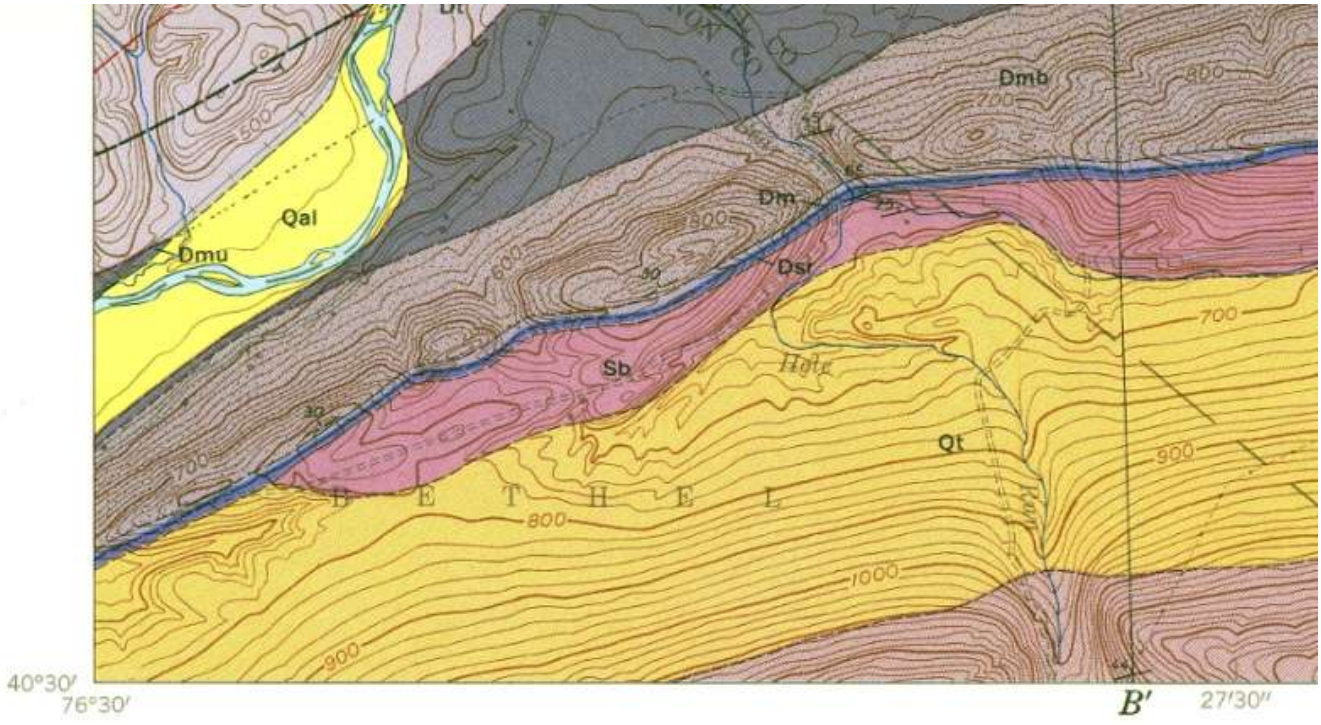


8. Closely spaced contour lines represent a steep slope (higher gradient), while widely spaced contour lines represent a gentle slope (lower gradient).

The image below is of a portion of a topographic map, answer the following questions using this map.



The Stream, Bear Hole Run, has many contour lines that cross it. Where they cross, the contour lines form Vs. The stream flows in the opposite direction of the stream.



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7.4: Geologic Map Units

The core data of any geologic map lies in the distribution of map units, each having distinctive physical characteristics. A fundamental map unit is a **Formation**, a rock unit with unique characteristics (such as color, texture, rock type, and/or age) that can be distinguished from other units in the field. A formation must be mappable, covering a large enough area to be shown on a geologic map. A lesser unit is called a **Member**, and several members combined can make a formation. Alternatively, several similar formations can be joined together into a larger **Group**, and groups can likewise be joined into a **Supergroup**. Rock formations are typically named after the locations (type localities) where the rock unit was first described or is best represented but can also refer to the composition of the rock unit.

Contacts delineate map unit boundaries. Solid line contacts mean that the location of the map unit boundary is relatively sharp or well-located. In contrast, a dashed contact may imply a gradational or less well-located boundary. Dotted contacts represent the inferred location of unit boundaries covered by overlying material. Other types of contacts depend on the geologic features present (like faults and intrusions). Contacts are not contour lines (which are related to elevation).

Map unit symbols

Map units always have labels or symbols to identify them. The exact symbols used on different maps can vary widely. Still, with most map unit symbols, the first letter is capitalized and typically represents the formation's geologic age (Figure 10.4). In contrast, the second (**lower case**) letter describes the formation name, rock composition, etc. For example, the Clinton Formation has the map unit symbol St, where "S" is the age of the formation (Silurian period, 438-408 million years ago), and "c" is for Clinton (named for a town in New York where several well-exposed outcrops of these rocks can be found). The symbols are printed on the geologic map and in the map unit explanation.

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			

Explanation of map units

All maps will have a description (names, ages, symbols, etc.) of the geologic map units, so the reader knows what the different units represent. Map units are listed from **oldest (at the bottom)** to **youngest (at the top)**. Brief descriptions of the map units and special map symbols are commonly present. Look at the explanation if you see a symbol on the map you do not recognize!

Correlation of map units

Maps may include a diagram showing the age relations of the various map units.

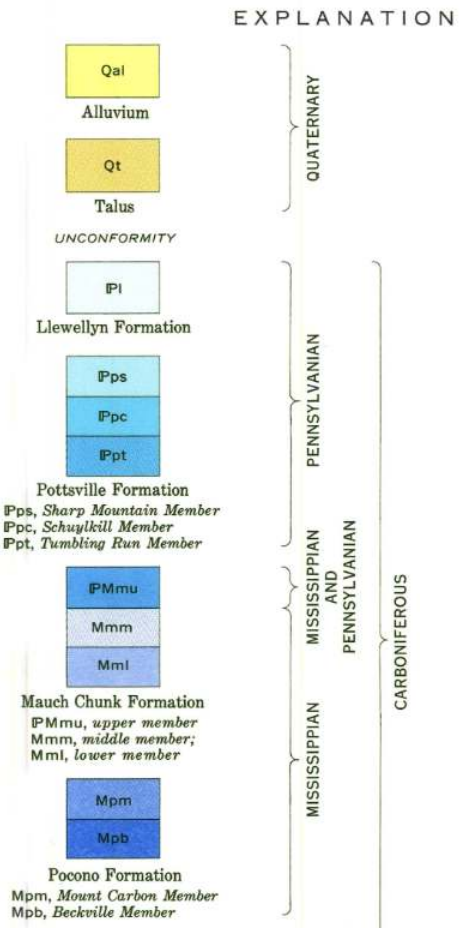
Geologic Cross Sections

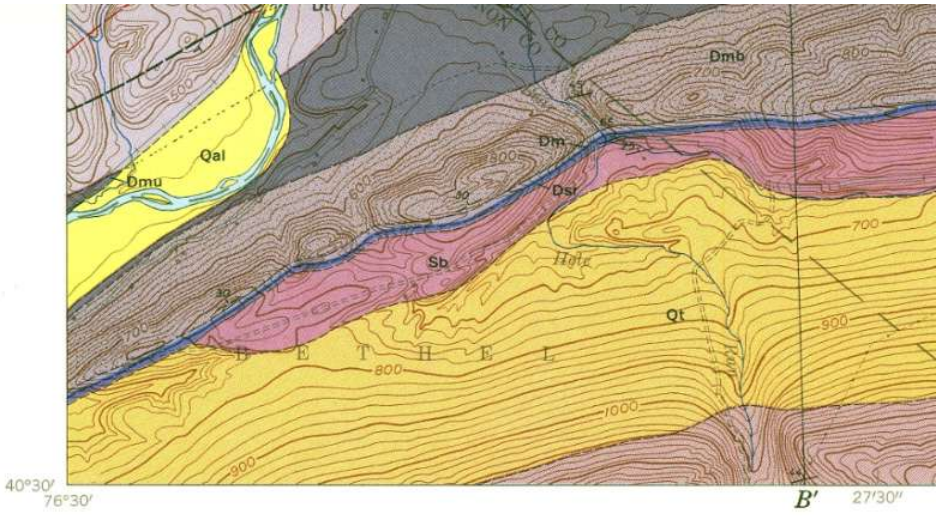
Maps may include one or more geologic cross sections. Topographic profiles that interpret the subsurface's geology are inferred from the orientation of rock units at the surface). These figures can be highly interpretive but are very helpful in visualizing the distribution of rock units below the surface. The locations of cross-section lines or endpoints are marked on the map. Cross sections may be drawn with some vertical exaggeration better to show the changes in topography across the section line.

Other data

Other map data may include photo imagery, map location, data tables, stratigraphic columns, etc.

Using the following portions of the map, answer the questions:



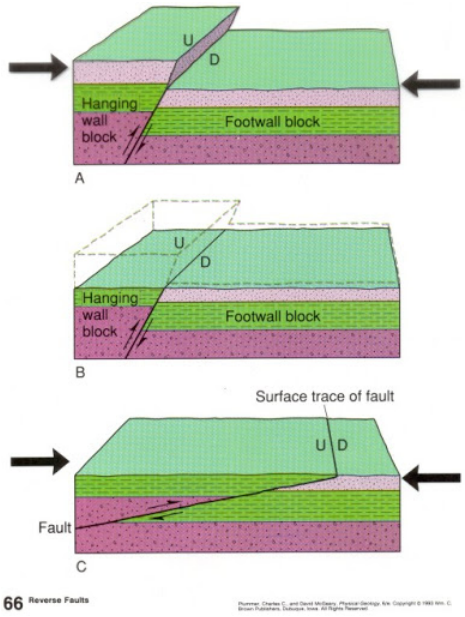


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7.5: Mapping Geologic Structures

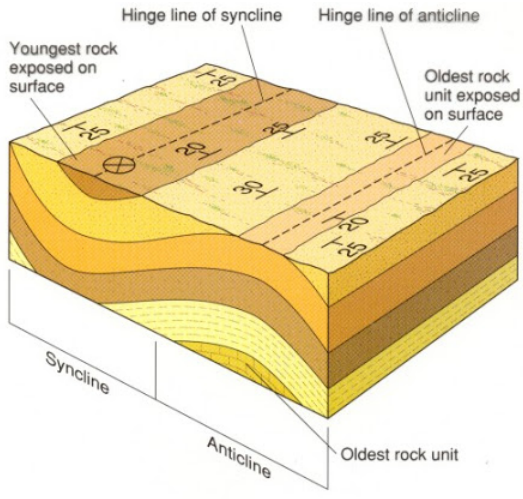
Below the surface of our planet, rocks experience substantial forces. These forces include **compression** (→ ←), **extension** (← →), and **shearing** (Yy). All of these forces deform rock. Not only can the forces bend rocks (creating a **fold**), they can even break them (called a **fault**). These forces help create the structure we observe in rocks.

Faults. Faults are breaks in rock layers where the movement has occurred. They are shown on a geologic map by black lines. The lines are solid where the fault is evident. The lines are dashed where the fault is not visible at the surface and must be inferred by a geologist. Most faults are roughly planes and, therefore, can be described using strike (direction) and dip (angle down from horizontal). Sometimes the side of the fault moved upward (U) or downward (D) relative to the other side is labeled.



66 Reverse Faults

Folds. Folds occur due to compression of ductile rocks. In sedimentary rocks, the axes of folds are also shown on geologic maps as lines. An *anticline* is drawn as a long line straddled by a two-headed arrow in which the arrowheads point away from one another. A *syncline* is drawn the same way with the arrowheads pointing toward one another. In rare cases, where sedimentary beds have been fully overturned by folding, a U-shaped symbol is used to give the strike and dip of the bed.



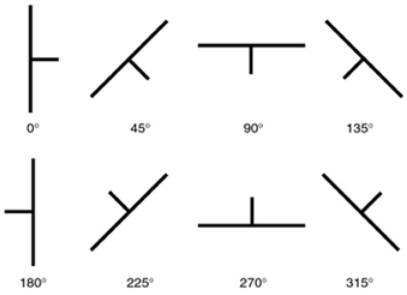
60 Block Diagram of Anticline and Syncline

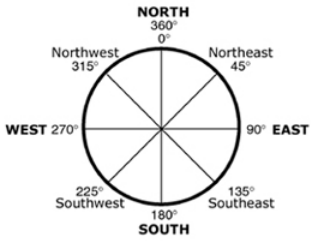
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Strike and Dip

When Earth materials undergo **stress**, the result is **strain** or **deformation**. Strain may vary from slightly warping of a sedimentary bed and complex flow textures in a metamorphic rock to the rupture of solid rocks along a fault. Geologists often have to visualize geologic structures in three dimensions (3-D) to understand how rocks got into their present position. They usually have to work from the limited clues they see on Earth's surface (or on geologic maps that depict the limited clues). Geologic maps show the rocks at Earth's surface and near-surface, not deep underground. They depict the area where each rock unit is exposed, how each is oriented, and where faulting occurs.

Small strike-and-dip symbols show Rock orientation on a map within the colored area for a given rock. These symbols indicate the general orientation of the units and whether a bed is horizontal or tilted.





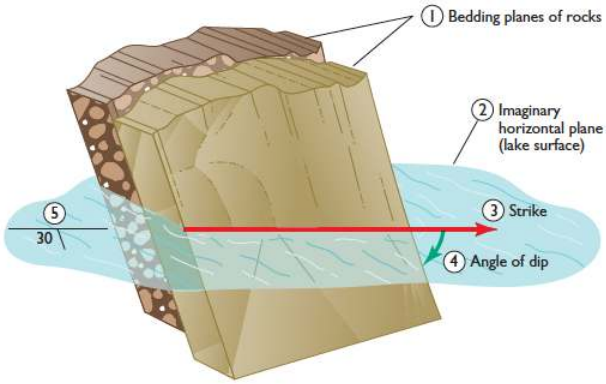
Consider the sandstone bed ① in the figure below. This sandstone was deposited horizontally but later was tilted to the angle you see in the figure. The original surfaces of the sandstone layer from its **bedding planes**.

Add a horizontal plane to show how the rocks were originally oriented. This can be thought of as the surface of a lake ② .

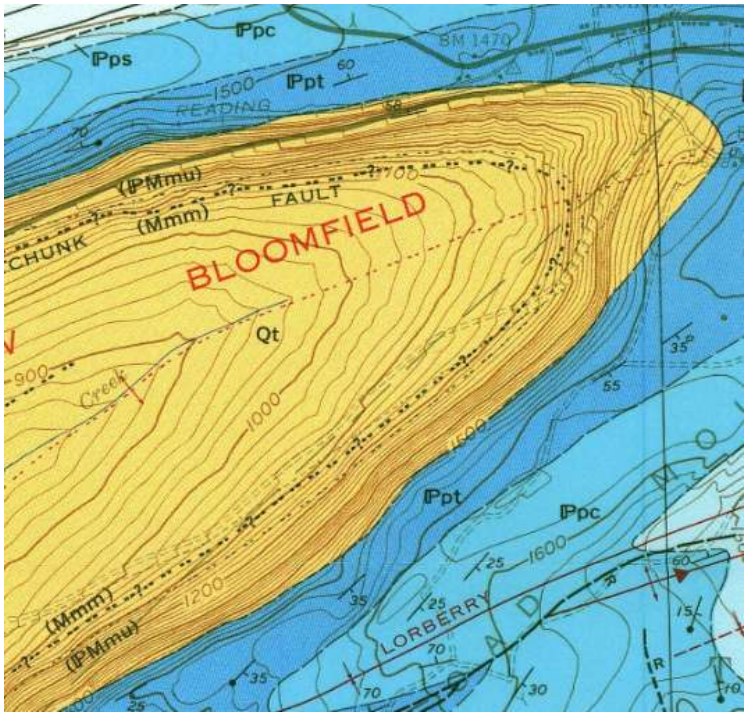
The imaginary line where the horizontal plane intersects the rock's bedding plane (the red arrow) is called the **strike** ③ . The strike line has a compass direction on Earth's surface (like northeast).

The dip angle indicates how steeply the sedimentary bed dips relative to the horizontal plane. The dip is measured *downward* from the horizontal plane (lake surface), as shown by the green arrow ④ . This angle is expressed in degrees (for example, 30°).

The strike and dip describe the three-dimensional orientation of the sedimentary bed. This is recorded on a geologic map using the strike-and-dip symbol ⑤ . The longer line is the strike direction, the shorter line is the dip direction, and the number is the dip angle in degrees (down from the horizontal).



Looking back at the Pine Grove map:



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CHAPTER OVERVIEW

8: Taphonomy - The Science of Death and Decay

8.1: What Is A Fossil?

8.2: Modes of Fossil Preservation

8.3: Trace Fossils- Studies in Scurrying, Scraping, and Slithering

Thumbnail: A model of a fossil of a giant toothed platypus from the Riversleigh fossil fields. About a metre long and more than twice the size of platypus today. (Unsplash License; David Clode via Unsplash)

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8.1: What Is A Fossil?



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Fossils

Fossils are evidence of ancient life. Fossils are found in sedimentary rocks. Here are ways in which they are preserved (We will go into depth about these in the next chapter):

1. Shells and coral – the shells of animals can be preserved in the rock. This is common in limestone formed at coral reefs.
2. Bones – the bones of animals are hard enough to be preserved in soft sedimentary rocks such as mudstone (shale), sandstone, and limestone.
3. Petrified wood by mineral replacement – some materials, such as the carbon in wood, can be replaced by other minerals such as silica (SiO_2). The original structure can still be seen after the replacement process.
4. Impression – some materials can leave behind just the impression. Leaves leave behind a carbon film, while bony fish can leave the entire fish impression behind.
5. Constructed mounds left behind by the organism. – some animals create structures, like an ant hill, that can be preserved.
6. Molds or casts – are cavities that can be filled, forming the original shape of the missing material.

7. Amber (tree sap) – organisms and material can get trapped in sab or tar and be preserved.



Figure Above: How are fossils preserved? They are preserved as shells and hard parts, bones, through replacement, as a cast or mold, as a thin carbon film, as impressions, within amber, and through constructed features left by the organism.

Most creatures are never preserved because preservation usually requires hard parts and rapid burial after death. If a creature remains on the surface, it will physically or chemically decompose or can be eaten by predators. Preservation is easier beneath the sea than on land because objects tend to get buried more rapidly, and it's a low oxygen environment.

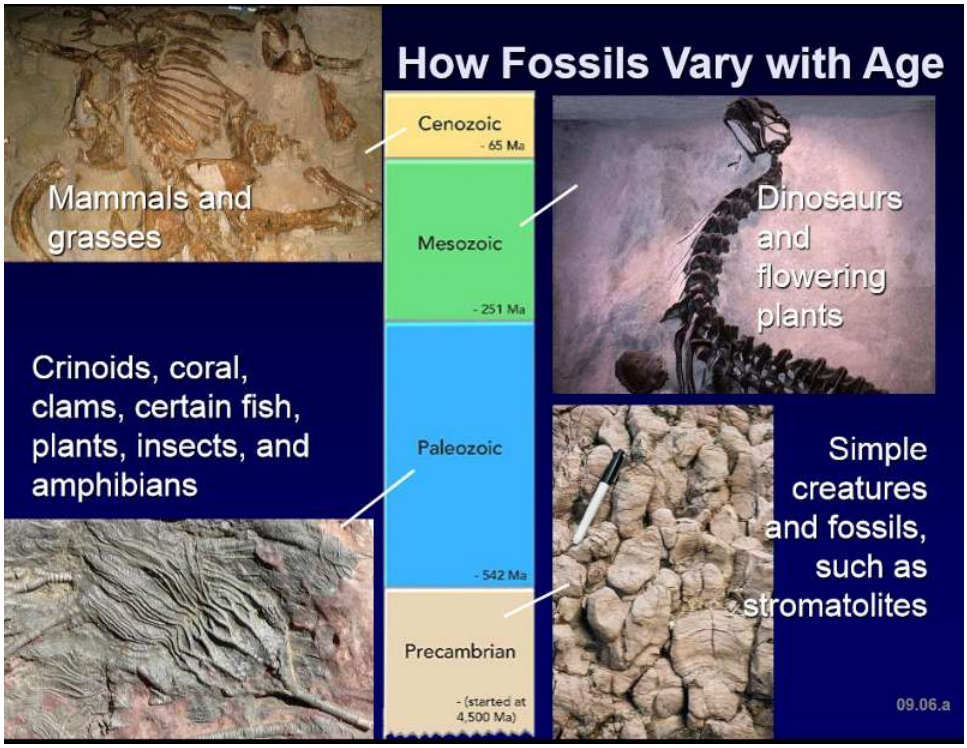


Figure Above: How fossils vary with age. Mammals and grasses are from the Cenozoic (65 Myr to the present). Dinosaurs and flowering plants are from the Mesozoic (251 to 65 Myr). Crinoids, coral, clams, some fish, plants, insects, and amphibians are from

the Paleozoic (542 to 251 Myr). Simple creatures and fossils such as bacteria and stromatolites are from the Precambrian (life started around 3.5 billion years ago through 542 Myr)

- Cenozoic: (Recent Life) The age of mammals because they became the dominant life form.
- Mesozoic: (Middle Life) The age of dinosaurs because they dominated during this time. The end of the Mesozoic marks the extinction of the dinosaurs.
- Paleozoic: (Ancient Life) The age of fish. Hard-shelled invertebrates developed along with vertebrates. During this time, plants, insects, amphibians, and reptiles came into existence. At the end is the Great dying.
- Precambrian: The largest of the periods. Ninety percent of earth's history is here. Only simple life forms such as bacteria, algae, and stromatolites exist.

What factors determine whether a species survives or becomes extinct? Almost all species that ever existed are now extinct. Below is a list of why species become extinct.

- Environmental Niche: Animals and plants have certain environmental settings (niches). Organisms can sometimes take advantage of new changes to niches, but others are very picky. If that niche becomes threatened due to changes in the organism population or climate, the creature will become extinct.
- Climate and Habitat: Changes in climate or environmental disease can wipe out animals. Earth history is marked by extreme changes in climate and species loss.
- Reproductive Strategy: How a plant or animal reproduces can be advantageous– or not depending on how the environment changes. For example, some creatures, like Pandas today, have trouble reproducing naturally.
- Adaptability: The more adaptable a species is, the more likely it will survive changes. Some species are not adaptable to climate changes because they are specifically evolved for certain climates (i.e., consider polar bears in the face of climate change).
- Competition: Competition among species for sparse resources can lead to life or death.
- Predators and Prey: Becoming a new food source for some other creature is never good!

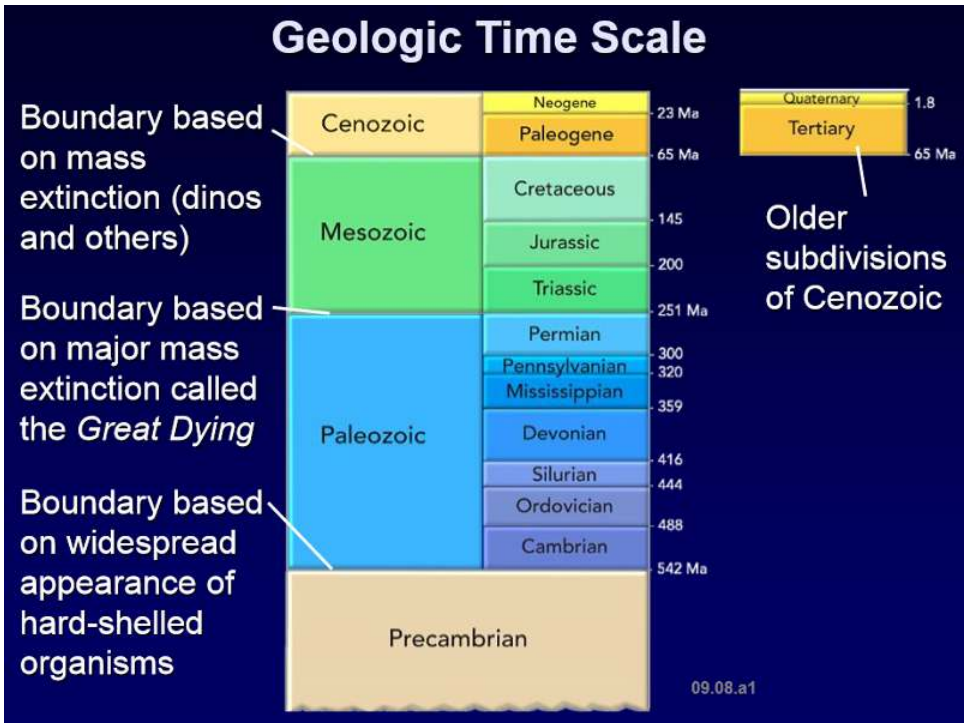


Figure Above: The Geologic Time Scale. See below for detail on the three main eras.

Mesozoic – Cenozoic: This boundary was based on the mass extinction of the dinosaurs at the end of the Mesozoic. The boundary is called the KT boundary because it separates the Cretaceous (K) and Tertiary periods.

Paleozoic – Mesozoic: This boundary was based on a mass extinction called the Great Dying, which occurred at the end of the Paleozoic. 1/3 of all the marine life died.

Precambrian – Paleozoic: This boundary was based on the widespread appearance of hard-shelled organisms at the beginning of the Cambrian period.

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8.2: Modes of Fossil Preservation

Unaltered remains



Wooly mammoth (Mammathus primigenius) hair.

An amazing diversity of fossils exist in a relatively unaltered state. While there may be some compression of the fossil material due to the burden of burial sediments above, or separation of individual shells or bones, unaltered remains are skeletal components preserved in their original mineral composition. It is important to have a general sense of what the expected mineralogy of a fossil organism should be to determine whether another mode of preservation could apply to the specimen.

Common fossil mineralogy	**Common organisms**
Calcite ($CaCO_3$)	Foraminifera, coccolithophores, sponges, tabulate and rugose corals, brachiopods, bryozoans, trilobites, mollusks, echinoderms.
Aragonite ($CaCO_3$)	Modern (scleractinian) corals, mollusks
Quartz (SiO_2)	Radiolarians, diatoms, sponges
Hydroxyapatite $Ca_5(PO_4)_3$	Vertebrate bone and teeth

Soft tissues, such as skin, feathers, or fur, may be preserved in some exceptional cases. This type of preservation does not typically occur within sediments but in materials like amber or glacial ice.

Permineralization

Permineralization occurs when open spaces within porous tissues are filled in with mineral crystals after the skeletal material of the fossil has been buried. Groundwater containing dissolved minerals passes through the buried fossil, and minerals grow in the available space when that water evaporates. Typical infilling minerals include quartz, calcite, and iron oxides, all common cementing minerals for sedimentary rocks.



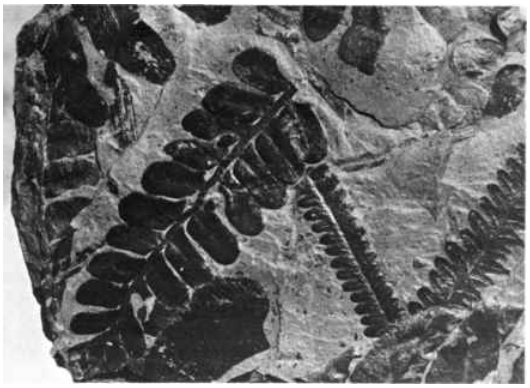
Permineralized wood from the late-Cretaceous Hell Creek Formation of eastern Montana. Note the preservation of the fine internal growth rings. Yellow pencil provides a sense of scale; each black bar on the pencil is 1 cm long.

Permineralization is a common mode of preservation for vertebrate bones and wood but can also occur in some shells. In the case of petrifaction (“petrified” wood), minerals fill the porous spaces in the plant tissues and replace the organic material of trees.

Permineralized fossils maintain a high degree of detail, as they preserve both internal structures and the three-dimensional shape of the original organism. Because crystal growth occurs in the prior open spaces within the skeletal material, permineralized fossils tend to be denser and heavier than their original counterparts.

Carbonization

When soft tissues are quickly buried in quiet water, they are subjected to increased pressure and rising temperatures as part of the lithification process. These conditions affect the remaining organic material, resulting in a process that releases chemically volatile compounds. When the volatiles are released, a dark-colored film of chemically stable carbon is left behind. Intricate details of delicate soft tissues may be preserved, including individual cell structure in plants, fish scales, insect wings, and even internal organs.



Plant fossil found near coal seams from the Pennsylvanian period; Terre Haute, Indiana.

Replacement



Pyritized trilobite, Triarthrus eatoni, Upper Ordovician, New York. The mineral pyrite, FeS2, has replaced the hard parts that made up the trilobite.

The increased pressure and temperature conditions in sediments' deep burial liberate trapped fluids containing mobile chemical ions. The fluids interact with buried fossils and may completely dissolve the original mineral or organic material, replacing it with different ions that precipitate from the circulating fluids. Common replacement minerals include silica and pyrite. The decay of organic material requires bacterial support, which can reduce sulfur and precipitate the iron sulfide mineral, pyrite. Often, the original organism's detail can be preserved via replacement.

Recrystallization

Recrystallization is an important process that preserves fossils made of aragonite, a common but chemically unstable mineral over geologic time. Many marine invertebrates (molluscs, modern corals) have a metabolism that allows them to crystallize aragonite as their skeletal material. Aragonite is a polymorph of $CaCO_3$ (meaning that it and a few other minerals are made of this material), and it sometimes appears as a beautiful "mother of pearl" on the inside shells.



Recrystallized ammonite, shown split open. These are typically sold to amateur and professional collectors alike.

But outside the organism's living tissue, aragonite is less chemically stable than calcite, so it spontaneously recrystallizes to make calcite after burial. This calcite "version" of the shell is more likely to last long enough to enter the geological record and be discovered by geologists. There is no change to the chemical formula of the mineral during recrystallization, just the same atoms reconfiguring into a more stable crystal structure. This preserves the overall shape of the fossil, but recrystallization can destroy some finer details as new crystals form.

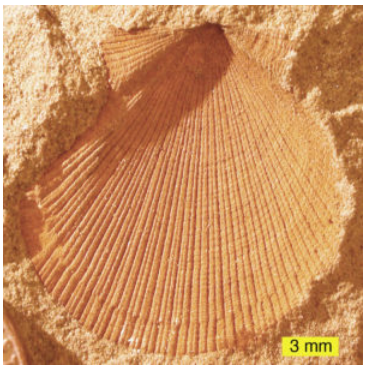
Internal molds



An internal mold of a snail. Note shell is dissolved away, and only infilling sediment remains.

When the interior space of an organism is buried and filled with sediment that is then lithified, the interior shape is preserved as an internal mold. Imagine making a sand castle at the beach, filling a plastic bucket with wet sand, then inverting it onto the ground and lifting the bucket away. The shape of the inside of the bucket is preserved through the bucket itself is gone–that's an internal mold. In geology, this preservation mode is usually found in bivalves, brachiopods, or snails with a single open chamber where sediment can get in and work its way into the entire space. This preservation mode can reveal lovely impressions of internal features like muscle scars. Still, since most mollusks (i.e., bivalves like clams) are more readily identified using external ornamentation, the loss of the original shell can make things tricky for paleontologists.

External molds



External mold of a bivalve from the Logan Formation, Lower Carboniferous, Ohio

An external mold is an impression of an organism's outside shape. This preservation commonly occurs when a fossil is pressed into soft sediment. The original hard parts of the organism may then be transported away by currents, or after burial and lithification, the original hard parts may dissolve. When external molds are discovered, the observer sees a version of the exterior of that organism, but it is not a perfect replica because it shows negative relief. Negative relief means that on the impression, raised areas represent indentations on the original fossil, while depressed areas represent raised areas. If you have ever walked along a beach, turned, and examined the footprints you left behind, these are negative impressions of your feet. If that sand could lithify instantly, it would represent an external mold of your foot.



Dinosaur trackways, Dinosaur Ridge, Colorado.

Fossil trackways, a type of trace fossil, are a type of external mold. Some very famous dinosaur tracks are preserved in this way. Most often, external molds are created by hard shells or bones. However, some external molds of dinosaur skin have been discovered, meaning that even soft tissues can be preserved via this mode under the right conditions.

Casts



Sue, the most complete (and famous) Tyrannosaurus rex fossil at the Field Museum, Chicago.

Casts are replicas of organisms, essentially recreating their original shape. First, an organism must be preserved as an external mold, or perhaps open burrows preserved as trace fossils. The impressions or other cavities are then filled in with sediment, producing a three-dimensional form that mimics the form of the original fossil. Imagine if you took some clay and pressed it into the external mold of the trilobite above. Once hardened, you could pop that trilobite out and have a cast of the original fossil, mimicking its original form, and presented in positive relief. Some details are typically lost when a cast is created but can still provide critical information about the organism.

Interestingly, many fossil skeletons on display at museums are primarily casts. This doesn't mean the fossils are fakes since the casts are created based on the original fossils. Casts are used for many reasons: the original fossils can be hundreds of pounds of permineralized bone, making it a challenge to mount the specimens, the originals may be fragile and require a controlled environment for preservation, and it is a way to share specimens among museums, or the fossils may currently be under study.

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8.3: Trace Fossils- Studies in Scurrying, Scraping, and Slithering

The previous section focused on body fossils, which preserve evidence of the actual organism. However, another type of fossil that helps paleontologists understand how ancient creatures lived is called a trace fossil. Trace fossils are preserved evidence of a creature in stone, such as footprints, burrows, precipitated structures, and even the animal's excrement (called coprolites)!

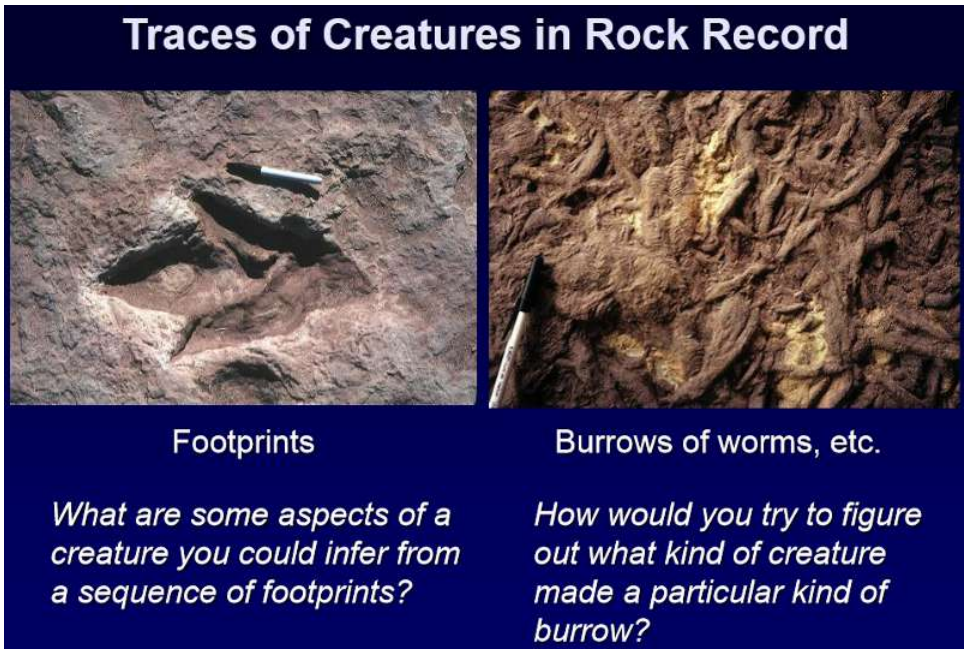


Figure Above: Traces of creatures in the rock record. Footprints: What aspects of a creature could you infer from a sequence of footprints? Burrow of worms, etc.: How would you figure out what kind of creature made a particular burrow?

Trace fossils can yield interesting perspectives on how an organism lived and interacted with its environment. After all, if someone excavated your fossilized bones, would that tell your entire story? Would your bones and how they were buried say anything about what you were doing at any time? Likewise, non-human organisms scurry, burrow, step, jump, run, walk, slither, and grow in various ways. Traces of these actions are preserved in sediment and are studied under the branch of paleontology called ichnology.

Earliest trace fossils



Stromatolite fossil from Wyoming. Image from Wikimedia Commons

Some of the earliest trace fossils are found in rocks around 3.5 billion years old and preserved as structures referred to as stromatolites. These are easily visible structures that have a dome-like shape. In them, wafer-thin mats of photosynthetic microbial mats (cyanobacteria) alternate with fine calcite mud. Initially flat, they grow increasingly round through time, though the relief and size vary tremendously. Stromatolites require certain conditions to grow: (1) sunlight and shallow water, and (2) the absence of grazing predators such as snails.



Image of modern-day stromatolites

Stromatolites are still forming today in places like Shark Bay of western Australia and tidal channels in the Bahamas, both locations where predatory grazing by snails is absent. These environments' high salinities and powerful scouring tides keep the snails away. The cyanobacteria that made the ancient stromatolites are not preserved but are evidenced by the layers of sediment that stuck to the cyanobacterial mats. Since the Archean, evolution has led to more complex organisms using ever-changing forms of movement. New and unique forms of trace fossils emerged over time, including horizontal and vertical burrows and, eventually, footprints.

Classification of trace fossils

Trace fossils are often classified by a method that focuses on the organism's behavior (see table).

Trace fossil classifications *(after Seilacher, 1967)*	
Associated behavior	Example
Movement/Trail	 Cruziana trace fossil
Resting	 Rusophycus trace fossil
Dwelling	 Thalassinoides trace fossil
Grazing	 Helminthoposis trace fossil
Feeding burrows	 Zoophycos trace fossil

Note: Coprolites, or fossilized feces, are also considered traces but do not have a formal classification.

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CHAPTER OVERVIEW

9: Geologic Time and Relative Dating

9.1: Lateral Continuity, Superposition, and Inclusions

9.2: Original Horizontality and Cross-Cutting Relationships

9.3: Relative Dating and Unconformities

9.4: Fossils in Sedimentary Sequences

9.5: Some Final Practice

Thumbnail: Eroded sedimentary rocks in Bryce Canyon National Park, Utah, USA. (Public Domain; Brigitte Werner via Wikicommons)

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9.1: Lateral Continuity, Superposition, and Inclusions

The deposition of rock layers, particularly sedimentary layers, in a sequence where one layer is atop the other layer and so on leads to a sequence of rock layers that geologists can interpret. A geologist can look at these layers and determine which was deposited first, which layer is the youngest, what type of environment existed at the time to form the rock layer, and where there is missing time in the sequence due to weathering and erosion, and so on. Interpreting the stacks upon stacks of rock layers in the Geologic Record is one of the fundamental things geologists can do to establish the history of our planet and the life that has lived on it. How do they do it? Let's start by learning fundamental rules for "reading" a stack of rock layers!

First, the **principle of lateral continuity** states that layers of sediment initially extend laterally in all directions; in other words, they are laterally continuous. As a result, rocks that are otherwise similar, but are now separated by a valley or other erosional feature, can be assumed to be originally continuous. Layers of sediment do not extend indefinitely; rather, the limits can be recognized and are controlled by the amount and type of sediment available and the size and shape of the sedimentary basin. As long as sediment is transported to an area, it will eventually be deposited. However, as the amount of material lessens from the source, the layer of that material will become thinner.

Often, coarser-grained material can no longer be transported to an area because the transporting medium has insufficient energy to carry it to that location. In its place, the particles that settle from the transporting medium will be finer-grained, and there will be a lateral transition from coarser- to finer-grained material. The lateral variation in sediment within a stratum is known as sedimentary facies. If sufficient sedimentary material is available, it will be deposited up to the limits of the sedimentary basin. Often, the sedimentary basin is within rocks very different from the sediments being deposited. In those cases, the lateral limits of the sedimentary layer will be marked by an abrupt change in rock type.

Principle of Superposition



Most sediment is laid down in layers where it eventually gets buried, compacted, and cemented to form a sedimentary rock. If the sedimentary rocks are no longer horizontal, then we can assume the tilting event happened after the layers originally formed.

The **law of superposition** (or the **principle of superposition**) is a key axiom based on observations of natural history that is a foundational principle of sedimentary stratigraphy and so of other geology-dependent natural sciences:

Sedimentary layers are deposited in a time sequence, with the oldest on the bottom and the youngest on the top.

The law was formulated in the 17th century by the Danish scientist Nicolas Steno. When combined with the related principle of faunal succession, the law of superposition provides a very powerful tool for dating rocks and strata.

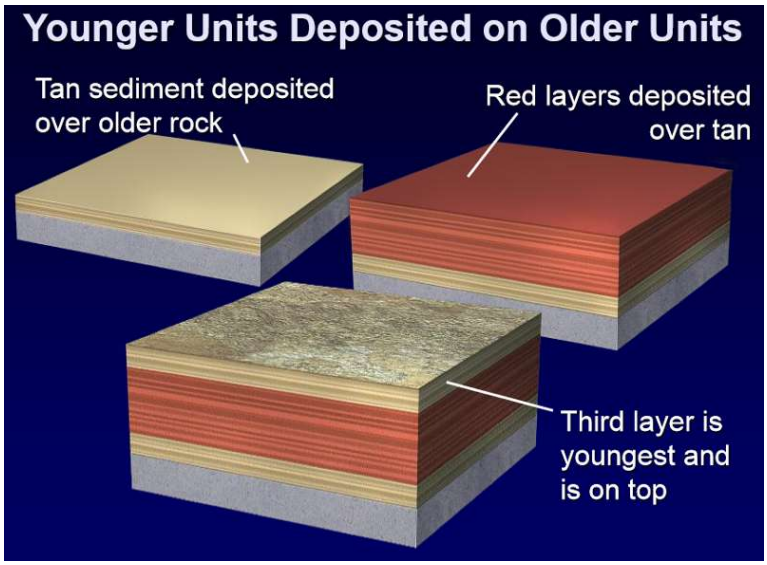


Figure Above: Younger units are deposited on older units. Tan sediment is deposited over older gray rock, and red layers are deposited over tan rock.

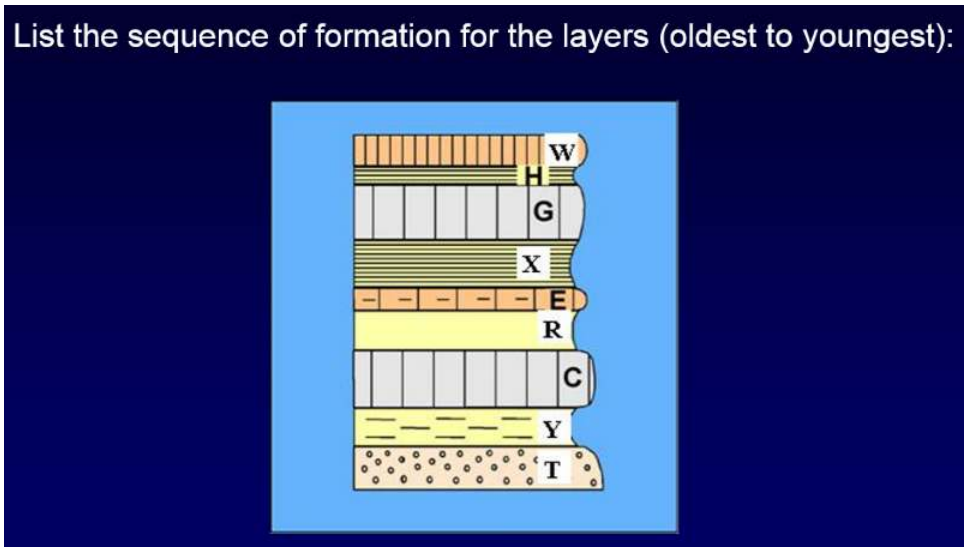
In a set of strata, the youngest layers must be those of the top layer, and the oldest must lie on the bottom. This is because the youngest layer was deposited after the oldest layers, determining their place in the layers. Since the oldest was deposited first, it is on the bottom and vice versa.

Principle of Inclusions

Steno stated another, more general principle in this way:

If a solid body is enclosed on all sides by another solid body, of the two bodies that one first became hard which, in the mutual contact, expresses on its own surface the properties of the other surface.

In other words: a solid object will cause any solids that form around it later to conform to its shape. Steno could show by this reasoning that fossils and crystals must have solidified before the host rock that contains them was formed.



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9.2: Original Horizontality and Cross-Cutting Relationships

The Principle of Original Horizontality was proposed by the Danish geological pioneer Nicholas Steno (1638–1686). This principle states that layers of sediment are originally deposited horizontally under the action of gravity. The principle is important to the analysis of folded and tilted strata. From these observations is derived the conclusion that the Earth has not been static and that great forces have been at work over long periods of time, further leading to the conclusions of the science of plate tectonics; that movement and collisions of large plates of the Earth's crust is the cause of folded strata.

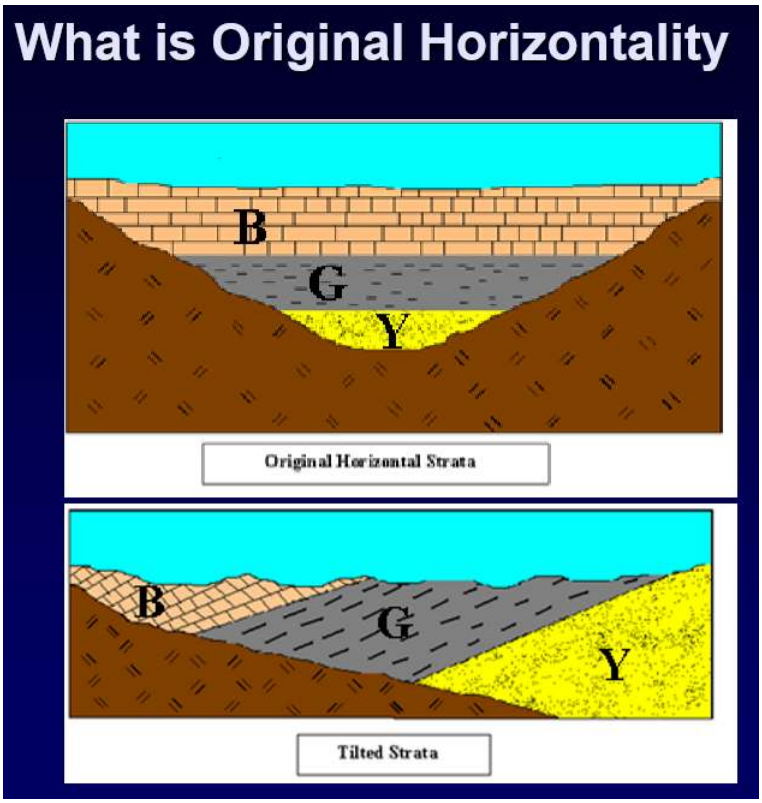


Figure Above: What is original horizontality? When encountering tilted strata, un-tilt them by imagining what they would be like if they were rotated so that they were all horizontal and parallel to one another.

As one of Steno's Laws, the Principle of Original Horizontality served well in the early days of geological science. However, it is now known that not all sedimentary layers are deposited purely horizontally. For instance, coarser-grained sediments such as sand may be deposited at angles of up to 15 degrees. The internal friction between grains prevents them from slumping to a lower angle without additional reworking or effort. This is known as the angle of repose, and a prime example is the surface of sand dunes. Similarly, sediments may drape over a pre-existing inclined surface: these sediments are usually deposited conformably to the pre-existing surface.

Also, sedimentary beds may pinch out along strike, implying that slight angles existed during their deposition. Thus the Principle of Original Horizontality is widely, but not universally, applicable in the study of sedimentology, stratigraphy, and structural geology. Sedimentary rocks are originally laid down in horizontal layers. If the sedimentary rocks are tilted, the tilting event happens after the layers originally formed. To determine which layer is the youngest, un-tilt the layers back to horizontal in the shortest direction of rotation.

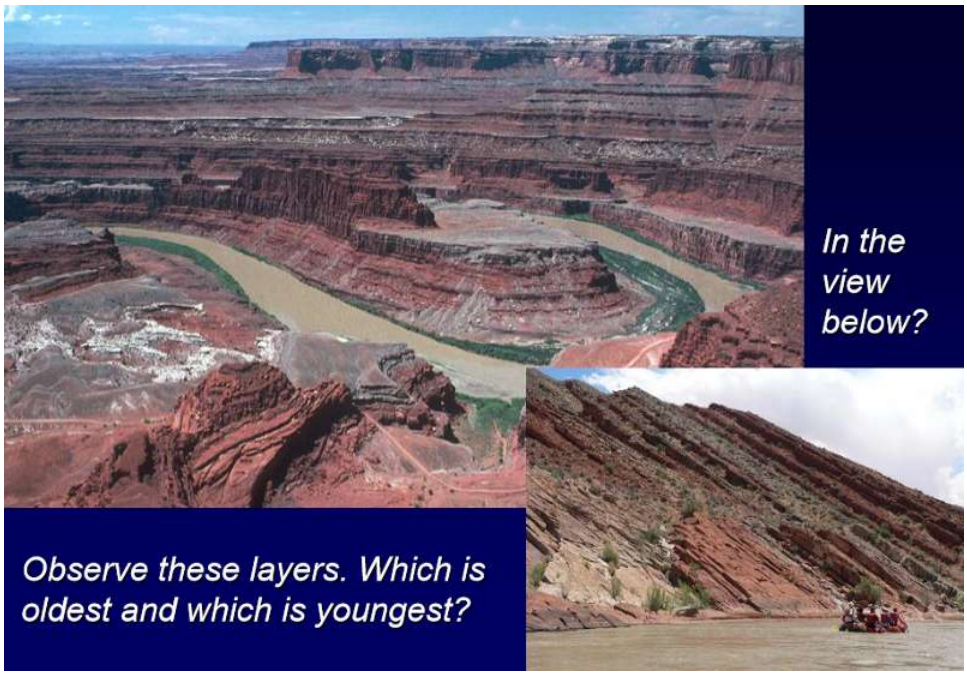


Image Above: Which layers are youngest and which layers are oldest? See the text below for a description and an answer.

When looking at the tilted layers in the lower right image you should notice that the rocks are tilted to the right. This means the layers on the right side of the image are younger and get progressively older as you move to the left.

Principle of Crosscutting Relationships

Originally developed by James Hutton in *Theory of the Earth* (1795) and embellished by Charles Lyell in *Principles of Geology* (1830), the principle of **cross-cutting relationships** states that the geologic feature which cuts another is the younger of the two features.

There are several basic types of cross-cutting relationships:

- Structural relationships may be faults or fractures cutting through older rock.
- Intrusional relationships occur when an igneous pluton or dike is intruded into pre-existing rocks.
- Stratigraphic relationships may be an erosional surface (or unconformity) cuts across older rock layers, geological structures, or other geological features.
- Sedimentological relationships occur where currents have eroded or scoured older sediment in a local area to produce, for example, a channel filled with sand.
- Paleontological relationships occur where the animal activity or plant growth produces truncation. This happens, for example, when animal burrows penetrate pre-existing sedimentary deposits.
- Geomorphological relationships may occur where a surficial feature, such as a river flowing through a gap in a rock ridge. In a similar example, an impact crater excavates into a subsurface layer of rock.

Cross-cutting relationships may be compound in nature. For example, if an unconformity truncated a fault, and that unconformity was cut by a dike. Based upon such compound cross-cutting relationships, it can be seen that the fault is older than the unconformity, which in turn is older than the dike. Using such rationale, the sequence of geological events can be better understood. When a rock layer or fault cuts across another rock layer or fault, the thing doing the cutting must be younger than the object it cuts.

Inclusions: When pieces of rock or material are included within a rock layer, the included pieces must be older than the rock they are included in.

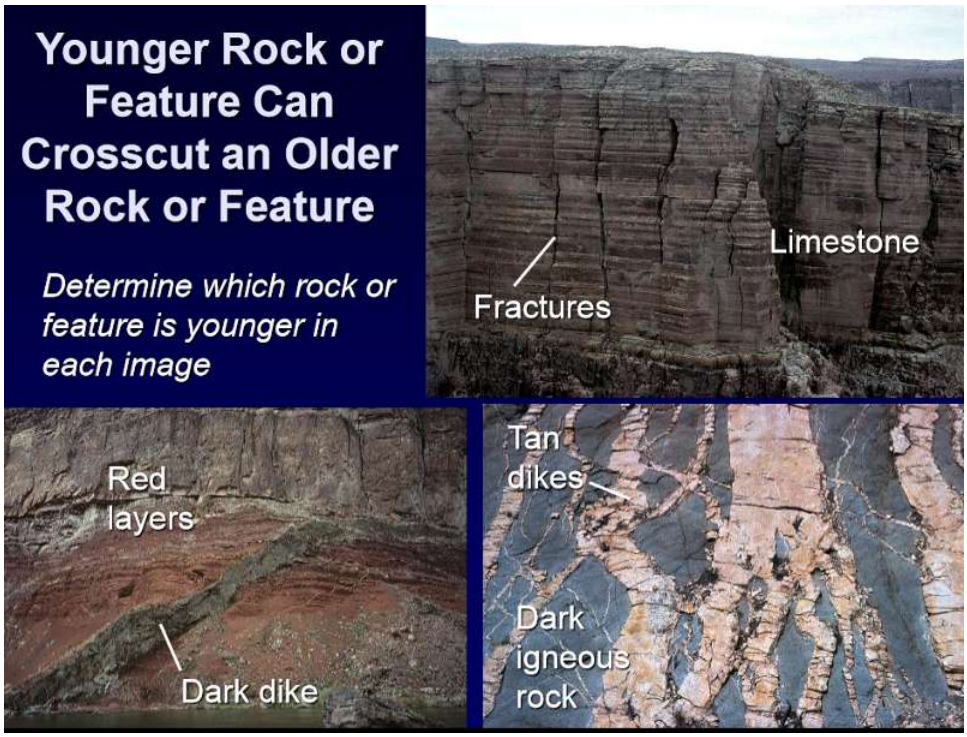


Figure Above: When features are cut across layers of rock, these features must be younger than the rock they are cutting across. Such features include volcanic dikes, sills, faults and fractures.

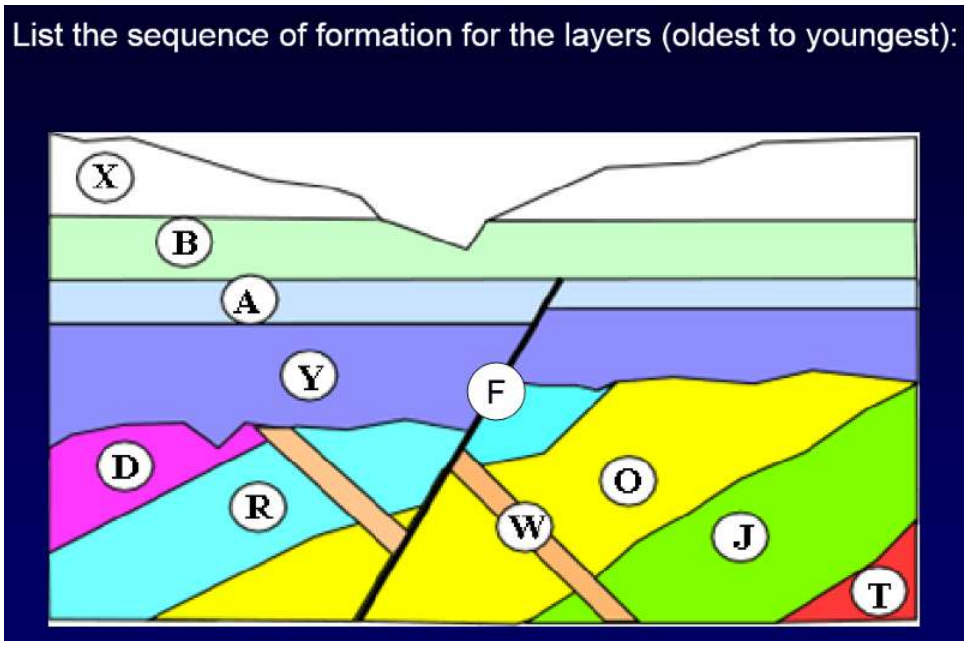
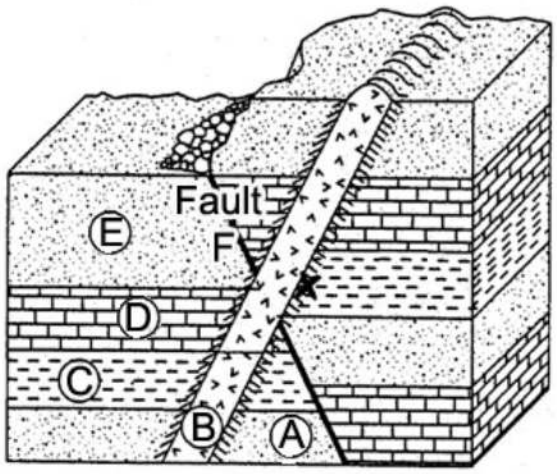


Figure Above: Layer T is at the bottom, and above it is layer J. Above layer J is layer O. Above layer O is layer R. Above layer R is layer D. Layers T through D are tilted. A volcanic dike, W, cuts across layers T through D. Above layer D is layer Y. Above layer Y is layer A. Cutting across layers T through A is fault F. Above layer A is layer B. Above layer B is layer X.



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9.3: Relative Dating and Unconformities

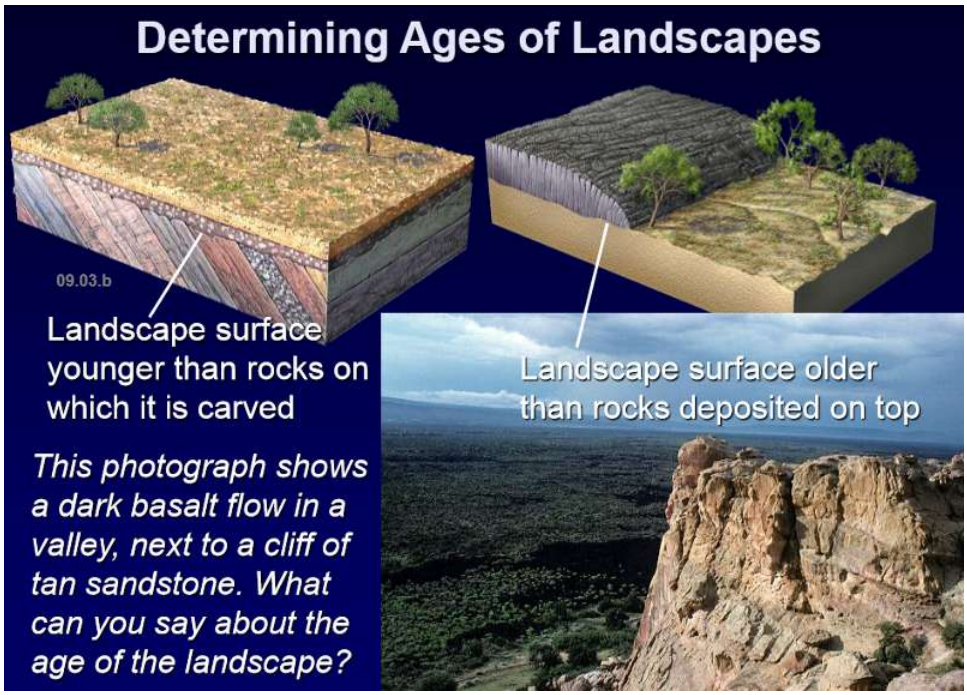


Figure Above: Determining ages of landscapes. A rock sequence on which a current landscape is depositing rocks is always older than the landscape.

How do we use rock layers to determine the age of layers?

When we examine the first drawing above we can ask ourselves what happened between the tilted layers on the bottom and the flat layers on top? The tilted layer was eroded flat before the new layer was deposited on top. So the sequence of events would be:

1. Sedimentary rocks are deposited in flat layers.
2. The sedimentary rocks are tilted through some kind of event and then eroded flat.
3. A new layer was deposited on top of the tilted layers forming a disconformity between the two groups of rocks.

If you can date the sediments in both the tilted layers and the flat top layer then you can get an estimate of when the tilting event occured.

Lava flows are excellent for dating the age of rocks through radioactive decay but you can also use them to establish relative ages of the rocks above and below the lava flows. If the lava flow in the second image above is 5 million years old then you know the rocks below the the lava flow must be older than 5 million.

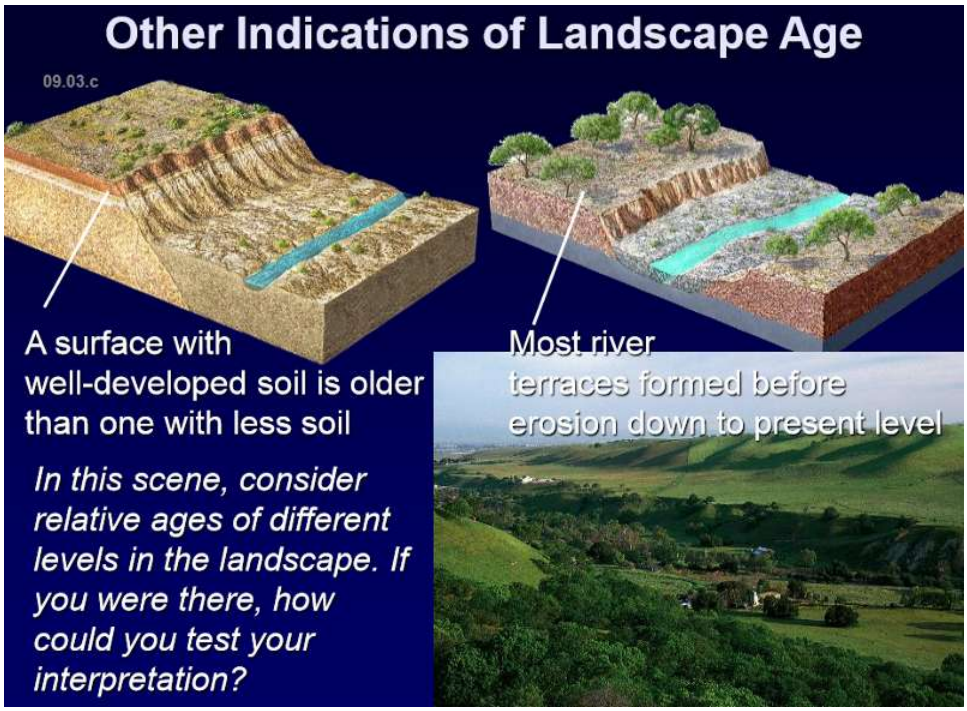


Figure Above: Other indicators of landscape age. A surface with well-developed soil is older than one with less soil. The elevation of a river terrace indicates the old elevation of the river before the river eroded it down.

A landscape surface develops more soil if it remains undisturbed by erosion and deposition. A surface with a thick layer of soil indicates old age. Rivers cut their way through landscapes leaving behind terraces (the old elevation of the river). Geologists can use terraces to establish relative ages for when the river eroded the landscape. The highest terrace is the oldest.

Unconformities

Unconformities represent a gap in time when the landscape is being eroded away. Keep in mind that we know about the geologic history of the Earth because of the rocks we can see and examine. If the rocks are missing because of erosion and weathering then we have no record of what existed during that time period. Unconformities are a part of landscape evolution. In the image above a gray limestone is deposited under the sea. The sea regresses and the limestone bed is folded into an anticline. The top of the fold is eroded flat. A conglomerate layer is deposited on top of the limestone and erosion surface. An angular unconformity is formed.The unconformity represent a time of erosion and a gap in geologic time history.

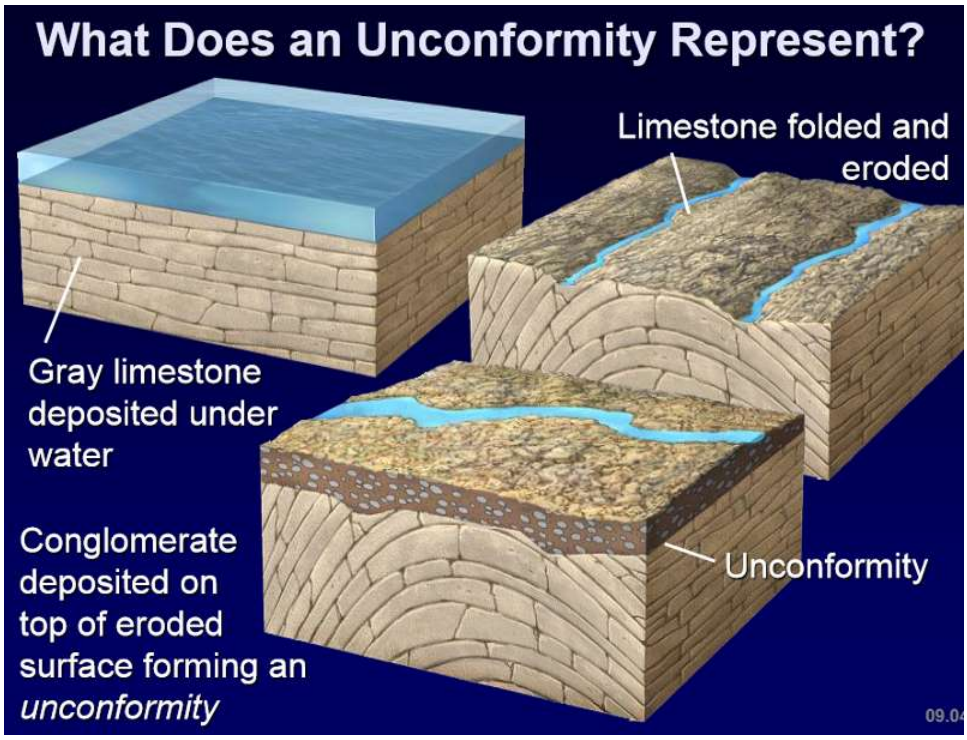
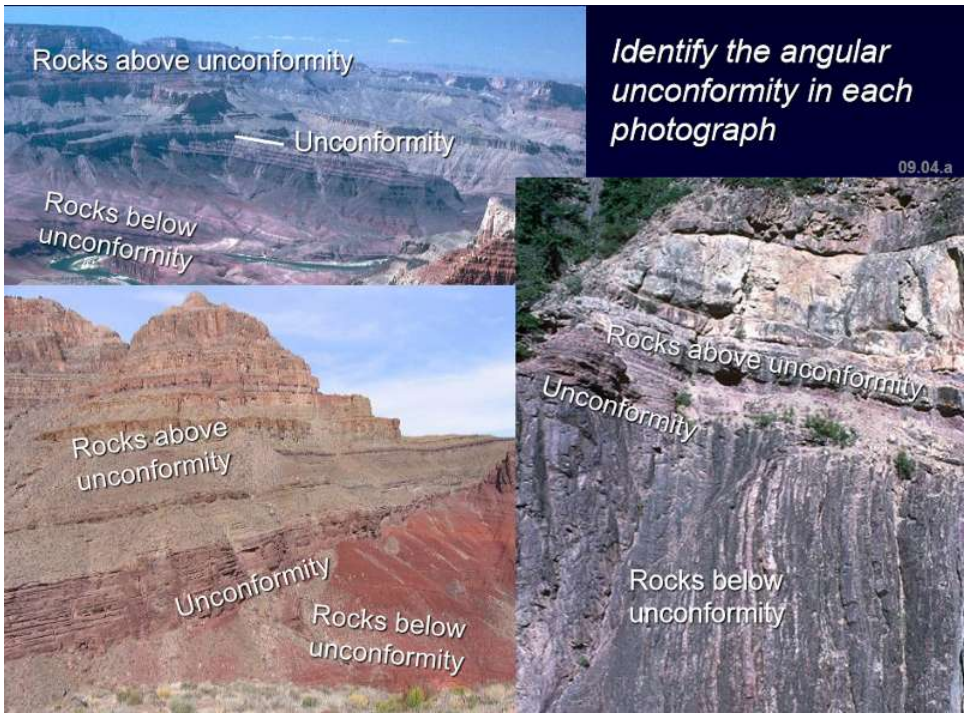


Figure Above: What does an unconformity represent? If a gray limestone is first deposited underwater, it might later become folded and much later have the overlying sediment on it eroded away by rivers as the area experiences a regression. Although the limestone is folded, a horizontal layer of conglomerate may be deposited directly over it. The missing time from when the sediment was eroded away between the limestone and the conglomerate is an unconformity.



In the Figure Above are some examples of unconformities.

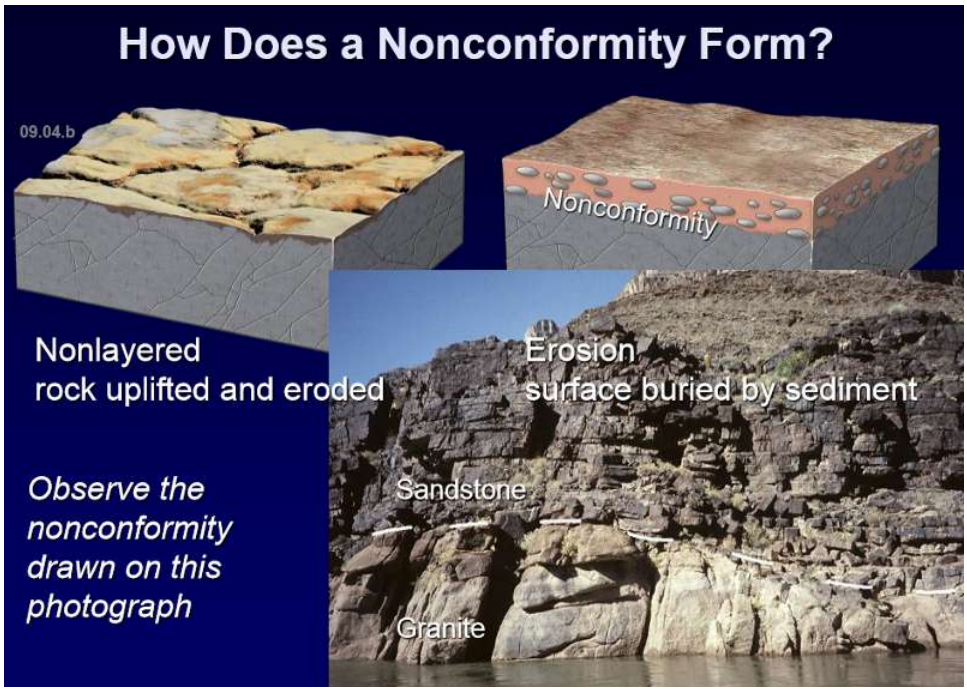


Figure Above: How does a nonconformity form? Nonlayered rock (such as igneous basement rock) is uplifted and eroded. A sedimentary layer might be deposited directly over the basement rock. The nonconformity is specifically resulting from the igneous rock being below the erosional surface.

When metamorphic or igneous rock is below the erosional surface then the type of unconformity created is called a nonconformity. This is different from an intrusion contact, where magma melts its way into a rock and solidifies because there is evidence of erosion at the nonconformity line.

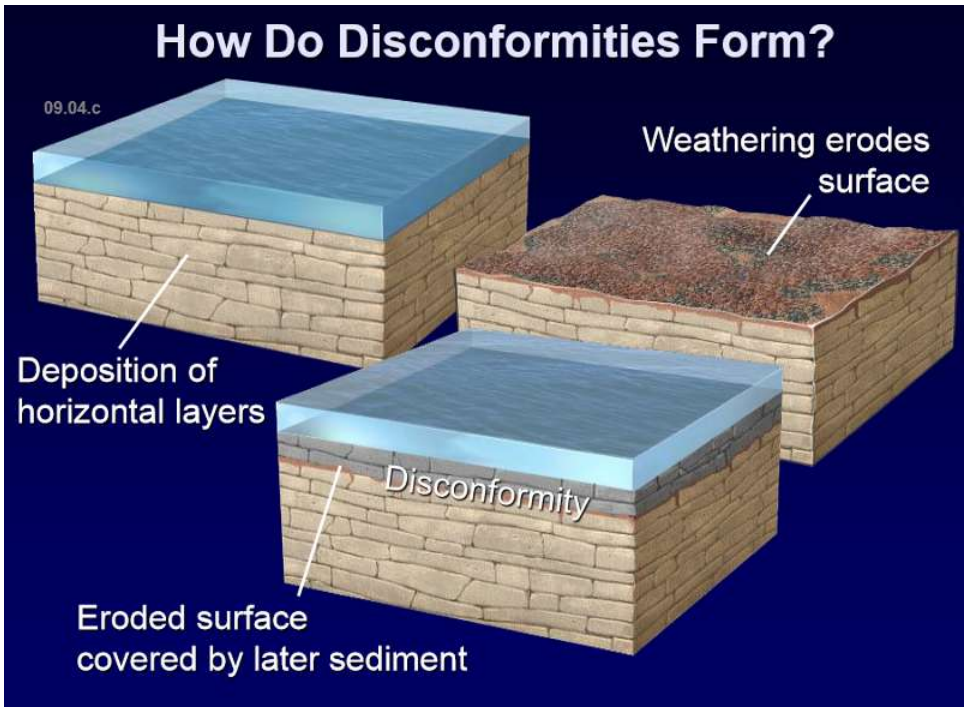


Figure Above: How do disconformities form? Horizontal sedimentary layers (limestone) form. Weathering erodes the surface. The eroded surface is covered by a new sedimentary rock (in this example, more limestone)

In a disconformity, flat-lying sedimentary rocks are deposited. The sedimentary rock is eroded for some period of time. After some time, the area changes from a time of erosion to a time of deposition and a new sedimentary layer is deposited on top of the old layer. The boundary between the two layers is the disconformity. In an angular unconformity, the layers are tilted or folded during the middle stage.



Figure Above: Example of a Disconformity between a Devonian limestone that is light gray, and a Mississippian Limestone that is dark gray.



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9.4: Fossils in Sedimentary Sequences

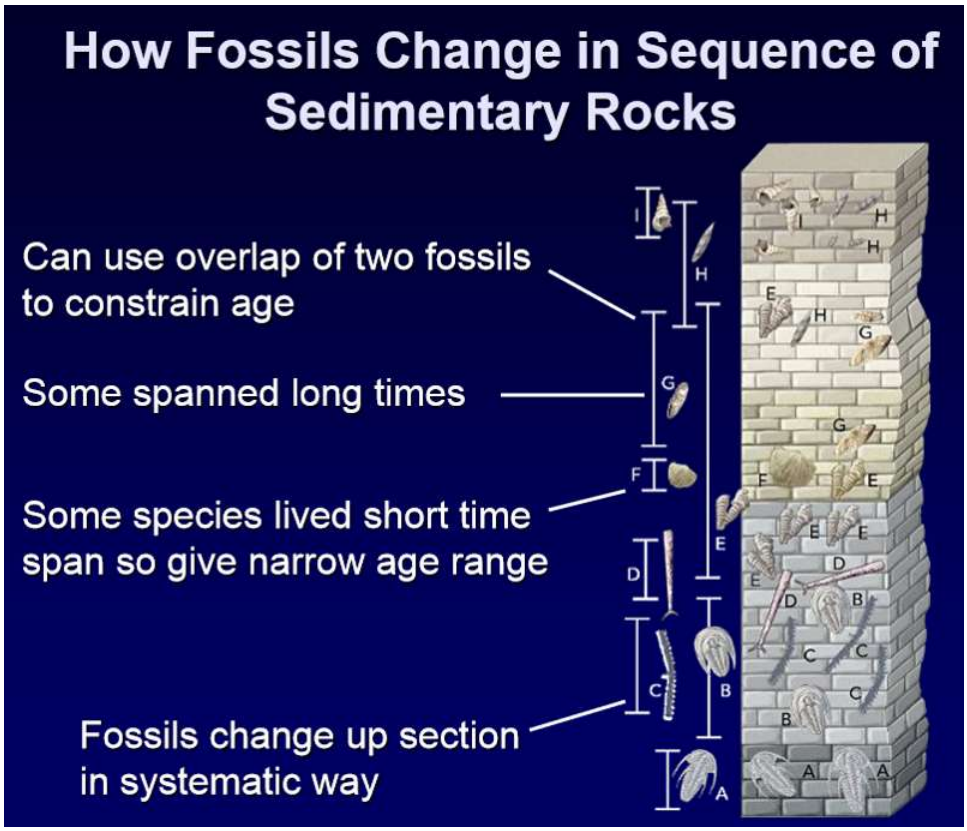


Figure Above: How fossils change in the sequence of sedimentary rocks. Can use the overlap of two fossils to constrain age. Some fossils spanned long times. Some species lived short timespans so they give a narrow age range of the strata. Fossils change up sections in a systematic way.

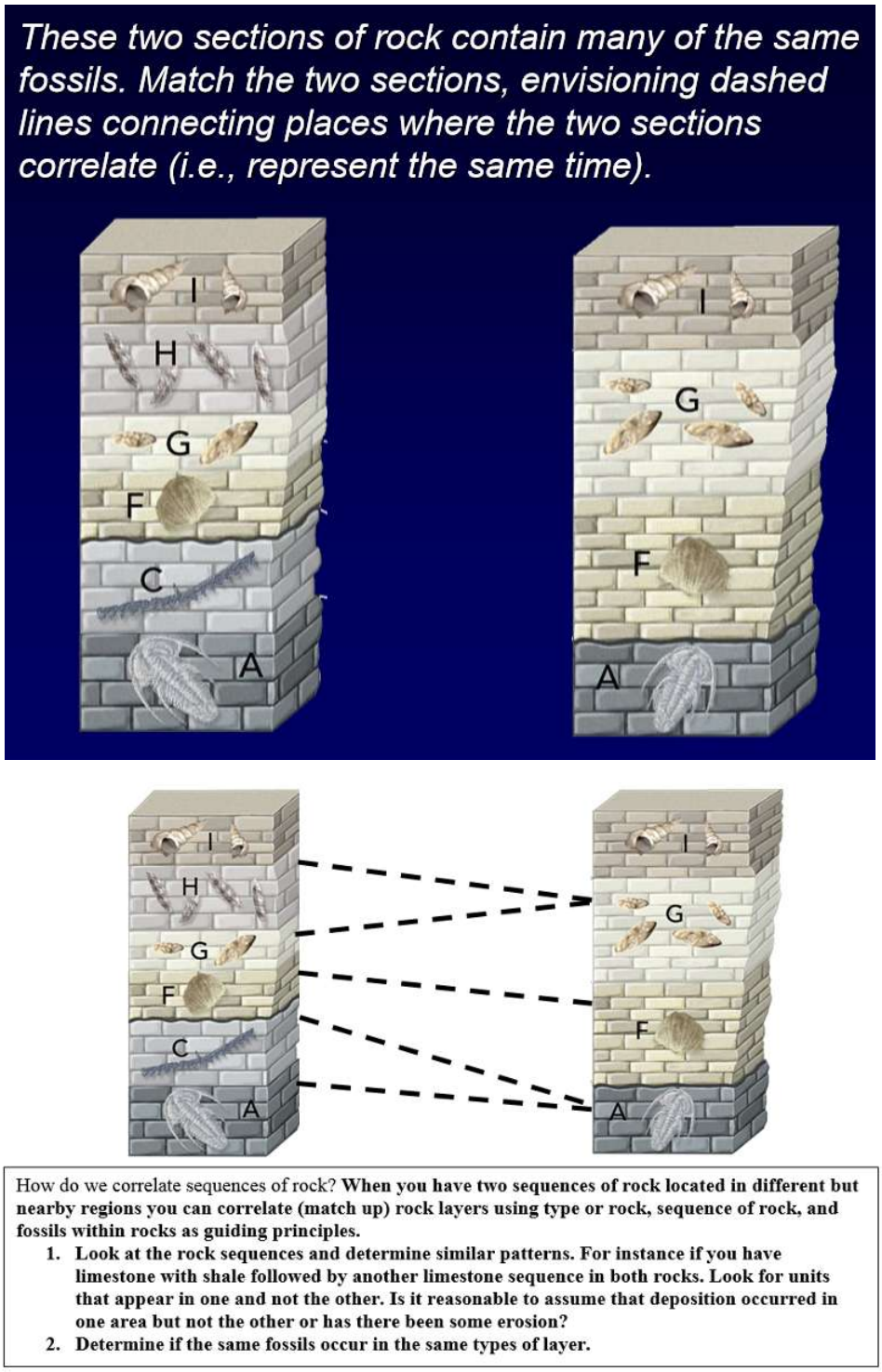
Sedimentary Sequences

Fossils change through time, which can be seen in the sequence of sedimentary rocks.

1. We can use the overlap of two fossils to constrain age. The two fossils together only occur in a narrow time range (I & G).
2. Some spanned long times and represented organisms that remained unchanged for long periods of time. These fossils are less useful for assigning ages than a fossil that lived for a short time period.
3. Some species only live a short time span, are widespread and abundant, and are easily identifiable. They are referred to as INDEX FOSSILS and are very useful for dating time periods.
4. Fossils change up sections in a systematic way. Changes in body design or shape help to distinguish one variety from another. Trilobite (A) is the main index fossil for early Paleozoic.

The **principle of faunal succession**, also known as the **law of faunal succession**, is based on the observation that sedimentary rock strata contain fossilized flora and fauna and that these fossils succeed each other vertically in a specific, reliable order that can be identified over wide horizontal distances. For example, a fossilized Neanderthal bone will never be found in the same stratum as a fossilized Megalosaurus because neanderthals and megalosaurus lived during different geological periods, separated by many millions of years.

This principle allows for strata to be identified and dated by the fossils found within. This principle, which received its name from the English geologist William Smith, is important in determining the relative age of rocks and strata. The fossil content of rocks, together with the law of superposition, helps to determine the time sequence in which sedimentary rocks were laid down. Theories of evolution explain the observed faunal and floral succession preserved in rocks, which are the facts on which the understanding of evolution is based.



How do we correlate sequences of rock? **When you have two sequences of rock located in different but nearby regions you can correlate (match up) rock layers using type or rock, sequence of rock, and fossils within rocks as guiding principles.**

1. **Look at the rock sequences and determine similar patterns. For instance if you have limestone with shale followed by another limestone sequence in both rocks. Look for units that appear in one and not the other. Is it reasonable to assume that deposition occurred in one area but not the other or has there been some erosion?**
2. **Determine if the same fossils occur in the same types of layer.**

Figure Above: How do we correlate sequences of rock? When you have two sequences of rock located in different but nearby regions you can correlate (match up) rock layers using the type of rock, sequence of rock, and fossils within the rocks as guiding principles.

1. Look at the rock sequences and determine similar patterns. For instance, if you have limestone with shale followed by another limestone sequence in both rocks. Look for units that appear in one and not the other. Is it reasonable to assume that deposition occurred in one area but not the other, or has there been some erosion? 2. Determine if the samefossils occur in the same types of layer.

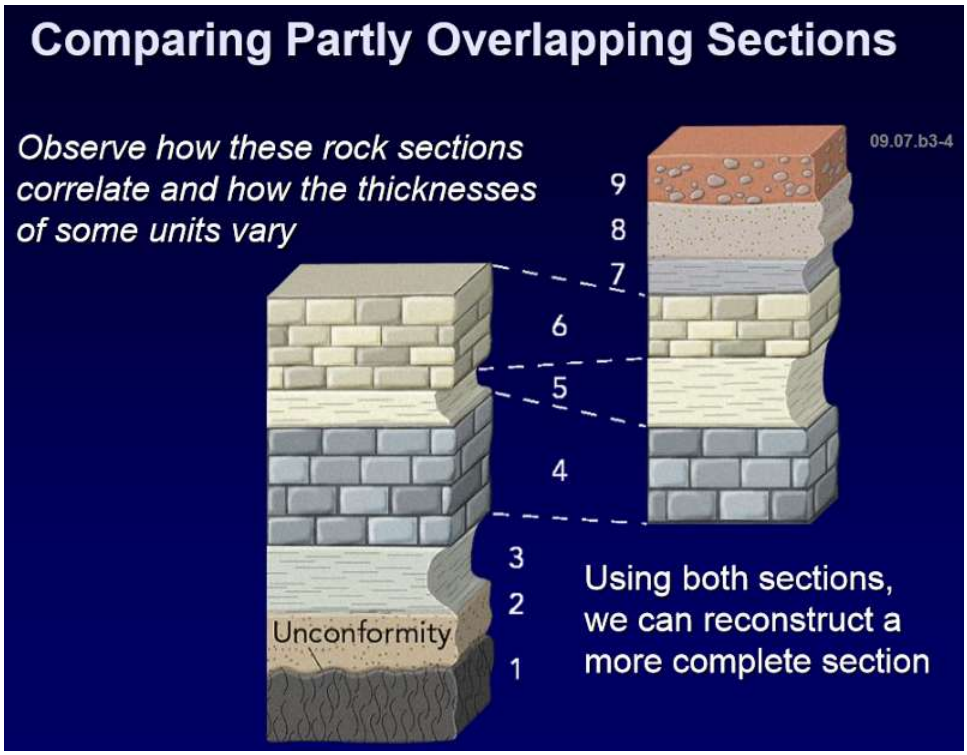


Figure Above: Comparing partly overlapping sections. Observe how these rock sections correlate and how the thicknesses of some units vary. Using both sections, we can reconstruct a complete section.

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9.5: Some Final Practice

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CHAPTER OVERVIEW

10: The PreCambrian

10.1: The Hadean Eon (4600-4000 Ma)

10.2: The Archean Eon (4000-2500 Ma)

10.3: The Proterozoic Eon (2500-541 Ma)

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10.1: The Hadean Eon (4600-4000 Ma)

The Hadean Eon, named after the Greek god and ruler of the underworld, Hades, is the oldest eon and dates from 4.6-4.0 billion years ago. This time represents Earth's earliest history, during which the planet was characterized by a partially molten surface, volcanism, and asteroid impacts. Several mechanisms made the newly forming Earth incredibly hot: gravitational compression, radioactive decay, and asteroid impacts. Most of this initial heat still exists inside the Earth. The Hadean was originally defined as the birth of the planet occurring 4.0 billion years ago and preceding the existence of many rocks and life forms. However, geologists have dated minerals at 4.4 billion years, with evidence that liquid water was present. There is possibly even evidence of life existing over 4.0 billion years ago. However, the most reliable record for early life, the microfossil record, starts at 3.5 billion years ago.

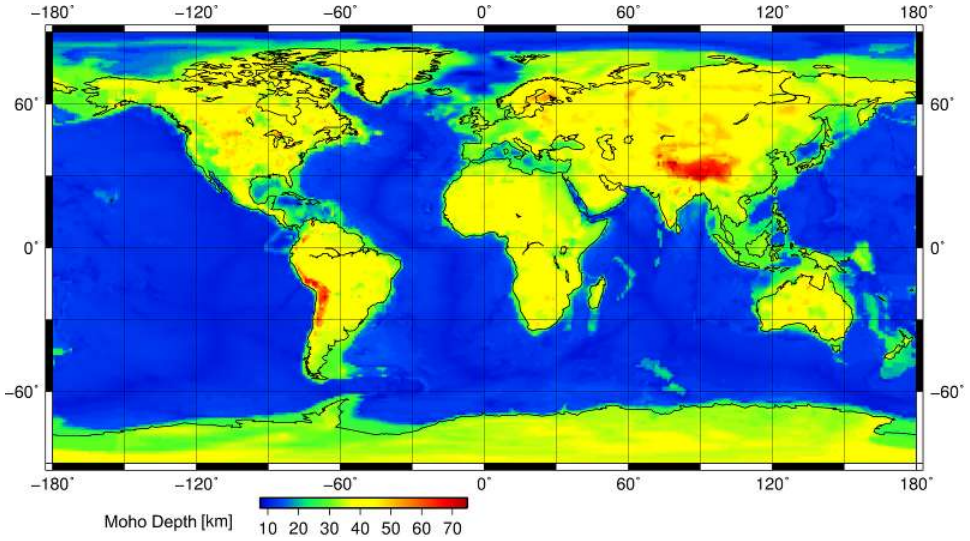


Origin of the Earth's Crust

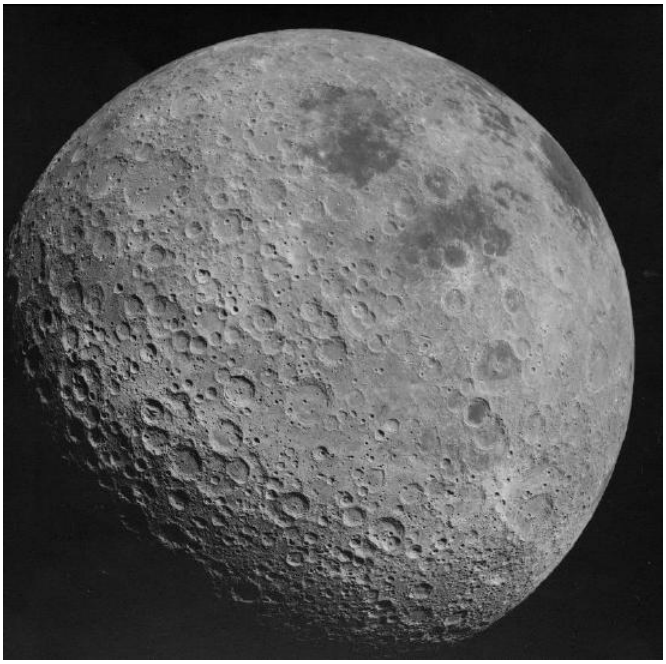
As Earth cooled from its molten state, minerals crystallized and settled, separating minerals based on density and creating the crust, mantle, and core. The earliest Earth was chiefly molten material and would have been rounded by gravitational forces, so it resembled a ball of lava floating in space. As the outer part of the Earth slowly cooled, the high melting-point minerals formed solid slabs of early crust. These slabs were probably unstable and easily reabsorbed into the liquid magma until the Earth cooled enough to allow numerous larger fragments to form a thin primitive crust.

Scientists generally assume this crust was oceanic and mafic in composition and littered with impacts, much like the Moon's current crust. There is still some debate over when plate tectonics started, which would have led to continental and oceanic crust formation. Regardless of this, as Earth cooled and solidified, less dense felsic minerals floated to the surface of the Earth to form the crust. In contrast, the denser mafic and ultramafic materials sank to form the mantle and the highest-density iron and nickel sank into the core. This differentiated the Earth from a homogenous planet into a heterogeneous one with layers of felsic crust, mafic crust, ultramafic mantle, and iron and nickel core.

Below is an image showing the depth to the Moho, which shows the thickness of the crust.



Origin of the Moon

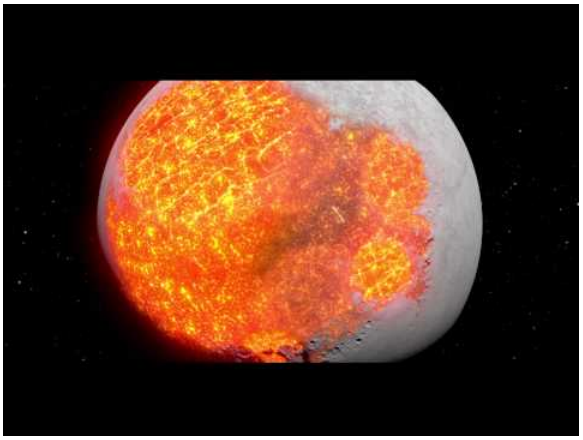


Several unique features of Earth's Moon have prompted scientists to develop the current hypothesis about its formation. The Earth and Moon are tidally locked, meaning that as the Moon orbits, one side always faces the Earth and the opposite side is not visible to us. Also, most importantly, the Earth's and Moon's chemical compositions show nearly identical isotope ratios and volatile content. Apollo missions returned from the Moon with rocks that allowed scientists to precisely compare Moon and Earth rocks. Other bodies in the solar system and meteorites do not share the same degree of similarity and show much higher variability. If the Moon and Earth formed together, this would explain why they are so chemically similar.

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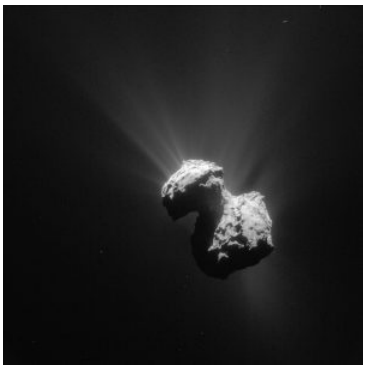
Many ideas have been proposed for the origin of the Moon: The Moon could have been captured from another part of the solar system and formed together with the Earth, or the Moon could have been ripped out of the early Earth. None of the proposed explanations can account for all the evidence. The currently prevailing hypothesis is the **giant-impact hypothesis**. It proposes a body about half of Earth's size must have shared at least parts of Earth's orbit and collided with it, resulting in a violent mixing and scattering of material from both objects. Both bodies would be composed of a combination of materials, with more of the lower density splatter coalescing into the Moon. This may explain why the Earth has a higher density and thicker core than the Moon.



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Origin of Water

Explanations for the origin of Earth's water vary. The volcanic outgassing hypothesis for Earth's water is that it originated from inside the planet and emerged via tectonic processes as vapor associated with volcanic eruptions. Since all volcanic eruptions contain some water vapor, at times more than 1% of the volume, these alone could have created Earth's surface water. Another likely source of water was from space. Comets are a mixture of dust and ice, with some or most of that ice being frozen water. Seemingly dry meteors can contain small but measurable amounts of water, usually trapped in their mineral structures.



During heavy bombardment periods later in Earth's history, its cooled surface was pummeled by comets and meteorites, which could be why so much water exists above ground. There isn't a definitive answer for what process is the source of ocean water. Earth's water isotopically matches water found in meteorites much better than that of comets. However, it is hard to know if Earth's processes could have changed the water's isotopic signature over the last 4-plus billion years. It is possible that all three sources contributed to the origin of Earth's water.

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10.2: The Archean Eon (4000-2500 Ma)

The Archean Eon lasted from 4.0 to 2.5 billion years ago and is named after the Greek word for beginning. This eon represents the beginning of the rock record. Although there is current evidence that rocks and minerals existed during the Hadean Eon, the Archean has a much more robust rock and fossil record.

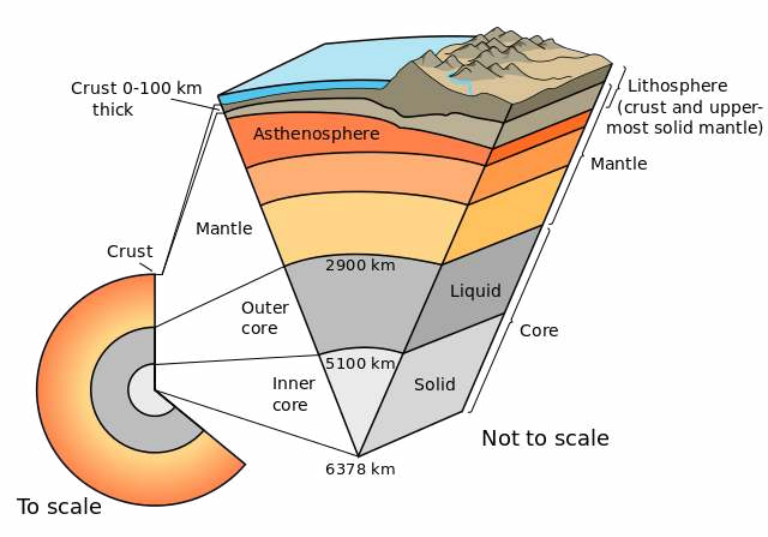


Late Heavy Bombardment

Objects were chaotically flying around at the start of the solar system, building the planets and moons. There is evidence that after the planets formed, about 4.1-3.8 billion years ago, a second large spike of asteroids and comets impacted the Earth and Moon in a **late heavy bombardment**. Meteorites and comets in stable or semi-stable orbits became unstable and started impacting objects throughout the solar system. In addition, this event is called the lunar cataclysm because most of the Moons craters are from this event. During the late heavy bombardment, the Earth, Moon, and all planets in the solar system were pummeled by material from the asteroid and Kuiper belts. Evidence of this bombardment was found within samples collected from the Moon.

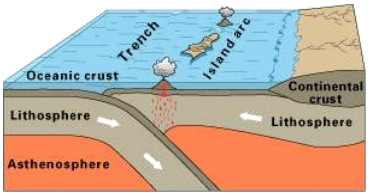
It is universally accepted that the solar system experienced extensive asteroid and comet bombardment at its start; however, some other process must have caused the second increase in impacts hundreds of millions of years later. A leading theory blames gravitational resonance between Jupiter and Saturn for disturbing orbits within the asteroid and Kuiper belts based on a similar process observed in the Eta Corvi star system.

Origin of the Continents



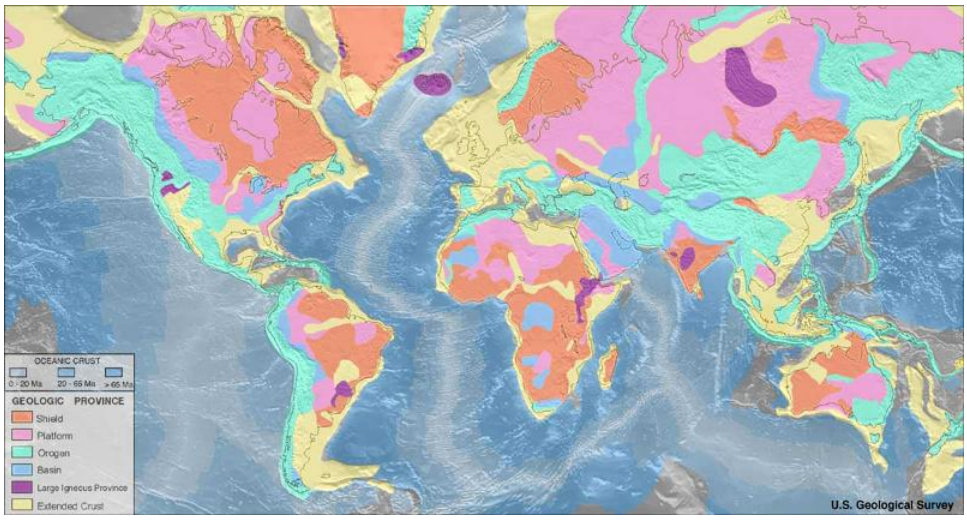
For plate tectonics to work as it currently, it must have continents. However, the easiest way to create continental material is via assimilation and differentiation of existing continents. This chicken-and-egg quandary over how continents were made in the first place is not easily answered because of the great age of continental material and how much evidence has been lost during tectonics and erosion. While the timing and specific processes are still debated, volcanic action must have brought the first continental material to the Earth's surface during the Hadean, 4.4 billion years ago. This model does not solve the problem of continent formation since magmatic differentiation seems to need a thicker crust.

Nevertheless, the continents formed by some incremental process during Earth's early history. The best idea is that density differences allowed lighter felsic materials to float upward and heavier ultramafic materials and metallic iron to sink. These density differences led to the layering of the Earth, the layers now detected by seismic studies. Early protocontinents accumulated felsic materials as developing plate-tectonic processes brought lighter material from the mantle to the surface.



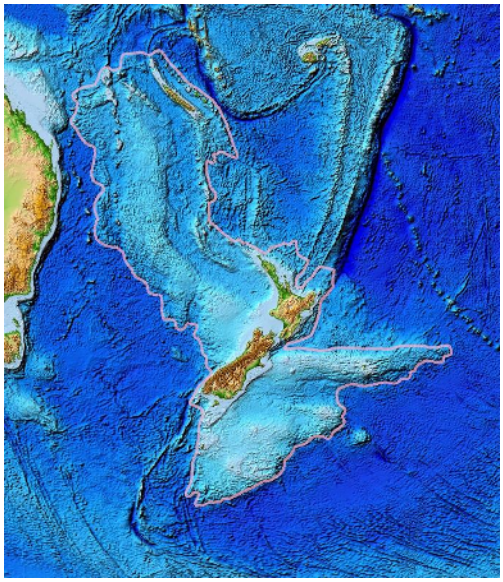
Oceanic-oceanic convergence

The first solid evidence of modern plate tectonics is found at the end of the Archean, indicating at least some continental lithosphere must have been in place. This evidence does not necessarily mark the starting point of plate tectonics; remnants of earlier tectonic activity could have been erased by the rock cycle.



The stable interiors of the current continents are called cratons and were mostly formed in the Archean Eon. A craton has two main parts: the **shield**, crystalline basement rock near the surface, and the **platform** of sedimentary rocks covering the shield. Most cratons have remained relatively unchanged, with most tectonic activity occurring around cratons instead of within them. Whether created by plate tectonics or another process, Archean continents gave rise to the Proterozoic continents that now dominate our planet.

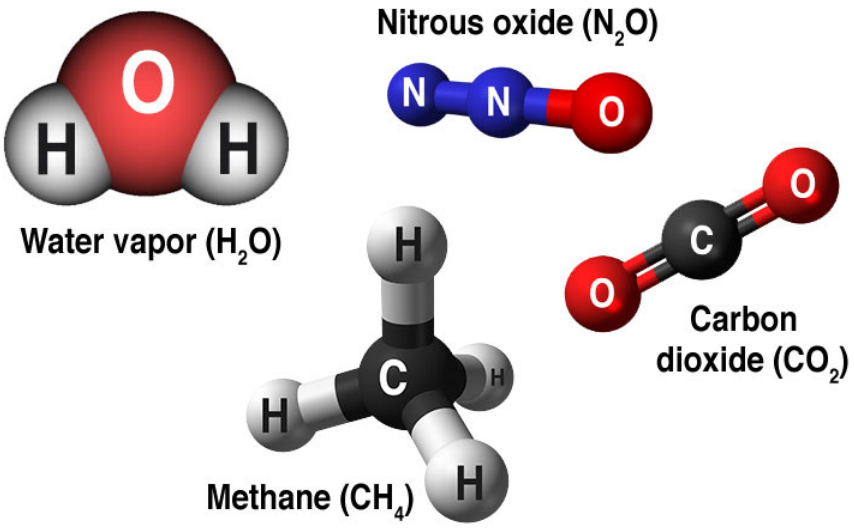
The general guideline for what constitutes a continent and differentiates oceanic from the continental crust is under some debate. At passive margins, continental crust grades into the oceanic crust at passive margins, making a distinction difficult. Even island-arc and hot-spot material can seem more closely related to continental crust than oceanic. Continents usually have a craton in the middle with felsic igneous rocks. There is evidence that submerged masses like Zealandia, which includes present-day New Zealand, would be considered a continent. A Continental crust that does not contain a craton is called a continental fragment, such as the island of Madagascar off the east coast of Africa.



First Life on Earth



Life most likely started during the late Hadean or early Archean Eons. The earliest evidence of life is chemical signatures, microscopic filaments, and microbial mats. Carbon found in 4.1 billion-year-old zircon grains have a chemical signature suggesting an organic origin. Other evidence of early life is 3.8-4.3 billion-year-old microscopic filaments from a hydrothermal vent deposit in Quebec, Canada. While the chemical and microscopic filaments evidence is not as robust as fossils, there is significant fossil evidence for life 3.5 billion years ago. These first well-preserved fossils are photosynthetic microbial mats, known as stromatolites, found in Australia.



Although the origin of life on Earth is unknown, hypotheses include a chemical origin in the early atmosphere and ocean, deep-sea hydrothermal vents, and delivery to Earth by comets or other objects. One hypothesis is that life arose from the chemical environment of the Earth's early atmosphere and oceans, which was very different than today. The oxygen-free atmosphere produced a reducing environment with abundant methane, carbon dioxide, sulfur, and nitrogen compounds. This is what the atmosphere is like on other bodies in the solar system.

Researchers in the famous Miller-Urey experiment simulated early Earth's atmosphere and lightning within a sealed vessel. After igniting sparks within the vessel, they discovered the formation of amino acids, the fundamental building blocks of proteins. In 1977, when scientists discovered an isolated ecosystem around hydrothermal vents on a deep-sea mid-ocean ridge, it opened the door for another explanation of the origin of life. The hydrothermal vents have a unique ecosystem of critters with chemosynthesis

as the foundation of the food chain instead of photosynthesis. The ecosystem derives energy from hot chemical-rich waters pouring out of underground towers. This suggests that life could have started on the deep ocean floor and derived energy from the heat from the Earth's interior via chemosynthesis. Scientists have since expanded the search for life to more unconventional places, like Jupiter's icy moon Europa.

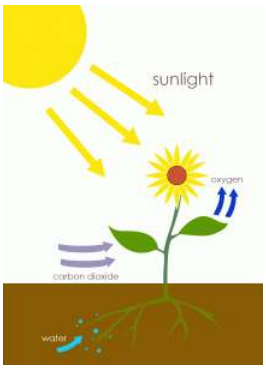
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Another possibility is that life or its building blocks came to Earth from space, carried aboard comets or other objects. Amino acids, for example, have been found within comets and meteorites. This intriguing possibility also implies a high likelihood of life existing elsewhere in the cosmos.

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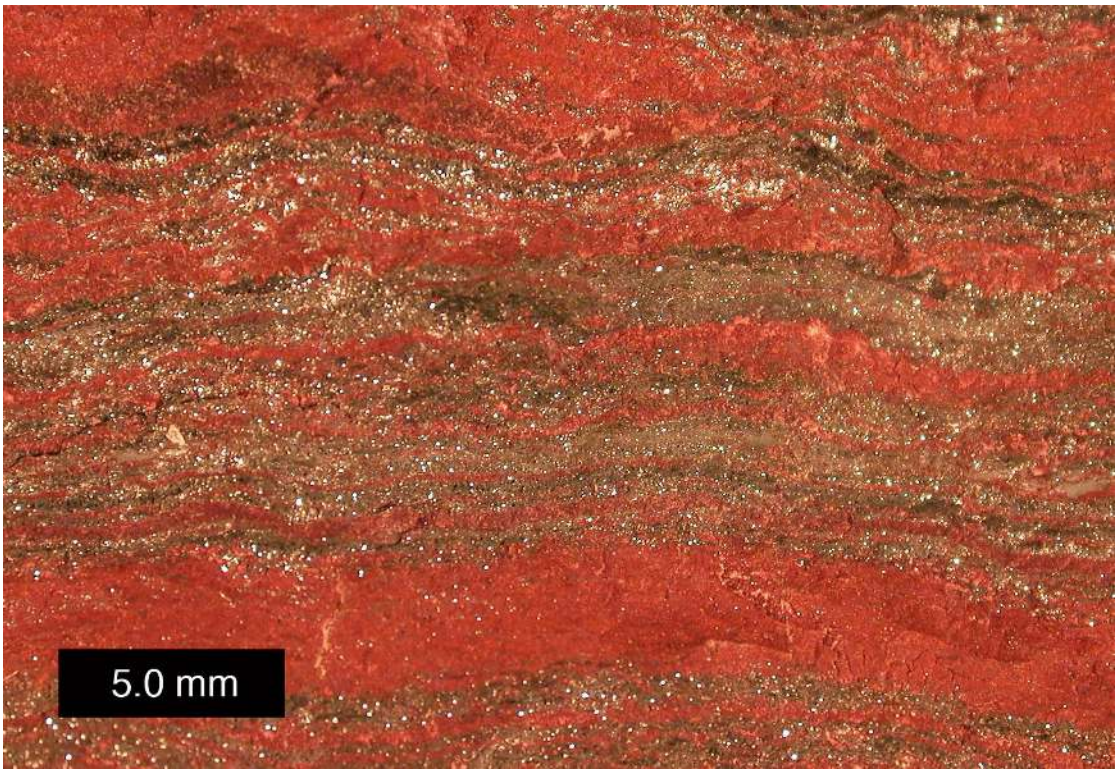
10.3: The Proterozoic Eon (2500-541 Ma)

The **Proterozoic Eon**, meaning “earlier life,” comes after the Archean Eon and ranges from 2.5 billion to 541 million years old. During this time, most of the central parts of the continents had formed, and plate tectonic processes had started. Photosynthesis by microbial organisms, such as single-celled cyanobacteria, had been slowly adding oxygen to the oceans. As cyanobacteria evolved into multicellular organisms, they completely transformed the oceans and the atmosphere by adding massive amounts of free oxygen gas (O_2) and initiating the **Great Oxygenation Event** (GOE).



Photosynthetic bacteria, like plants today, take energy from sunlight and carbon dioxide to produce free oxygen as a waste product.

This drastic environmental change decimated the anaerobic bacteria, which could not survive in the presence of free oxygen. On the other hand, aerobic organisms could thrive in ways they could not earlier. An oxygenated world also changed the chemistry of the planet in significant ways. For example, iron remained in solution in the non-oxygenated environment of the earlier Archean Eon. In chemistry, this is known as a reducing environment. Once the environment was oxygenated, iron combined with free oxygen to form solid precipitates of iron oxide, such as the mineral hematite or magnetite. These precipitates accumulated into large mineral deposits with red chert known as banded-iron formations dated at about 2 billion years.

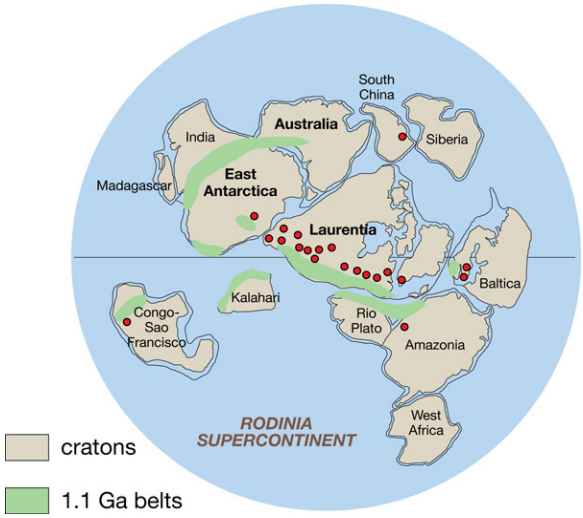


The banded iron formation, or BIF, forms when reduced iron in the water rusts to an oxidized red color in between silica minerals to give a banded appearance. These rocks herald the increase of free oxygen in the ocean and atmosphere.

The formation of iron oxide minerals and red chert in the oceans lasted a long time and prevented oxygen levels from increasing significantly since precipitation took the oxygen out of the water and deposited it into the rock strata. As oxygen continued to be produced and mineral precipitation leveled off, dissolved oxygen gas eventually saturated the oceans and bubbled out into the atmosphere. Oxygenation of the atmosphere is the biggest event that distinguishes the Archean and Proterozoic environments.

In addition to changing mineral and ocean chemistry, the GOE is also tabbed as triggering Earth's first glaciation event around 2.1 billion years ago, the Huron Glaciation. Free oxygen reacts with methane in the atmosphere to produce carbon dioxide. Carbon dioxide and methane are called greenhouse gases because they trap heat within the Earth's atmosphere, like the insulated glass of a greenhouse. Methane is a more effective insulator than carbon dioxide, so as the proportion of carbon dioxide in the atmosphere increased, the greenhouse effect decreased, and the planet cooled.

Rodinia



The supercontinent of Rodinia, which formed on the Earth approximately 1 billion years ago.

By the Proterozoic Eon, lithospheric plates had formed and were moving according to plate tectonic forces that were similar to current times. As the moving plates collided, the ocean basins closed to form a **supercontinent** called **Rodinia**. The supercontinent formed about 1 billion years ago and broke up about 750 to 600 million years ago, at the end of the Proterozoic. One of the resulting fragments was a continental mass called **Laurentia** that would later become North America. Geologists have reconstructed Rodinia by matching and aligning ancient mountain chains, assembling the pieces like a jigsaw puzzle, and using paleomagnetic to orient to the magnetic north.

The disagreements over these complex reconstructions are exemplified by geologists proposing at least six different models for the breakup of Rodinia to create Australia, Antarctica, parts of China, and the Tarim craton north of the Himalayas, Siberia, or the Kalahari craton of eastern Africa. This breakup created many shallow-water, biologically favorable environments that fostered the evolutionary breakthroughs marking the start of the next eon, the Phanerozoic.



Modern-day examples of stromatolites, or colonies of microbial mats that are photosynthetic, which are found in Shark Bay, Western Australia.

Early life in the Archean and earlier is poorly documented in the fossil record. Based on chemical evidence and evolutionary theory, scientists propose this life would have been single-celled photosynthetic organisms, such as the cyanobacteria that created **stromatolites**. Cyanobacteria produce free oxygen in the atmosphere through photosynthesis. Cyanobacteria, archaea, and bacteria are **prokaryotes** —primitive organisms made of single cells that lack cell nuclei and other organelles.



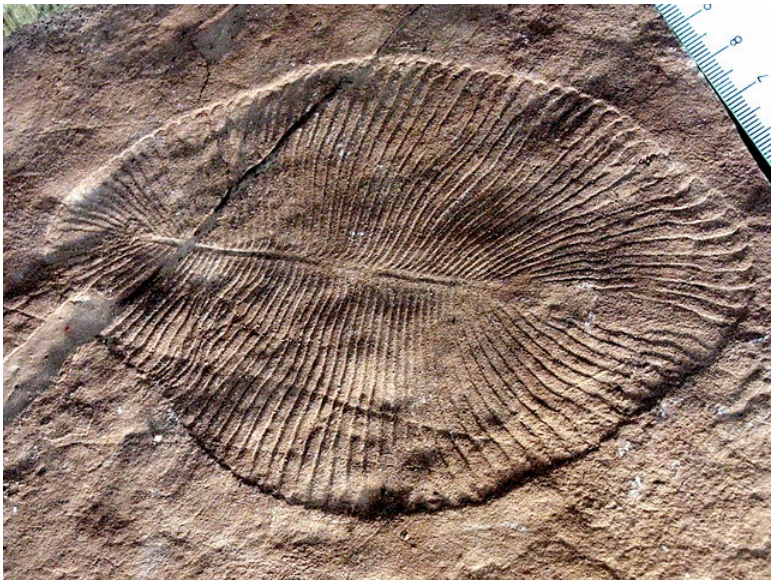
Fossilized cross-section of stromatolites

A large evolutionary step occurred during the Proterozoic Eon with the appearance of **eukaryotes** around 2.1 to 1.6 billion years ago. Eukaryotic cells are more complex, having nuclei and organelles. Nuclear DNA is capable of more complex replication and regulation than that of prokaryotic cells. The organelles include mitochondria for producing energy and chloroplasts for photosynthesis. The eukaryote branch in the tree of life gave rise to fungi, plants, and animals.

Another important event in Earth's biological history occurred about 1.2 billion years ago, when eukaryotes invented sexual reproduction. Sharing genetic material from two reproducing individuals, male and female, greatly increased genetic variability in

their offspring. This genetic mixing accelerated evolutionary change, increasing complexity among individual organisms and ecosystems.

Proterozoic land surfaces were barren of plants and animals, and geologic processes actively shaped the environment differently because land surfaces were not protected by leafy and woody vegetation. For example, rain and rivers would have caused erosion at much higher rates on land surfaces devoid of plants. This resulted in thick accumulations of pure quartz sandstone from the Proterozoic Eon, such as the extensive quartzite formations in the core of the Uinta Mountains in Utah.



The bizarre Ediacaran soft-bodied fossil of a genus of species now called Dickinsonia.

Fauna during the Ediacaran Period, 635.5 to 541 million years ago, is known as the **Ediacaran fauna** and offers a first glimpse at the diversity of ecosystems that evolved near the end of the Proterozoic. These soft-bodied organisms were among the first multicellular life forms and probably were similar to jellyfish or worm-like. Ediacaran fauna did not have hard parts like shells and was not well preserved in the rock records. However, studies suggest they were widespread in the Earth's oceans. Scientists still debate how many species were evolutionary dead-ends that became extinct and how many were ancestors of modern groupings. The transition of soft-bodied Ediacaran life to life forms with hard body parts occurred at the end of the Proterozoic and the beginning of the Phanerozoic Eons. This evolutionary explosion of biological diversity dramatically changed scientists' understanding of the history of life on Earth.

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CHAPTER OVERVIEW

11: The Paleozoic Era

11.1: Paleozoic Tectonics and Paleogeography

11.2: Paleozoic Evolution

11.3: The Permian Mass Extinction

Thumbnail: Trilobite. (Public domian; Heinrich Harder via Wikicommons)

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11.1: Paleozoic Tectonics and Paleogeography

What Did the World Map Look Like 541 to 252 Million Years Ago??

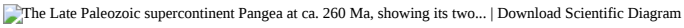


During the Paleozoic Era, sea levels rose and fell four times. With each sea-level rise, most of North America was covered by a shallow tropical ocean. Evidence of these submersions is the abundant marine sedimentary rocks such as limestone with fossils, corals, and ooids. Widespread unconformities document extensive sea-level falls. Today, the midcontinent has extensive marine sedimentary rocks from the Paleozoic. Western North America has thick layers of marine limestone on block faulted mountain ranges such as Mt. Timpanogos near Provo, Utah.



The supercontinent of Pangea, which was assembled around 300 million years ago.

The assembly of supercontinent **Pangea**, sometimes Pangaea, was completed by the late Paleozoic Era. The name Pangea was originally coined by Alfred Wegener and means “all land.” This supercontinent existed when all the major continents were grouped by a series of tectonic events, including subduction island-arc accretion, continental collisions, and ocean-basin closures. The Appalachian Mountains are the erosional remnants of these mountain-building events in North America. Surrounding Pangea was a global ocean basin known as the Panthalassa. Continued plate movement extended the ocean into Pangea through rifting, forming a large bay called the Tethys Sea that eventually divided the land mass into two smaller supercontinents, Laurasia and Gondwana. Laurasia consisted of Laurentia and Eurasia, and Gondwana comprised the remaining continents of South America, Africa, India, Australia, and Antarctica.



The breakup of Pangea around 260 Ma, shows the two smaller supercontinents, Laurasia and Gondwana, along with the Panthalassa Ocean and Tethys Ocean. Can you find where the Appalachian mountains were 260 million years ago?

While the east coast of North America was tectonically active during the Paleozoic Era, the west coast remained mostly inactive as a passive margin during the early Paleozoic. The western edge of the North American continent was near the present-day Nevada-Utah border and was an expansive shallow continental shelf near the paleo equator. By the end of the Paleozoic Era, the east coast of North America had a very high mountain range due to continental collision and the creation of Pangea. North America's west coast had smaller, isolated volcanic highlands associated with island arc accretion. During the Mesozoic Era, the size of the mountains on either side of North America would flip, with the west coast being a more tectonically active plate boundary and the east coast changing into a passive margin after the breakup of Pangea.

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11.2: Paleozoic Evolution

The beginning of the Paleozoic Era is marked by the first appearance of hard body parts like shells, spikes, teeth, and scales; and the appearance in the rock record of most animal phyla known today. Most basic animal body plans appeared in the rock record during the Cambrian Period. This sudden appearance of biological diversity is called the **Cambrian Explosion.** Scientists debate whether this sudden appearance is more from a rapid evolutionary diversification due to a warmer climate following the late Proterozoic glacial environments, better preservation and fossilization of hard parts, or artifacts of a more complete and recent rock record. For example, animals may have been diverse during the Ediacaran Period, setting the stage for the Cambrian Explosion. Still, they lacked hard body parts and would have left few fossils behind. Regardless, the Cambrian Period, 541-485 million years ago, marked the appearance of most animal phyla.



The diverse arthropods/shelled creatures found in the Burgess shale.

One of the best fossil sites for the Cambrian Explosion was discovered in 1909 by Charles Walcott (1850-1927) in the **Burgess Shale formation** in western Canada. The Burgess Shale is a **Lagerstätte** (LAGER-STAT-EH), a site of exceptional fossil preservation that includes impressions of soft body parts. This discovery allowed scientists to study Cambrian animals in immense detail because soft body parts are not normally preserved and fossilized.

Other Lagerstätte sites of similar age in China and Utah have allowed scientists to form a detailed picture of Cambrian biodiversity. The biggest mystery surrounds animals that do not fit existing lineages and are unique to that time. This includes many famous fossilized creatures: the first compound-eyed trilobites; *Wiwaxia*, a creature covered in spiny plates; *Hallucigenia*, a walking worm with spikes; *Opabinia*, a five-eyed arthropod with a grappling claw; and *Anomalocaris*, the alpha predator of its time, complete with grasping appendages and circular mouth with sharp plates. Most notably appearing during the Cambrian is an important ancestor to humans. A segmented worm called *Pikaia* is thought to be the earliest ancestor of the **Chordata** phylum that includes vertebrates and animals with backbones. The Chordata phylum contains all animals with spines, including humans!



Anomalacaris canadensis. Reconstruction at MUSE - Science Museum in Trento (CC BY-SA 3.0)

Anomalocaris, the apex predator of the Cambrian oceans.

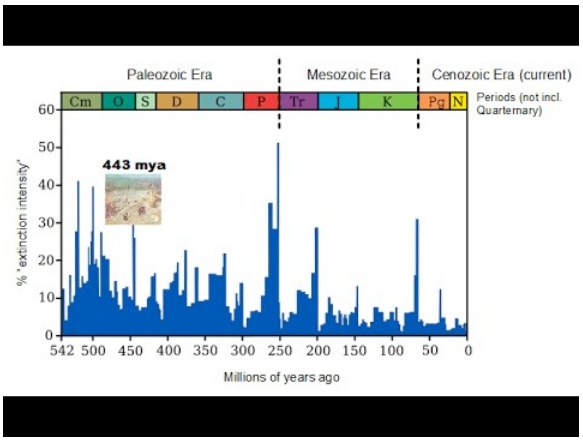
By the end of the Cambrian, mollusks, brachiopods, nautiloids, gastropods, graptolites, echinoderms, and trilobites covered the sea floor. Although most animal phyla appeared by the Cambrian, the biodiversity at the family, genus, and species level were low until the Ordovician Period. During the **Great Ordovician Biodiversification Event**, vertebrates and invertebrates (animals without a backbone) became more diverse and complex at the family, genus, and species levels.

The cause of the rapid speciation event is still debated. Still, some likely causes are a combination of warm temperatures, expansive continental shelves near the equator, and more volcanism along the mid-ocean ridges. Some have shown evidence that an asteroid breakup event and consequent heavy meteorite impacts correlate with this diversification event. The additional volcanism added nutrients to ocean water, helping support a robust ecosystem. Many life forms and ecosystems that would be recognizable in current times appeared at this time. Mollusks, corals, and arthropods, in particular, multiplied to dominate the oceans.

One important evolutionary advancement during the Ordovician Period was reef-building organisms, mostly colonial coral. Corals took advantage of the ocean chemistry, using calcite to build large structures that resembled modern reefs like the Great Barrier Reef off the coast of Australia. These reefs housed thriving ecosystems of organisms that swam around, hid in, and crawled over them. Reefs are important to paleontologists because of their preservation potential, massive size, and in-place ecosystems. Few other fossils offer more diversity and complexity than reef assemblages.

Ordovician Mass Extincition

According to evidence from glacial deposits, a small ice age caused sea levels to drop and led to major mass extinction by the end of the Ordovician. This is the earliest of five **mass extinction** events documented in the fossil record. During this mass extinction, an unusually large number of species abruptly disappear in the fossil record (see video).

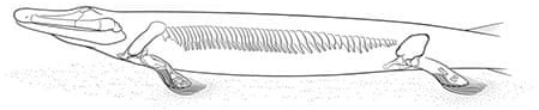


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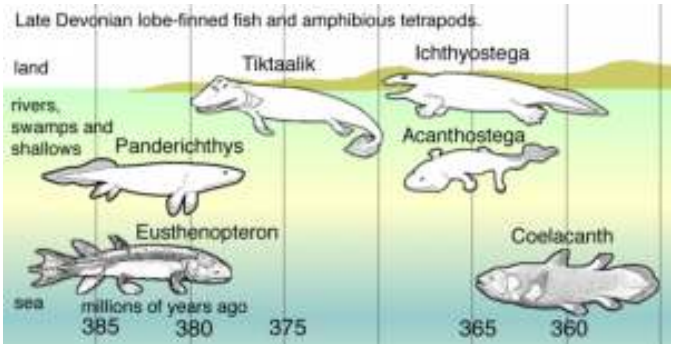
Life bounced back during the Silurian period. The major evolutionary event was the development of the forward pair of gill arches into jaws, allowing fish new feeding strategies and opening up new ecological niches. This period also included the start of armored fishes, known as the placoderms. In addition to fish and jaws, Silurian rocks provide the first evidence of terrestrial or land-dwelling plants and animals. The first vascular plant, *Cooksonia*, had woody tissues, pores for gas exchange, and veins for water and food transport. Insects, spiders, scorpions, and crustaceans began to inhabit moist, freshwater terrestrial environments.

The Devonian Period is called the "Age of Fishes" due to the rise in plated, jawed, and lobe-finned fishes. The lobe-finned fishes, which were related to the modern lungfish and coelacanth, are important for their eventual evolution into tetrapods, four-limbed vertebrate animals that can walk on land. The first lobe-finned land-walking fish, named *Tiktaalik*, appeared about 385 million years ago and served as a transitional fossil between fish and early tetrapods. Though Tiktaalik was a fish, it also had some tetrapod structures.



Illustrated skeleton of the lobe-finned fish Tiktaalik.

Several fossils from the Devonian are more tetrapod-like than fish, but these weren't fully terrestrial. The first fully terrestrial tetrapod arrived in the Mississippian (early Carboniferous) period. By the Mississippian (early Carboniferous) period, tetrapods had evolved from a common tetrapod ancestor into two main groups, amphibians, and amniotes. The amphibians could breathe air and live on land but still needed water to nurture their soft eggs. The first reptile (an amniote) could live and reproduce entirely on land with hard-shelled eggs that wouldn't dry out.



Evolution of lobe-finned fishes to land amphibians, such as Ichthyostega

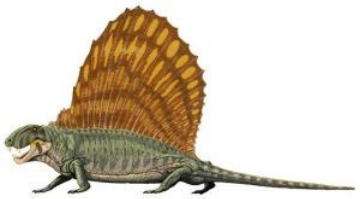
Land plants had also evolved into the first trees and forests. Toward the end of the Devonian, another mass extinction event occurred. This extinction, while severe, is the least temporally defined, with wide variations in the timing of the event or events. Reef-building organisms were the hardest hit, leading to dramatic changes in marine ecosystems.

The next period called the Carboniferous (North American geologists have subdivided this into the Mississippian and Pennsylvanian periods), saw the highest levels of oxygen ever known, with forests (e.g., ferns, club mosses) and swamps dominating the landscape. This helped cause the largest arthropods ever, like the millipede *Arthropleura*, at 2.5 meters (6.4 feet) long!



Arthropleura, one of the giant insects of the Carboniferous period, measured over 6 feet long. This period's very high oxygen levels, brought by immense vegetation, allowed arthropods to grow to large sizes.

It also saw the rise of a new group of animals, the reptiles. The evolutionary advantage reptiles have over amphibians is the amniote egg (egg with a protective shell, i.e., think chicken eggs), which allows them to rely on non-aquatic environments for reproduction. This widened the terrestrial reach of reptiles compared to amphibians. This booming life, especially plant life, created cooling temperatures as carbon dioxide was removed from the atmosphere. By the middle Carboniferous, these cooler temperatures led to an ice age (called the Karoo Glaciation) and less-productive forests. The reptiles fared much better than the amphibians, leading to their diversification. This glacial event lasted into the early Permian.

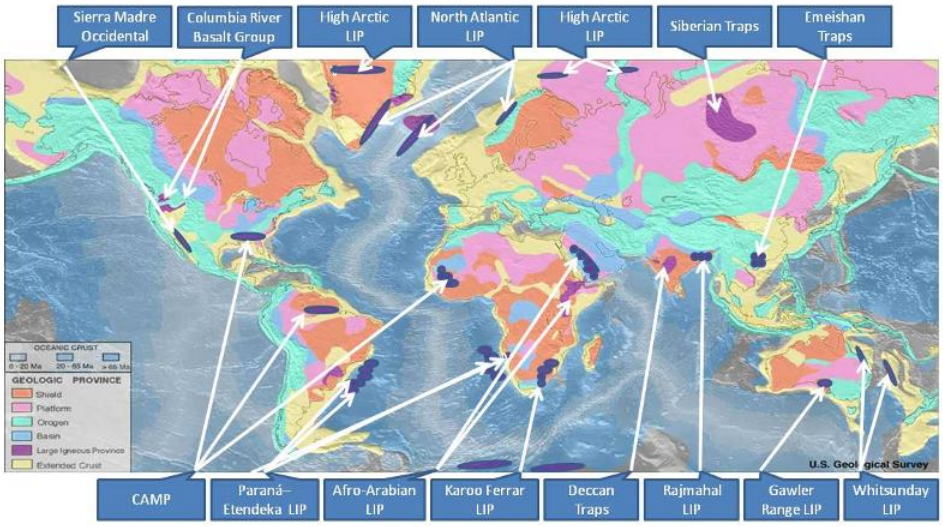


The synapsid reptile Dimetrodon lived in the Permian period. This reptile was thought to regulate its body temperature using the enormous sail on its back.

By the Permian, with Pangea assembled, the supercontinent led to a dryer climate and even more diversification and domination by the reptiles. The groups that developed in this warm climate eventually radiated into dinosaurs. Another group, known as the synapsids, eventually evolved into mammals. Synapsids, including the famous sail-backed *Dimetrodon*, are commonly confused with dinosaurs. Pelycosaurs (of the Pennsylvanian to early Permian like *Dimetrodon, above)* are the first group of synapsids exhibiting mammalian characteristics such as well-differentiated dentition incisors, highly developed canines in lower and upper jaws and cheek teeth, premolars and molars. Starting in the late Permian, the second group of synapsids, called the therapsids (or mammal-like reptiles) evolve, and became the ancestors of mammals.

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11.3: The Permian Mass Extinction



The locations of flood basalt volcanism that affected the world throughout its history. In particular, the Siberian traps during the Permian period were the world's largest flood basalts/volcanism.

The largest mass extinction marks the end of the Paleozoic era in earth's history. The Paleozoic era had two smaller mass extinctions, but these were not as large as the **Permian Mass Extinction**, also known as the Permian-Triassic Extinction Event. It is estimated that 96% of marine species and 70% of land-dwelling (terrestrial) vertebrates went extinct. Many famous organisms, like sea scorpions and trilobites, were never seen again in the fossil record.

What caused such a widespread extinction event? The exact cause is still debated, though the leading idea relates to extensive volcanism associated with the **Siberian Traps**, one of the largest deposits of flood basalts known on Earth, dating to the time of the extinction event. The eruption size is estimated at over 3 million cubic kilometers, a continuous eruption approximately 4,000,000 times larger than the famous 1980 Mt. St. Helens eruption in Washington.

The unusually large volcanic eruption would have contributed many toxic gases, aerosols, and greenhouse gasses into the atmosphere. Further, some evidence suggests that the volcanism burned vast coal deposits, releasing methane (a greenhouse gas) into the atmosphere. As discussed, greenhouse gases cause the climate to warm. This extensive addition of greenhouse gases from the Siberian Traps may have caused a runaway greenhouse effect that rapidly changed the climate, acidified the oceans, disrupted food chains, disrupted carbon cycling, and caused the largest mass extinction.

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CHAPTER OVERVIEW

12: The Mesozoic Era

12.1: Mesozoic Tectonics and Paleogeography

12.2: Mesozoic Evolution

12.3: The K-T Extinction

Thumbnail: Iconic “Berlin specimen” Archaeopteryx lithographica fossil from Germany.

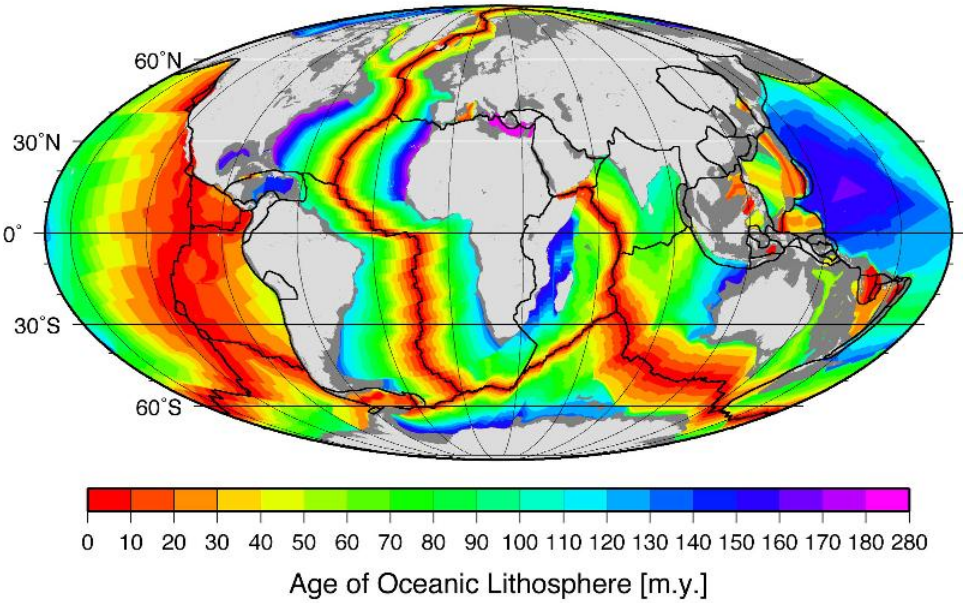
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12.1: Mesozoic Tectonics and Paleogeography



Animation showing Pangea breaking up

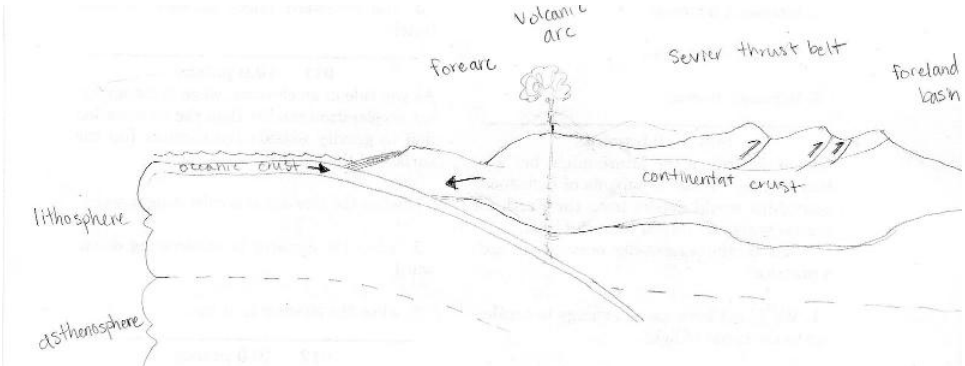
Pangea started breaking up (in a region that would become eastern Canada and the United States) around 210 million years ago in the Late Triassic. Clear evidence includes the age of the sediments in the Newark Supergroup rift basins, the Palisades sill of the eastern part of North America, and the age of the Atlantic ocean floor. Due to sea-floor spreading, the oldest rocks on the Atlantic floor are along the coast of northern Africa and the east coast of North America, while the youngest are along the mid-ocean ridge.



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

This age pattern shows how the Atlantic Ocean opened as the young Mid-Atlantic Ridge began to create the seafloor. This means the Atlantic ocean started opening and was first formed here. The southern Atlantic opened next, with South America separating from central and southern Africa. Last (after the Mesozoic ended) was the northernmost Atlantic, with Greenland and Scandinavia parting ways. The breaking points of each rifted plate margin eventually turned into the passive plate boundaries of the east coast of the Americas today.

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Sketch of the major features of the Sevier Orogeny (also known as a subduction zone leading to mountain building).

In western North America, a plate boundary had started with the subduction of ancient oceanic crust, controlling most of the tectonics of that region in the Mesozoic. Another possible collision with oceanic crust created the Sonoman Orogeny (orogeny= mountain belt) in Nevada during the latest Paleozoic to the Triassic. In the Jurassic, another collision caused the Nevadan Orogeny, a large volcanic belt similar to the Andes mountains today. The Sevier Orogeny followed in the Cretaceous, mainly a volcanic arc to the west and a thin-skinned fold and thrust belt to the east, meaning stacks of shallow faults and folds built up the topography. Many of the structures in the Rocky Mountains today date from this orogeny.



The Cretaceous Interior Seaway in the mid-Cretaceous.

Tectonics influenced one more important geographic feature in North America: the Cretaceous Western Interior Foreland Basin, which flooded during high sea levels forming the **Cretaceous Interior Seaway**. Subduction of Farallon Plate, an oceanic plate connected to the Pacific Plate (seen today as remnants such as the Juan de Fuca Plate, off the coast of the Pacific Northwest), occurred on the West coast of Northern America. Subduction was shallow at this time because a very young, hot, and less dense portion of the Farallon plate was subducted. This shallow subduction caused a down warping in the central part of North America, which created the basin that flooded to become an interior sea.

High sea levels due to shallow subduction, increasing seafloor spreading and subduction rates, high temperatures, and melted ice also contributed to the high sea levels. These factors allowed a shallow epicontinental seaway that extended from the Gulf of Mexico to the Arctic Ocean to divide North America into two separate land masses, Laramidia to the west and Appalachia to the east, for 25 million years. Many of the coal deposits in Utah and Wyoming formed from swamps along the shores of this seaway. By the end of the Cretaceous, cooling temperatures caused the seaway to regress.

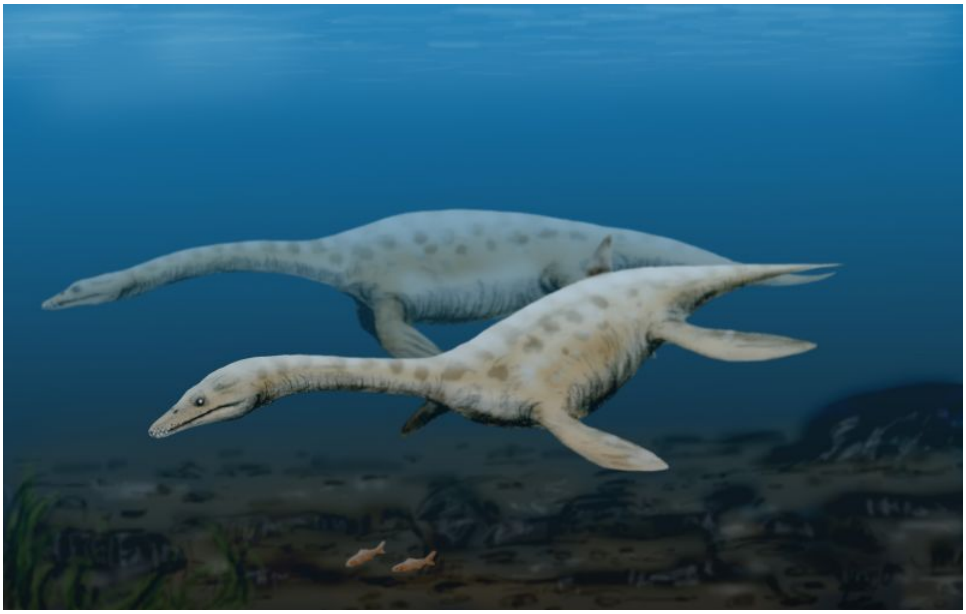
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12.2: Mesozoic Evolution



A Mesozoic scene from the late Jurassic.

The Mesozoic era is dominated by reptiles, specifically the dinosaurs. The Triassic saw devastated ecosystems that took over 30 million years to fully re-emerge after the Permian Mass Extinction. The first appearance of many modern groups of animals that would later flourish occurred at this time. This includes frogs (amphibians), turtles (reptiles), marine ichthyosaurs and plesiosaurs (marine reptiles), mammals, and archosaurs. The archosaurs ("ruling reptiles") include ancestral groups that went extinct at the end of the Triassic, the flying pterosaurs, crocodilians, and the dinosaurs. Archosaurs, like the placental mammals after them, occupied all major environments: terrestrial (dinosaurs), in the air (pterosaurs), aquatic (crocodilians), and even fully marine habitats (marine crocodiles). The pterosaurs, the first vertebrate group to take flight, like the dinosaurs and mammals, start small in the Triassic.



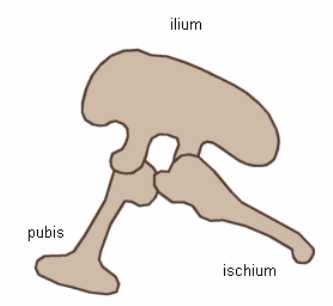
A drawing of the early plesiosaur Agustasaurus from the Triassic of Nevada.

At the end of the Triassic, another mass extinction event occurred, the fourth major mass extinction in the geologic record. This was perhaps caused by the Central Atlantic Magmatic Province flood basalt. The end-Triassic extinction made certain lineages go extinct and helped spur the evolution of survivors like mammals, pterosaurs (flying reptiles), ichthyosaurs/plesiosaurs/mosasaurs (marine reptiles), and dinosaurs.

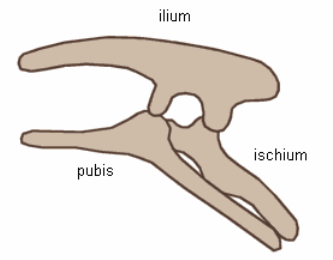


Reconstruction of the small (<5″) Megazostrodon, one of the first animals considered a true mammal.

Mammals, as previously mentioned, got their start from a reptilian synapsid ancestor, possibly in the late Paleozoic. Mammals stayed small, in mainly nocturnal niches, with insects being their largest prey. The development of warm-blooded circulation and fur may have responded to this lifestyle.



The open structure of a saurischian hip is similar to a lizard's.



The closed structure of an ornithischian hip is similar to a bird's.

In the Jurassic, previously common species flourished due to a warmer and more tropical climate. The dinosaurs were relatively small animals in the Triassic period of the Mesozoic but became truly massive in the Jurassic. Dinosaurs are split into **two groups**

based on their hip structure, i.e., the orientation of the pubis and ischium bones in relationship to each other. These orientations are referred to as the “reptile-hipped” **saurischians** and the “bird-hipped” **ornithischians**. However, this idea has recently been questioned by a new idea for dinosaur lineage.

The bird-hipped dinosaurs include many herbivorous species, such as the armored Stegosaurus, Ankylosaurus, the horned Triceratops, and the duck-billed Hadrosaurs.

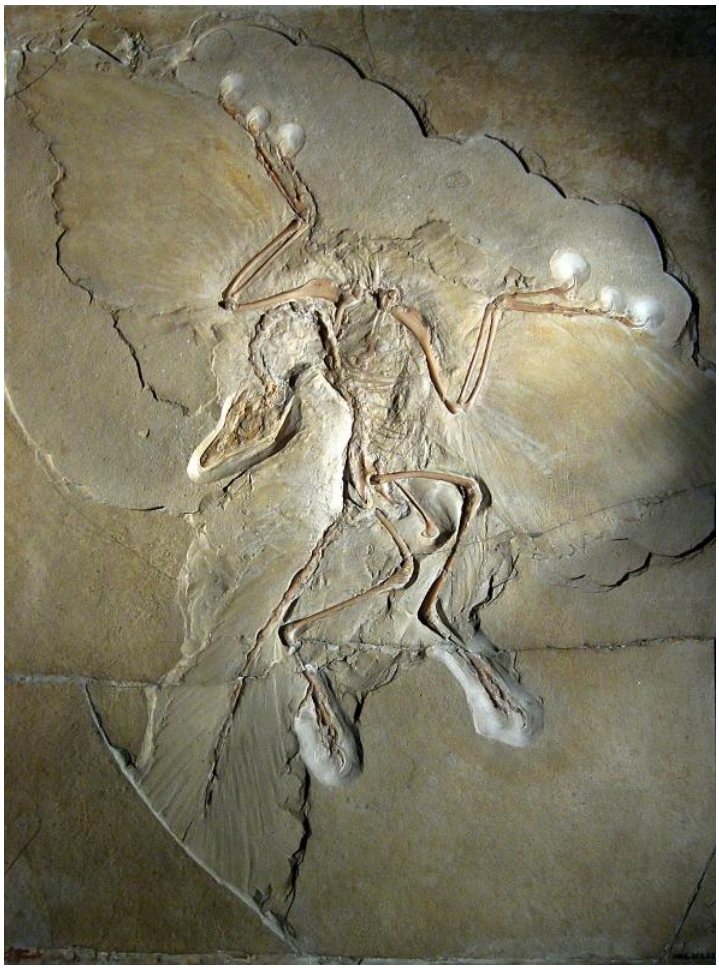
The lizard-hipped dinosaurs were divided into two groups – the herbivorous sauropods, such as the 100-ton, long-necked brachiosaurus and apatosaurus, and the more vicious and carnivorous theropods, such as Tyrannosaurus Rex and Velociraptor. Strangely enough, modern birds have evolved from theropods, with the “lizard-hip” rather than dinosaurs that possessed the “bird-hip.”

The major adaptive advantage dinosaurs had was changes in the hip and ankle bones, tucking the legs under the body for improved locomotion as opposed to the semi-erect gait of crocodiles or the sprawling posture of reptiles.



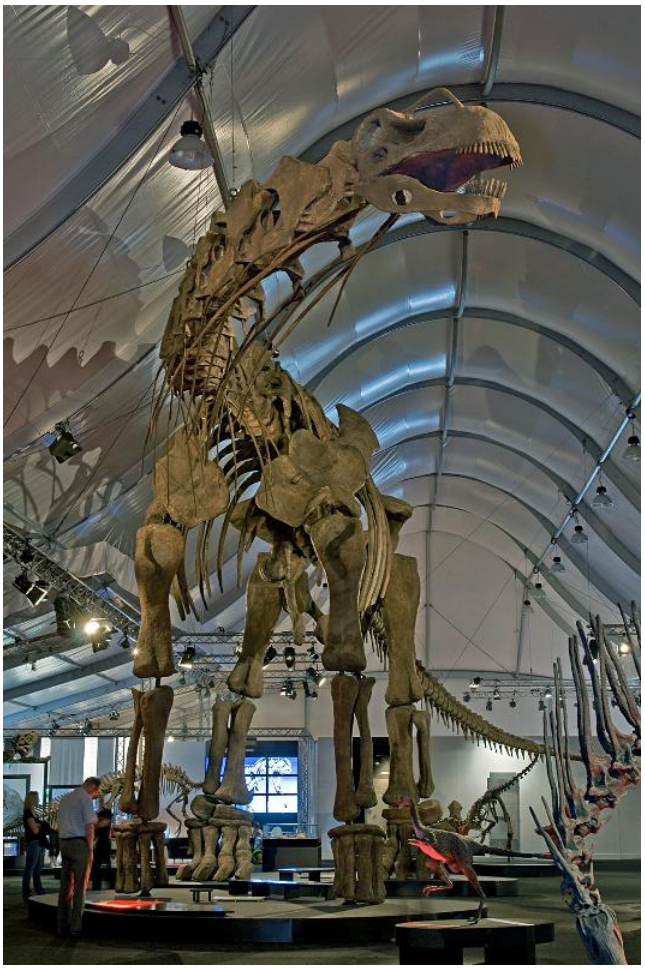
Therizinosaurs, like Beipiaosaurus (shown in this restoration), are known for their enormous hand claws.

There is a paucity of dinosaur fossils from the Early and Middle Jurassic, but by the Late Jurassic, they were dominating the planet.



Iconic “Berlin specimen” Archaeopteryx lithographica fossil from Germany.

The pterosaurs grew and diversified in the Jurassic, and another notable aerial organism developed and thrived in the Jurassic: birds. When *Archaeopteryx* was found in the Solnhofen Lagerstätte of Germany, a seeming dinosaur-bird hybrid, it started the conversation on the origin of birds. The idea that birds evolved from dinosaurs occurred very early in the history of research into evolution, only a few years after Darwin’s *On the Origin of Species*. This study used a remarkable fossil of Archeopteryx, which had visible feathers preserved, from a transitional animal between dinosaurs and birds. Small meat-eating theropod dinosaurs were likely the branch that became birds due to their similar features. A significant debate still exists over how and when powered flight evolved. Some have stated a running-start model, while others have favored a tree-leaping gliding model or even a semi-combination: flapping to aid climbing.



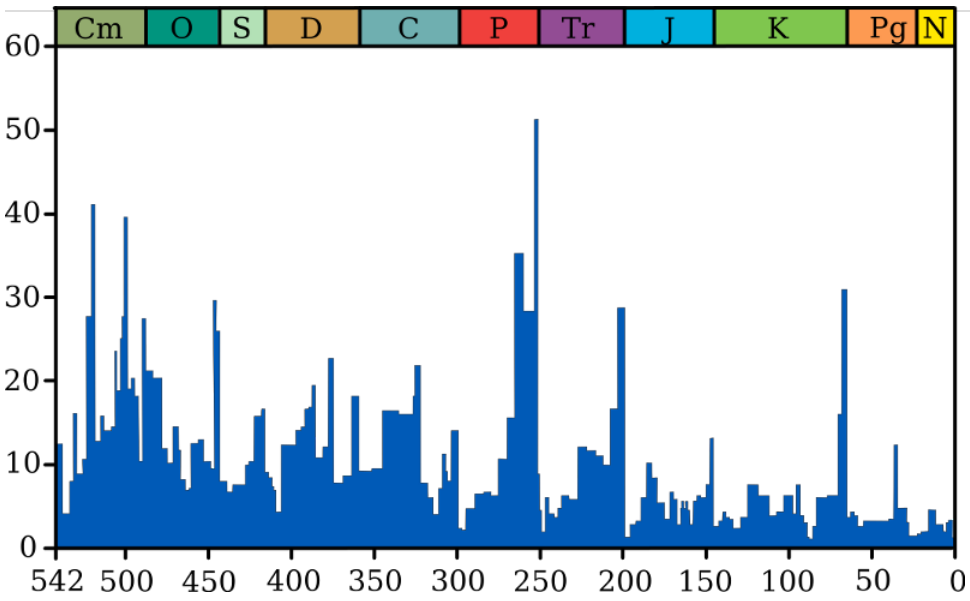
Reconstructed skeleton of Argentinosaurus, from Naturmuseum Senckenberg in Germany.

The Cretaceous saw a further diversification, specialization, and domination of the dinosaurs and other fauna. One of the biggest changes on land was the transition to angiosperm-dominated flora. Angiosperms, plants with flowers and seeds, originated in the Cretaceous, switching many plains to grasslands by the end of the Mesozoic. By the end of the period, they had replaced gymnosperms (evergreen trees) and ferns as the dominant plant in the world's forests. Haplodiploid eusocial insects (bees and ants) are descendants of Jurassic wasp-like ancestors that co-evolved with the flowering plants during this period.

The breakup of Pangea not only shaped our modern world's geography but biodiversity at the time as well. Throughout the Mesozoic, animals on the isolated, now separated island continents (formerly parts of Pangea) took strange evolutionary turns. This includes giant titanosaurian sauropods (*Argentinosaurus*) and theropods (*Giganotosaurus*) from South America.

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12.3: The K-T Extinction



Graph of the rate of extinction. Note the large spike at the end of the Cretaceous (labeled as K).

Similar to the end of the Paleozoic era, the Mesozoic Era ended with the K-Pg Mass Extinction (previously known as the K-T Extinction) 66 million years ago. This extinction event was likely caused by a large bolide (an extraterrestrial impactor such as an asteroid, meteoroid, or comet) that collided with earth. Ninety percent of plankton species, 75% of plant species, and all the dinosaurs went extinct.



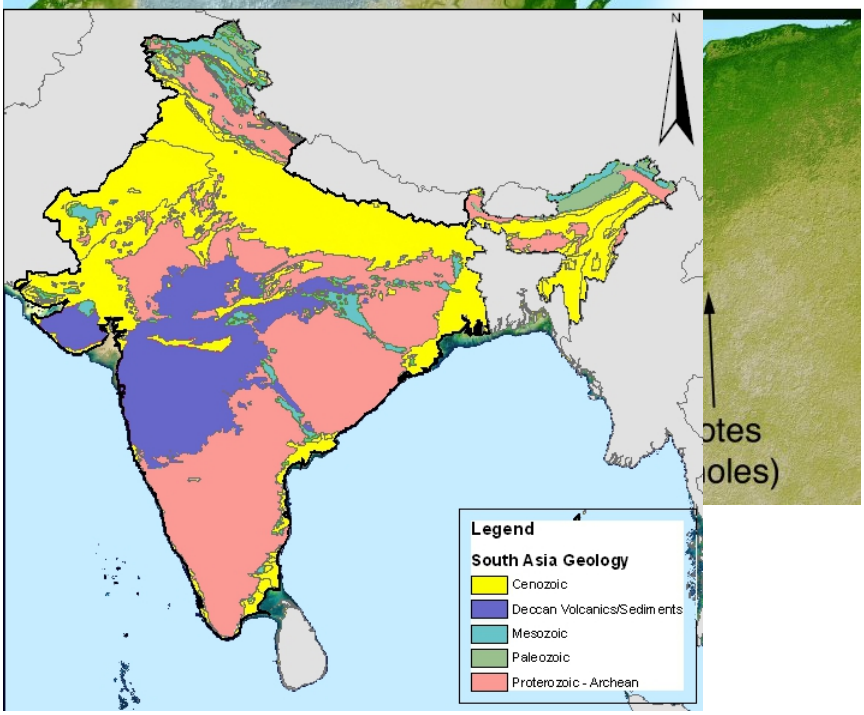
Artist's depiction of an impact event

One of the strongest pieces of evidence comes from the element iridium. Quite rare on Earth and more common in meteorites, it has been found worldwide in higher concentrations at a particular layer of rock that formed at the time of the K-T boundary. Soon other scientists started to find evidence to back up the claim. Melted rock spheres, a special type of "shocked" quartz called stishovite found only at impact sites, were found in many places worldwide. The huge impact created a strong thermal pulse that could be responsible for global forest fires, strong acid rains, a corresponding abundance of ferns, the first colonizing plants after a forest fire, enough debris thrown into the air to significantly cool temperatures afterward, and a 2-km high tsunami inferred from deposits found from Texas to Alabama.



The land expression of the Chicxulub crater. The other side of the crater is within the Gulf of México.

Still, with all this evidence, one large piece remained missing: the crater where the bolide impacted. It was not until 1991 that the crater was confirmed using petroleum company geophysical data. Even though it is the third largest confirmed crater on Earth at roughly 180 km wide, the Chicxulub Crater was hard to find due to being partially underwater and partially obscured by the dense forest canopy of the Yucatan Peninsula. Coring of the center of the impact, called the peak ring, contained granite, indicating the impact was so powerful that it lifted basement sediment from the crust several miles toward the surface. In 2010, an international team of scientists reviewed 20 years of research and blamed the extinction on the impact.



Geology of India, showing purple as Deccan Traps-related rocks.

With all of this information, it seems like the case would be closed. However, other events at this time could have partially aided the demise of so many organisms. For example, sea levels are known to slowly decrease during the K-T event, which is tied to marine extinctions. However, any study on gradual vs. sudden changes in the fossil record is flawed due to the incomplete nature of the fossil record. Another big event at this time was India's Deccan Traps flood basalt volcanism. At over 1.3 million cubic kilometers of material, it was a large source of material hazardous to ecosystems at the time. It has been suggested as at least partially responsible for the extinction. Some have found the impact and eruptions too much of a coincidence and even linked the two.

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CHAPTER OVERVIEW

13: The Cenozoic Era

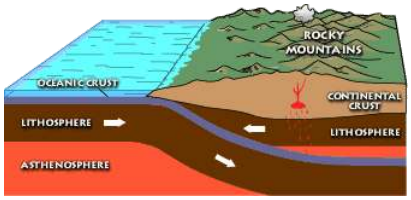
13.1: Cenozoic Tectonics and Paleogeography

13.2: Cenozoic Evolution

13.3: Origins of Humans and the Anthropocene

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13.1: Cenozoic Tectonics and Paleogeography



In the Cenozoic, the Earth's tectonic plates moved into more familiar places, with the biggest change being the closing of the Tethys Sea with collisions such as the Alps, Zagros, and Himalaya. This collision started about 57 million years ago and continues today. Maybe the most significant tectonic feature that occurred in the Cenozoic of North America was the conversion of the west coast of California from a convergent boundary subduction zone to a transform boundary. Subduction off the western United States coast, which had occurred throughout the Mesozoic, had continued in the Cenozoic.

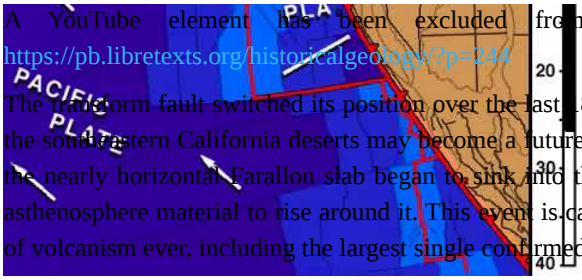
After the Sevier Orogeny in the late Mesozoic, a subsequent orogeny called the Laramide Orogeny occurred in the early Cenozoic. The Laramide was thick-skinned, different than the Sevier Orogeny. It involved deeper crustal rocks and produced bulges that would become mountain ranges like the Rockies, Black Hills, Wind River Range, Uinta Mountains, and the San Rafael Swell. Instead of descending directly into the mantle, the subducting plate shallowed out. It moved eastward beneath the continental plate affecting the overlying continent hundreds of miles east of the continental margin and building high mountains. This occurred because the subducting plate was so young and near the spreading center, and the density of the plate was therefore low, and subduction was hindered.



As the mid-ocean ridge itself started to subduct, the relative motion changed. Subduction caused a relative convergence between the subducting Farallon plate and the North American plate. On the other side of the mid-ocean ridge from the Farallon plate was the Pacific plate, which was moving away from the North American plate. Thus, as the subduction zone consumed the mid-ocean ridge, the relative movement became a transform boundary instead of convergent (View the video below to understand better how this occurred), becoming the San Andreas Fault System. As the San Andreas grew, it caused east-west directed extensional forces to spread over the western United States, creating the **Basin and Range province (where we are in Arizona!).** These forces built a system of mountains and valleys in the Western United States.



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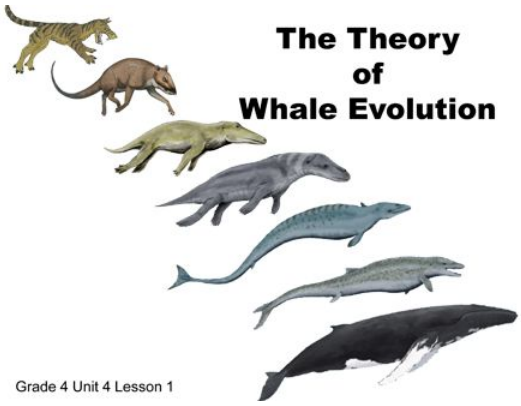


The transform fault switched its position over the last 18 million years, twisting the mountains around Los Angeles. New faults in the southeastern California deserts may become a future San Andreas-style fault. During this switch from subduction to transform, the nearly horizontal Farallon slab began to sink into the mantle. This caused magmatism as the subducting slab sank, allowing asthenosphere material to rise around it. This event is called the Oligocene ignimbrite flare-up, one of the most significant periods of volcanism ever, including the largest single confirmed eruption, the 5000 cubic kilometer Fish Canyon Tuff.

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13.2: Cenozoic Evolution

Five groups of early mammals in the fossil record, based primarily on fossil teeth, are the hardest bone in vertebrate skeletons. For this text, the most important group is the Eupantotheres, which diverge into the two main groups of mammals, the marsupials (*Sinodelphys*) and placentals or eutherians (*Eomaia*) in the Cretaceous and then diversified in the Cenozoic. The marsupials dominated on the isolated island continents of South America and Australia, and many went extinct in South America after the introduction of placental mammals. Some well-known mammal groups have been highly studied with interesting evolutionary stories in the Cenozoic. For example, horses started small with four toes, grew larger, and had just one toe.



Cetaceans (marine mammals like whales and dolphins) started on land from small bear-like (mesonychids) creatures in the early Cenozoic and gradually took to the water. For example, whales from the Eocene Period (56-34 million years ago) bear vestigial back limbs near their tails, indicating their ancestral past on land. An interesting example of a whale during this period is the large and carnivorous Basilosaurus ("king lizard"), which measured up to 65 feet long and had small hind limbs.

What would Basilosaurus Look Like Today?

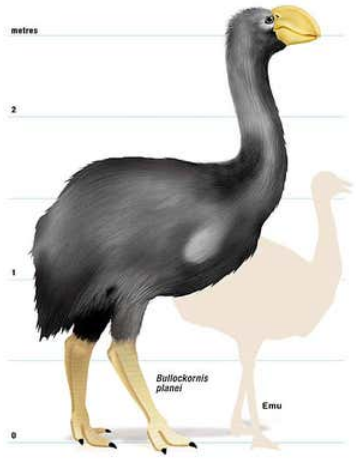


Another frightening undersea predator of the Cenozoic was the Megalodon ("big tooth"), a gigantic shark that lived during the Miocene and Pliocene, approximately 23 to 3.6 million years ago. This shark was approximately two to three times the size of a great white shark!



Larger animals were not simply limited to the seas in the early to mid-Cenozoic. Gigantic land mammals ruled the continents in Pliocene and Pleistocene epochs (5.4 million to 11,700 years ago). There were mammals we commonly identify in pop cultures, such as the saber-toothed tiger, *Smilodon*, or the wooly mammoth or mastodons. However, other mammals were also larger – the giant beaver, *Castoroides*, the wooly rhino, *Coelodonta*, large cattle such as the auroch, and giant deer (*Megaloceros*). Other gigantic animals included sloths (*Megatherium* in South America was a whopping three tons!) and birds such as penguins (*Palaeeudyptinae*) and ducks (the demon duck of doom, *Bullockornis*).





The demon duck of doom, standing far taller than today's modern emus.

Many gigantic animals, megafauna, died off approximately 50,000 to 2,000 years ago. There are two main reasons why we do not have as many large mammals today. The first reason is a shift in climate; the climate in the Pliocene and Pleistocene experienced an ice age, which transitioned into a warmer interglacial period at the beginning of the Quaternary period, in which we currently live. This shift caused a lot of plants to go extinct, so the herbivorous large mammals also died off.

However, it is also not a coincidence that humans began to evolve around the time most of these large mammals went extinct. Scientists hypothesize that humans may have overhunted some of these mammals while already suffering a food shortage. These two effects combined may have led to the downfall of many of the large mammals.

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13.3: Origins of Humans and the Anthropocene

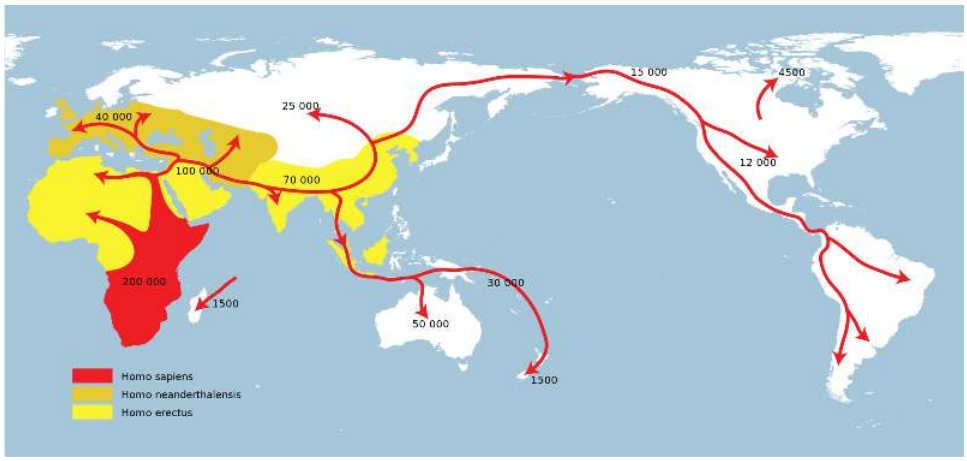
No study of evolution has been more studied than human evolution. **Hominids**, the name for human-like primates, started in eastern Africa several million years ago.



"Lucy", of the species Australopithecus afarensis, one of the first bipedal ancestors of humans.

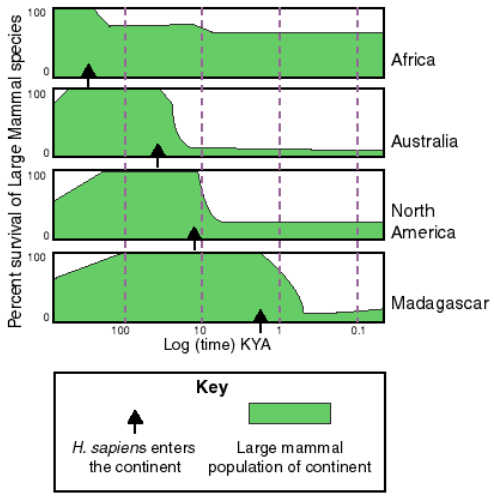
The first critical event in this story is an environmental change from jungle to more of a savanna, probably caused by changes in Indian Ocean circulation. While bipedalism is known to have evolved before this shift, it is generally believed that our bipedal ancestors (like *Australopithecus*) had an advantage by covering ground more easily in a more open environment compared to their non-bipedal evolutionary cousins. There is also a growing body of evidence, including the famous "Lucy" fossil of an Australopithecine, that our early ancestors lived in trees.

Arboreal animals usually demand a high intelligence to navigate through a three-dimensional world. From this lineage, humans evolved, using endurance running to acquire more resources and possibly even hunt. This can explain many uniquely human features, from our long legs, strong Achilles, lack of lower gut protection, and our wide range of running efficiencies.



Now that the hands are freed up, the next big step is a large brain. There have been arguments from a switch to more meat-eating, cooking with fire, tool use, and even the construct of society itself to explain this increase in brain size. Regardless, this increased cognitive power allowed humans to reign as their ancestors moved out of Africa and explored the world, ultimately entering the Americas through land bridges like the Bering Land Bridge. The details of this worldwide migration and the different branches of the hominid evolutionary tree are very complex and best reserved for their course.

Anthropocene and Extinction

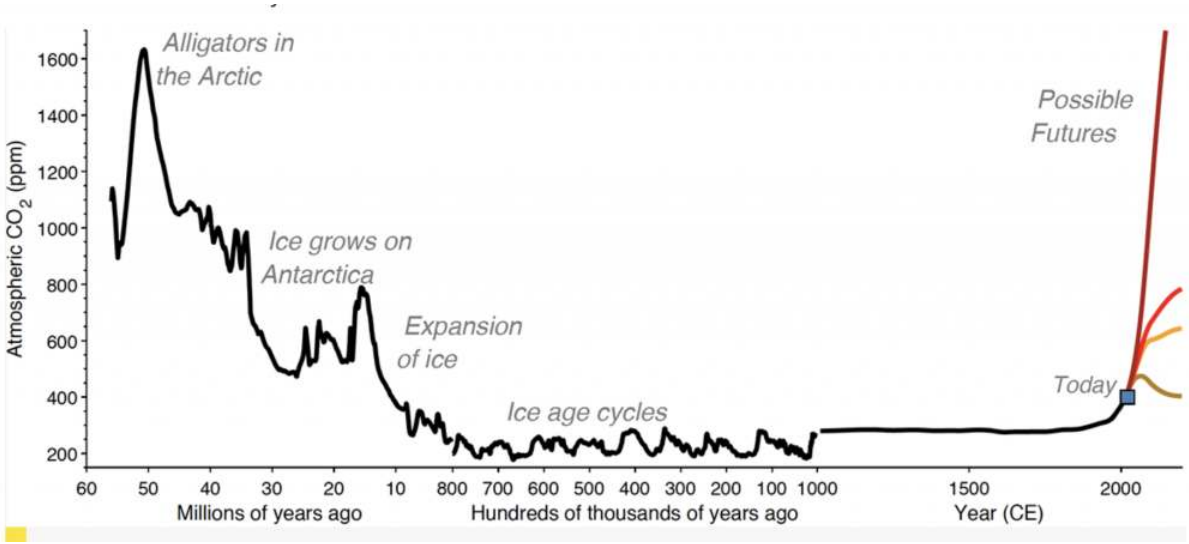


Humans have influenced the Earth, its ecosystems, and climate. Yet, human activity cannot explain all of the changes that have occurred in the recent past. The start of the Quaternary period, the last and current period of the Cenozoic, is marked by the start of our current ice age 2.58 million years ago. Ice sheets advanced and retreated during this period, most likely due to Milankovitch cycles, which are natural shifts in the Earth's climate due to the Earth's eccentric orbit around the Sun and tilt. Miniature temperate or colder periods within overall warm or ice ages occur due to a Milankovitch cycle every 23,000 years.

At this time of the Ice Age, various cold-adapted megafauna emerged (like giant sloths, saber-tooth cats, and woolly mammoths), and most of them went extinct as the Earth warmed from the most recent glacial maximum. A long-standing debate is over the cause of these and other extinctions. Is climate warming to blame, or were they caused by humans? Certainly, we know of the recent human extinction of animals like the dodo or passenger pigeon. Can we connect modern extinctions to extinctions in the recent past? If so, there are several ideas as to how this happened. Possibly the most widely accepted and oldest is the hunting/overkill hypothesis. The idea behind this hypothesis is that humans hunted large herbivores for food. Carnivores could not find food, and human arrival times in locations are often tied to increased extinction rates.



Modern human impact on the environment and the Earth is unquestioned. Many scientists are starting to suggest that the rise of human civilization ended and/or replaced the Holocene epoch and defined a new geologic time interval: the **Anthropocene**. Evidence for this change includes extinctions, increased tritium (hydrogen with two neutrons) due to nuclear testing, rising pollutants like carbon dioxide, increased warming of the climate, more than 200 never-before-seen mineral species that have occurred only in this epoch, materials such as plastic and metals which will be long lasting ” fossils ” in the geologic record, and large amounts of earthen material moved.



Atmospheric CO_2 over the past 66 million years. Fifty million years ago, it was so warm that fossilized alligators were found in the Arctic. The last time current CO_2 levels were this high was 3 million years ago. Possible futures be conservative or extreme. Image by James Rae.

The biggest scientific debate on this topic is the starting point. Some say that humans' invention of agriculture would be recognized in geologic strata, which should be the starting point, around 12,000 years ago. Others link the start of the industrial revolution and the subsequent addition of vast amounts of carbon dioxide to the atmosphere. Either way, the idea is that alien geologists visiting Earth in the distant future would easily recognize the impact of humans on the Earth as the beginning of a new geologic period.

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