HISTORICAL GEOLOGY



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Historical Geology

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This text was compiled on 05/19/2026

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CHAPTER OVERVIEW

1: What is Historical Geology?

Who do you know that tells great stories?

A friend, a grandparent, a sibling, a teacher, an author, a director, a singer... Gifted storytellers come in many forms, and tell their tales in many styles.

Now imagine all the stories you have not heard. Imagine the stories you could hear if you could speak every language in the world, and sit across the table from people with fundamentally different experiences, speaking of ancient times and exotic places. Their stories might change the way you look at the world. They might change the way you live your life.

The planet you live on has some stories to tell. Earth has had 4.5 billion years to rack up experiences that range from ordinary to unimaginably violent. Through a variety of rock-forming processes, Earth has written her autobiography. Historical Geology is the science of reading that autobiography. But in order to read it, we first need to understand the language in which it is recorded.

We must learn to “speak Rock.”

1.1: Stories from stone

1.2: Sequences of strata

1.3: Fossil vocabulary

1.4: Magma, metamorphism, and mountains

1.5: Earth's story is your story

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1.1: Stories from stone

What stories has the Earth written in the language of Rock? Consider this cobble, found in Montana:



Figure 1.1.1: What is the history of a rock like this? What did it take for this thing to exist? (Callan Bentley photo.)

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?

As an exercise in practicing decoding Rock, let's translate: The smooth tan material is limestone. It originally formed as inorganic deposits of carbonate mud, laid down (1) in calm water, as evidenced from its very fine grain size. It sat still long enough to stick together into a semi-coherent mass. But then it must have been savagely ripped up to make (2) mud-chip clasts & entombed in (3) quartz sand. The event that was energetic enough to tear into the carbonate mud and rip it up, as well as energetic enough to transport in coarse particles of sand. It must have been a storm, perhaps a hurricane. The threat of more violent storms diminished as the sediment was buried and sank further and further underground, smothered beneath the crushing weight of thousands of feet of sediment piled atop it. Then the carbonate mud chunks and surrounding quartz sand (4) lithified into rock: No longer loose sediment, it was transformed through compaction and cementation into solid, coherent rock: limestone clasts in quartz sandstone. After it was solid rock, it broke. Some stress, perhaps imparted by the wrenching compression of mountain building, snapped the bonds holding the rock together. This made fractures, but those fractures were soon healed by the precipitation of the mineral quartz, making (5) thin quartz veins. This is something that must have taken place deep underground, but in order for us to pick it up and hold it, we know it must thereafter have been (6) uplifted. It made its way back to Earth's surface, and this particular chunk was broken off from its source outcrop. Finally, it experienced (7) differential weathering, etching away the limestone more readily than the quartz sandstone. This gives the sample its distinctive texture.

The rock is no longer just a rock. By speaking its language, we have turned it into a book, with seven chapters.



Figure 1.1.2: Points of annotation showing the 7-chapter history of this one rock sample. (Callan Bentley photo.)

This is the power of Historical Geology: it allows curious humans to interact with inert rock, and coax stories from that rock — vital tales of the planet's ancient past. Being able to decode the stories of rocks, hidden in plain sight, is an ability akin to an acquired superpower. As Spiderman gained his powers after being bitten by a radioactive spider, you are about to be bitten by the Historical Geology bug. What you learn will change the way you look at your world.

Like the best stories from our favorite storytellers, what we learn resonates with us, changes us. The Earth's many tales help us understand our present, live our lives more meaningfully, and prepare for a future we can better comprehend. The perspective granted by speaking Rock helps us live better lives, and look to the future with hope and confidence.

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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 a 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 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? Shallower conditions are often marked by a change in the fossil fauna. Tectonic drift can be marked by changes in the paleomagnetic inclination. There are 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.

Here is another example, from Inuyama, Japan:

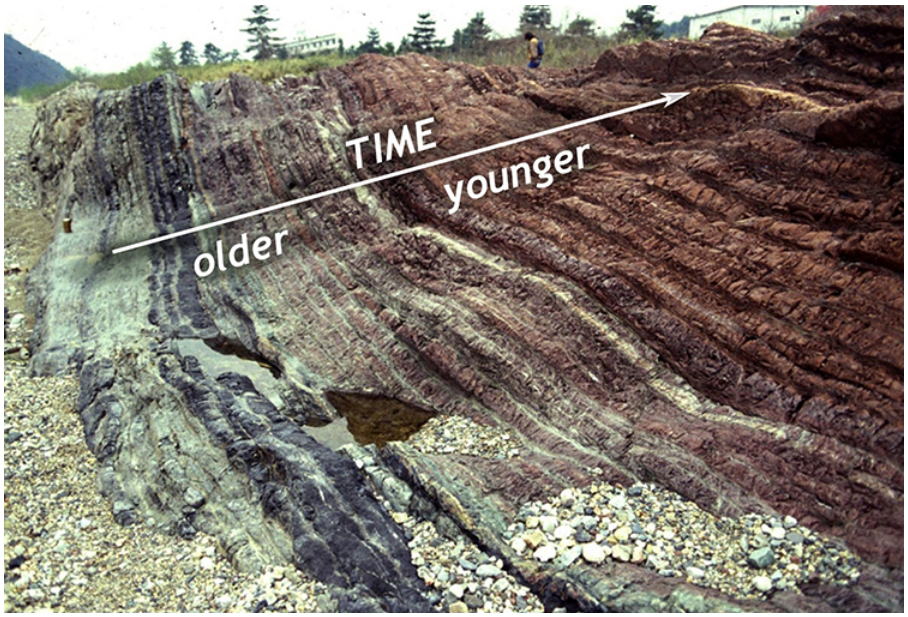


Figure 1.2.2: Outcrops in Inuyama, Japan, show a Triassic sequence of sedimentary strata that record changes in deep ocean oxygen levels after the end-Permian mass extinction event. The strata have been tectonically rotated due to accretion with the Japanese mainland. (Photo by Yukio Isozaki; reproduced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

Here, 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 suggest 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.



Figure 1.3.1: 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.



Figure 1.3.2: 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.

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.



Figure 1.3.3: You'll never have a pet Anomalocaris, because you live at the wrong time in Earth history. (Photo illustration mashup orchestrated by Callan Bentley.)

Like a digital watch on a medieval knight, it would be anachronistic to find an Ordovician fossil in Cretaceous strata. 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. Click on it and spin it around to see its distinctive shape.



Figure 1.4.1: Igneous is a dialect of Rock with elegant structure and fiery intonation. (3D model by Marissa Dudek & Callan Bentley)

The pattern you observe in the 3D model 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 it must have been constructed by a conscious mind. 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 own hasty retreat, knocking the causeway to pieces as he fled. The site is therefore known as the Giant's Causeway.

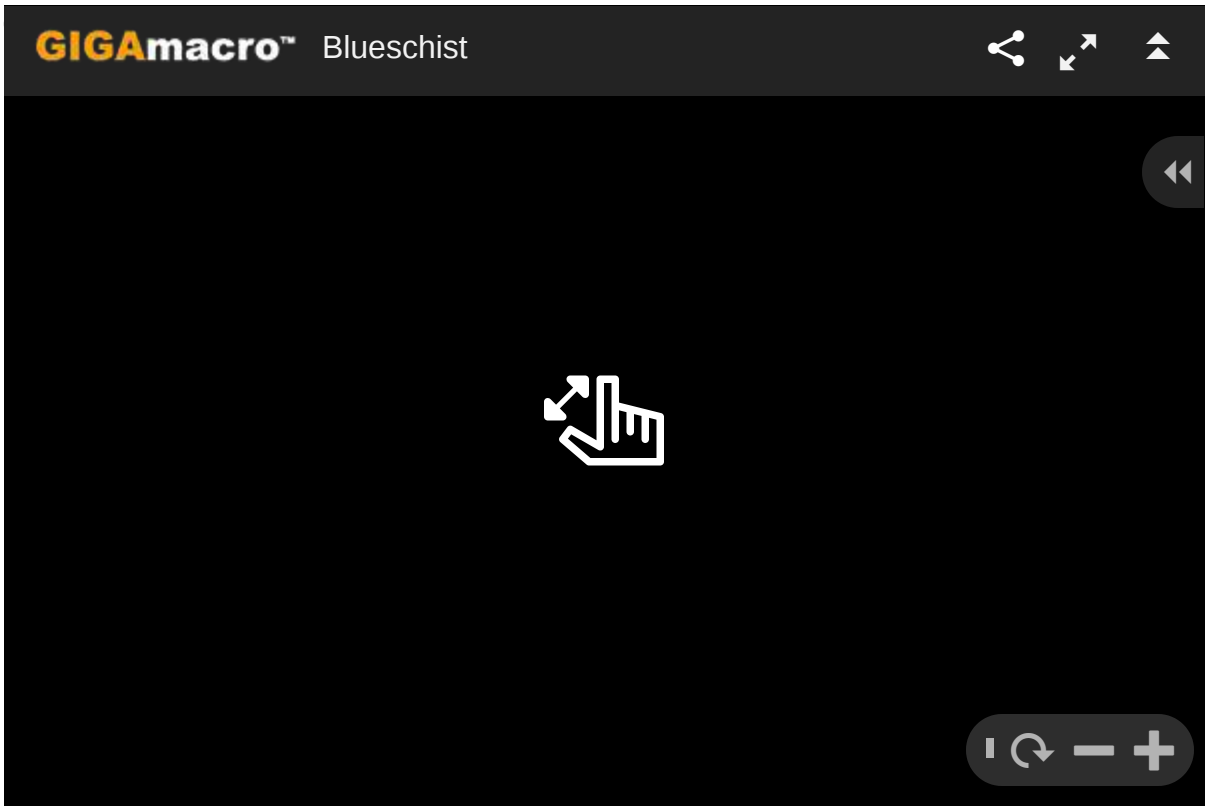
Geologists have a different way of explaining the Giant's Causeway, of course. Because they speak Rock rather than Giant, they note that on a landscape scale, this lava flow is one of many that erupted in a vast region between 60 and 50 million years ago, smothering the local landscape as they flowed and pooled. The volcanic activity was not limited to Ireland and Scotland. It also produced basalt in the Faroe Islands, Norway, and Greenland. Collectively, all of these sites are 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 prior to the various landmasses separating 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 urgency 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 history.

On the metamorphic front, consider this blueschist:



Blueschist is a metamorphic rock which is blue and flaky. But like all metamorphic rocks, it didn't start off 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 any more 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 coming back from the Moon, the blueschist has come back 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 tells stories too, and Earth has picked up her share of scars through the years. For the most part, these "scars" are tectonic ***structures*** like folds, faults, and shear zones. Consider this photograph of the Andes Mountains in southern Argentina:



Figure 1.4.2: 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 having **formed** it.

Stories can be extracted not only from the sequence of sedimentary layers in this case, but also from their contortion, their orientation, their elevation, and their shape. 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.



Figure 1.5.1: 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 breath 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: A Brief History of Earth

Based on An Introduction to Geology, Chapter 8: Earth History

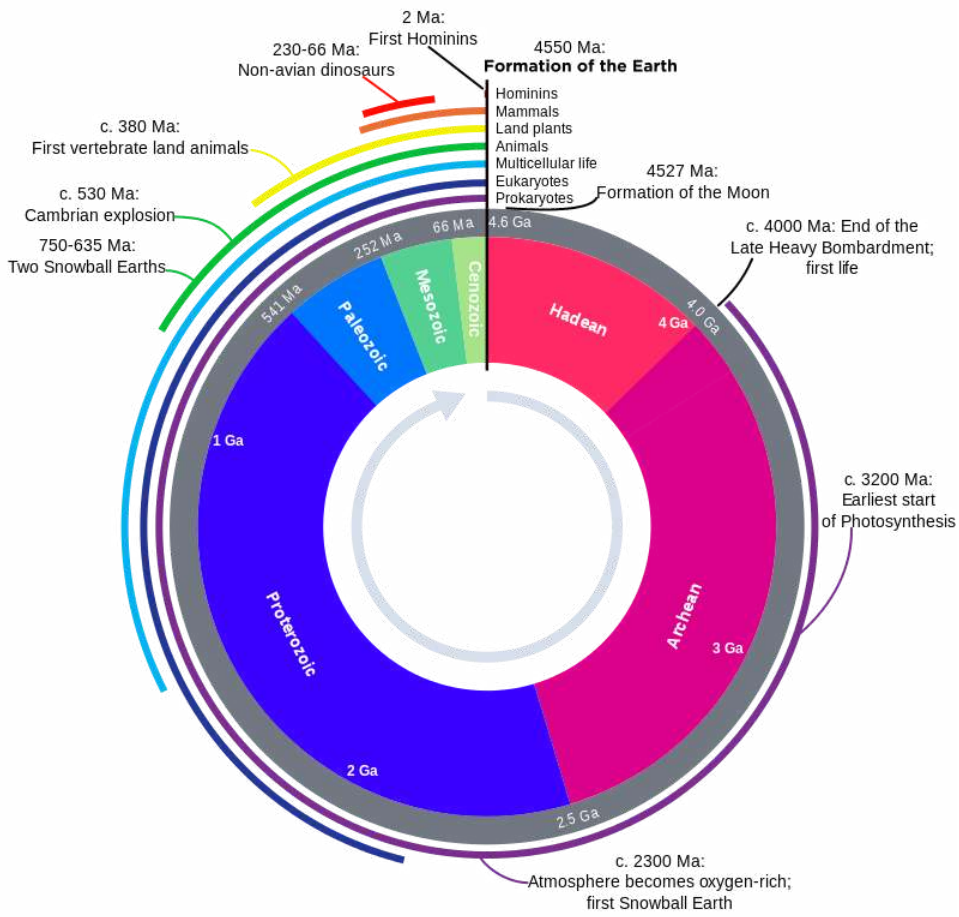


Figure 2.1: Geologic time on Earth, represented circularly, to show the individual time divisions and important events. Ga=billion years ago, Ma=million years ago.

Entire courses and careers have been based on the wide-ranging topics covering Earth's history. Throughout the long history of Earth, change has been the norm. Looking back in time, an untrained eye would see many unfamiliar life forms and terrains. The main topics studied in Earth history are paleogeography, paleontology, and paleoecology and paleoclimatology—respectively, past landscapes, past organisms, past ecosystems, and past environments. This chapter will cover (briefly) the origin of the universe and the 4.6 billion year history of Earth. It will act as a guide, linking out to other chapters, case studies, and sections in this book.

2.1: Origin of the Universe

2.2: Origin of the Solar System - The Nebular Hypothesis

2.3: Hadean Eon

2.4: Archean Eon

2.5: Proterzoic Eon

2.6: Phanerozoic Eon - Paleozoic Era

2.7: Phanerozoic Eon - Mesozoic Era

2.8: Phanerozoic Eon - Cenozoic Era

2.9: References

2.S: Summary

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2.1: Origin of the Universe



Figure 2.1.1: The Hubble Deep Field. This image, released in 1996, is a composite long-exposure picture of one of the darkest parts of the night sky. Every light on this image that does not have diffraction spikes is believed to be an entire galaxy, with hundreds of billions of stars, demonstrating the immense size and scope of the universe.

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

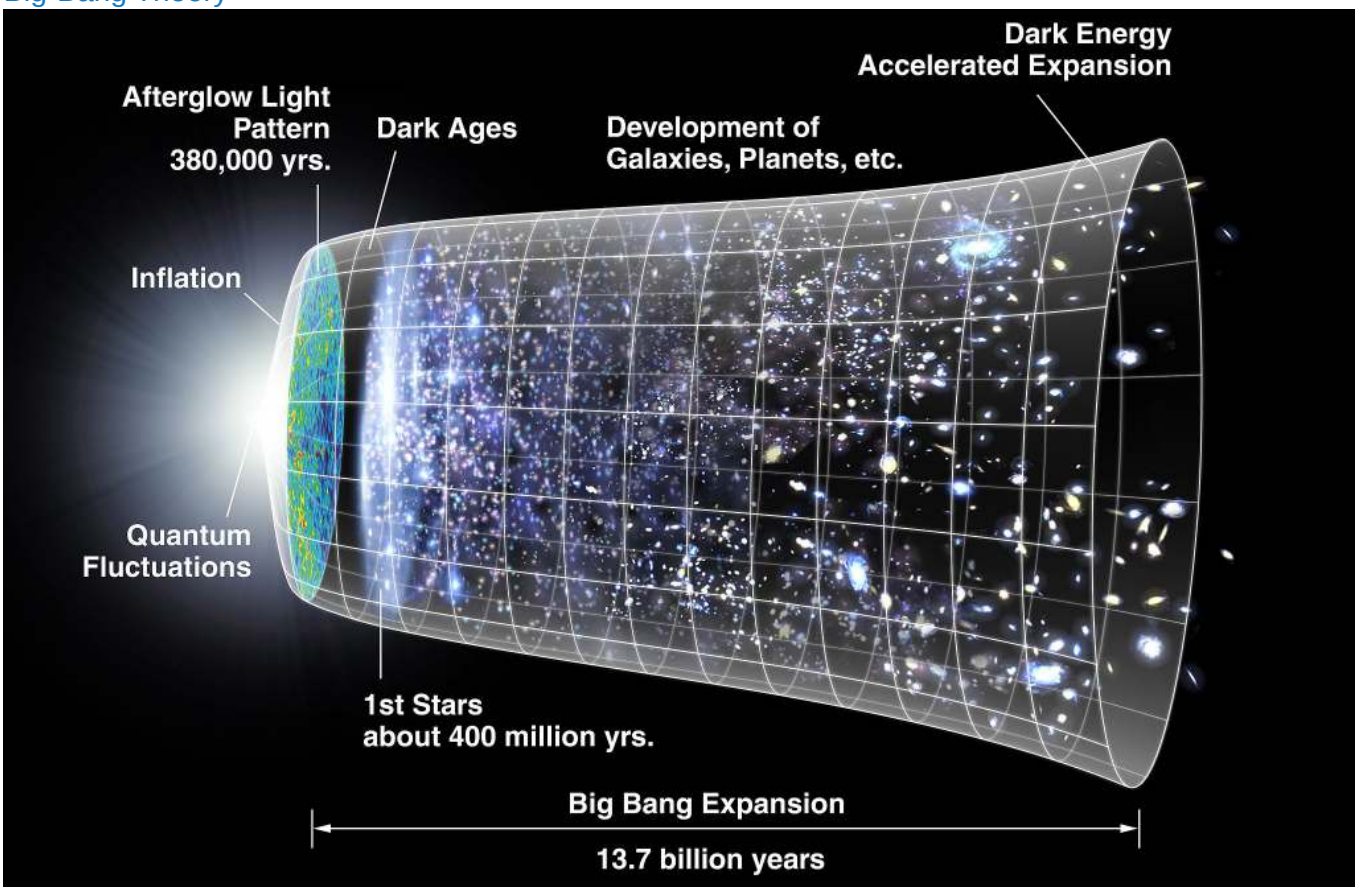


Figure 2.1.2: Timeline of expansion of the universe

The mysterious details of events prior to and during the origin of the universe are subject to great scientific debate. The prevailing idea about how the universe was created is called the **big-bang theory**. Although the ideas behind the big-bang theory feel almost mystical, they are supported by Einstein's theory of general relativity. Other scientific evidence, grounded in empirical observations, supports the big-bang theory.

The big-bang theory proposes the universe was formed from an infinitely dense and hot core of material. The bang in the title suggests there was an explosive, outward expansion of all matter and space that created atoms. Spectroscopy confirms that hydrogen makes up about 74% of all matter in the universe. Since its creation, the universe has been expanding for 13.8 billion years and recent observations suggest the rate of this expansion is increasing.

SPECTROSCOPY

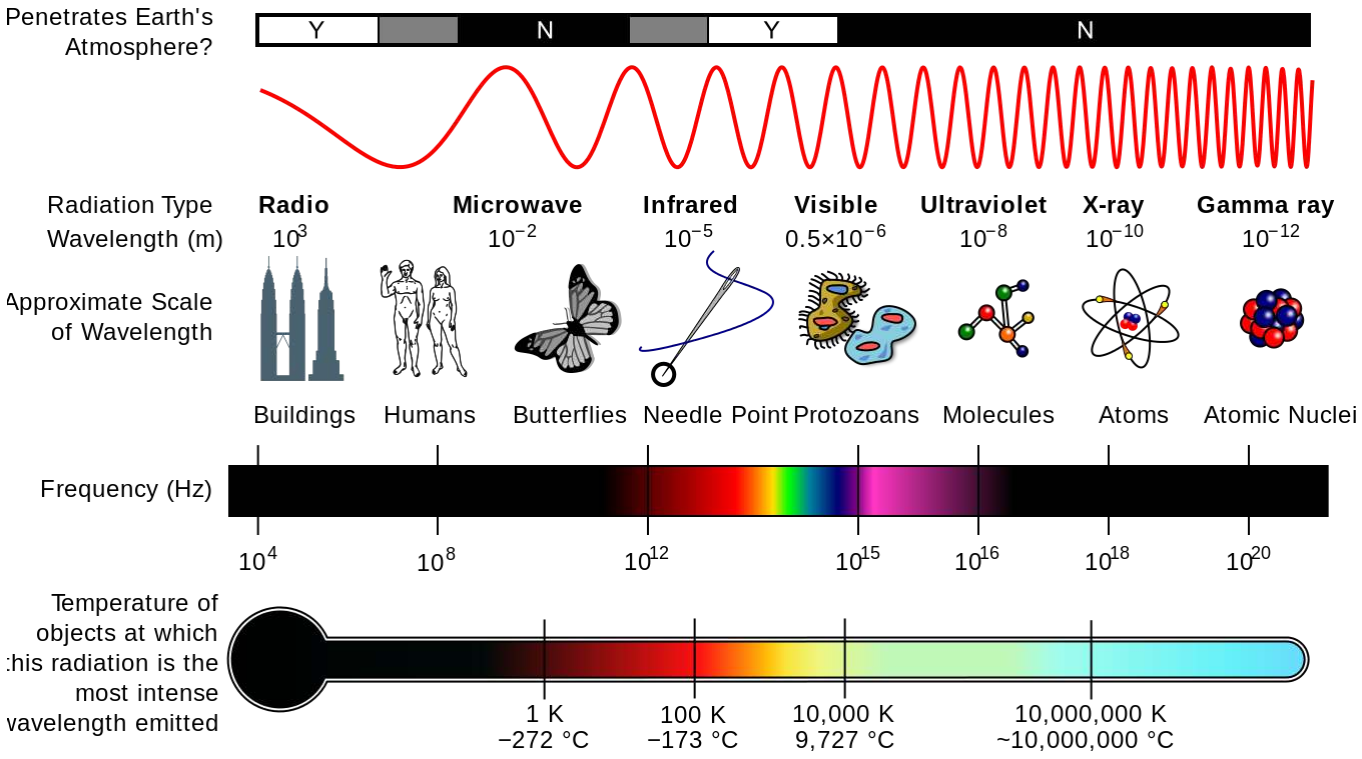


Figure 2.1.3: The electromagnetic spectrum and properties of light across the spectrum.

Spectroscopy is the investigation and measurement of spectra produced when materials interacts with or emits electromagnetic radiation. *Spectra* is the plural for *spectrum* which is a particular wavelength from the **electromagnetic spectrum**. Common spectra include the different colors of visible light, X-rays, ultraviolet waves, microwaves, and radio waves. Each beam of light is a unique mixture of wavelengths that combine across the spectrum to make the color we see. The light wavelengths are created or absorbed inside atoms, and each wavelength signature matches a specific element. Even white light from the Sun, which seems like an uninterrupted continuum of wavelengths, has gaps in some wavelengths. The gaps correspond to elements present in the Earth's atmosphere that act as filters for specific wavelengths. These missing wavelengths were famously observed by Joseph von Fraunhofer (1787–1826) in the early 1800s, but it took decades before scientists were able to relate the missing wavelengths to atmospheric filtering . Spectroscopy shows that the Sun is mostly made of hydrogen and helium. Applying this process to light from distant stars, scientists can calculate the abundance of elements in a specific star and visible universe as a whole. Also, this spectroscopic information can be used as an interstellar speedometer.

REDSHIFT



Figure 2.1.4: Click to animate. This animation demonstrates how the Doppler effect is heard as a car moves. The waves in front of the car are compressed together, making the pitch higher. The waves in the back of the car are stretched, and and the pitch gets lower.

The **Doppler effect** is the same process that changes the pitch of the sound of an approaching car or ambulance from high to low as it passes. When an object emits waves, such as light or sound, while moving toward an observer, the wavelengths get compressed. In sound, this results in a shift to a higher pitch. When an object moves away from an observer, the wavelengths are extended, producing a lower pitched sound. The Doppler effect is used on light emitted from stars and galaxies to determine their speed and direction of travel. Scientists, including Vesto Slipher (1875–1969) and Edwin Hubble (1889–1953) , examined galaxies both near and far and found that almost all galaxies outside of our galaxy are moving away from each other, and us. Because the light

wavelengths of receding objects are extended, visible light is shifted toward the red end of the spectrum, called a **redshift**. In addition, Hubble noticed that galaxies that were farther away from Earth also had the greater amount of redshift, and thus, the faster they are traveling away from us. The only way to reconcile this information is to deduce the universe is still expanding. Hubble's observation forms the basis of big-bang theory.



COSMIC MICROWAVE BACKGROUND RADIATION

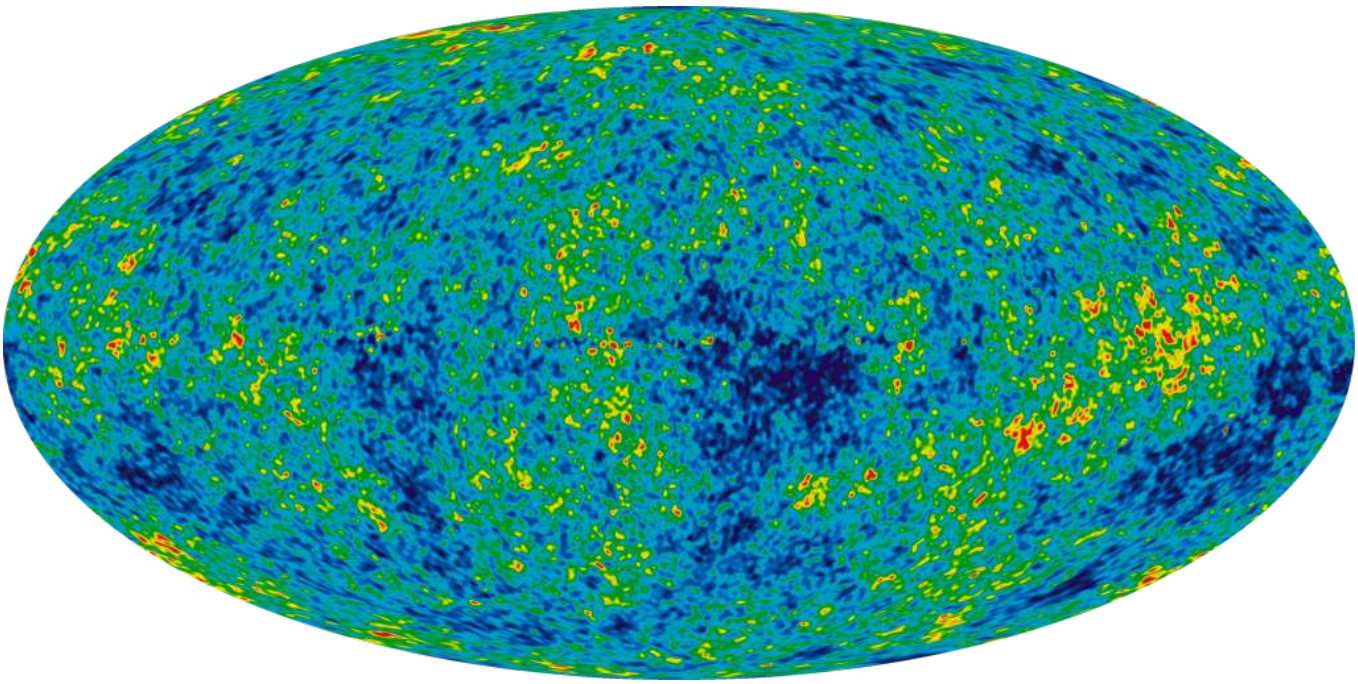


Figure 2.1.5: Heat map, showing slight variations in background heat, which is related to cosmic background radiation.

Another 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.

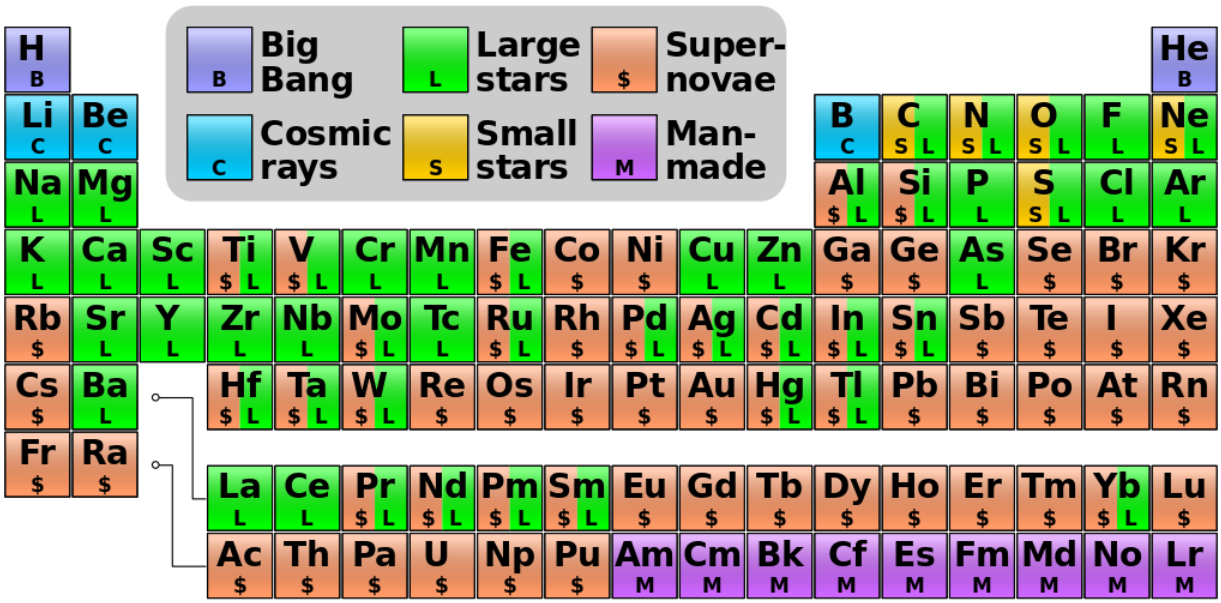


Figure 2.1.6: Origin of the elements on the periodic table, showing the important role the star life cycle plays.

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

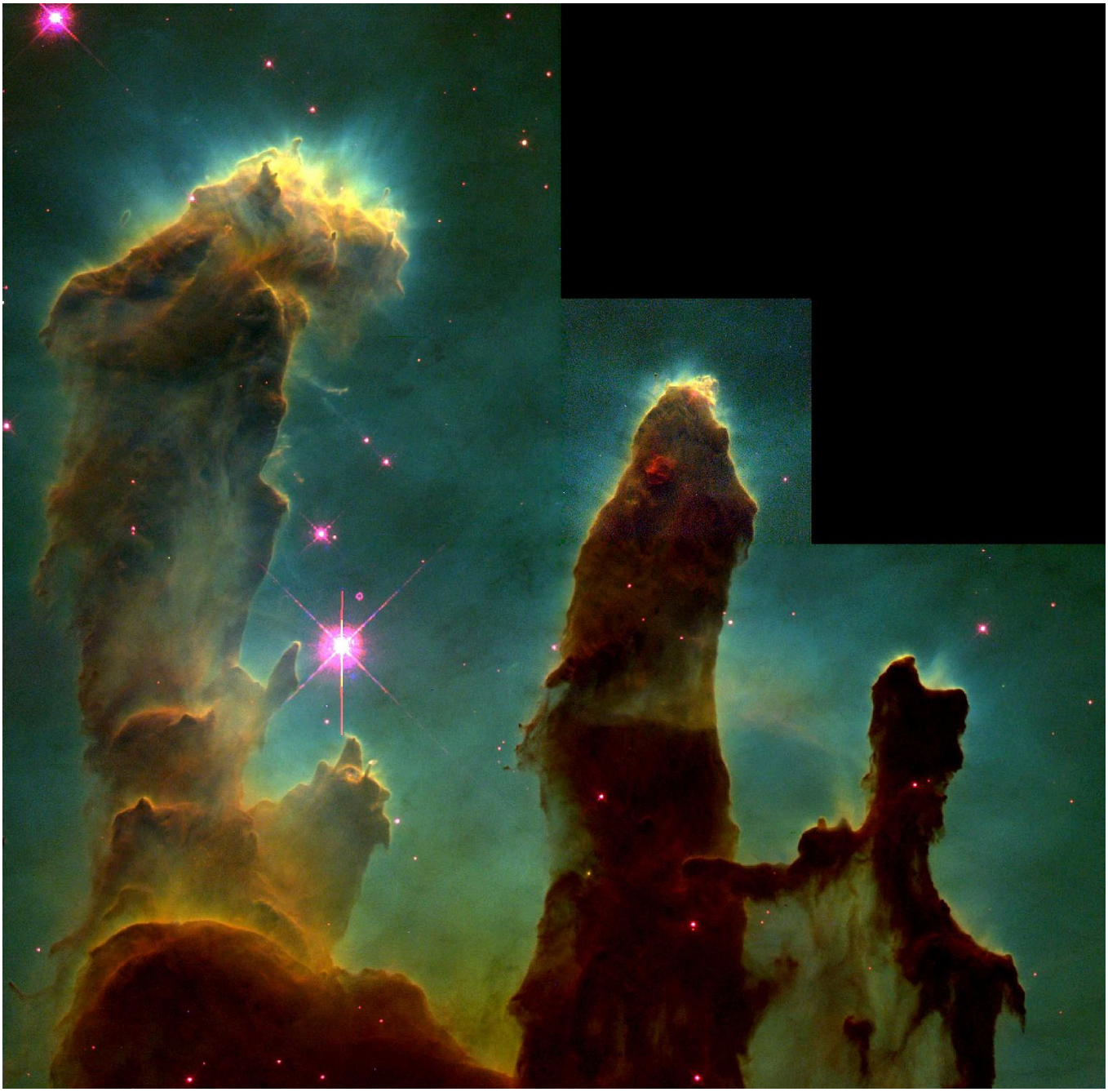


Figure 2.1.7: Section of the Eagle Nebula known as “The Pillars of Creation.”

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 disk shape. In the center of this disk shape a new star is born under the force of gravity. The spinning whirlpool concentrates material in the center, and the increasing gravitational forces collect even more mass. Eventually, the immensely concentrated mass of material reaches a critical point of such intense heat and pressure it initiates fusion.

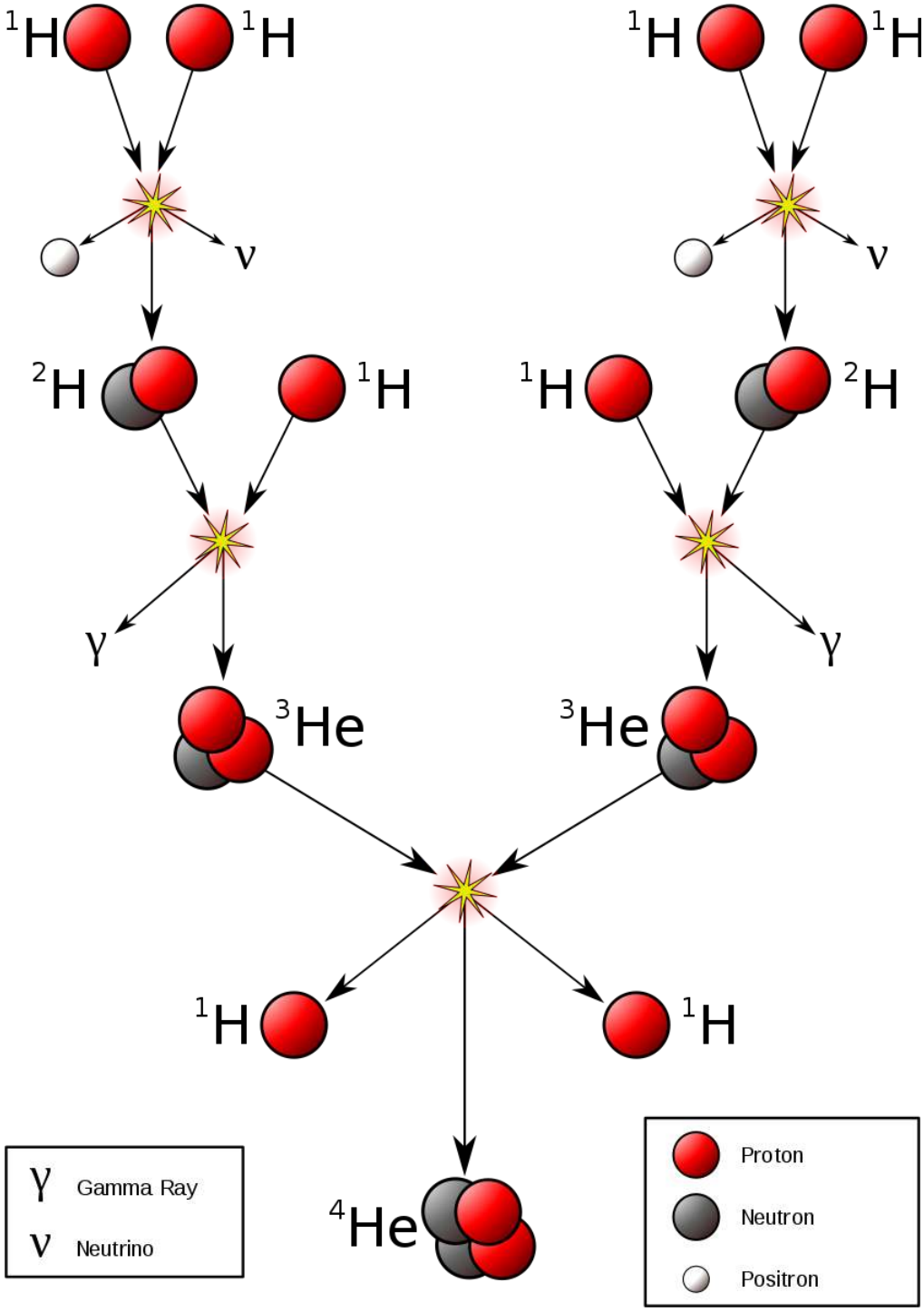


Figure 2.1.8: General diagram showing the series of fusion steps that occur in the sun.

Fusion is not a chemical reaction. **Fusion** is a nuclear reaction in which two or more nuclei, the centers of atoms, are forced together and combine creating a new larger atom. This reaction gives off a tremendous amount of energy, usually as light and solar radiation. An element such as hydrogen combines or fuses with other hydrogen atoms in the core of a star to become a new element, in this case, helium. Another product of this process is energy, such as solar radiation that leaves the Sun and comes to the Earth as light and heat. Fusion is a steady and predictable process, which is why we call this the main phase of a star's life. During its main phase, a star turns hydrogen into helium. Since most stars contain plentiful amounts of hydrogen, the main phase may last billions of years, during which their size and energy output remains relatively steady.

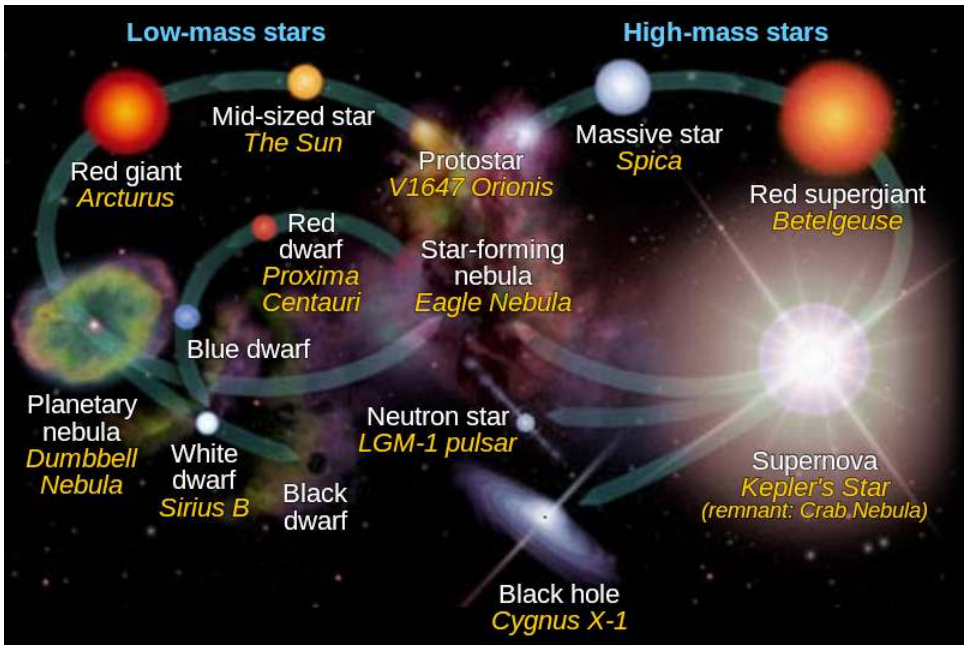


Figure 2.1.9: Two main paths of a star's life cycle, depending on mass.

The giant phase in a star's life occurs when the star runs out of hydrogen for fusion. If a star is large enough, it has sufficient heat and pressure to start fusing helium into heavier elements. This style of fusion is more energetic and the higher energy and temperature expand the star to a larger size and brightness. This giant phase is predicted to happen to our Sun in another few billion years, growing the radius of the Sun to Earth's orbit, which will render life impossible. The mass of a star during its main phase is the primary factor in determining how it will evolve. If the star has enough mass and reaches a point at which the primary fusion element, such as helium, is exhausted, fusion continues using new, heavier elements. This occurs over and over in very large stars, forming progressively heavier elements like carbon and oxygen. Eventually, fusion reaches its limit as it forms iron and nickel. This progression explains the abundance of iron and nickel in rocky objects, like Earth, within the solar system. At this point, any further fusion absorbs energy instead of giving it off, which is the beginning of the end of the star's life.

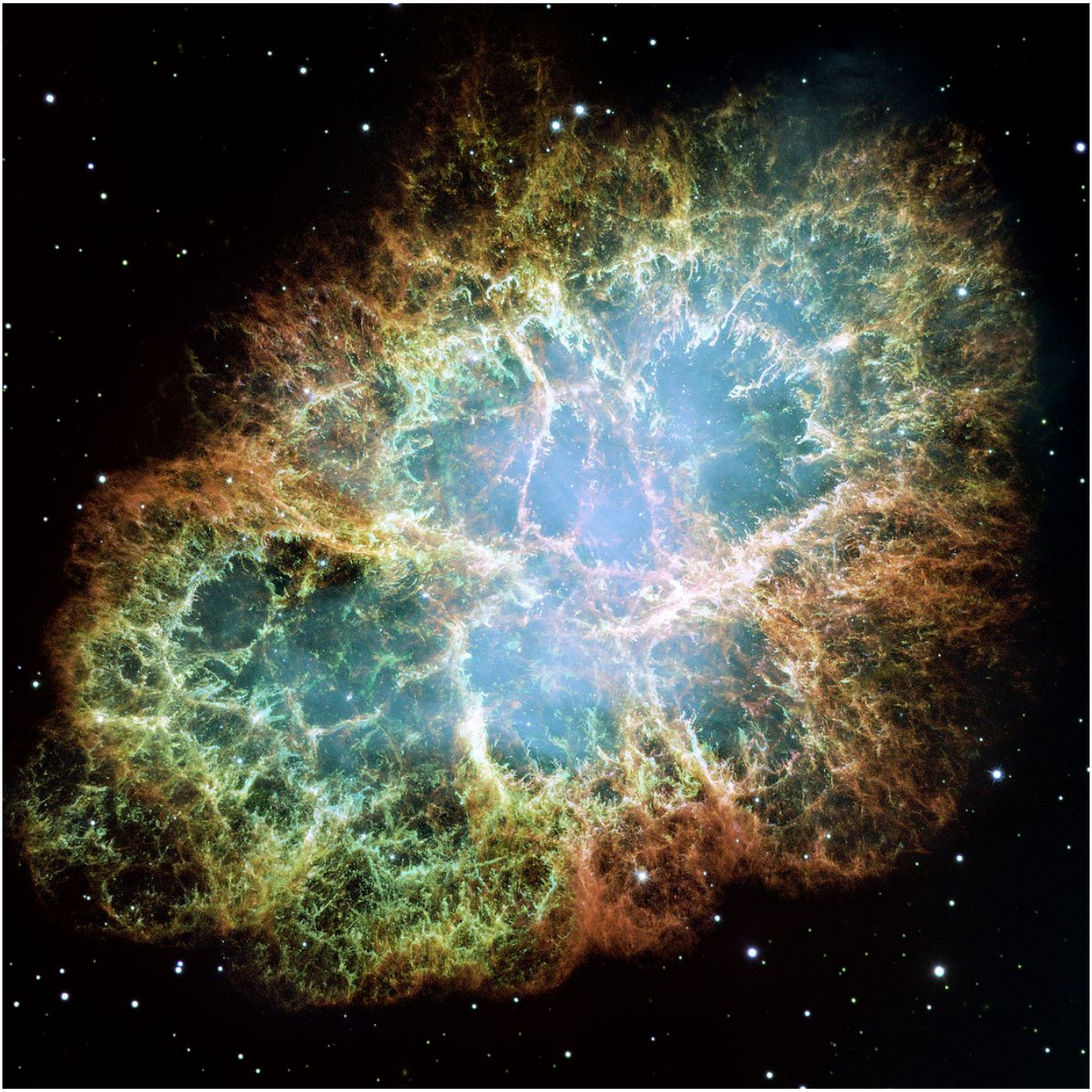


Figure 2.1.10: Hubble space telescope image of the Crab Nebula, the remnants of a supernova that occurred in 1054 C.E.

The death of a star can range from spectacular to other-worldly (see figure). Stars like the Sun form a planetary nebula, which comes from the collapse of the star's outer layers in an event like the implosion of a building. In the tug-of-war between gravity's inward pull and fusion's outward push, gravity instantly takes over when fusion ends, with the outer gasses puffing away to form a nebula. More massive stars do this as well but with a more energetic collapse, which starts another type of energy release mixed with element creation known as a supernova. In a **supernova**, the collapse of the core suddenly halts, creating a massive outward-propagating shock wave. A supernova is the most energetic explosion in the universe short of the big bang. The energy release is so significant the ensuing fusion can make every element up through uranium.

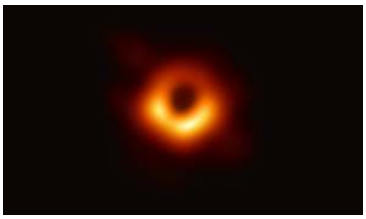


Figure 2.1.11: A black hole and its shadow have been captured in an image for the first time in 2019, a historic feat by an international network of radio telescopes called the Event Horizon Telescope (Source: NASA)

The death of the star can result in the creation of white dwarfs, neutron stars, or black holes. Following their deaths, stars like the Sun turn into white dwarfs.

White dwarfs are hot star embers, formed by packing most of a dying star's mass into a small and dense object about the size of Earth. Larger stars may explode in a supernova that packs their mass even tighter to become neutron stars. Neutron stars are so dense that protons combine with electrons to form neutrons. The largest stars collapse their mass even further, becoming objects so dense that light cannot escape their gravitational grasp. These are the infamous black holes and the details of the physics of what occurs in them are still up for debate.

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2.2: Origin of the Solar System - The Nebular Hypothesis



Figure 2.2.1: Small protoplanetary discs in the Orion Nebula

Our solar system formed as 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.

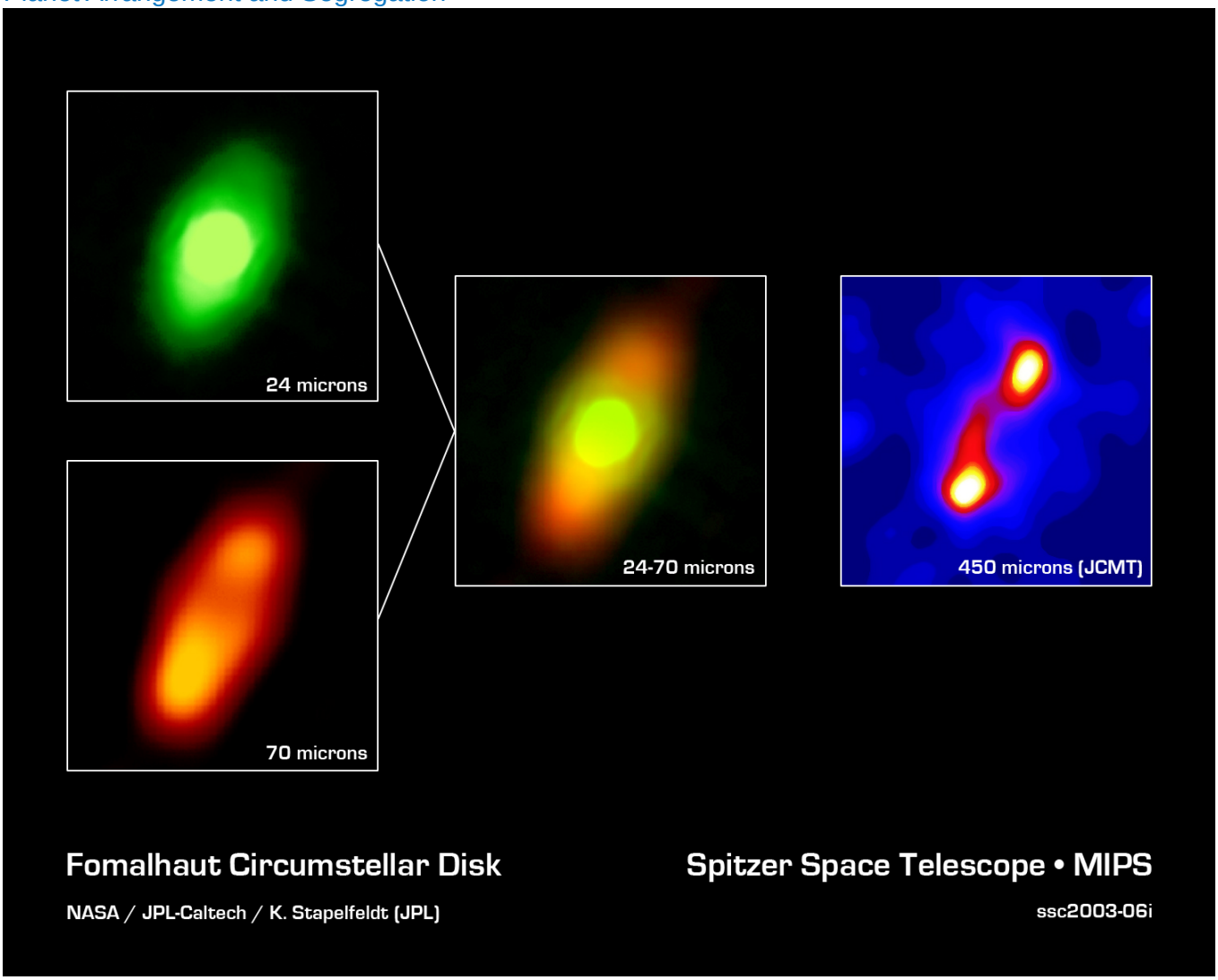


Figure 2.2.2: This disk is asymmetric, possibly because of a large gas giant planet orbiting relatively far from the star.

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 the condensation of lighter gaseous molecules such as methane, ammonia, carbon dioxide, and water. This temperature differentiation resulted in the inner four planets of the solar system becoming rocky, and the outer four planets becoming gas giants.

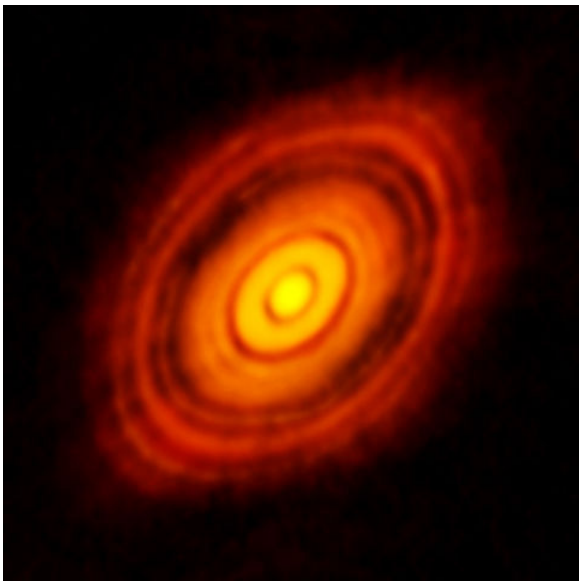


Figure 2.2.3: Image by the ALMA telescope of HL Tauri and its protoplanetary disk, showing grooves formed as planets absorb material in the disk.

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.

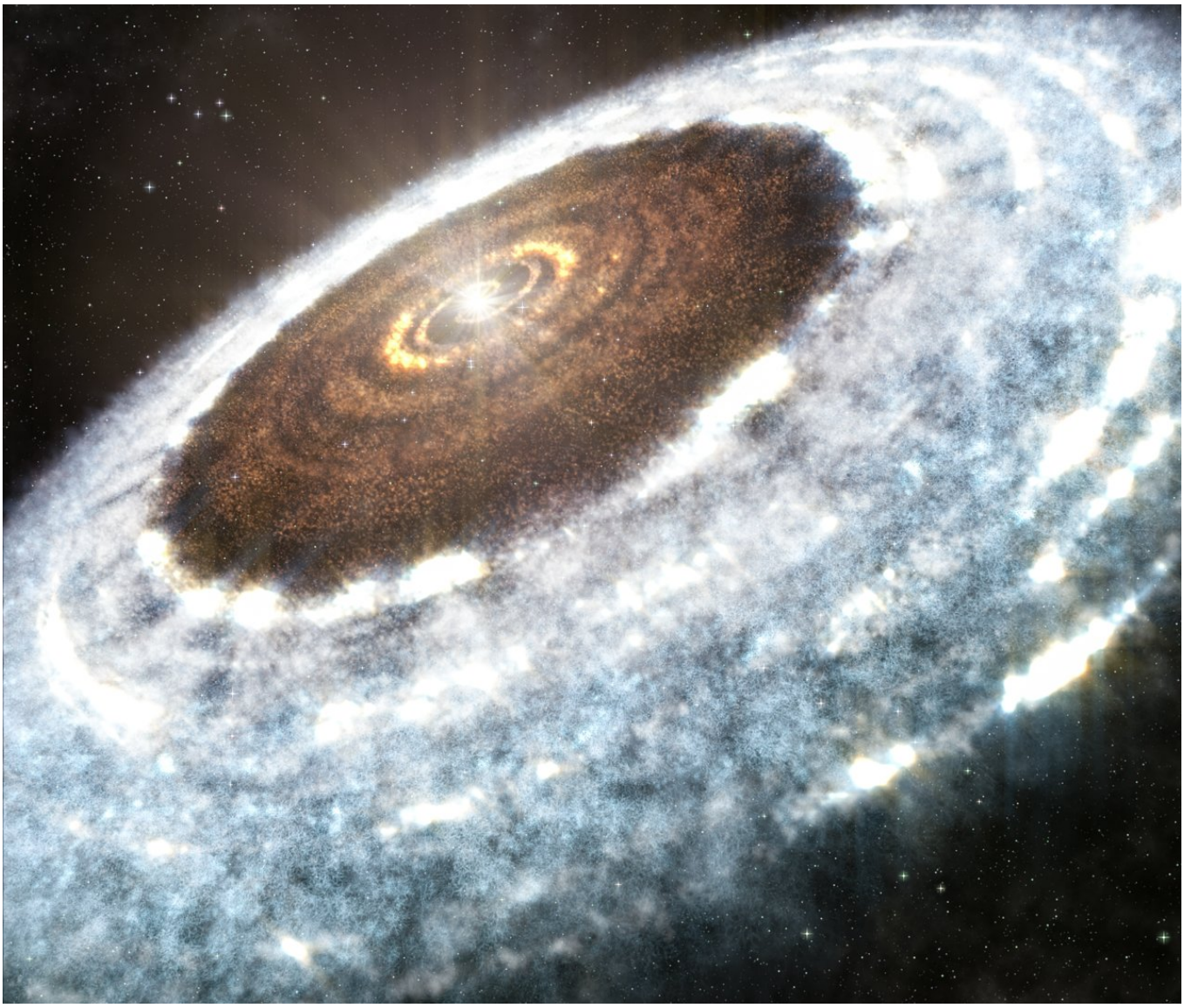


Figure 2.2.4: This artist's impression of the water snowline around the young star V883 Orionis, as detected with ALMA.

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 weaker gravity of the smaller planets.



Figure 2.2.5: 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. 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** that currently impact the Earth. Study of asteroids and meteorites help geologist to determine the age of Earth and the composition of its core, mantle, and crust. Jupiter's gravity may also explain Mars' smaller mass, with the larger planet consuming material as it migrated from the inner to outer edge of the solar system.

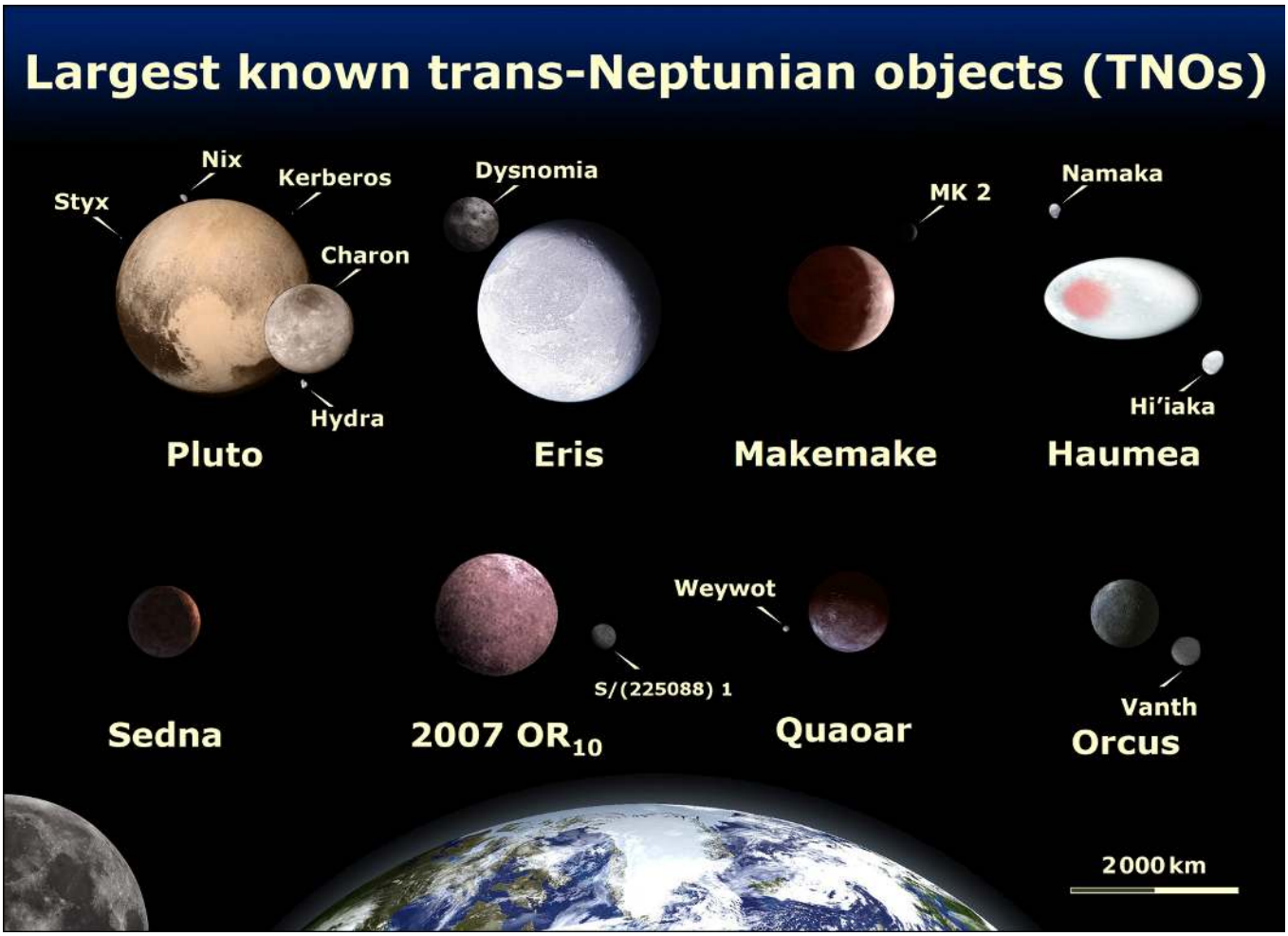


Figure 2.2.6: Eight largest objects discovered past Neptune.

The outermost part of the solar system is known as the Kuiper belt, which is a scattering of rocky and icy bodies. Beyond that is the Oort cloud, a zone filled with small and dispersed ice traces. These two locations are where most comets form and continue to orbit, and objects found here have relatively irregular orbits compared to the rest of the solar system. Pluto, formerly the ninth planet, is located in this region of space. The XXVIth General Assembly of the International Astronomical Union (IAU) stripped Pluto of planetary status in 2006 because scientists discovered an object more massive than Pluto, which they named Eris. The IAU decided against including Eris as a planet, and therefore, excluded Pluto as well. The IAU narrowed the definition of a planet to three criteria: 1) enough mass to have gravitational forces that force it to be rounded, 2) not massive enough to create fusion, and 3) large enough to be in a cleared orbit, free of other planetesimals that should have been incorporated at the time the planet formed. Pluto passed the first two parts of the definition, but not the third. Pluto and Eris are currently classified as dwarf planets.

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2.3: Hadean Eon

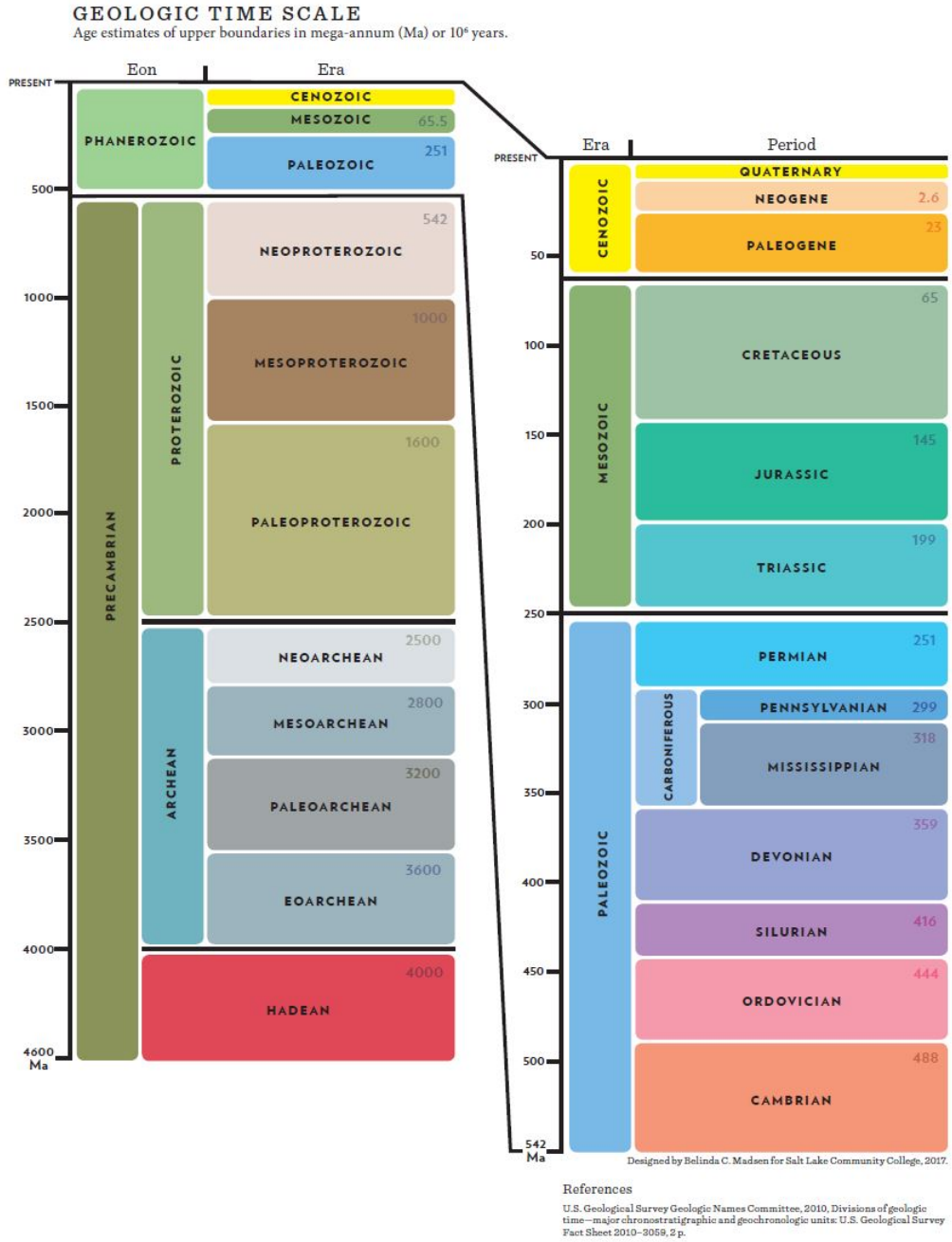


Figure 2.3.1: Geologic Time Scale with ages shown

Geoscientists use the geological time scale to assign relative age names to events and rocks, separating major events in Earth's history based on significant changes as recorded in rocks and fossils. This section summarizes the most notable events of each major time interval. For a breakdown on how these time intervals are chosen and organized, see An Introduction to Geology: Chapter 7.

The Hadean Eon, named after the Greek god and ruler of the underworld Hades, is the oldest eon and dates from 4.5–4.0 billion years ago.



Figure 2.3.2: Artist's impression of the Earth in the Hadean.

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.

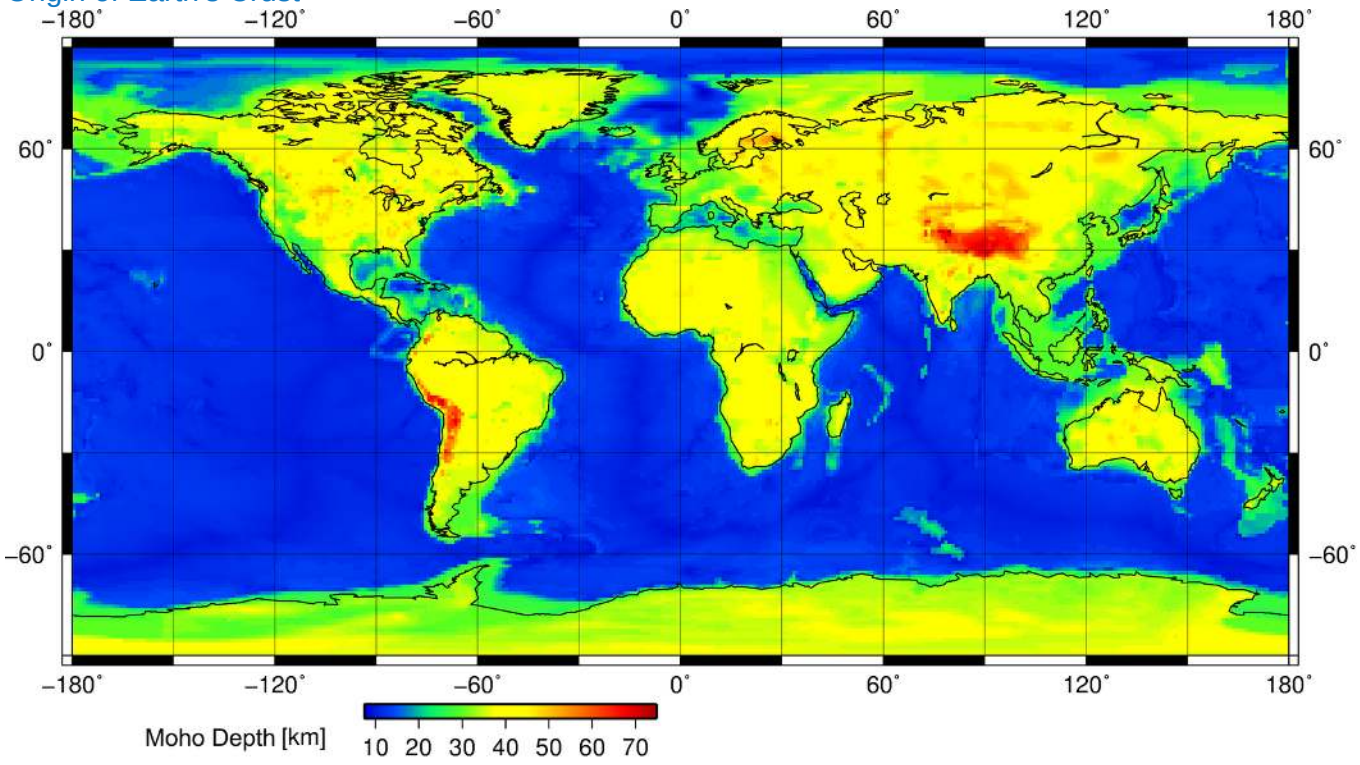


Figure 2.3.3: The global map of the depth of the moho, or thickness of the crust.

As Earth cooled from its molten state, minerals started to crystallize and settle resulting in a separation of minerals based on density and the creation of the crust, mantle, and core. The earliest Earth was chiefly molten material and would have been rounded by gravitational forces so it resembled a ball of lava floating in space. As the outer part of the Earth slowly cooled, the high melting-point minerals (see Bowen's Reaction Series in An Introduction to Geology: Chapter 4) formed solid slabs of early crust. These slabs were probably unstable and easily reabsorbed into the liquid magma until the Earth cooled enough to allow numerous larger fragments to form a thin primitive crust. Scientists generally assume this crust was oceanic and mafic in composition, and littered with impacts, much like the Moon's current crust. There is still some debate over when plate tectonics started, which would have led to the formation of continental and felsic crust. Regardless of this, as Earth cooled and solidified, less dense felsic minerals floated to the surface of the Earth to form the crust, while the denser mafic and ultramafic materials sank to form the mantle and the highest-density iron and nickel sank into the core. This differentiated the Earth from a homogenous planet into a heterogeneous one with layers of felsic crust, mafic crust, ultramafic mantle, and iron and nickel core.

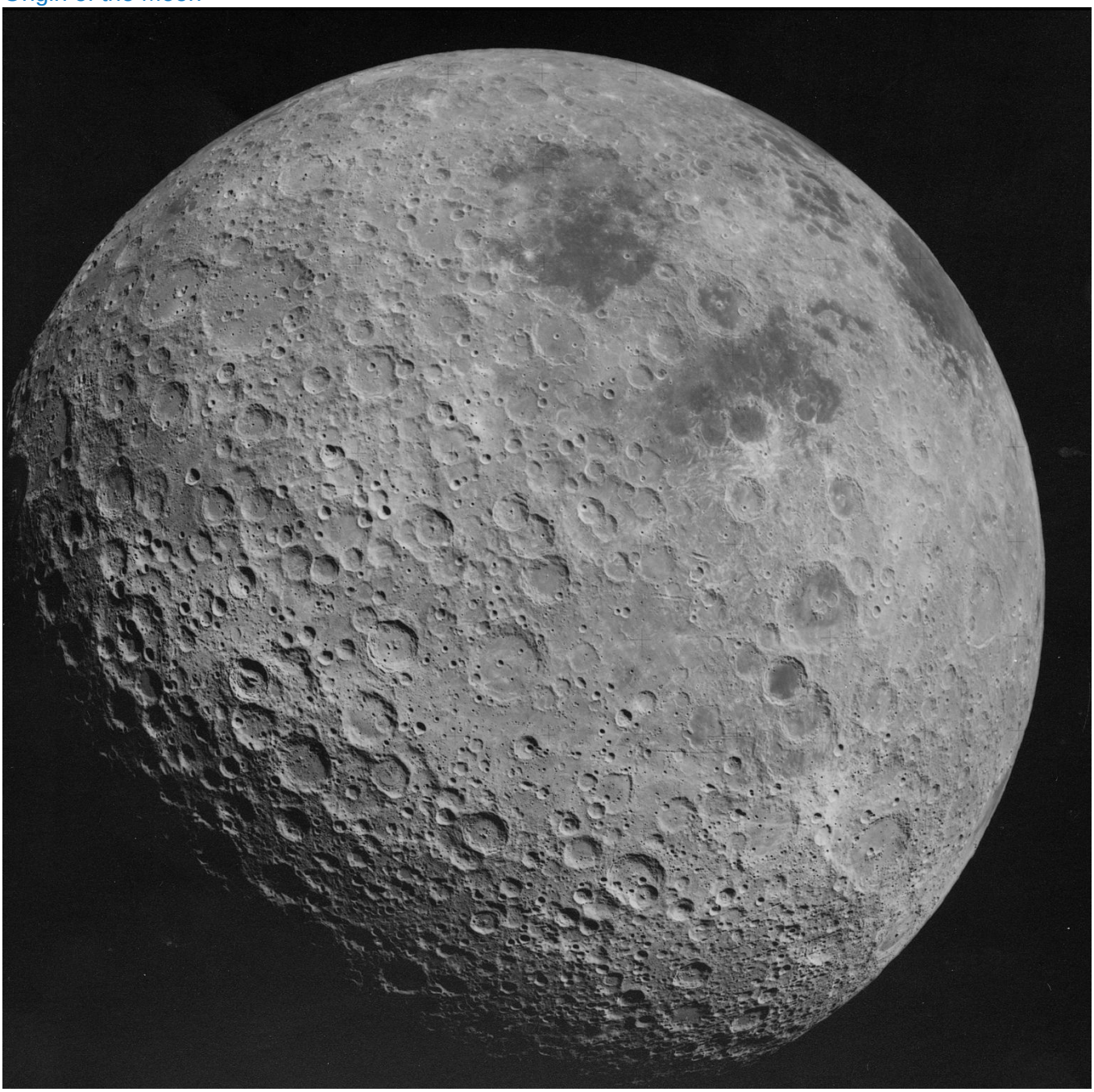


Figure 2.3.4: Dark side of the Moon

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



Figure 2.3.5: Artist's concept of the giant impact from a Mars-sized object that could have formed the moon.

Many ideas have been proposed for the origin of the Moon: The Moon could have been captured from another part of the solar system and formed in place together with the Earth, or the Moon could have been ripped out of the early Earth. None of 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.



Computer simulation of the evolution of the Moon (2 minutes).

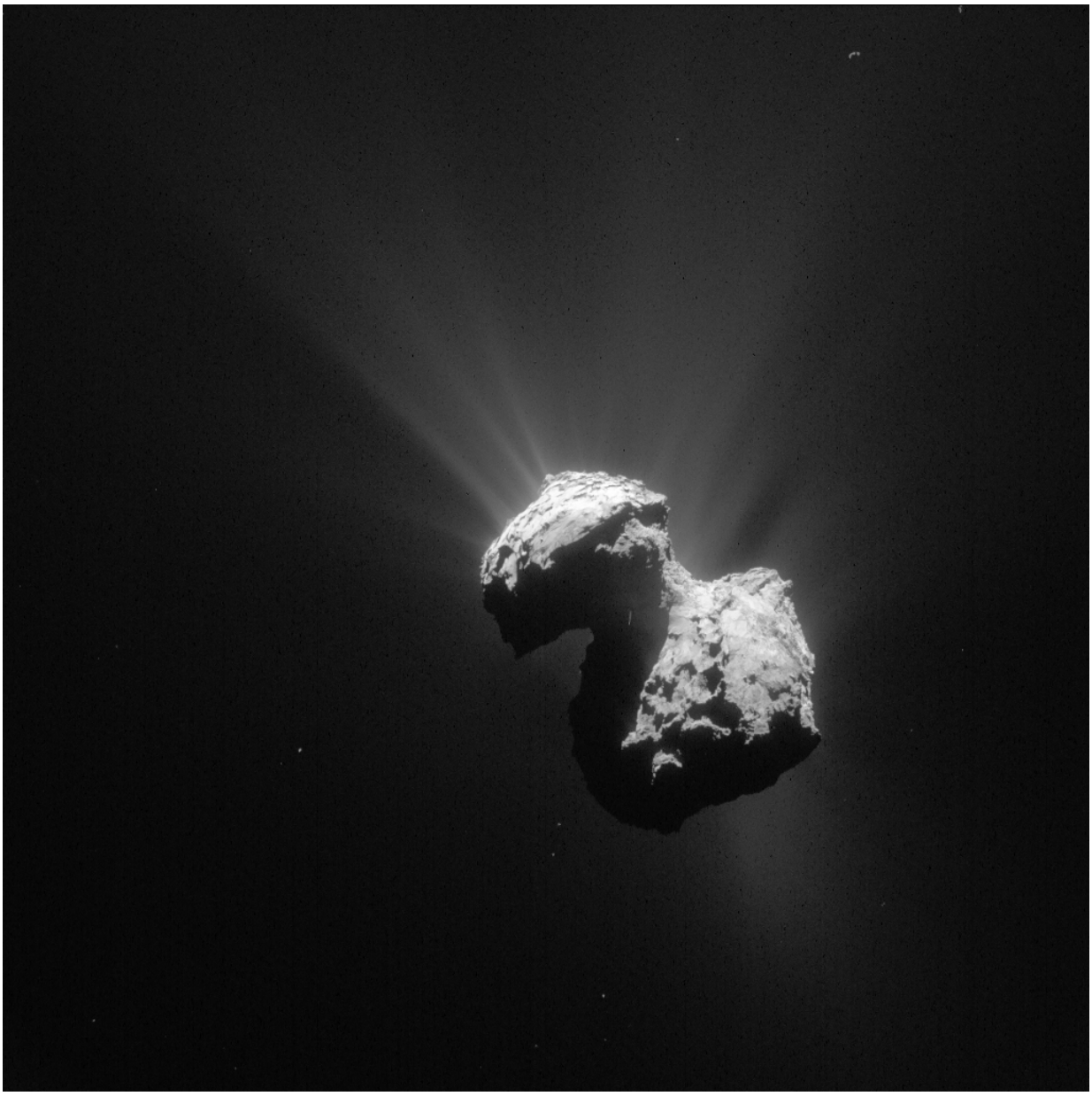


Figure 2.3.6: Water vapor leaves comet 67P/Churyumov–Gerasimenko.

Explanations for the origin of Earth's water include volcanic outgassing, comets, and meteorites. The volcanic outgassing hypothesis for the origin of Earth's water is that it originated from inside the planet, and emerged via tectonic processes as vapor associated with volcanic eruptions. 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 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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2.4: Archean Eon



Figure 2.4.1: Artist's impression of the Archean.

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

Late Heavy Bombardment

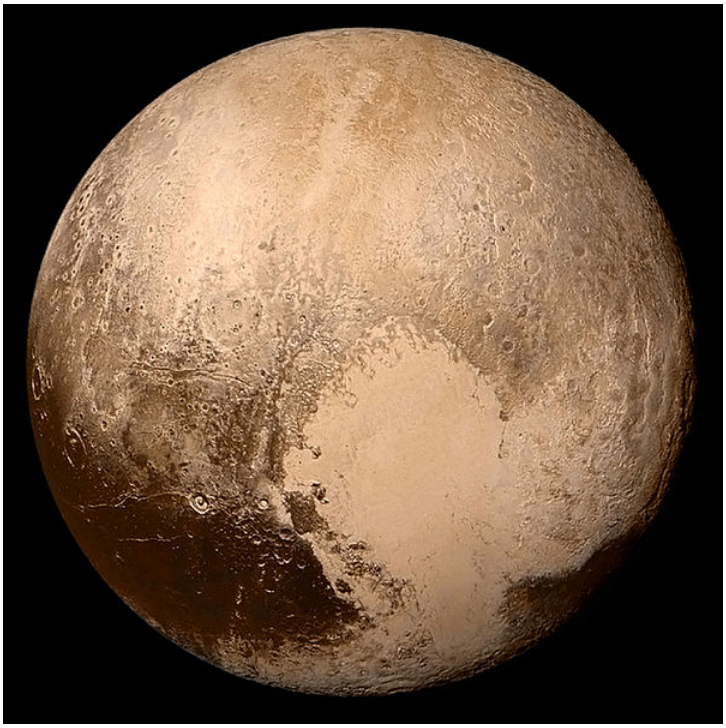


Figure 2.4.2: 2015 image from NASA's New Horizons probe of Pluto. The lack of impacts found on the Tombaugh Regio (the heart-shaped plain, lower right) has been inferred as being younger than the Late Heavy Bombardment and the surrounding surface due to its lack of impacts.

Objects were chaotically flying around at the start of the solar system, building the planets and moons. There is evidence that after the planets formed, about 4.1–3.8 billion years ago, a second large spike of asteroid and comet impacted the Earth and Moon in an

event called 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 **late heavy bombardment**, the Earth, Moon, and all planets in the solar system were pummeled by material from the asteroid and Kuiper belts. Evidence of this bombardment was found within samples collected from the Moon.

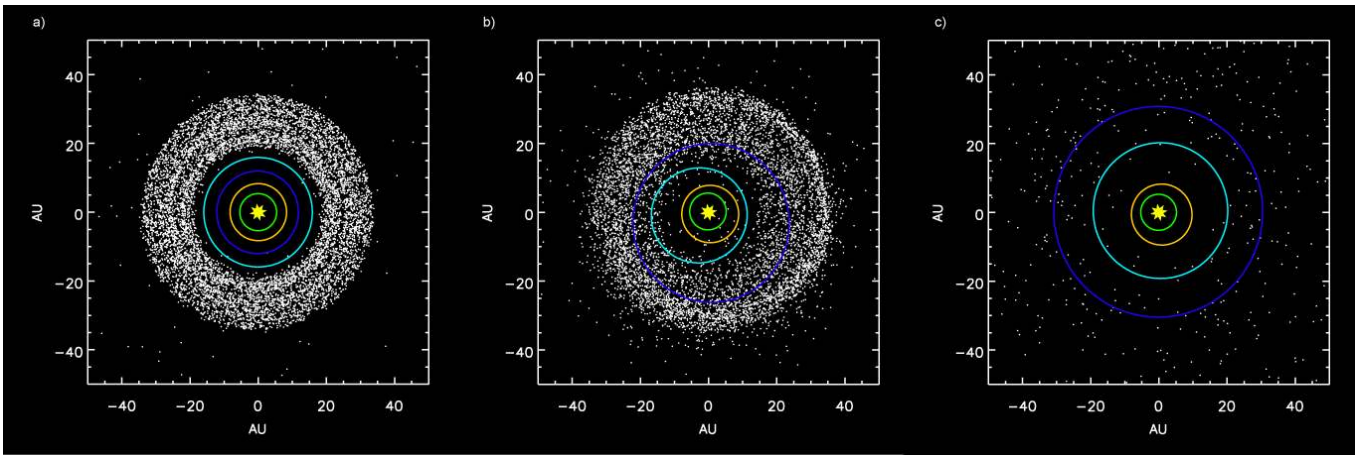


Figure 2.4.3: Simulation of before, during, and after the late heavy bombardment.

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

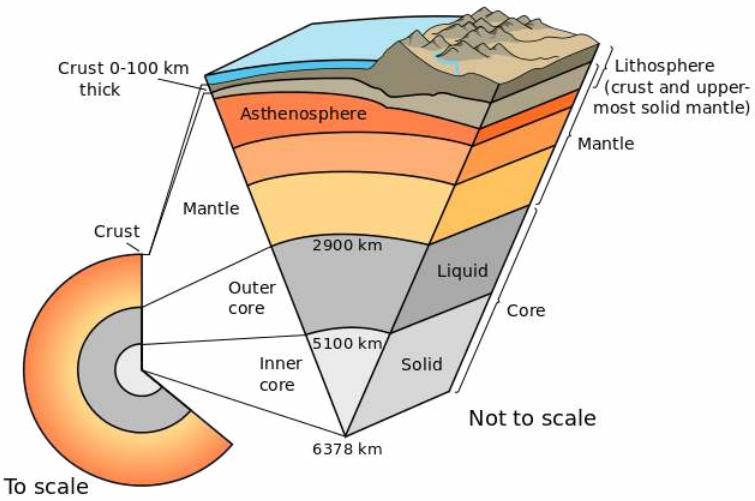


Figure 2.4.4: The layers of the Earth. Physical layers include lithosphere and asthenosphere; chemical layers are crust, mantle, and core.

In order for plate tectonics to work as it does currently, it necessarily must have continents. However, the easiest way to create continental material is via assimilation and differentiation of existing continents (see An Introduction to Geology: Chapter 4). 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 thicker crust. Nevertheless, the continents formed by some incremental process during the early history of Earth. The best idea is that density differences allowed lighter felsic materials to float upward and heavier ultramafic materials and metallic iron to sink. These

density differences led to the layering of the Earth, the layers that are now detected by seismic studies. Early protocontinents accumulated felsic materials as developing plate-tectonic processes brought lighter material from the mantle to the surface.

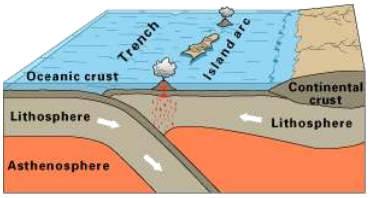


Figure 2.4.5: Subduction of an oceanic plate beneath another oceanic plate, forming a trench and an island arc. Several island arcs might combine and eventually evolve into a continent.

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

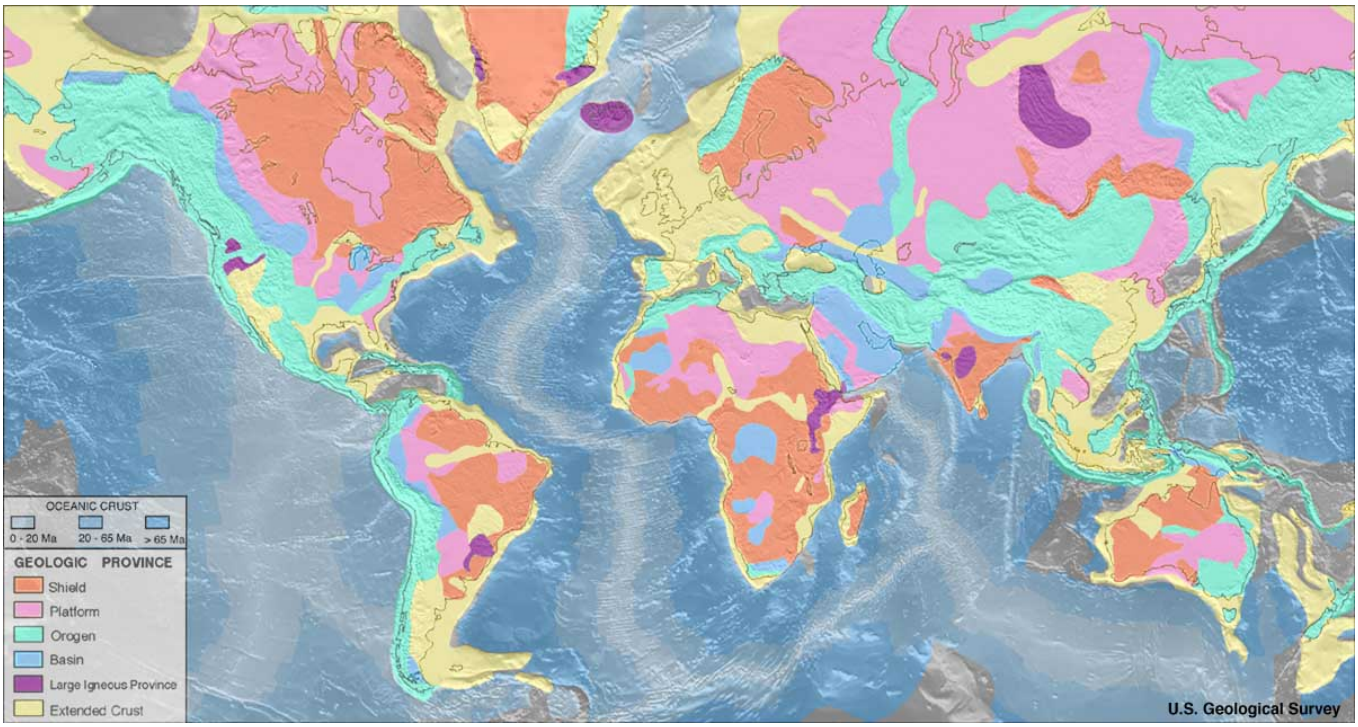


Figure 2.4.6: Geologic provinces of Earth. Cratons are pink and orange.

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

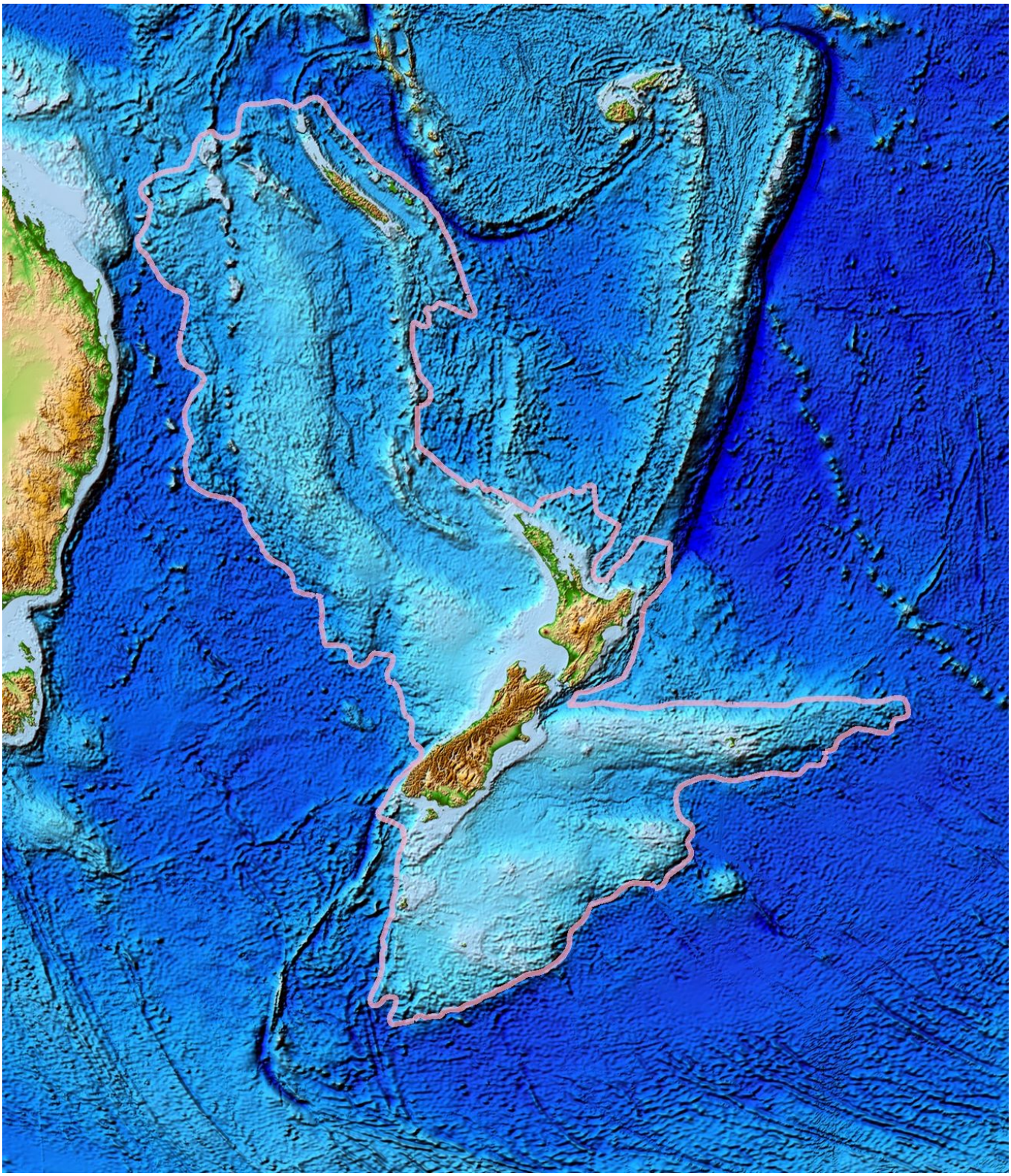


Figure 2.4.7: The continent of Zealandia

The general guideline as to what constitutes a continent and differentiates oceanic from continental crust is under some debate. At passive margins, continental crust grades into 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, that includes present-day New Zealand, would

be considered a continent. 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



Figure 2.4.8: Fossils of microbial mats from Sweden

Life most likely started during the late Hadean or early Archean Eons. The earliest evidence of life are 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 are 3.8–4.3 billion-year-old microscopic filaments from a hydrothermal vent deposit in Quebec, Canada. While the chemical and microscopic filaments evidence is not as robust as fossils, there is significant fossil evidence for life at 3.5 billion years ago. These first well-preserved fossils are photosynthetic microbial mats, called stromatolites, found in Australia.

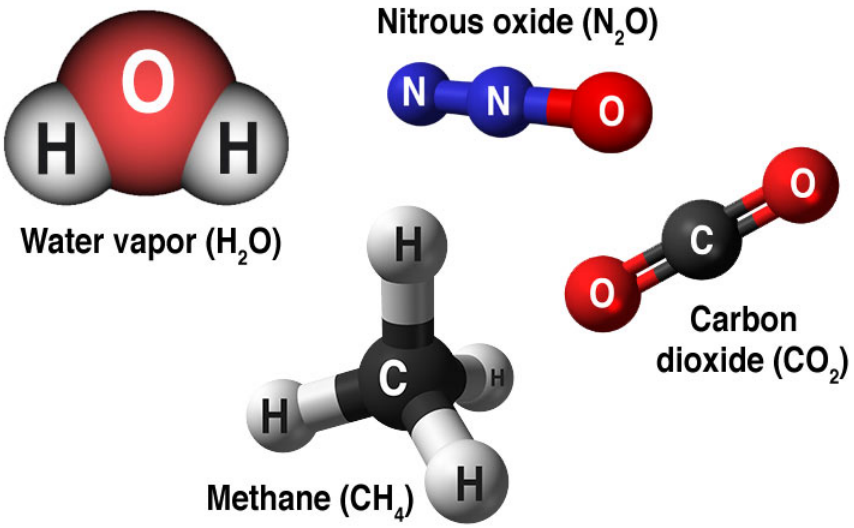


Figure 2.4.9: Greenhouse gases were more common in Earth's early atmosphere.

Although the origin of life on Earth is unknown, hypotheses include a chemical origin in the early atmosphere and ocean, deep-sea hydrothermal vents, and delivery to Earth by comets or other objects. One hypothesis is that life arose from the chemical environment of the Earth's early atmosphere and oceans, which was very different than today. The oxygen-free atmosphere produced a reducing environment with abundant methane, carbon dioxide, sulfur, and nitrogen compounds. This is what the atmosphere is like on other bodies in the solar system. In the famous Miller-Urey experiment, researchers simulated early Earth's atmosphere and lightning within a sealed vessel. After igniting sparks within the vessel, they discovered the formation of amino acids, the fundamental building blocks of proteins. In 1977, when scientists discovered an isolated ecosystem around hydrothermal vents on a deep-sea mid-ocean ridge (see An Introduction to Geology: Chapter 4), 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 is deriving its energy from hot chemical-rich waters pouring out of underground towers. This suggests that life could have started on the deep ocean floor and derived energy from the heat from the Earth's interior via chemosynthesis. Scientists have since expanded the search for life to more unconventional places, like Jupiter's icy moon Europa.

Animation of the original Miller-Urey 1959 experiment that simulated the early atmosphere and created amino acids from simple elements and compounds.

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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2.5: Proterzoic Eon

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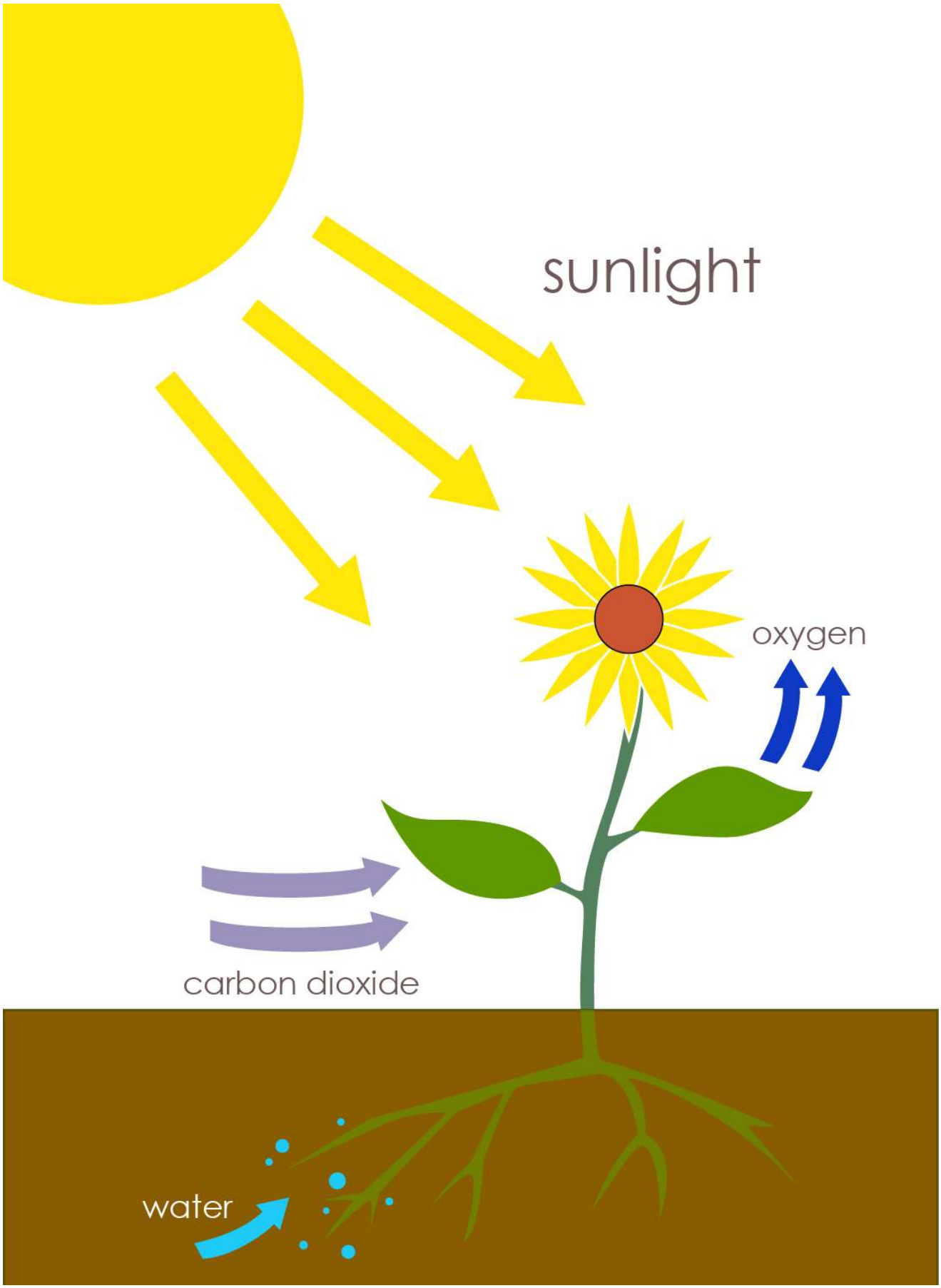

sunlight
oxygen
carbon dioxide
water

Figure 2.5.1: Diagram showing the main products and reactants in photosynthesis. The one product that is not shown is sugar, which is the chemical energy that goes into constructing the plant, and the energy that is stored in the plant which is used later by the plant or by animals that consume the plant.

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. Cyanobacteria were incorporated into larger collaborations through endosymbiosis, making algae and then plants, and the spread of photosynthesis allowed them to completely transformed the oceans and later the atmosphere by adding massive amounts of free oxygen gas (O_2) and initiated what is called the **Great Oxygenation Event (GOE)**. 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, which are dated at about 2 billion years.

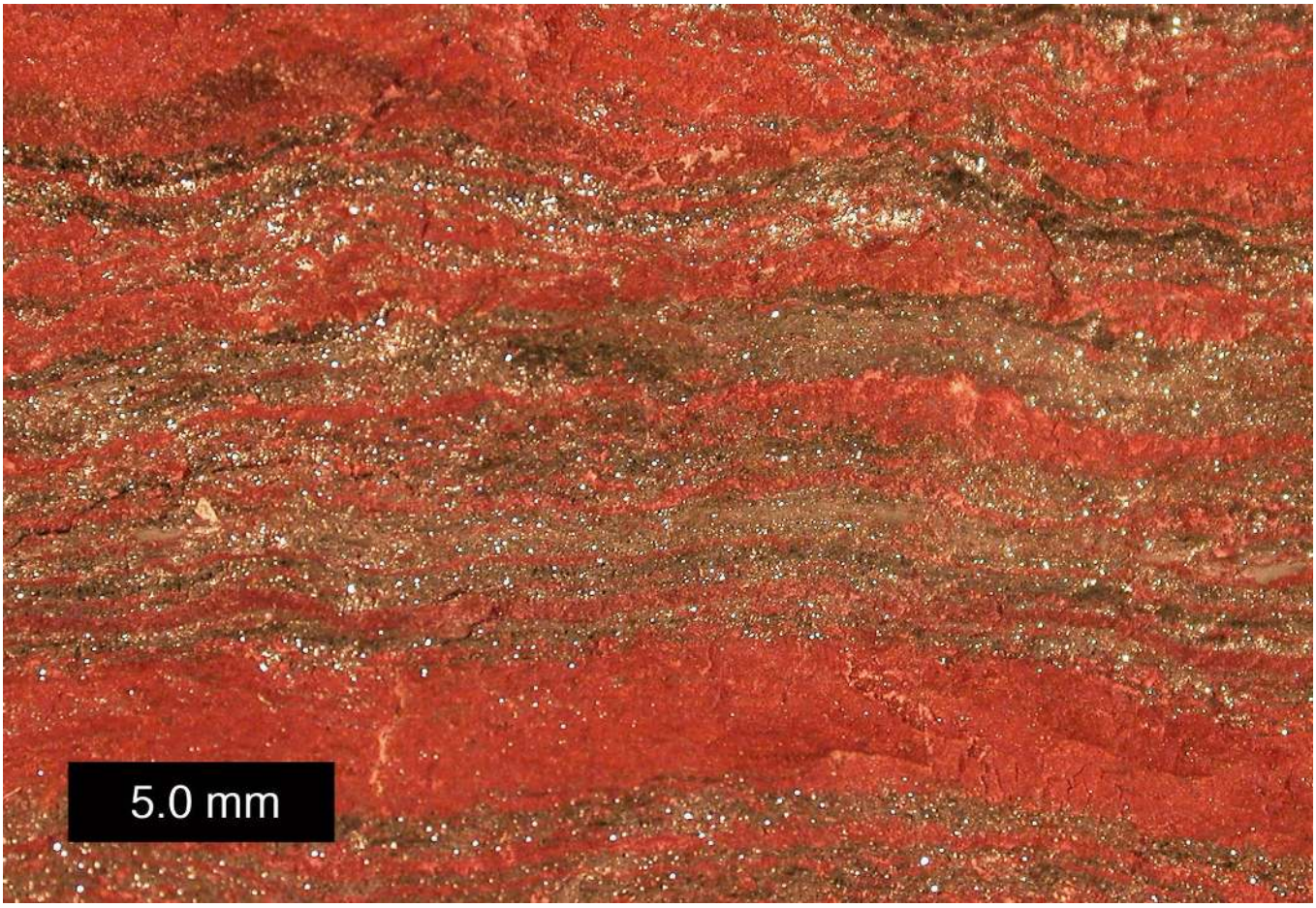


Figure 2.5.2: Alternating bands of iron-rich and silica-rich mud, formed as oxygen combined with dissolved iron.

The formation of iron oxide minerals and red chert (see figure) 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 started bubbling out into the atmosphere. Oxygenation of the atmosphere is the single 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 reacted 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

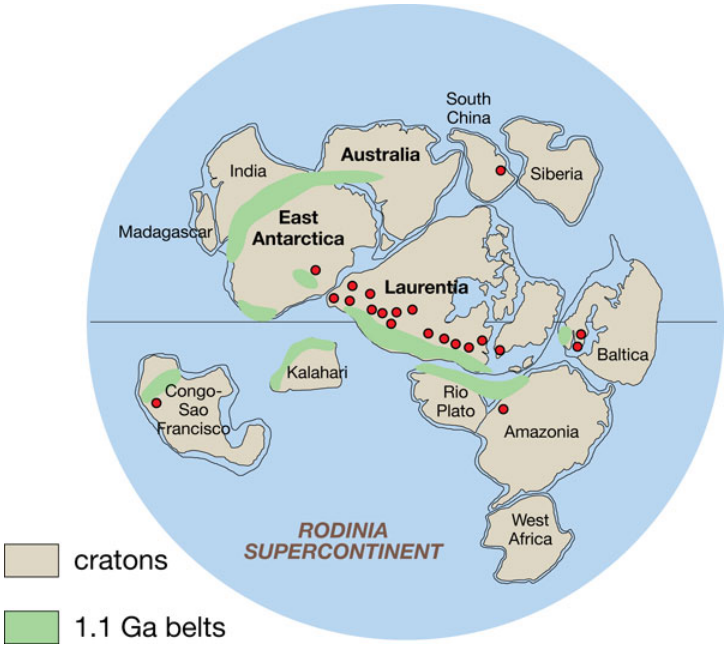


Figure 2.5.3: One possible reconstruction of Rodinia 1.1 billion years ago. (John Goodge, modified from Dalziel (1997).)

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 paleomagnetics to orient to magnetic north.

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



Figure 2.5.4: Modern cyanobacteria (as stromatolites) in Shark Bay, 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 produced 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.



Figure 2.5.5: Fossil stromatolites in Saratoga Springs, New York.

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. The nuclear DNA is capable of more complex replication and regulation than that of prokaryotic cells. The organelles include mitochondria for producing energy and chloroplasts for photosynthesis. The eukaryote branch in the tree of life gave rise to fungi, plants, and animals.

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, contributing to more complexity among individual organisms and within ecosystems (see An Introduction to Geology: Chapter 7).

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.

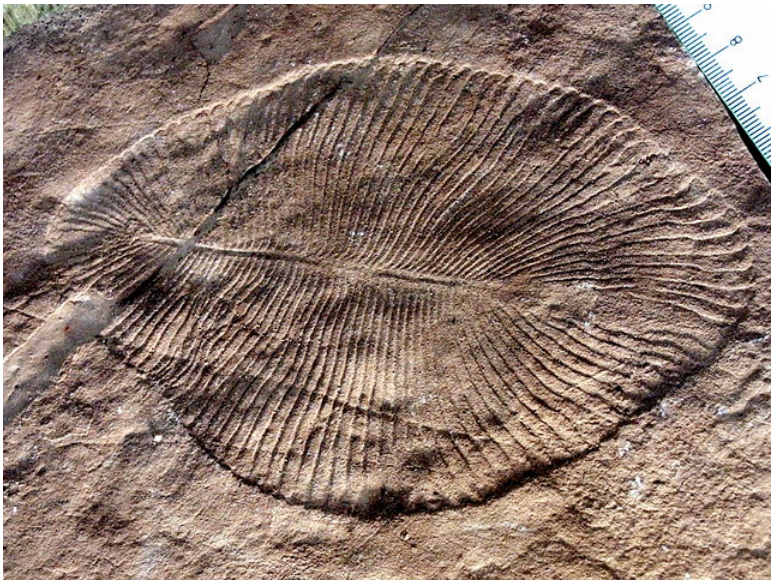


Figure 2.5.6: Dickinsonia, a typical Ediacaran fossil.

Fauna during the Ediacaran Period, 635.5 to 541 million years ago, are known as the **Ediacaran fauna**, and offer 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 were 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 beginning of the Phanerozoic Eons. This evolutionary explosion of biological diversity made a dramatic difference in scientists' ability to understand the history of life on Earth.

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2.6: Phanerozoic Eon - Paleozoic Era



Figure 2.6.1: The trilobites had a hard exoskeleton, and were an early arthropod, the same group that includes modern insects, crustaceans, and arachnids.

The **Phanerozoic** Eon is the most recent, 541 million years ago to today, and means “visible life” because the Phanerozoic rock record is marked by an abundance of fossils. Phanerozoic organisms had hard body parts like claws, scales, shells, and bones that were more easily preserved as fossils. Rocks from the older Precambrian time are less commonly found and rarely include fossils because these organisms had soft body parts. Phanerozoic rocks are younger, more common, and contain the majority of extant fossils. The study of rocks from this eon yields much greater detail. The Phanerozoic is subdivided into three eras, from oldest to

youngest they are Paleozoic (“ancient life”), Mesozoic (“middle life”), and Cenozoic (“recent life”) and the remaining three chapter headings are on these three important eras.

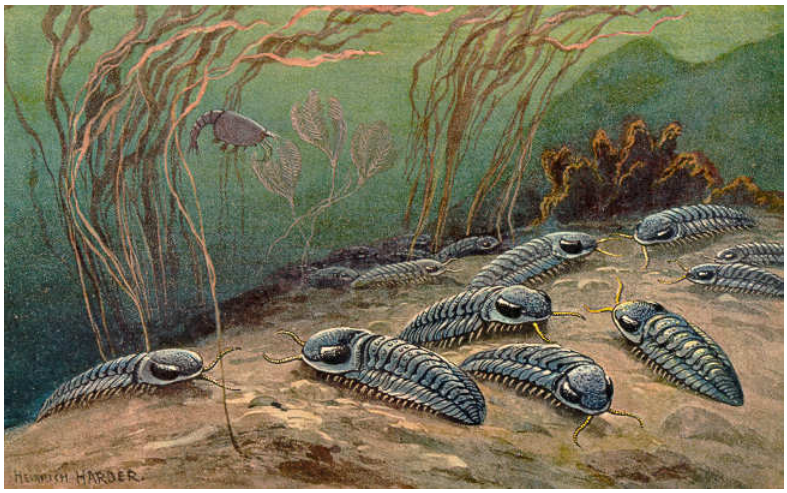


Figure 2.6.2: Trilobites, by Heinrich Harder, 1916.

Life in the early **Paleozoic Era** was dominated by marine organisms but by the middle of the era plants and animals evolved to live and reproduce on land. Fish evolved jaws and fins evolved into jointed limbs. The development of lungs allowed animals to emerge from the sea and become the first air-breathing tetrapods (four-legged animals) such as amphibians. From amphibians evolved reptiles with the amniotic egg. From reptiles evolved an early ancestor to birds and mammals and their scales became feathers and fur. Near the end of the Paleozoic Era, the Carboniferous Period had some of the most extensive forests in Earth’s history. Their fossilized remains became the coal that powered the industrial revolution

Paleozoic Tectonics and Paleogeography



Figure 2.6.3: Laurentia, which makes up the North American craton.

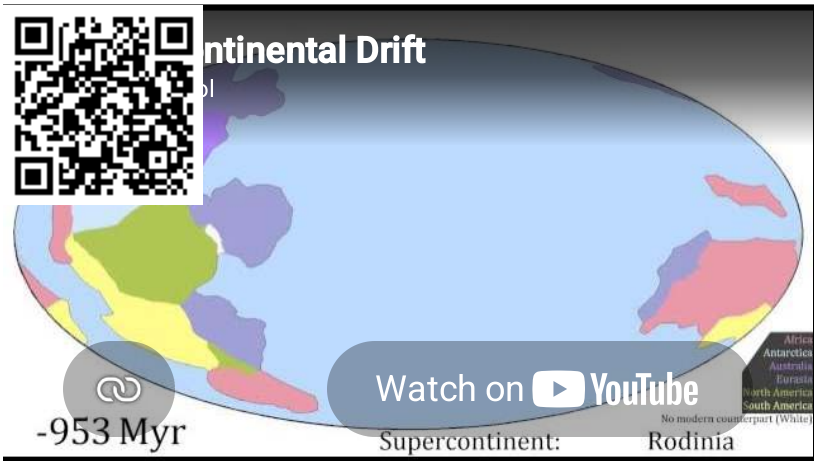
During the Paleozoic Era, sea-levels rose and fell four times. With each sea-level rise, the majority of North America was covered by a shallow tropical ocean. Evidence of these submersions are the abundant marine sedimentary rocks such as limestone with fossils corals and ooids. Extensive sea-level falls are documented by widespread unconformities. Today, the midcontinent has

extensive marine sedimentary rocks from the Paleozoic and western North America has thick layers of marine limestone on block faulted mountain ranges such as Mt. Timpanogos near Provo, Utah.



Figure 2.6.4: A reconstruction of Pangaea, showing approximate positions of modern continents.

The assembly of supercontinent **Pangea**, sometimes spelled Pangaea, was completed by the late Paleozoic Era. The name Pangea was originally coined by Alfred Wegener and means “all land.” Pangea is the when all of the major continents were grouped together as one by a series of tectonic events including subduction island-arc accretion, and continental collisions, and ocean-basin closures. In North America, these tectonic events occurred on the east coast and are known as the Taconian, Acadian, Caledonian, and Alleghanian orogenies. 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, 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 consisted of the remaining continents of South America, Africa, India, Australia, and Antarctica.



Animation of plate movement the last 3.3 billion years. Pangea occurs at the 4:40 mark.

While the east coast of North America was tectonically active during the Paleozoic Era, the west coast remained mostly inactive as a passive margin during the early Paleozoic. The western edge of North American continent was near the present-day Nevada-Utah border and was an expansive shallow continental shelf near the paleoequator. However, by the Devonian Period, the Antler orogeny started on the west coast and lasted until the Pennsylvanian Period. The Antler orogeny was a volcanic island arc that was accreted onto western North America with the subduction direction away from North America. This created a mountain range on the west coast of North American called the Antler highlands and was the first part of building the land in the west that would eventually make most of California, Oregon, and Washington states. By the late Paleozoic, the Sonoma orogeny began on the west coast and was another collision of an island arc. The Sonoma orogeny marks the change in subduction direction to be toward North America with a volcanic arc along the entire west coast of North America by late Paleozoic to early Mesozoic Eras.

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. The west coast of North America had smaller and 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.



Figure 2.6.5: Anomalocaris reconstruction by the MUSE science museum in Italy.

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. That is, 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 as a result of 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, fauna may have been diverse during the Ediacaran Period, setting the state for the Cambrian Explosion, but they lacked hard body parts and would have left few fossils behind. Regardless, during the Cambrian Period 541–485 million years ago marked the appearance of most animal phyla.

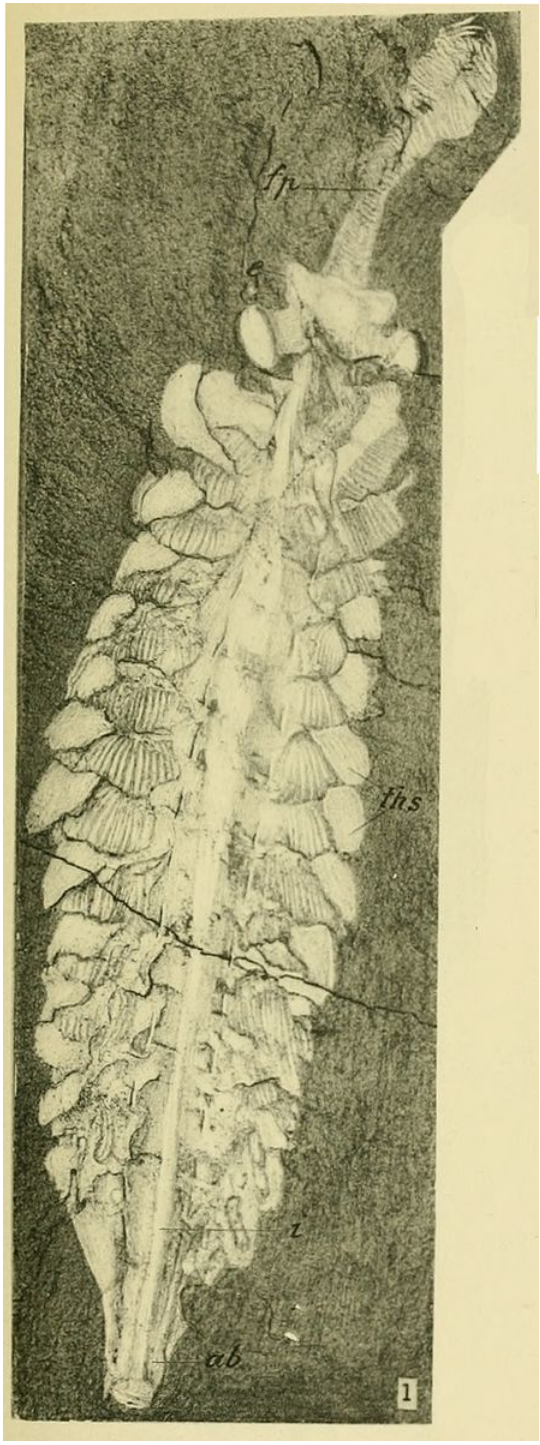


Figure 2.6.6: Original plate from Walcott's 1912 description of Opabinia, with labels: fp = frontal appendage, e = eye, ths = thoracic somites, i = intestine, ab = abdominal segment.

One of the best fossil sites for the Cambrian Explosion was discovered in 1909 by Charles Walcott (1850–1927) in the Burgess Shale in western Canada. The **Burgess Shale** is a **Lagerstätte**, 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 scientist 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, animals with backbones.



Figure 2.6.7: A modern coral reef.

By the end of the Cambrian, mollusks, brachiopods, nautiloids, gastropods, graptolites, echinoderms, and trilobites covered the sea floor. Although most animal phyla appeared by the Cambrian, the biodiversity at the family, genus, and species level was low until the Ordovician Period. During the **Great Ordovician Biodiversification Event**, vertebrates and invertebrates (animals without backbone) became more diverse and complex at family, genus, and species level. The cause of the rapid speciation event is still debated but some likely causes are a combination of warm temperatures, expansive continental shelves near the equator, and more volcanism along the mid-ocean ridges. Some have shown evidence that an asteroid breakup event and consequent heavy meteorite impacts correlate with this diversification event. The additional volcanism added nutrients to ocean water helping support a robust ecosystem. Many life forms and ecosystems that would be recognizable in current times appeared at this time. Mollusks, corals, and arthropods in particular multiplied to dominate the oceans.



Figure 2.6.8: Guadalupe Mountains National Park in Texas is made of a giant fossil reef complex.

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.

According to evidence from glacial deposits, a small ice age caused sea-levels to drop and led to a major mass extinction by the end of the Ordovician. 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).



4-minute video describing mass extinctions.



Figure 2.6.9: The armor-plated fish (placoderm) Bothriolepis panderi from the Devonian of Russia.

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

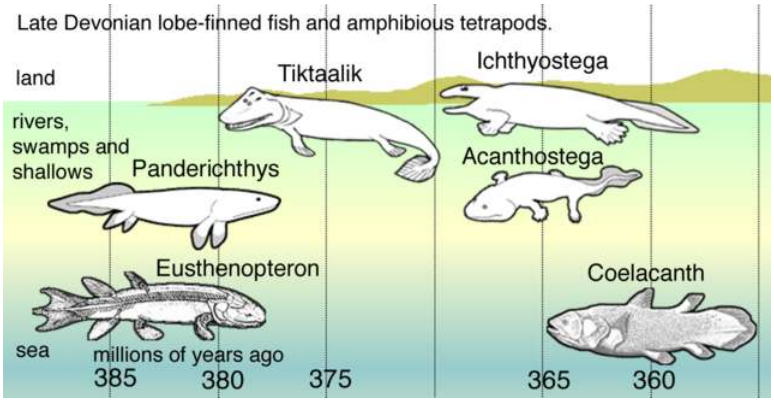


Figure 2.6.10: Several different types of fish and amphibians that led to walking on land.

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

Land plants had also evolved into the first trees and forests. 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.

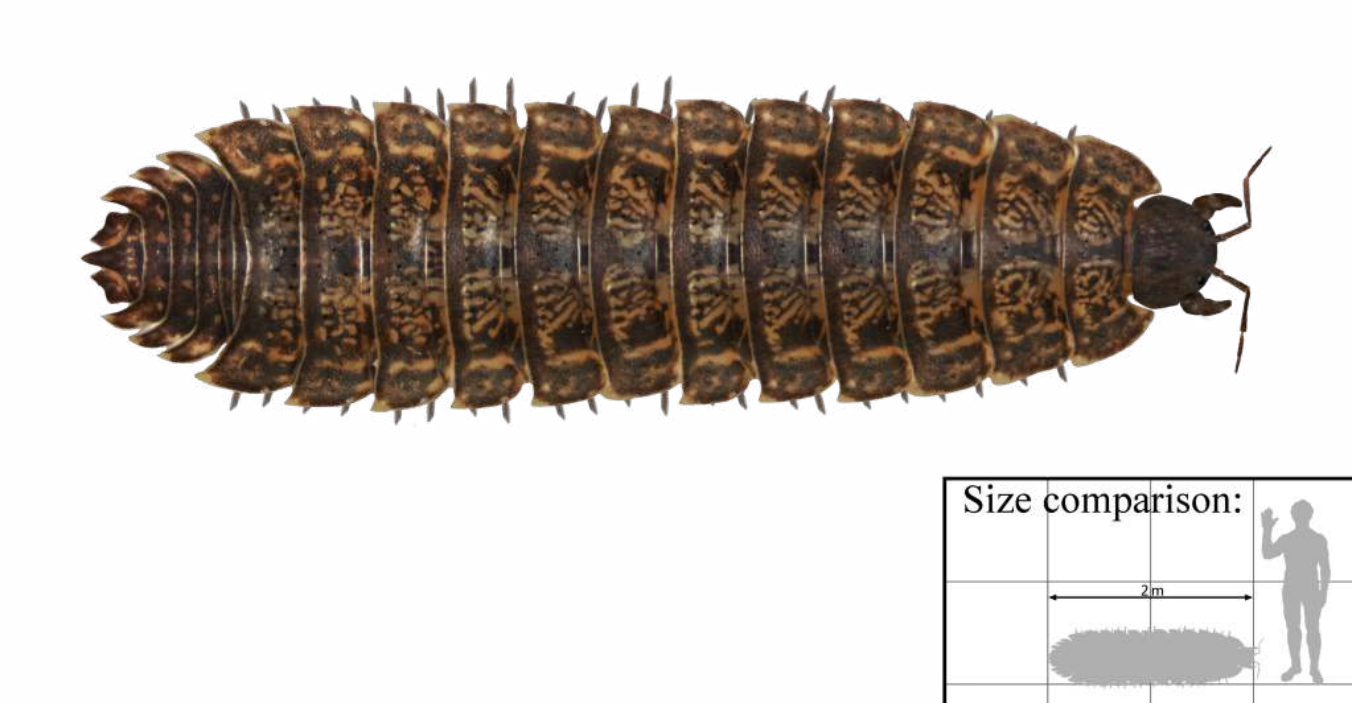


Figure 2.6.11: A reconstruction of the giant arthropod (insects and their relatives) Arthropleura.

The next time period, called the Carboniferous (North American geologists have subdivided this into the Mississippian and Pennsylvanian periods), saw the highest levels of oxygen ever known, with forests (e.g., ferns, club mosses) and swamps dominating the landscape. This helped cause the largest arthropods ever, like the millipede Arthropleura, at 2.5 meters (6.4 feet) long! It also saw the rise of a new group of animals, the reptiles. The evolutionary advantage that reptiles have over amphibians is the amniote egg (egg with a protective shell), which allows them to rely on non-aquatic environments for reproduction. This widened the terrestrial reach of reptiles compared to amphibians. This booming life, especially plant life, created cooling temperatures as carbon dioxide was removed from the atmosphere. 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.

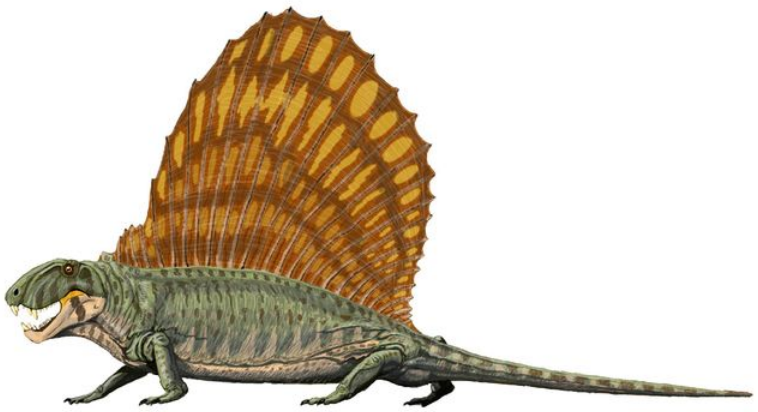


Figure 2.6.12: Reconstruction of Dimetrodon.

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*) are the first group of synapsids that exhibit the beginnings of mammalian characteristics such as well-differentiated dentition: incisors, highly developed canines in lower and

upper jaws and cheek teeth, premolars and molars. Starting in the late Permian, a second group of synapsids, called the therapsids (or mammal-like reptiles) evolve, and become the ancestors to mammals.

PERMIAN MASS EXTINCTION

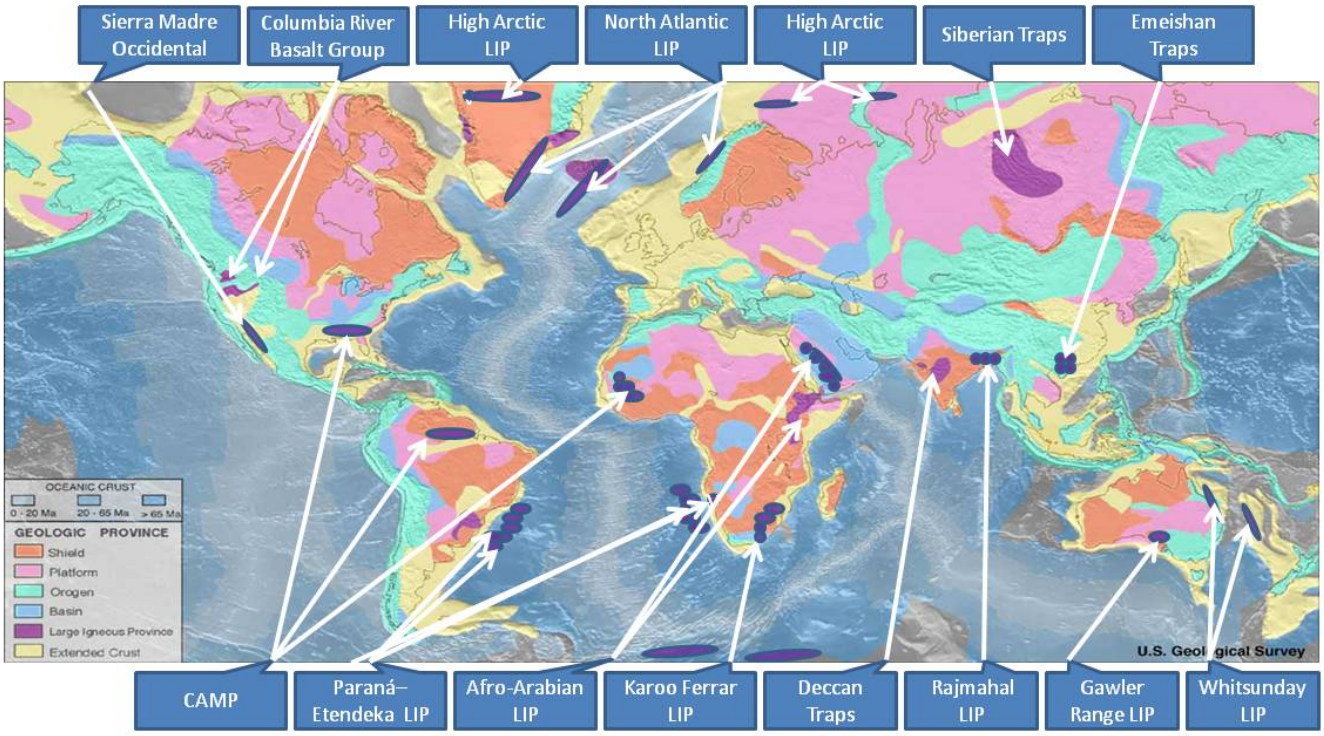


Figure 2.6.13: Map of global flood basalts. Note the largest is the Siberian Traps.

The end of the Paleozoic era is marked by the largest mass extinction in earth history. The Paleozoic era had two smaller mass extinctions, but these were not as large as the **Permian Mass Extinction**, also known as the Permian-Triassic Extinction Event. It is estimated that up to 96% of marine species and 70% of land-dwelling (terrestrial) vertebrates went extinct. Many famous organisms, like sea scorpions and trilobites, were never seen again in the fossil record. What caused such a widespread extinction event? The exact cause is still debated, though the leading idea relates to extensive volcanism associated with the **Siberian Traps**, which are one of the largest deposits of flood basalts known on Earth, dating to the time of the extinction event. The eruption size is estimated at over 3 million cubic kilometers that is approximately 4,000,000 times larger than the famous 1980 Mt. St. Helens eruption in Washington. The unusually large volcanic eruption would have contributed a large amount of 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 in An Introduction to Geology: Chapter 15, 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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2.7: Phanerozoic Eon - Mesozoic Era

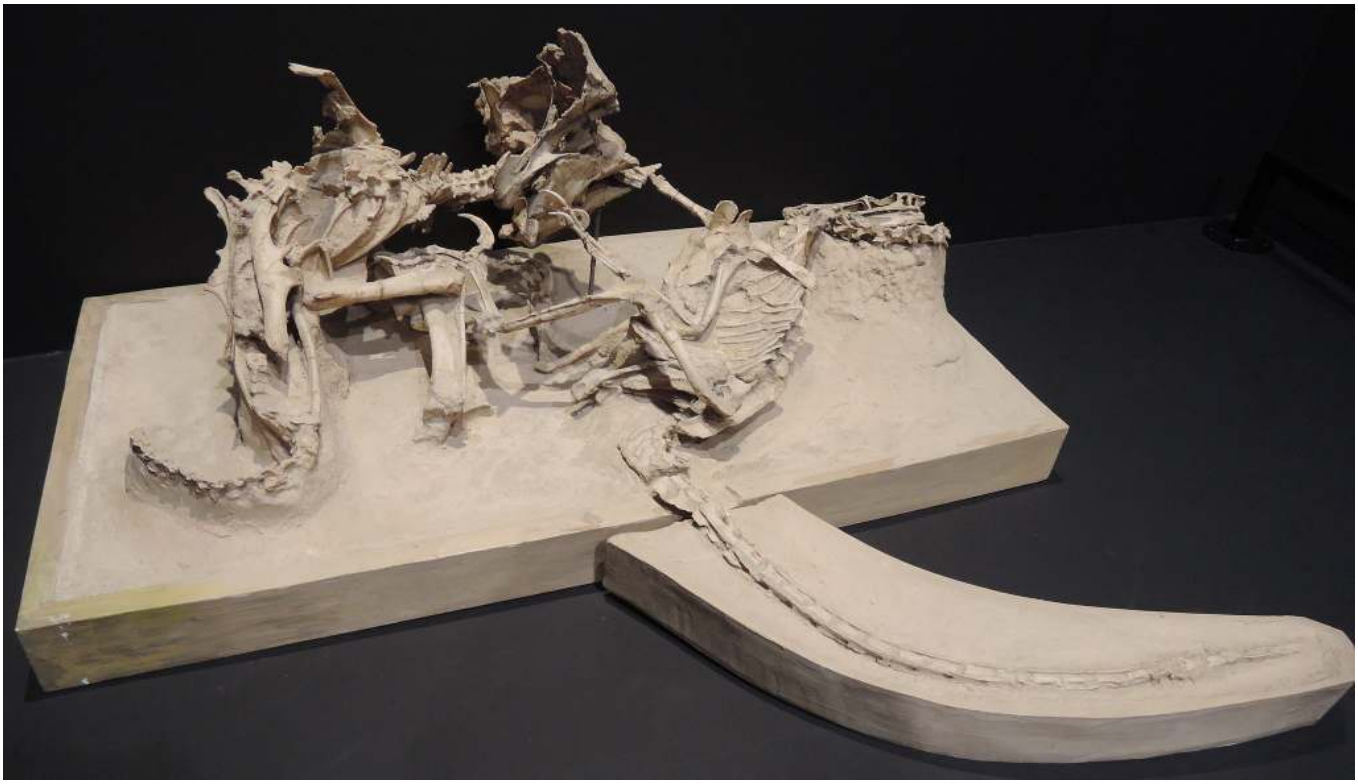


Figure 2.7.1: Perhaps the greatest fossil ever found, a velociraptor attacked a protoceratops, and both were fossilized mid sequence.

Following the Permian Mass Extinction, the **Mesozoic** ("middle life") was from 252 million years ago to 66 million years ago. As Pangea started to break apart, mammals, birds, and flowering plants developed. The Mesozoic is probably best known as the age of reptiles, most notably, the dinosaurs.

Mesozoic Tectonics and Paleogeography



Figure 2.7.2: Animation showing Pangea breaking up

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

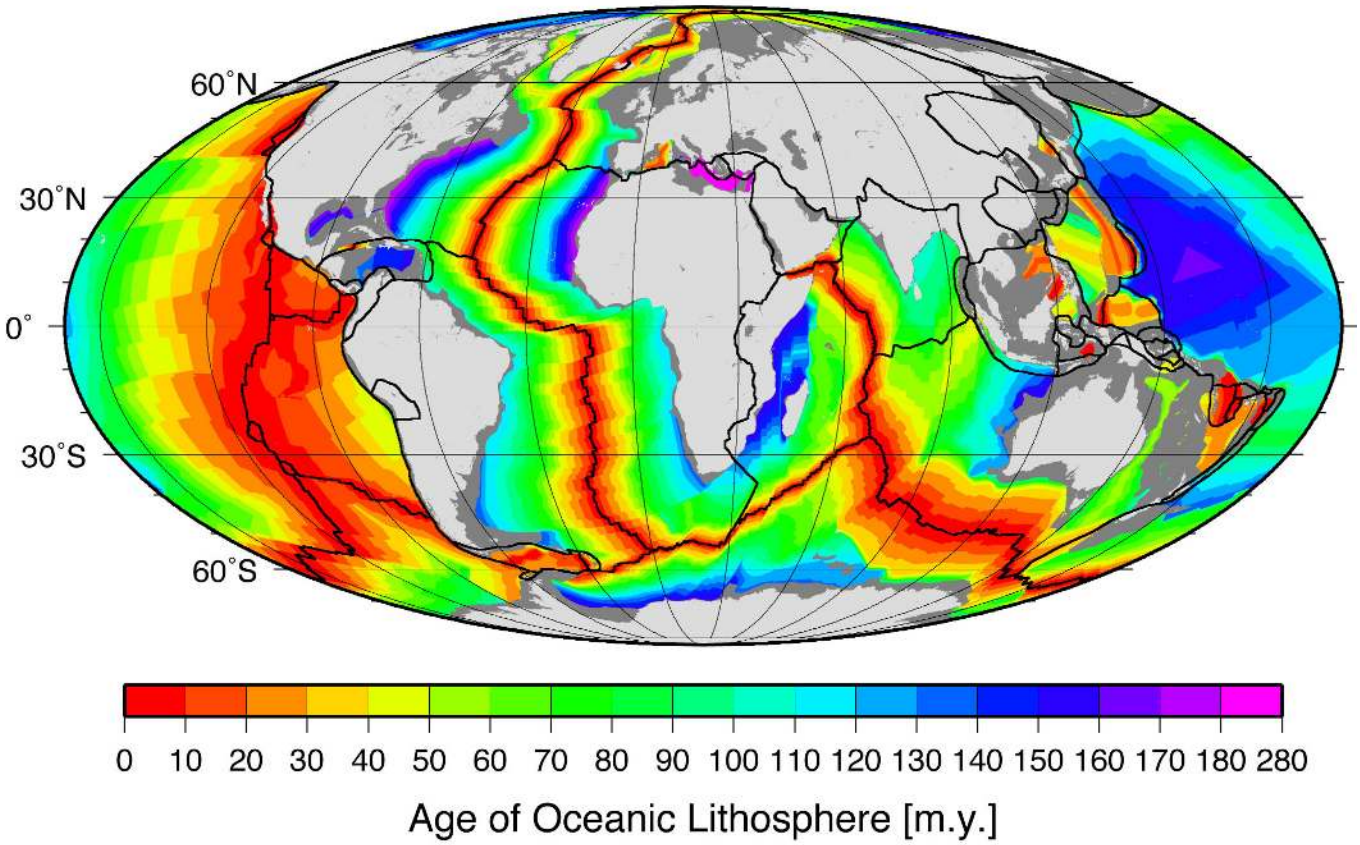


Figure 2.7.3: Age of oceanic lithosphere, in millions of years. Notice the differences in the Atlantic Ocean along the coasts of the continents.

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 (happening after the Mesozoic ended) was the northernmost Atlantic, with Greenland and Scandinavia parting ways. The breaking points of each rifted plate margin eventually turned into the passive plate boundaries of the east coast of the Americas today.

Video of Pangea breaking apart and plates moving to their present locations. By Tanya Atwater.

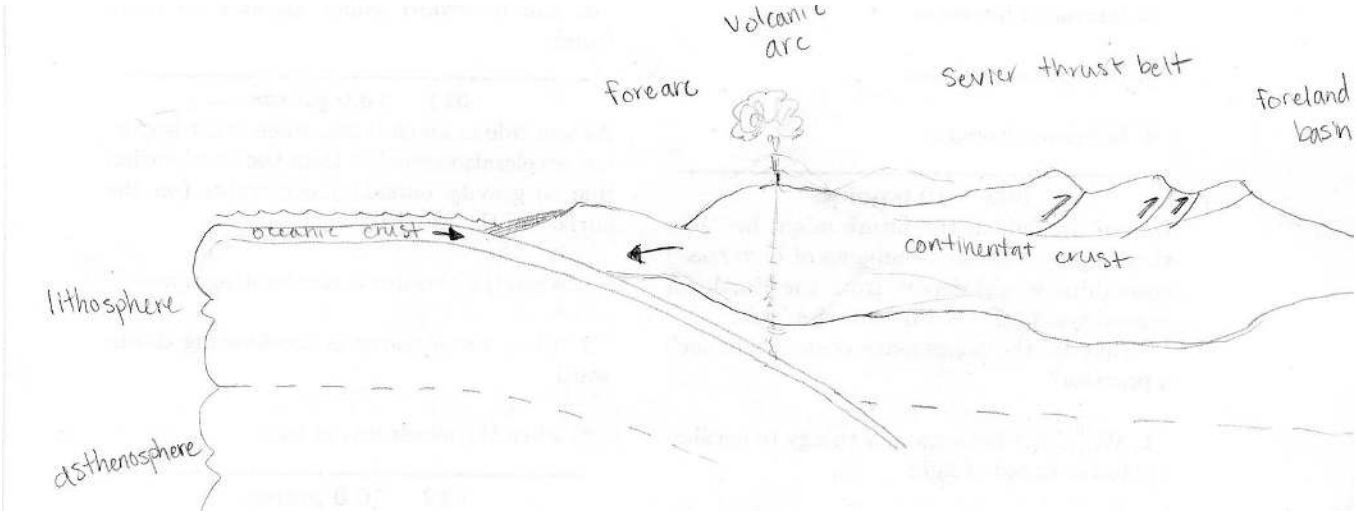


Figure 2.7.4: Sketch of the major features of the Sevier Orogeny.

In western North America, an active plate margin had started with subduction, controlling most of the tectonics of that region in the Mesozoic. Another possible island-arc collision created the Sonoman Orogeny in Nevada during the latest Paleozoic to the Triassic. In the Jurassic, another island-arc collision caused the Nevadan Orogeny, a large Andean-style volcanic arc and thrust belt. The Sevier Orogeny followed in the Cretaceous, which was mainly a volcanic arc to the west and a thin-skinned fold and thrust belt to the east, meaning stacks of shallow faults and folds built up the topography. Many of the structures in the Rocky Mountains today date from this orogeny.



Figure 2.7.5: The Cretaceous Interior Seaway in the mid-Cretaceous.

Tectonics had an influence in one more important geographic feature in North America: the Cretaceous Western Interior Foreland Basin, which flooded during high sea levels forming the **Cretaceous Interior Seaway**. Subduction from the west was the Farallon Plate, an oceanic plate connected to the Pacific Plate (seen today as remnants such as the Juan de Fuca Plate, off the coast of the Pacific Northwest). Subduction was shallow at this time because a very young, hot and less dense portion of the Farallon plate was subducted. This shallow subduction caused a downwarping in the central part of North America. High sea levels due to shallow subduction, and increasing rates of seafloor spreading and subduction, high temperatures, and melted ice also contributed to the high sea levels. 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.



Figure 2.7.6: A Mesozoic scene from the late Jurassic.

The Mesozoic era is dominated by reptiles, and more 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 the archosaurs. The archosaurs ("ruling reptiles") include ancestral groups that went extinct at the end of the Triassic, as well as the flying pterosaurs, crocodilians, and the dinosaurs. Archosaurs, like the placental mammals after them, occupied all major environments: terrestrial (dinosaurs), in the air (pterosaurs), aquatic (crocodilians) and even fully marine habitats (marine crocodiles). The pterosaurs, the first vertebrate group to take flight, like the dinosaurs and mammals, start small in the Triassic.

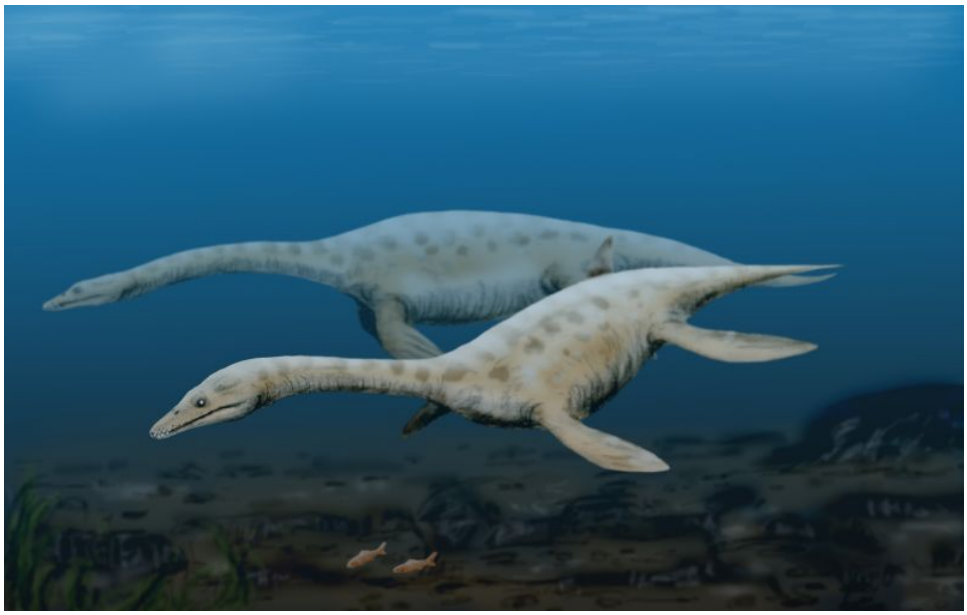


Figure 2.7.7: A drawing of the early plesiosaur Agustasaurus from the Triassic of Nevada.

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.



Figure 2.7.8: Reconstruction of the small (<5″) Megazostrodon, one of the first animals considered to be a true mammal.

Mammals, as previously mentioned, got their start from a reptile-like synapsid ancestor possibly in the late Paleozoic. True mammals did not evolve until the Triassic, and after the extinction, mammals generally stayed small, in mainly nocturnal niches, with insects being their largest prey. The development of warm-blooded circulation and fur may have been a response to this lifestyle.

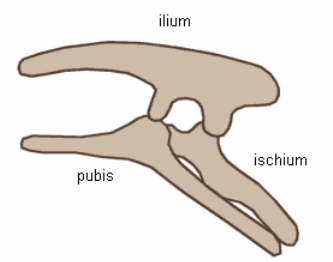


Figure 2.7.9: Closed structure of a ornithischian hip, which is similar to a birds.

In the Jurassic, species that were previously common, flourished due to a warmer and more tropical climate. The dinosaurs were relatively small animals in the Triassic period of the Mesozoic, but became truly massive in the Jurassic. Dinosaurs are split into two groups based on their hip structure, i.e. orientation of the pubis and ischium bones in relationship to each other. This is referred to as the “reptile hipped” saurischians and the “bird hipped” ornithischians. This has recently been brought into question by a new idea for dinosaur lineage.

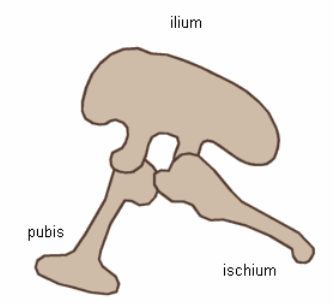


Figure 2.7.10: Open structure of a saurischian hip, which is similar to a lizards.

Most of the dinosaurs of the Triassic were saurischians, but all of them were bipedal. The major adaptive advantage dinosaurs had was changes in the hip and ankle bones, tucking the legs under the body for improved locomotion as opposed to the semi-erect gait of crocodiles or the sprawling posture of reptiles. In the Jurassic, limbs (or a lack thereof) were also important to another group of reptiles, leading to the evolution of *Eophis*, the oldest snake.



Figure 2.7.11: 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. The saurischians diversified into the giant herbivorous (plant-eating) long-necked sauropods weighing up to 100 tons and bipedal carnivorous theropods, with the possible exception of the *Therizinosaurs*. All of the ornithischians (e.g *Stegosaurus*, *Iguanodon*, *Triceratops*, *Ankylosaurus*, *Pachycephhlosaurus*) were herbivorous with a strong tendency to have a “turtle-like” beak at the tips of their mouths.



Figure 2.7.12: Iconic “Berlin specimen” Archaeopteryx lithographica fossil from Germany.

The pterosaurs grew and diversified in the Jurassic, and another notable arial organism developed and thrived in the Jurassic: birds. When *Archeopteryx* 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* 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 in climbing.



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Figure 2.7.13: Reconstructed skeleton of Argentinosaurus, from Natu[illegible]ncke[illegible]ermany.

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, which are plants with flowers and seeds, had 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 from Jurassic wasp-like ancestors that co-evolved with the flowering plants during this time period. The breakup of Pangea not only shaped our modern world's geography, but biodiversity at the time as well. Throughout the Mesozoic, animals on the isolated, now separated island continents (formerly parts of Pangea), took strange evolutionary turns. This includes giant titanosaurian sauropods (*Argentinosaurus*) and theropods (*Giganotosaurus*) from South America.

K-T EXTINCTION

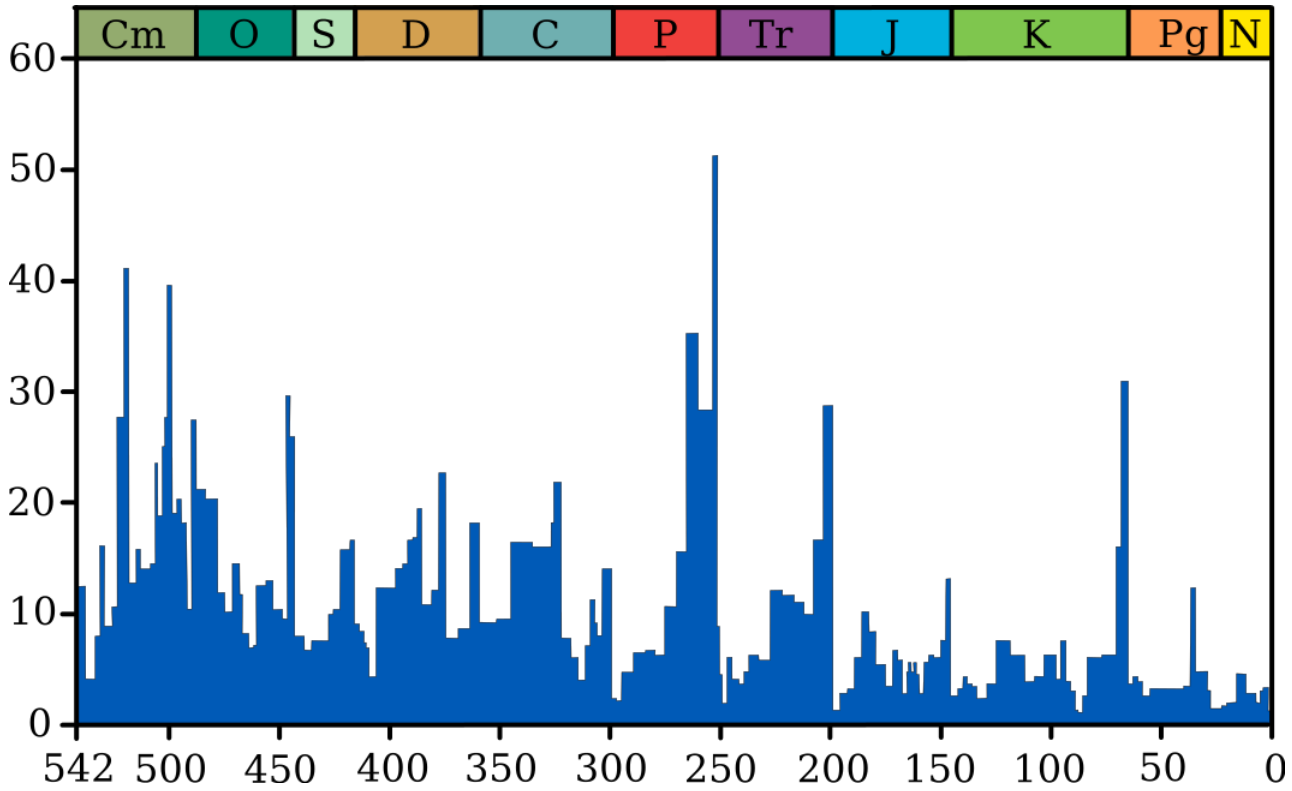


Figure 2.7.14: Graph of the rate of extinctions. Note the large spike at the end of the Cretaceous (labeled as K).

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 at this time.



Figure 2.7.15: 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 all over the world 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, that only is found at impact sites, was found in many places around the world. The huge impact created a strong thermal pulse that could be responsible for global forest fires, 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.

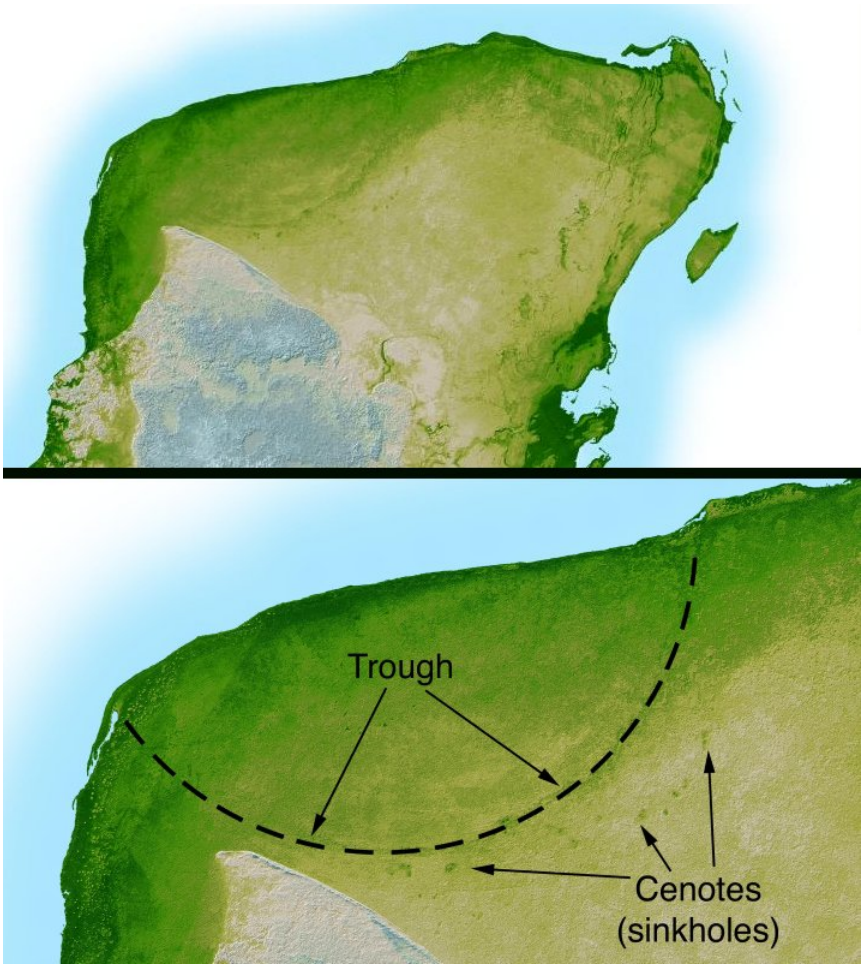


Figure 2.7.16: The land expression of the Chicxulub crater. The other side of the crater is within the Gulf of México.

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 impact for the extinction.

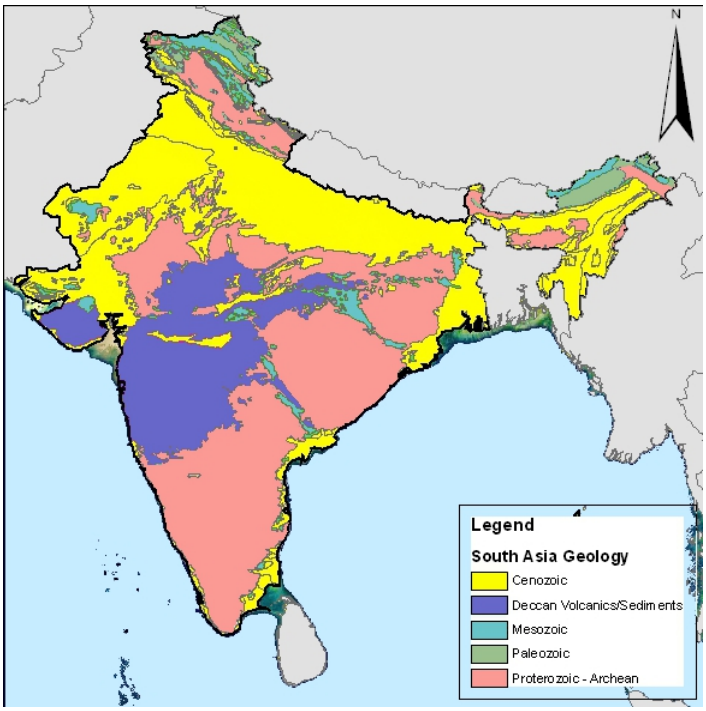


Figure 2.7.17: Geology of India, showing purple as Deccan Traps-related rocks.

With all of this information, it seems like the case would be closed. However, there are other events at this time which could have partially aided the demise of so many organisms. For example, sea levels are known to be slowly decreasing at the time of the K-T event, which is tied to marine extinctions, though 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 the **Deccan Traps** flood basalt volcanism in India. At over 1.3 million cubic kilometers of material, it was certainly a large source of material hazardous to ecosystems at the time, and it has been suggested as at least partially responsible for the extinction. Some have found the impact and eruptions too much of a coincidence, and have even linked the two together.

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2.8: Phanerozoic Eon - Cenozoic Era

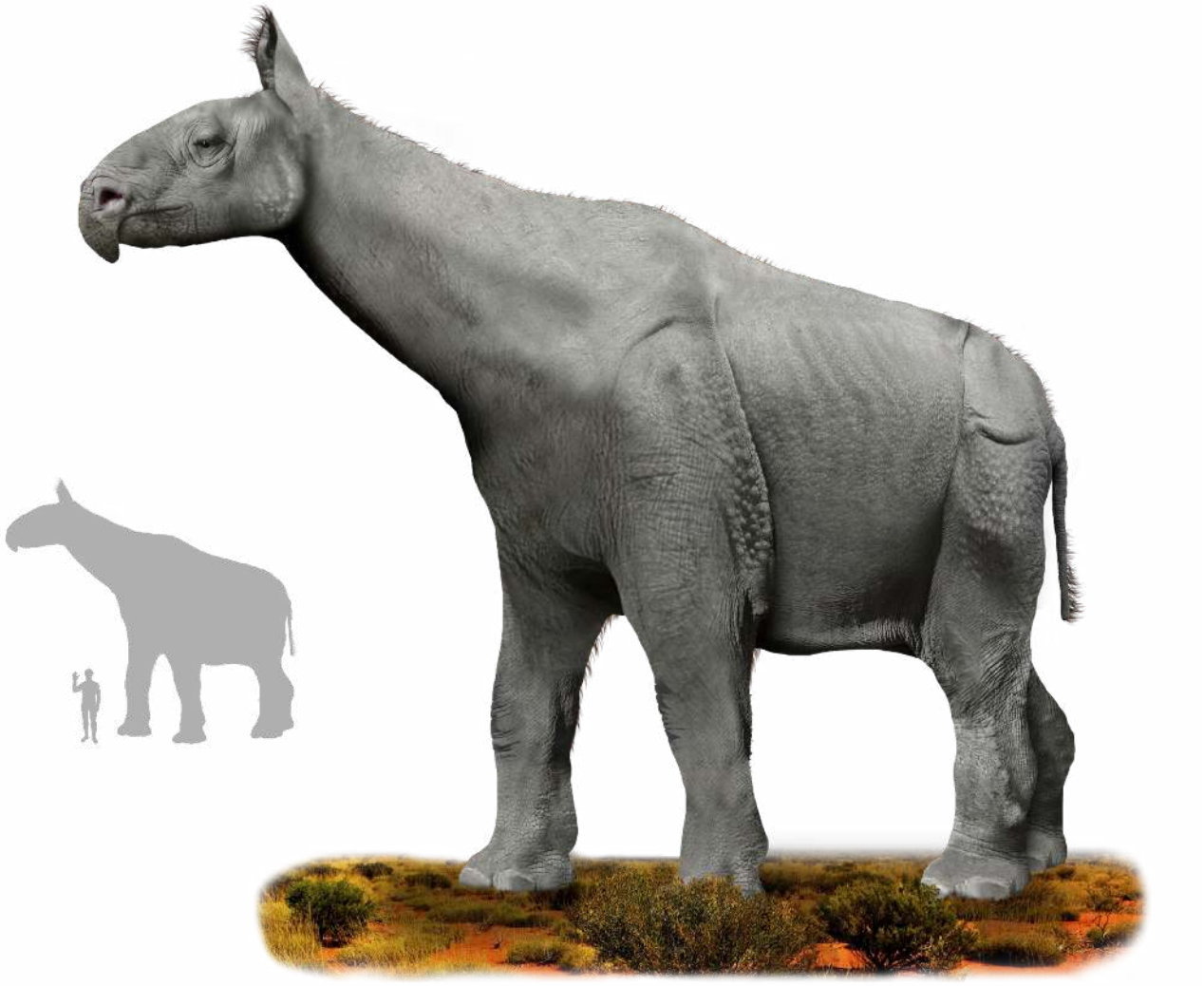
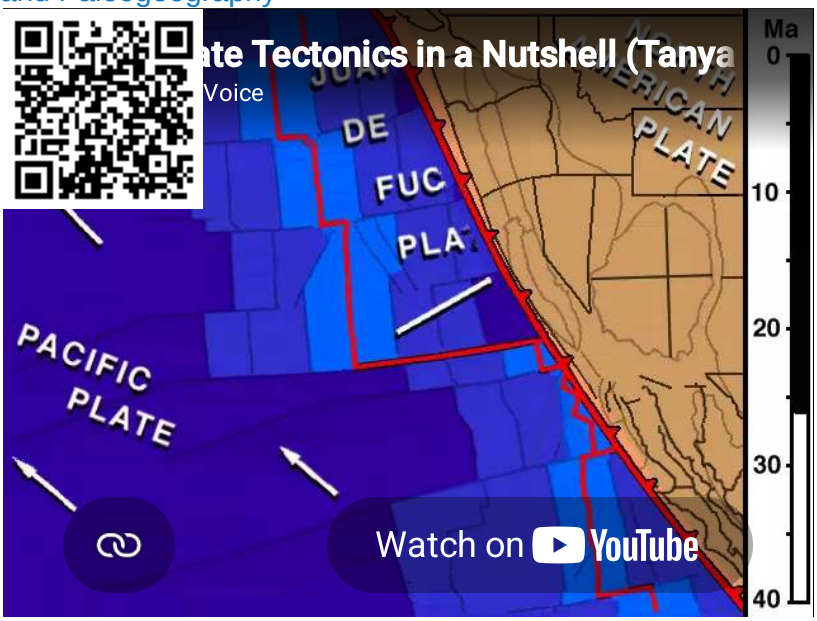


Figure 2.8.1: Paraceratherium, seen in this reconstruction, was a massive (15-20 ton, 15 foot tall) ancestor of rhinos.

The **Cenozoic**, meaning “new life,” is known as the age of mammals because it is in this era that mammals came to be a dominant and large life form, including human ancestors. Birds, as well, flourished in the open niches left by the dinosaur’s demise. Most of the Cenozoic has been relatively warm, with the main exception being the ice age that started about 2.558 million years ago and (despite recent warming) continues today. Tectonic shifts in the west caused volcanism, but eventually changed the long-standing subduction zone into a transform boundary.



Animation of the last 38 million years of movement in western North America. Note, that after the ridge is subducted, convergent turns to transform (with divergent inland).

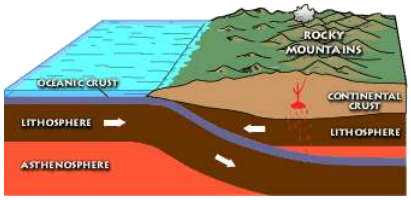


Figure 2.8.2: Shallow subduction during the Laramide Orogeny.

In the Cenozoic, the plates of the Earth moved into more familiar places, with the biggest change being the closing of the Tethys Sea with collisions such as the Alps, Zagros, and Himalaya, a collision that started about 57 million years ago, and continues today. Maybe the most significant tectonic feature that occurred in the Cenozoic of North America was the conversion of the west coast of California from a convergent boundary subduction zone to a transform boundary. Subduction off the coast of the western United States, which had occurred throughout the Mesozoic, had continued in the Cenozoic. After the Sevier Orogeny in the late Mesozoic, a subsequent orogeny called the Laramide Orogeny, occurred in the early Cenozoic. The Laramide was thick-skinned, different than the Sevier Orogeny. It involved deeper crustal rocks, and produced bulges that would become mountain ranges like the 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 and moved eastward beneath the continental plate affecting the overlying continent hundreds of miles east of the continental margin and building high mountains. This occurred because the subducting plate was so young and near the spreading center and the density of the plate was therefore low and subduction was hindered.



Figure 2.8.3: Map of the San Andreas fault, showing relative motion.

As the mid-ocean ridge itself started to subduct, the relative motion had changed. Subduction caused a relative convergence between the subducting Farallon plate and the North American plate. On the other side of the mid-ocean ridge from the Farallon plate was the Pacific plate, which was moving away from the North American plate. Thus, as the subduction zone consumed the mid-ocean ridge, the relative movement became transform instead of convergent, which went on to become the San Andreas Fault System. As the San Andreas grew, it caused east-west directed extensional forces to spread over the western United States, creating the Basin and Range province. The transform fault switched position over the last 18 million years, twisting the mountains around Los Angeles, and new faults in the southeastern California deserts may become a future San Andreas-style fault. During this switch from subduction to transform, the nearly horizontal Farallon slab began to sink into the mantle. This caused magmatism as the subducting slab sank, allowing asthenosphere material to rise around it. This event is called the Oligocene ignimbrite flare-up, which was one of the most significant periods of volcanism ever, including the largest single confirmed eruption, the 5000 cubic kilometer Fish Canyon Tuff.

Cenozoic Evolution

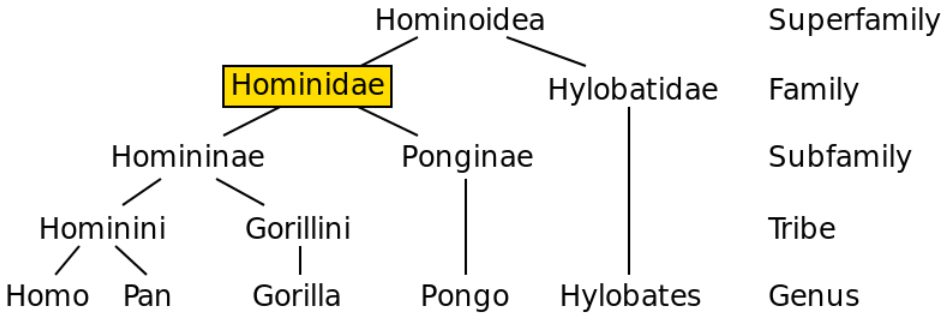


Figure 2.8.4: Family tree of Hominids (Hominadae).

There are five groups of early mammals in the fossil record, based primarily on fossil teeth, the hardest bone in vertebrate skeletons. For the purpose of this text, the most important group are the Eupantotheres, that diverge into the two main groups of mammals, the marsupials (like *Sinodelphys*) and placentals or eutherians (like *Eomaia*) in the Cretaceous and then diversified in the Cenozoic. The marsupials dominated on the isolated island continents of South America and Australia, and many went extinct in South America with the introduction of placental mammals. Some well-known mammal groups have been highly studied with interesting evolutionary stories in the Cenozoic. For example, horses started small with four toes, ended up larger and having 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 water. However, no study of evolution has been more studied than human evolution. **Hominids**, the name for human-like primates, started in eastern Africa several million years ago.



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Figure 2.8.5: Lucy skeleton from the Cleveland Natural History Museum, showing real fossil (brown) and reconstructed skeleton (white).

The first critical event in this story is an environmental change from jungle to more of a savanna, probably caused by changes in Indian Ocean circulation. While bipedalism is known to have evolved before this shift, it is generally believed that our bipedal ancestors (like *Australopithecus*) had an advantage by covering ground more easily in a more open environment compared to their non-bipedal evolutionary cousins. There is also a growing body of evidence, including the famous "Lucy" fossil of an Australopithecine, that our early ancestors lived in trees. Arboreal animals usually demand a high intelligence to navigate through a three-dimensional world. It is from this lineage that humans evolved, using endurance running as a means to acquire more resources and possibly even hunt. This can explain many uniquely human features, from our long legs, strong achilles, lack of lower gut protection, and our wide range of running efficiencies.

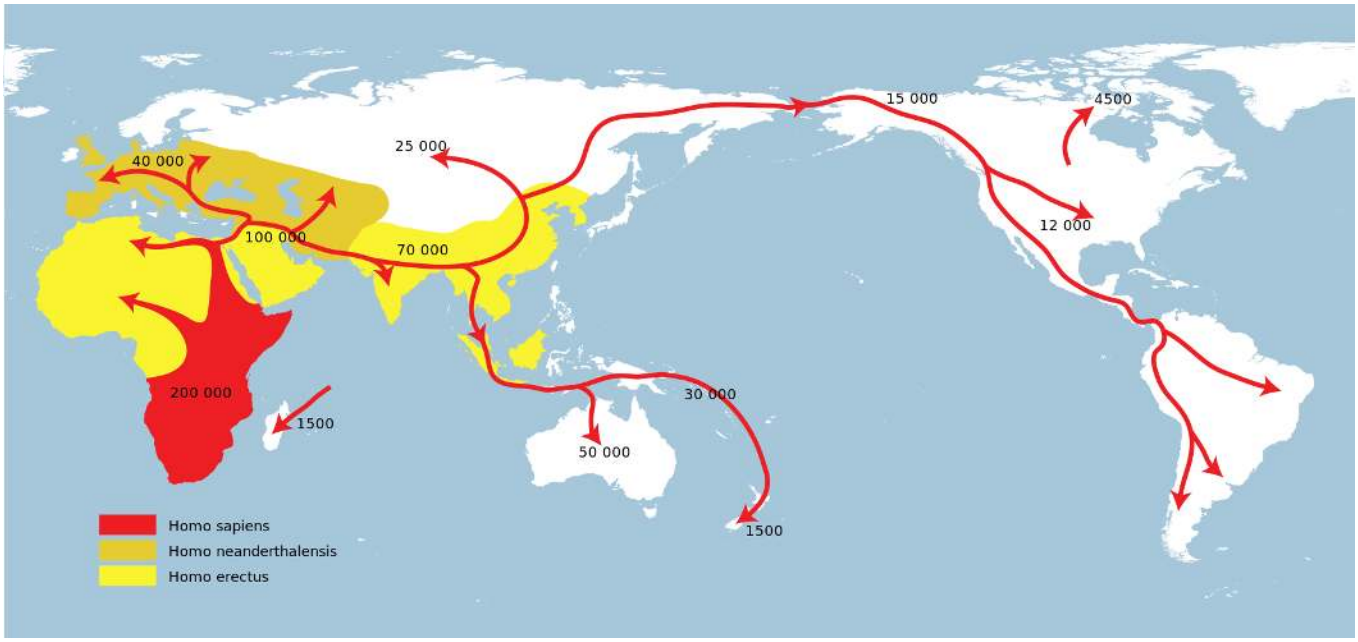


Figure 2.8.6: The hypothesized movement of the homo genus. Years are marked as to the best guess of the timing of movement.

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 of how, it was this increased cognitive power that allowed humans to reign as their ancestors moved out of Africa and explored the world, ultimately entering the Americas through land bridges like the Bering Land Bridge. The details of this worldwide migration and the different branches of the hominid evolutionary tree are very complex, and best reserved for its own course.

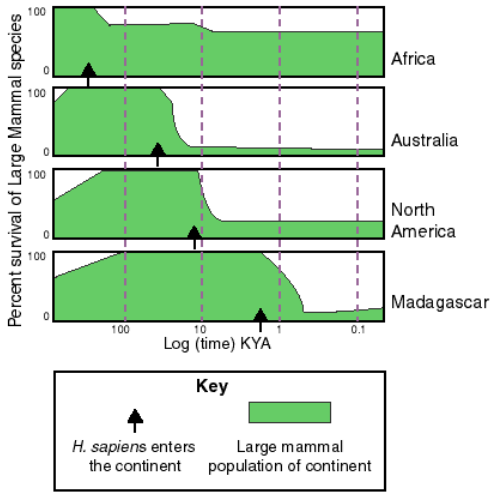


Figure 2.8.7: Graph showing abundance of large mammals and the introduction of humans.

Humans have had an influence on the Earth, its ecosystems and climate. Yet, human activity can not explain all of the changes that have occurred in the recent past. The start of the Quaternary period, the last and current period of the Cenozoic, is marked by the start of our current ice age 2.58 million years ago. During this time period, ice sheets advanced and retreated, most likely due to Milankovitch cycles (see An Introduction to Geology: Chapter 15). Also at this time, various cold-adapted megafauna emerged (like giant sloths, saber-tooth cats, and woolly mammoths), and most of them went extinct as the Earth warmed from the most recent glacial maximum. A long-standing debate is over the cause of these and other extinctions. Is climate warming to blame, or were they caused by humans? Certainly, we know of recent human extinctions of animals like the dodo or passenger pigeon. Can we connect modern extinctions to extinctions in the recent past? If so, there are several ideas as to how this happened. Possibly the most widely accepted and oldest is the hunting/overkill hypothesis. The idea behind this hypothesis is that humans hunted large herbivores for food, then carnivores could not find food, and human arrival times in locations has been shown to be tied to increased extinction rates in many cases.

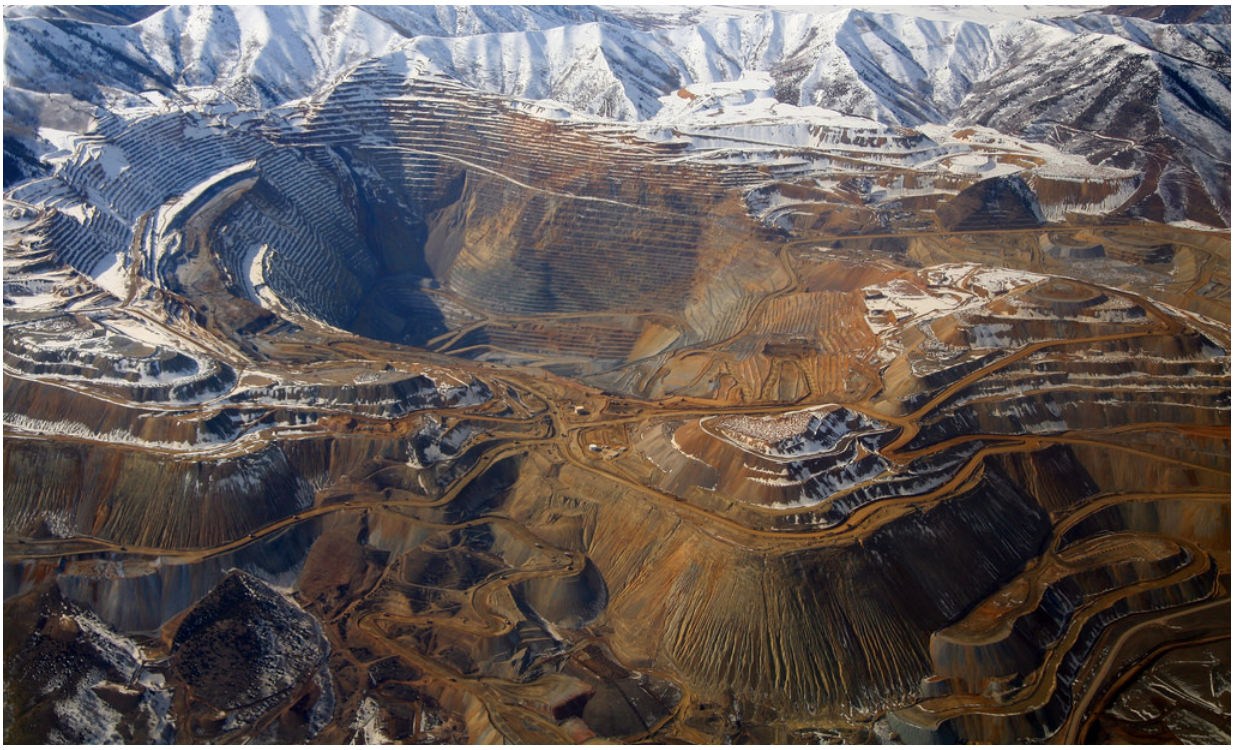


Figure 2.8.8: Bingham Canyon Mine, Utah. This open pit mine is the largest man-made removal of rock in the world.

Modern human impact on the environment and the Earth as a whole is unquestioned. In fact, many scientists are starting to suggest that the rise of human civilization ended and/or replaced the Holocene epoch and defines a new geologic time interval: the **Anthropocene**. Evidence for this change includes extinctions, increased tritium (hydrogen with two neutrons) due to nuclear testing, rising pollutants like carbon dioxide, more than 200 never-before seen mineral species that have occurred only in this epoch, materials such as plastic and metals which will be long lasting "fossils" in the geologic record, and large amounts of earthen material moved. The biggest scientific debate with this topic is the starting point. Some say that humans' invention of agriculture would be recognized in geologic strata and that should be the starting point, around 12,000 years ago. Others link the start of the industrial revolution and the subsequent addition of vast amounts of carbon dioxide in the atmosphere. 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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2.9: References

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2.S: Summary

The changes that have occurred since the inception of Earth are vast and significant. From the oxygenation of the atmosphere, the progression of life forms, the assembly and deconstruction of several supercontinents, to the extinction of more life forms than exist today, having a general understanding of these changes can put present change into a more rounded perspective.

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CHAPTER OVERVIEW

3: Earth as a System

Earth Systems Through Time

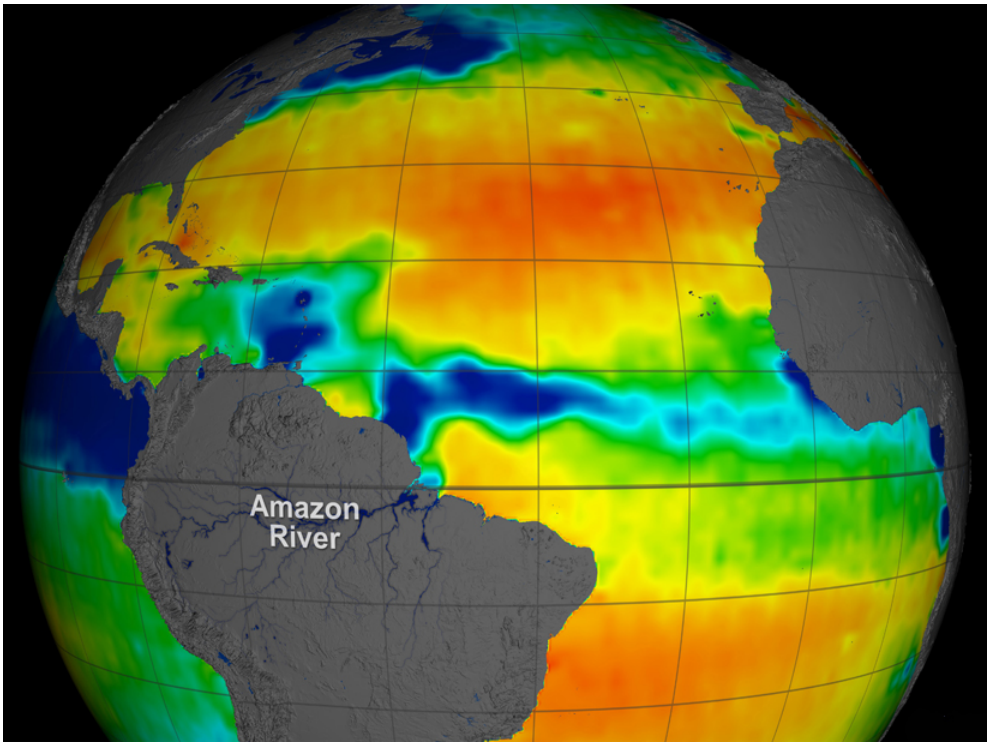


Figure 3.1: Salinity data from NASA's AQUARIUS instrument, dated 27 February 2013. Salinity data provides important insights into the planet's water cycle. (CC-BY 2.0; NASA)

Key Concepts

After reading this chapter, students should be able to:

- Identify the main "spheres" of the Earth.
- Discuss how the Earth's spheres interact through flows of energy and materials.
- Identify ways that changes in one sphere can effect changes in other spheres and even lead to changes in dynamic equilibrium of the entire Earth system.
- Distinguish between amplifying and balancing feedbacks and provide an example or two of each.
- Discuss how remote sensing data is important in understand the complexity of the Earth system.
- Understand the importance of time scales for Earth systems research.

> "To everything, there is a season." – Ecclesiastes 3:1 and The Byrds

3.1: Introduction
3.2: Earth Systems Overview
3.3: The Symphony of the Spheres
3.4: Earth's Energy Budget
3.5: Earth Systems collide- Events and Interactions

3.6: Scales of Measurement- Geography
3.7: Scales of Measurement - Time
3.8: Ground-based instrumentation and satellite data
3.9: Further Reading

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3.1: Introduction

Throughout its entire 4.5 Ga history, every event in the history of our planet happened because something forced it to happen. An asteroid once caused an extinction. Extinctions led to novel new evolutionary pathways for origination. 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 really that simple?

Contained within our galactic system is a solar system powered by an average G type star. “On a mote of dust suspended in a sunbeam” (Carl Sagan) resides our Earth system. Like any system, it is composed of many moving parts, powered by flows of energy and sustained in a state of dynamic equilibrium. Simply put, a system is a naturally occurring group of interacting, interrelated, or interdependent elements that form a complex whole.



Figure 3.1.1: Earth systems science seeks to understand the interconnectedness of our planet’s air, water, land, and life (M. Ruzek 1999).

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 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 what are called biogeochemical cycles. Examples of these include the carbon and nitrogen cycles.

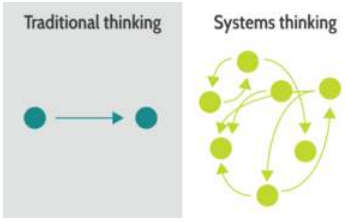


Figure 3.1.2: Systems thinking transcends linear approaches to problems and ways of understanding how things work (Image: Kindling.xyz)

Energy does not flow through these systems in a linear fashion. Its pathways are complex, taking advantage of the points where systems interact to transfer from one system to another and back again. In order to really gain a grasp of how our planet works today and how it has worked in the past (uniformitarianism), we need to use some of the principles of systems thinking. If we think in terms of systems, we begin to think in terms of cycles, feedbacks, forcing mechanisms, storage sinks, and flows of energy and material.

Remember the extinction caused by an asteroid mentioned above? While the asteroid (from the exosphere) did indeed cause the extinction, it was the effect that asteroid had on the hydrosphere, atmosphere, and lithosphere that led to the destruction that occurred in the biosphere. In systems thinking, the asteroid was a forcing mechanism that put the entire Earth system into a state of disequilibrium, past a tipping point of no return. While certainly some organisms were directly crushed by the impending rock from space, most were killed by the downstream effects as its energy rippled across the interconnected web of systems.

“To everything, there is a season”.

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3.2: Earth Systems Overview



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.

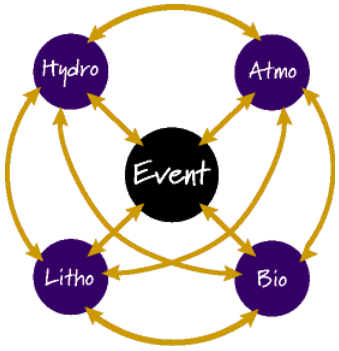


Figure 3.2.1: Earth system interactions due to an "Event". Events can be a wide variety of things that trigger interactions, or forcings, within the Earth system. (Wheeling Jesuit University/NASA Classrooms of the Future; reproduced with permission.)

Events have an effect on 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 in the immediate vicinity of 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 to one another. 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 are no shortage of such events!

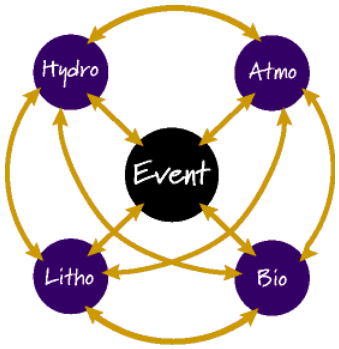
Earth System Forcing Events

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 leading 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 - 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 - surge of carbon and sulfur dioxide in atmosphere leads to lower pH in marine and freshwater environments.	Direct - Major influxes of volcanic gases and ash for extended periods of time.	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 large area; high-pressure metamorphism in zone of impact; Seismic energy release	Direct - Vaporization of large amounts of surface and groundwater from a local to global level, depending upon size and composition of 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 global extinction event
Earthquake	Direct - Local to regional changes in landscape near 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	Direct - Changes in regional evaporation	Direct - addition of carbon dioxide warms	Direct - Climate warming leads to

Forcing Event	Lithosphere/Geosphere	Hydrosphere	Atmosphere	Biosphere
	matter and black carbon to sedimentary deposits; Climate change leads to other downstream effects	rates due to warming; Changes in hydrologic cycle	the troposphere; Changes in weather patterns	biodiversity loss, extinction, etc.
Viral Pandemic Among Human Population	Indirect - Social isolation and pause in economic activity leads 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 human species

This table lists a few examples of forcing events in the Earth system and some of their direct influences on a given sphere or more. Some forcing events may not have a direct influence on every sphere, but only an indirect secondary effect.

Earth Systems Interactions - Quiz



Exercise 3.2.1

A volcanic eruption emits sulfur dioxide gas.



a. Between event and hydrosphere

b. Between event and atmosphere

c. Between event and biosphere

d. Between event and lithosphere/geosphere

Answer

b. Between event and atmosphere

Exercise 3.2.2

A volcanic eruption creates pyroclastic flows of super-heated gas and ash that rush downslope.



a. Between event and atmosphere

b. Between event and hydrosphere

c. Between event and lithosphere/geosphere

d. Between event and biosphere

Answer

d. Between event and biosphere

Exercise 3.2.3

When volcanoes erupt, rapid melting of snow and ice can lead to lahars, or mudflows.



a. Event to Atmosphere

b. Event to Biosphere

c. Event to Hydrosphere

d. Event to Lithosphere/Geosphere

Answer

c. Event to Hydrosphere

Exercise 3.2.4

Pyroclastic flows from a volcanic eruption event, in this case Mt. St. Helens in 1980, can have a downstream effect connecting the lithosphere/geosphere to...



a. Hydrosphere

b. Atmosphere

c. Biosphere

Answer

c. Biosphere

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3.3: The Symphony of the Spheres

Now that we have this background, it is time to turn our attention to the spheres themselves in more detail. In particular, the sections below will not only describe the four main spheres that have been mentioned so far, but will also add in three additional spheres. These are the:

- Exosphere
- Atmosphere
- Hydrosphere
- Cryosphere
- Lithosphere
- Biosphere
- Anthroposphere

Exosphere – Space Environment

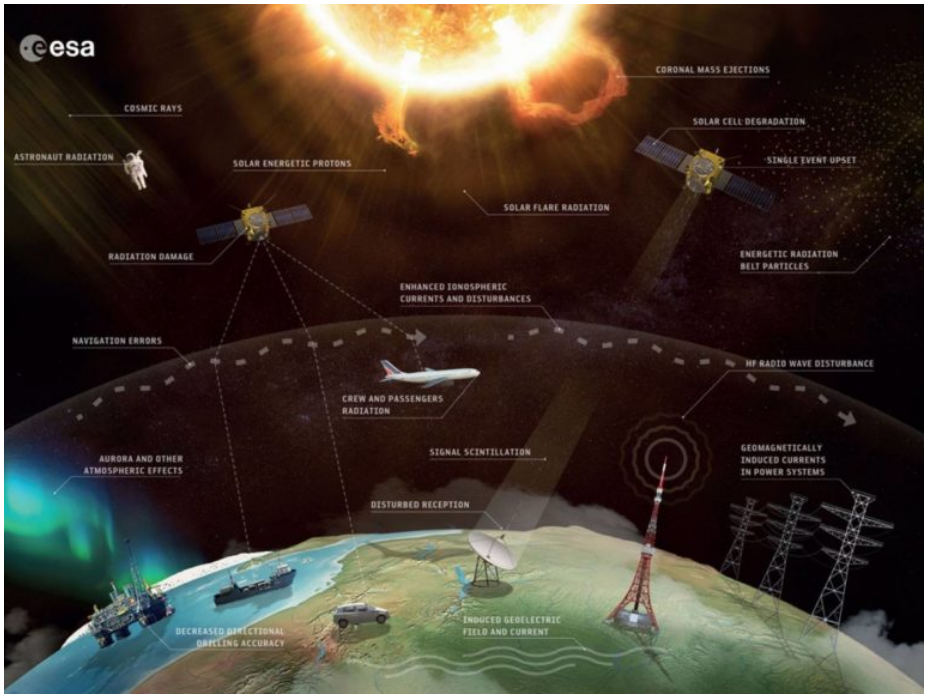


Figure 3.3.1: The exosphere, or the space environment in which our planet resides, has many effects on our planet. (ESA)

The exosphere is the space environment. Thought of from the perspective of the Earth, it is the location in the solar system, around our particular star, within the Milky Way galaxy, and so forth. All of the energy that powers the other systems on our planet, with the exception of the geosphere, comes from the Sun – outside our planet. This solar radiation, ranging from gamma radiation through long-wave radio waves, is critical for the normal operation of the biosphere, atmosphere, and hydrosphere. The exosphere is also the source of dangerous radiation in the form of galactic cosmic rays and solar particle events. Ultimately, the exosphere is a place hostile to life, but paradoxically critical to it also.

Atmosphere – Gaseous Earth (Troposphere, Greenhouse Gases, etc)

The atmosphere is the gaseous envelope that surrounds our planet. Our current atmosphere can be thought of as the third atmosphere our planet has had. The primordial atmosphere of the Hadean and outgassing processes eventually gave way to one dominated by volcanic gases and rich in CO_2. Once photosynthesis evolved around 3.8 Ba, the atmosphere would increasingly contain oxygen and nitrogen. Since about 600 Ma, the concentration of oxygen in the lower atmosphere (troposphere) has been pretty consistent with what we have today.

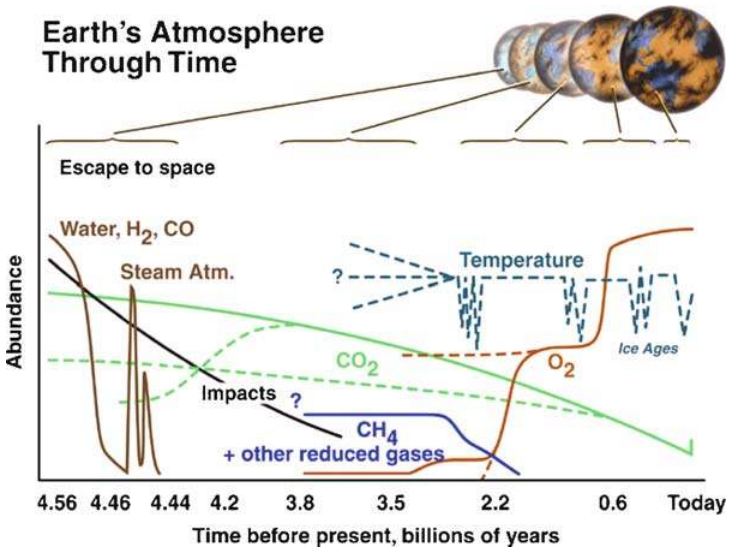


Figure 3.3.2: Evolutionary history of the atmosphere over Earth history. (NASA)

Our atmosphere is layered, with the densest portion at the bottom. This layer, the troposphere, is where most of the action occurs. It is where all human activity and weather occur. It is really the upper boundary of mountain growth at orogenic belts. It is affected on a daily basis by the uneven heating of the land and sea and the rotation of the Earth, creating what is called the planetary boundary layer. Above the troposphere is the stratosphere. In the stratosphere, exospheric UV radiation is absorbed by photochemically produced ozone. Ozone at this level of the atmosphere is very important for protecting life. Human activity, through the release of chlorofluorocarbon compounds via hairspray, refrigerants, and more had carved a hole in this critical layer. During the late 1980s, the world came together to ratify the Montreal Protocol. This banned these classes of chemicals and prevented the concentration of this important gas from deteriorating to an even more dangerous state.

Above the stratosphere lies the mesosphere (middle atmosphere) and thermosphere. As you ascend upward into the mesosphere, temperatures and pressure decrease. At the thermopause, the lower boundary of the thermosphere, pressure continues to drop but temperatures suddenly begin to rise sharply.

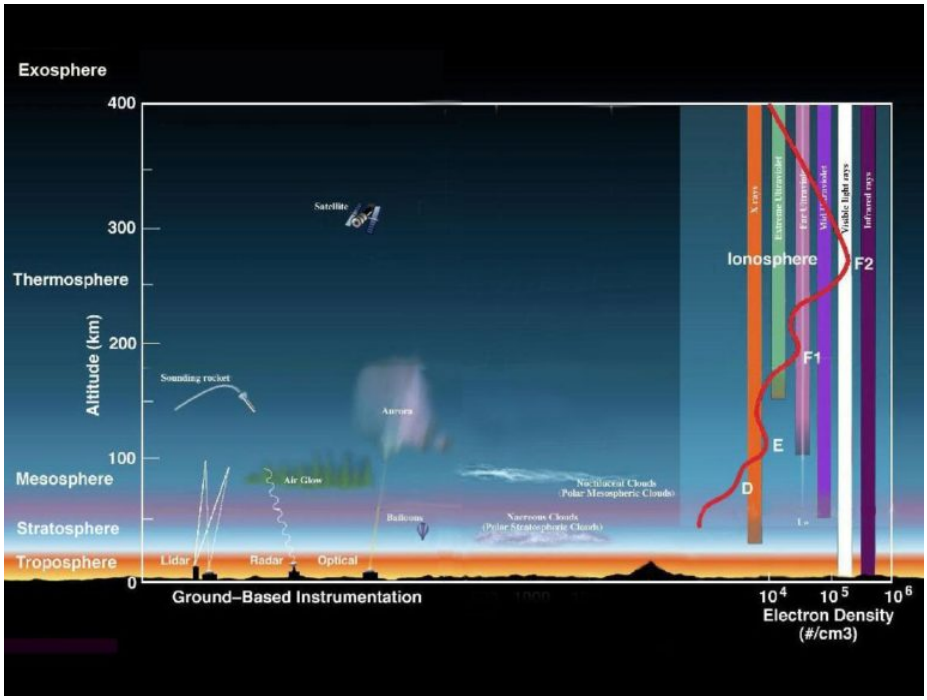
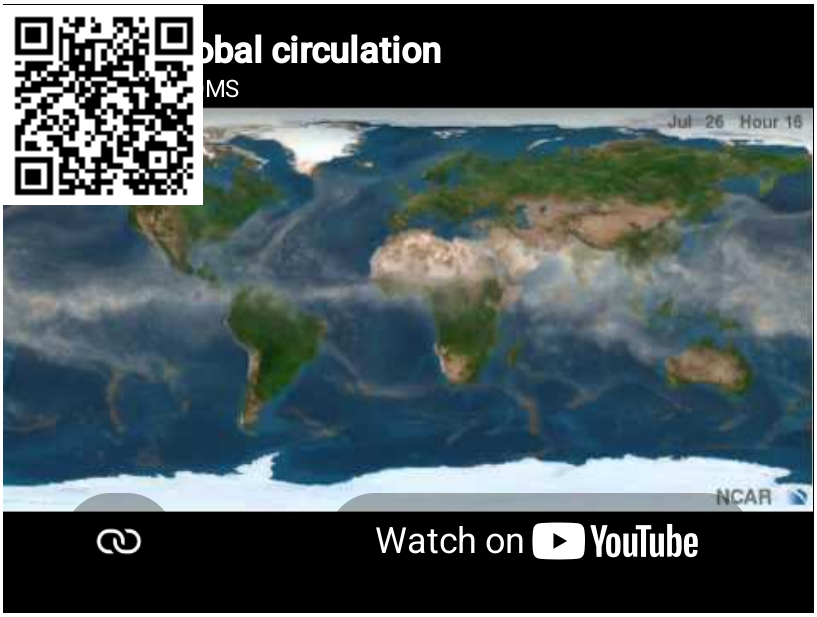


Figure 3.3.3: Layers of the Atmosphere (NASA)

The atmosphere plays a very important role in the Earth's climate. The most important of these is the concentration of gases in the troposphere that trap heat and prevent it from escaping to the exosphere. Solar radiation, incoming as shortwave infrared energy, will warm the surface (land and water) and then be re-radiated as longwave radiation. This gets trapped by greenhouse gases like water vapor, CO_2, CH_4, and NO_x. All of these are gases that are increasing in concentration due to human activity. This heat gets trapped in the atmosphere, which has a much lower specific heat capacity than water. Generally, this means that the hydrosphere ends up absorbing this heat in all of its uneven nature around the planet, leading to increases in evaporation in some places and precipitation in others. It also generally warms the oceans and land surfaces, leading to higher ocean and land temperatures.



The uneven heating of the atmosphere is very important for normal weather patterns, however. Rising warm air at the equator and descending cool air at the poles create the foundation for convection cells that circulate air between the equator and poles. The rotation of the Earth produces the Coriolis Effect, which deflects the movement of air in these cells in an easterly direction at the poles (Polar Cell), westerly direction in the mid-latitudes (Ferrell Cell), and again in an easterly direction in the tropics (Hadley Cell).

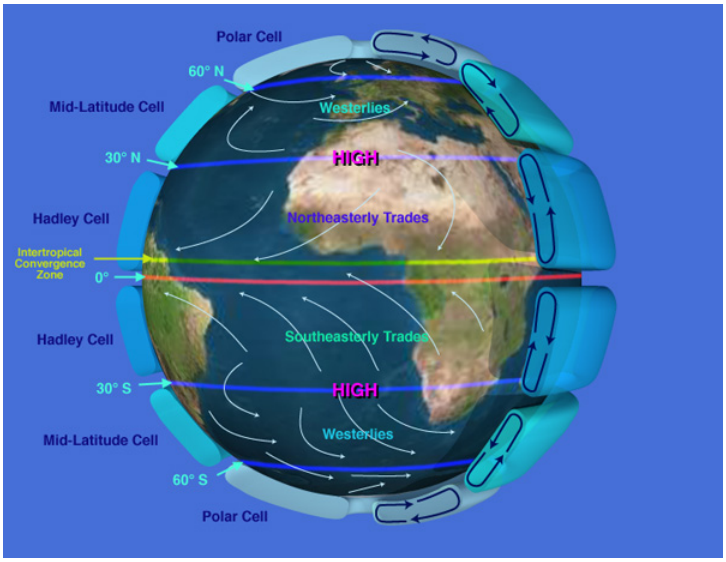


Figure 3.3.4: Atmospheric circulation patterns at equinox position (Spring or Autumn). Note the positions of Hadley Cells through Polar Cells by latitude and their deflection due to the Coriolis Effect. (NASA)

These latitudinal variations not only drive our weather, but they also lead to important gradients in the biosphere, called the latitudinal biodiversity gradients (LDB). Generally, biodiversity decreases from the tropics to the poles. This change in biodiversity

is directly related to temperature changes with latitude...today. It is important to note here that today's LDB is not the norm for Earth's past. In fact, during the Eocene epoch, it is very likely that the temperate regions of the planet were the most diverse. This is because it was so much hotter in the tropics.

Hydrosphere – Liquid Earth's Water (Fresh, Marine, Ground, etc.)

The hydrosphere makes up the liquid envelope of our planet. It consists not only of the oceans, but also all freshwater contained in lakes, rivers, and streams. While it is considered separately here, the cryosphere is a critical element of this sphere also, as it provides a significant storage sink for water.

Freshwater and marine systems have many similarities and differences. However, the key feature of all liquid water in the hydrosphere is its ability to regulate heat in various ways.

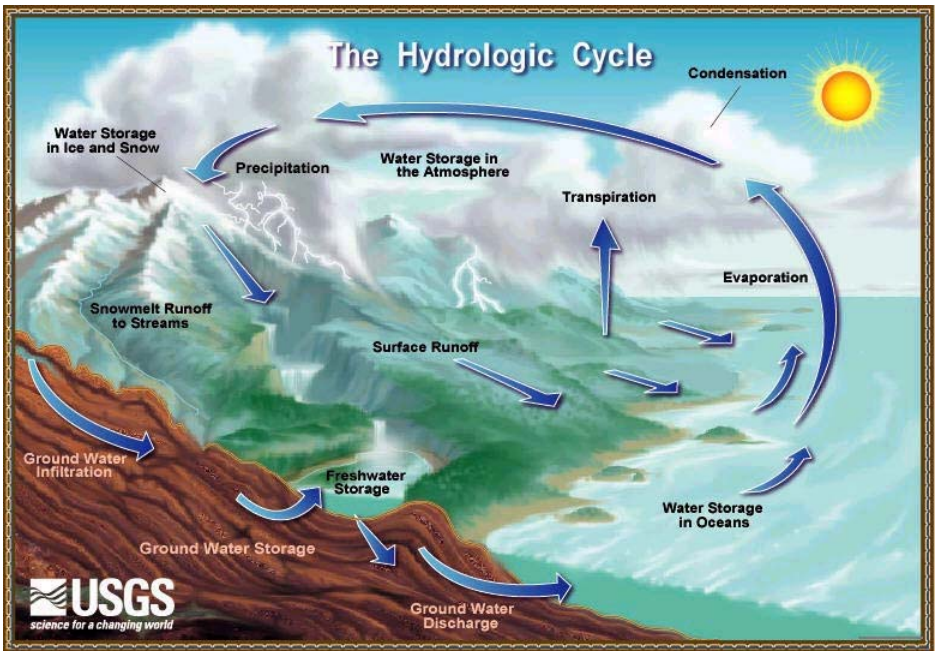


Figure 3.3.5: The hydrologic cycle and hydrosphere. (USGS graphic)

Water has a high heat capacity (specific heat). That is, it will absorb a great deal of heat energy before it gets hot itself. Because of this, water is a critical regulator of energy on the planet. 1 kg of water has to absorb 4,184 J of energy for its temperature to rise 1 ° C. By comparison with rock, in this case a special native mineral called copper, only 385 J are needed to accomplish the same task. Not only does this property make fish very happy, but it also means that excess heat from the atmosphere, the Sun, or the land during the day can be absorbed by water with very little effect. Ultimately, while the hydrosphere is driven by the exchange of heat itself (the hydrologic cycle), it also regulated the planet in the same way sweat helps regulate your body temperature during a workout.

Another critical heat regulating feature of the hydrosphere are ocean currents. Whether the continents are combined into a single supercontinent or split into what we have today, warm surface waters will move and exchange with cool nutrient-rich bottom waters. This exchange, or convection, produces currents. These rivers in the ocean move heat around the planet. Temperature is not the only physical property in play when it comes to marine currents, as the salinity of marine water can vary enough to produce density-driven currents. The combined thermohaline circulation patterns in the ocean regulate our climate.



Cryosphere – Frozen portion of the hydrosphere

The cryosphere is the solid portion of the hydrosphere. It is worth discussing separately from the hydrosphere because of the particular importance of ice in exchanges of heat and light between Earth's systems. Like water, ice has a heat regulation effect.



Figure 3.3.6: A view of the Antarctic cryosphere (NASA)

The presence of polar ice caps provides a very important counterweight to the extreme heat experienced in the mid-latitudes. The temperature differences between these regions drive the movement of air in the atmosphere, forming Hadley Cells (warm air rising and cool air descending) that, while deflected by the Earth's rotation and Coriolis Effect, drive our weather. Polar ice is also a critical water storage sink. As it melts, sea level rises and isostatic adjustment of continents occurs.

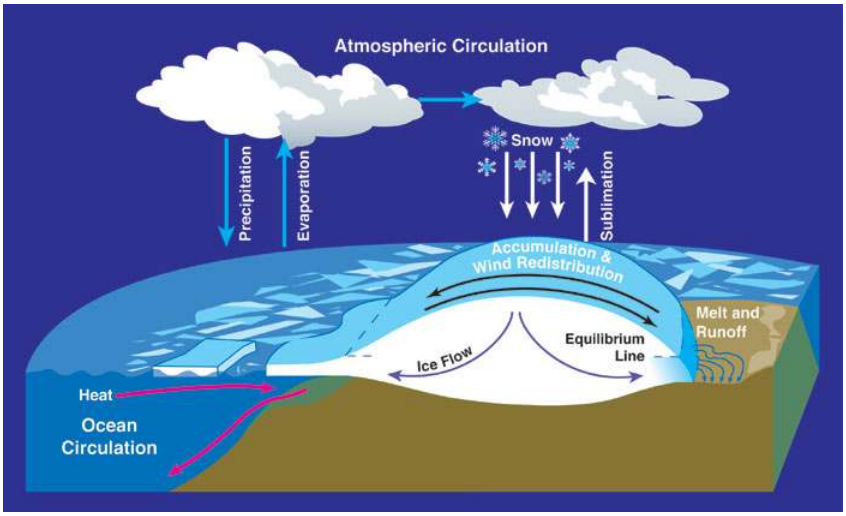


Figure 3.3.7: Earth's Cryosphere (NASA)

The Earth has not always had polar ice. Just 40 Ma ago, there was no ice at the poles because the planet's climate was so warm. With the rising of the Himalayas and Andes came increase silicate weathering. As mountains rise, this form of weathering pulls carbon dioxide out of the atmosphere and deposits the byproduct of that carbon into the oceans as bicarbonate ions. Eventually water began to be stored on land over Antarctica first, as a continental glacier. Later, ice would form over the northern polar regions in the sea.

The polar ice caps are only part of the story of the cryosphere. The ebb and flow of continental glaciers in places like Greenland and valley glaciers all over the world provide us with important visual thermometers for global temperature changes.

As glaciers and sea ice retreat, there are changes in the color distribution on the Earth's surface. Albedo, a measure of this, is highest on ice and lowest on ocean water. When sunlight hits ice, it reflects nearly 100% of the energy. When sunlight hits water, it acts nearly as a blackbody, absorbing nearly all of the incoming energy. As the climate warms, ice recedes, and the climate warms more. We will return to this later in the chapter.

Lithosphere/Geosphere – Solid Earth (Rock, including molten rock)

The geosphere makes up the solid portion of the planet. If you have taken a course in Physical Geology, you know all about the rocks and minerals that make up the various layers of our planet from the inner core outward to the crust. You can read in great detail about the geosphere in the Earth Materials chapters of this book on rocks and minerals.

Unique among the spheres, the geosphere produces its own energy. There are three sources. Radioactive decay of unstable elements within our planet's interior produces a great deal of heat energy. Another heat source comes from remnant primordial heat still yet to dissipate from the planet's formation. Finally, frictional heating plays a role, descending material rubs against other materials. Earth is hottest near the center and is cooler at the surface. This collective heat leads to the geothermal gradient, the increase in temperature as you descend further into the Earth. It also drives plate tectonics and resulting volcanism, earthquakes, and natural hazards.

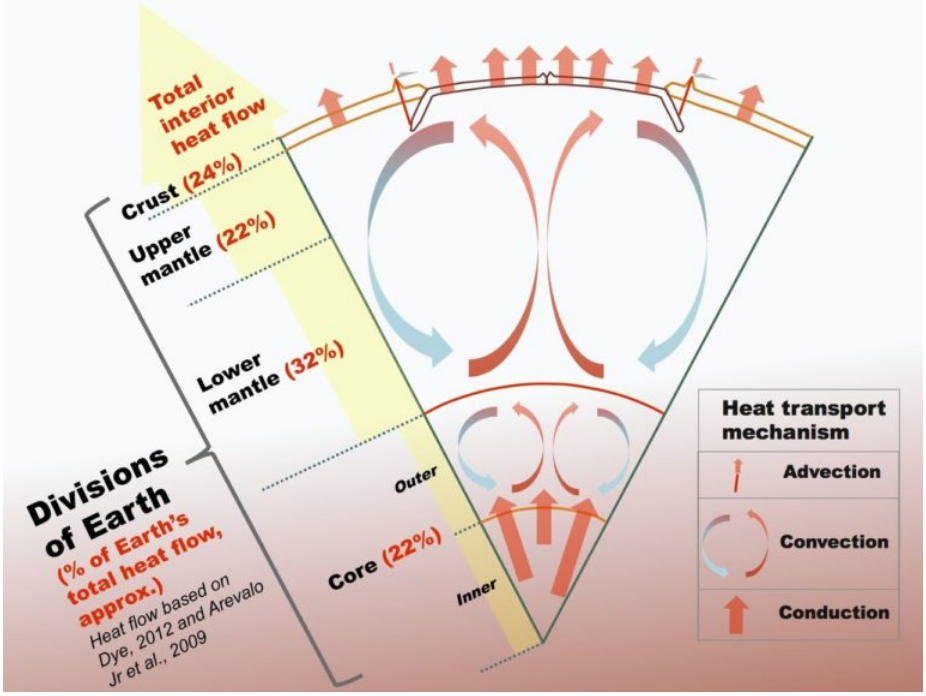


Figure 3.3.8: Cross-section of the Earth showing its main divisions and their approximate contributions to Earth's total internal heat flow to the surface, and the dominant heat transport mechanisms within the Earth (Wikimedia, Bkilli[1])

Because of this independent heat energy, the solid Earth is much more dynamic than any other terrestrial planet we know of. Mercury, Venus, and Mars are very different places with differing amounts of interior activity. Some natural satellites in our solar system likely also produce their own interior heat. Examples may include Saturn's natural satellite Titan. Earth, because of plate tectonics, is a much more active planet and one where the rocks are constantly interacting with and having an effect on the liquid and gaseous portions of the planet.

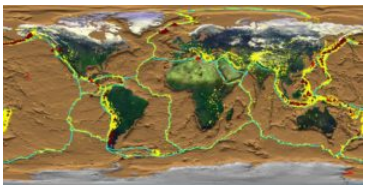


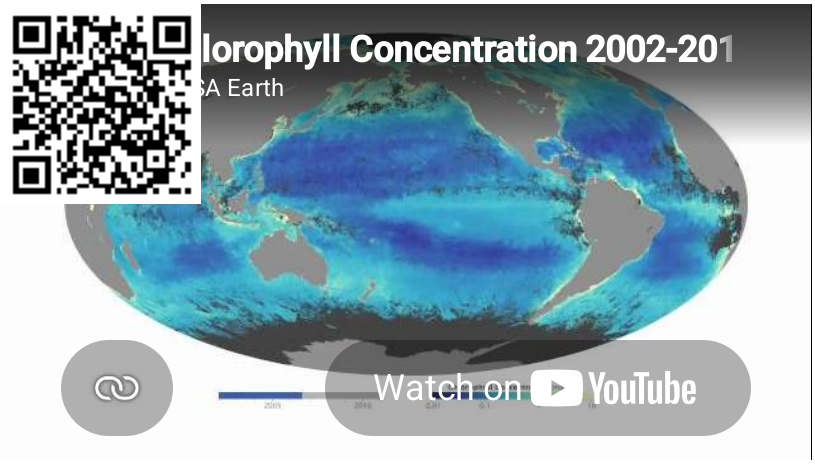
Figure 3.3.9: Crustal dynamics in the Earth's geosphere. Red triangles indicate volcanoes and yellow dots indicate earthquake epicenters. Blue lines are plate boundaries. (NASA SVS)

It is hard to pin down the most important geosphere elements for driving climatic change on the Earth. Volcanism is very likely most important, due to the input of gases like SO_2 into the atmosphere, which can have a cooling effect due to its ability to reflect solar radiation back to space. There is also airborne dust, sometimes referred to as loess and measured as a component of particular matter. This airborne dust has the ability to absorb heat and heat the atmosphere, causing warming in places where dust is moved from elsewhere.

Biosphere – Life

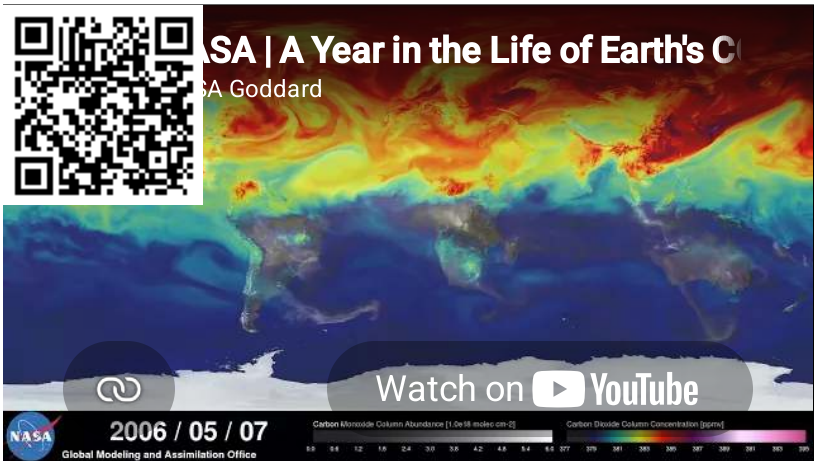
Without a tectonically active geosphere, an atmosphere with the right concentration of gases, and a hydrosphere made up so heavily of liquid water, there would likely be little to no biosphere on Earth. Unique among planets in our solar system and perhaps anywhere, Earth contains an amazing array of life. The origins of life on Earth are still somewhat elusive, but we do know that over time it has evolved to live in balance with the other spheres to the point where because the biosphere is such a critical influence on the other spheres, it can be considered on its own terms. Being located in the habitable zone (Goldilocks Zone) of a G-type star has its perks. Is Earth unique in this? A number of exoplanets have been identified that are located in the habitable zones around their stars, but as yet, no extraterrestrial life has been confirmed.

The video below provides a sense of life on Earth. You are viewing the ebb and flow of life forms in the oceans through the lens of chlorophyll from space, with data collected by the MODIS sensor. This data was collected by NASA between 2002 and 2010 and displays visually the intricate link between photosynthetic life and seasonality.



The biosphere plays a very important role in the climate, particularly when it comes to the regulation of greenhouse gases. Of course, the decomposition of organic matter is a major source of CH_4 and NO_x gases, aside from what humans contribute. But, the real effect is the planet-wide exchange of CO_2 and O_2 that is so critical to regulating temperature and providing breathable oxygen.

Check out the supercomputer simulation of CO_2, offered by NASA Goddard below. The red concentrations mostly represent CO_2, but you will see as the animation and narration progress that, during the southern hemisphere summer, CO concentrations from fires in Africa and South American increase significantly. As the animation progresses, note the role that the biosphere plays in the annual fluctuations of CO_2 in various places around the planet, but particularly in the Amazon and Congo basins.



3.8 Ba ago, the biosphere began to alter the Earth's atmosphere for its own purposes. The evolution of photosynthesis and the subsequent Great Oxygenation Event have left a long legacy that lives on today, in the form of algae and other plants that provide the base of the trophic levels so critical to our existence. Today, the biosphere is once again altering the Earth's atmosphere, but through the actions of a single species.

Anthroposphere – Human portion of the biosphere



Figure 3.3.10: Earth at night (NASA)

Humans are a force of nature. No single species in the history of life can claim such a mantle. Yet, humans are a part of the biosphere. Because of our outsized impact, it is useful and perhaps even appropriate to consider the human element of the planet independently from the other spheres. Thus, the anthroposphere. The human impact on the planet goes well beyond altering the atmosphere and climate, leading some scientists to consider whether the Earth itself has indeed entered a new geological epoch. As of this writing, the Anthropocene Working Group, a subcommittee of the Quaternary Working Group of the International Commission on Stratigraphy has officially voted to recommend several things. First, that the Anthropocene should be treated as a chronostratigraphic unit for the geologic time scale (Series/Epoch level). Second, that the mid-20th century should be the time marker for its start. If ratified by the International Union of the Geosciences, the Holocene Epoch will have ended around 1950 and we will be living, officially, in the Anthropocene.

Did I Get It? - Quiz

Exercise 3.3.1

Which of the Earth's spheres encompasses frozen water?

a. Hydrosphere

b. Cryosphere

c. Lithosphere

Answer

b. Cryosphere

Exercise 3.3.2

Which of the Earth's spheres do you see interacting here?



a. Biosphere, Atmosphere, Hydrosphere

b. Biosphere, Anthroposphere, Lithosphere

c. Biosphere, Lithosphere, Atmosphere

Answer

c. Biosphere, Lithosphere, Atmosphere

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3.4: Earth's Energy Budget

Energy from the Exosphere

The Sun powers our solar system. In addition to the Sun, some planets have a significant source of internal heat. Venus is one such planet, releasing its heat through volcanism. Saturn produces more internal heat than it receives from the Sun, given its great distance from the star it orbits. Earth is also one of these planets. In Earth's case, the radiation received from the Sun is the primary driver for the entire Earth system, including its climate.

When considering the energy that arrives from the Sun, it is worth studying the image below from NASA. Radiation from the Sun reaches the Earth system in a wide range of wavelengths spanning the entire electromagnetic spectrum. Much of this energy is blocked by the Earth's atmosphere, allowing only radio waves, visible light, short-wave infrared radiation, and some amount of UV radiation to penetrate to the surface of the planet. This selective admission is important for the functioning of life on Earth as we see it today. But, it has not always been this way. The Earth's atmosphere has evolved along with life, affecting life's evolution and also being changed by life itself. The types of radiation that are admitted to the surface have also changed over time.

Energy from the Sun flows through the Earth system in its many forms and powers the hydrologic and the biogeochemical cycles that all make life possible on our planet. Of the radiation that enters the atmosphere, the shortwave (UV/Visible/infrared) radiation is most important. UV radiation gives you "sunburn" and is carcinogenic over time. Your eyes are evolved to detect the violet to red wavelengths of visible light. Infrared radiation is what you experience as heat. Some of this is reflected by clouds, a notoriously difficult factor to model, given their variability. The rest of it hits the Earth's surface, being absorbed in places like land and ocean (low albedo/dark in color) and reflected in other places like ice and snowpack (high albedo/light in color).

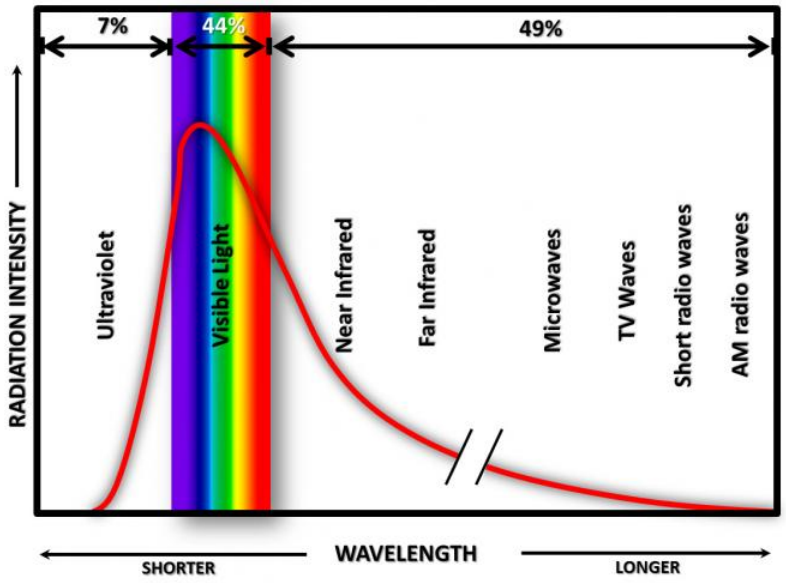


Figure 3.4.1: Full spectrum of electromagnetic energy from the Sun. Most incident radiation from the Sun ranges from visible light through radio waves. Near infrared is shortwave radiation and far infrared is longwave radiation. Visible light drives most biological processes on the Earth. (NASA)

After the energy is absorbed, it can be re-radiated in the form of longwave infrared radiation. This is still in the form if heat, but has lower energy (longer wavelength). As this radiation attempts to leave the atmosphere, much of it is absorbed by greenhouse gases, such as water vapor, carbon dioxide, methane, and nitrogen dioxide. This trapping of heat by gases is what is referred to as the greenhouse effect. And, this effect is critical for life on Earth.

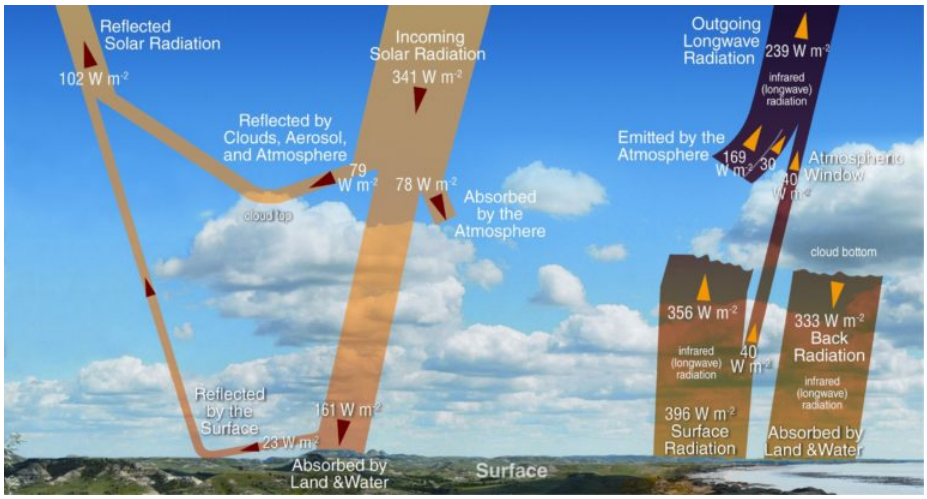


Figure 3.4.2: Earth's energy budget (NASA)

Energy from Earth's Interior

Unlike Mercury and Mars, who have a negative energy balance as they lose more heat over time than they create, Earth's interior experiences a balance. Since the formation of all of the planets, they have been cooling over time. Whether a planet is able to balance this heat loss with its own heat production is a function of the composition of its interior. In Earth's case (and for all terrestrial planets to varying degrees), it is producing much of its own interior heat through radioactive decay. The heat from this decay is caused by particles, emitted during decay of unstable isotopes, bouncing off of other particles. These collisions transform this kinetic energy into heat.

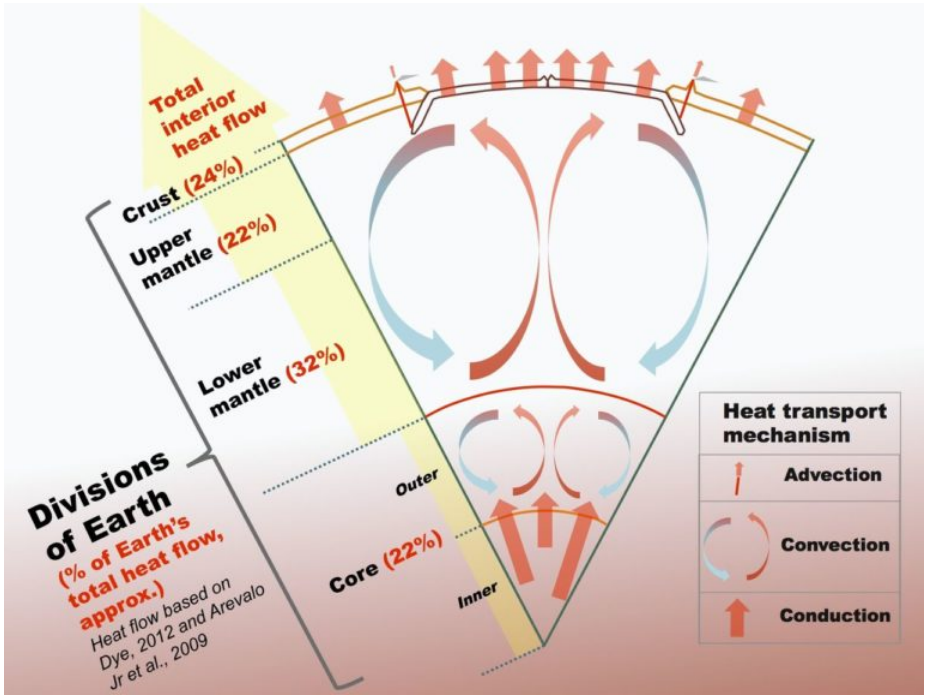


Figure 3.4.3: Cross-section of the Earth showing its main divisions and their approximate contributions to Earth's total internal heat flow to the surface, and the dominant heat transport mechanisms within the Earth (Wikimedia, Bkilli[1])

This heat is transported outward by conduction, convection, and advection, as seen in the figure below. Plate tectonics is driven by this internal heat, in addition to gravity.

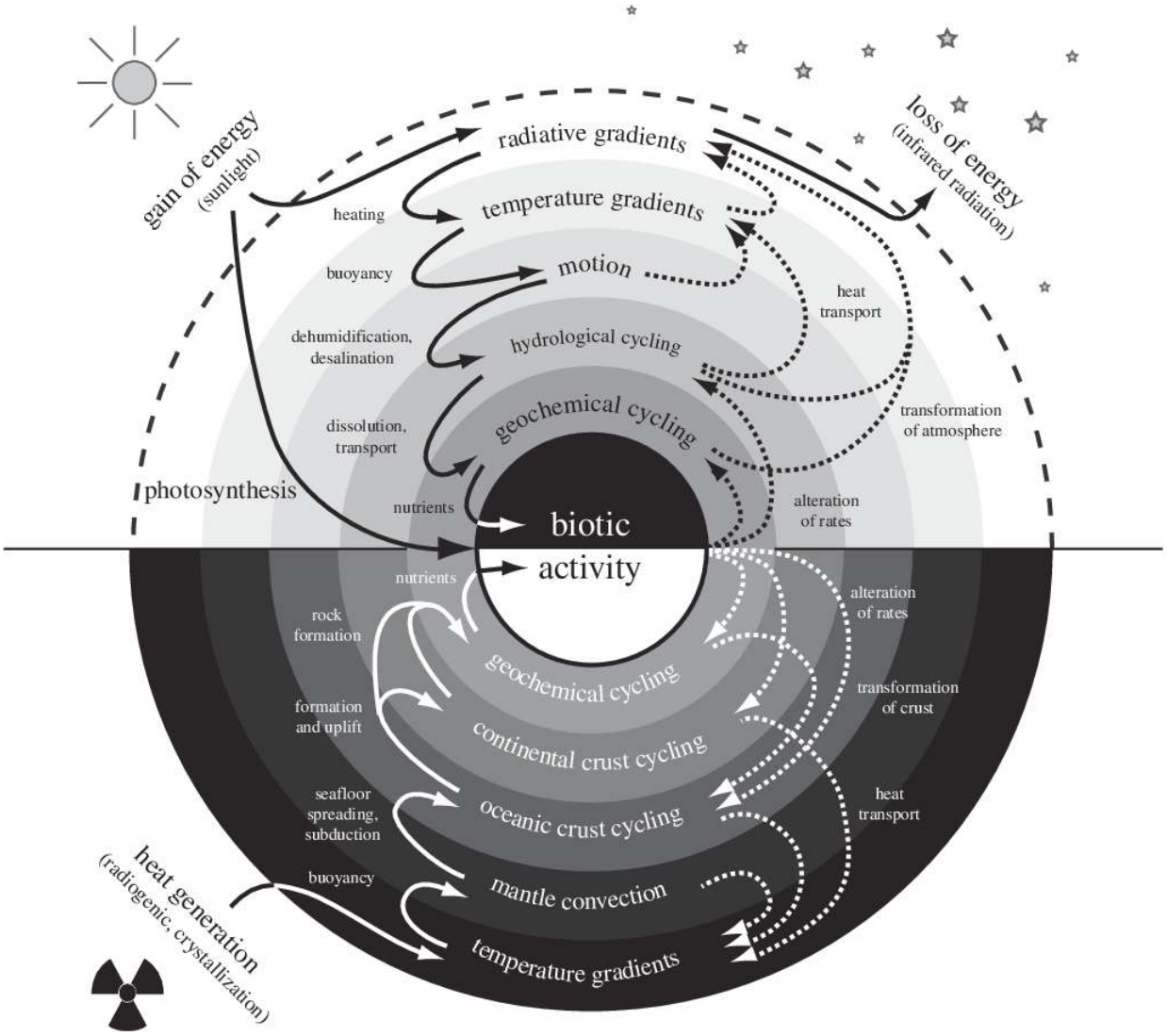


Figure 3.4.4: The Earth system's two main sources of energy and some of the transfers and flows of that energy, all of which contribute to biotic activity. Solid lines are modes of energy transfer and dissipation. Dotted lines are processes, or effects, created by these flows of energy. (Kleidon, (2012))

Through plate tectonics, this internal energy teams up with solar energy to help drive the biogeochemical cycling that provides important nutrients that sustain life, including carbon, hydrogen, nitrogen, oxygen, phosphorus, and sulfur. Because of this, energy from the Sun is only part of the story of our planet's geologic past. The combination of these sources of energy is critical for the function of Earth's systems. You can see examples of some of these important interactions in the figure below.

Geological Sources of Important Elements for Life

Table 3.4.1: Table of important elements for life. All of these elements are ultimately sourced from within the Earth's interior. Plate tectonic processes, driven by Earth's interior heat, drive one end of the biogeochemical cycling of these nutrients while the Sun's incoming energy drives them from above.

Elemental Nutrient	Primordial Source on Earth	Important Life Processes
Carbon	Cosmogenetic, Mantle, Volcanism	Basic element of organic chemistry, important nutrient for photosynthesis of glucose
Hydrogen	Cosmogenetic, Mantle, Volcanism	Basic element of organic chemistry, important nutrient for photosynthesis of glucose
Nitrogen	Cosmogenetic ammonia, Earth's atmosphere	Important for the creation of amino acids, enzymes
Oxygen	Cosmogenetic, Volcanism, Carbon Dioxide dissociation, Photodissociation in Atmosphere of Water	By-product of photosynthesis, Necessary for respiration and associated oxidation
Phosphorus	Phosphides in Earth's core, Inorganic minerals such as apatite and fluorite, ocean-floor sediments	"Spine" of DNA molecule, important micronutrient that helps store energy in cells via ATP
Sulfur	Cosmogenetic, Stored in Earth's interior, Volcanism	Allows for the synthesis of a greater variety of amino acids, Important nutrient for chemosynthesis

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3.5: Earth Systems collide- Events and Interactions

Systems exist in a state of dynamic equilibrium. Equilibrium is a balance. When it is dynamic, it means that when the state of equilibrium changes the system(s) adjust. Consider a seesaw with equal weight on both sides. To balance it, you would place the fulcrum in the center. Put two people on either end and then attempt to balance it and you would find that the position of the fulcrum would need to adjust. The system would need to adjust to the new reality.

Every event that occurs within the Earth system upsets a state of equilibrium. In the chapters in this book that deal with evolution, the term for this was “stasis.” When a volcano erupts, the stasis/equilibrium is changed and a new reality emerges for a time. In most cases, the new equilibrium is not much different than the prior situation. In some cases, a tipping point is reached. In such situations, a radically new equilibrium becomes the reality, eventually, and the overall change to the system is extreme. A good example of such a tipping point in the Earth’s past is the Chicxulub asteroid impact that occurred at the end of the Cretaceous Period, mentioned earlier in the chapter.

However, we need to define some systems thinking terms to help us enable our ability to communicate about events and interactions. In systems parlance, events are known as “forcings.” These are things that force changes to the dynamic equilibrium at the time. Anthropogenic greenhouse emissions are a forcing that today is causing the global temperature to warm.

Forcings

The events we have been discussing are referred to as forcings in systems parlance. A forcing event is an action that moves a system away from dynamic equilibrium, usually through an initial push within one system. In terms of climate change, anthropogenic additions of greenhouse gases to the atmosphere are forcing the climate to warm through additional trapping of greenhouse gases. This initial forcing, of course, leads to additional effects downstream.

In the Earth’s history, there have been plenty of examples of these. When photosynthesis evolved and led to the Great Oxygenation Event, the climate cooled which eventually led not only to the rapid radiation of photosynthetic organisms, but also the snowball Earth events of the Precambrian. During the late Devonian extinction, it is thought by some that massive expansion of vascular plants on land led to huge influxes of organic matter into the epeiric seas that existed at the time, leading to eutrophic conditions and mass extinction. In some situations, massive flood basalt events (end-Permian, end-Cretaceous) led to massive emissions of sulfur dioxide and carbon dioxide and subsequent climatic change.

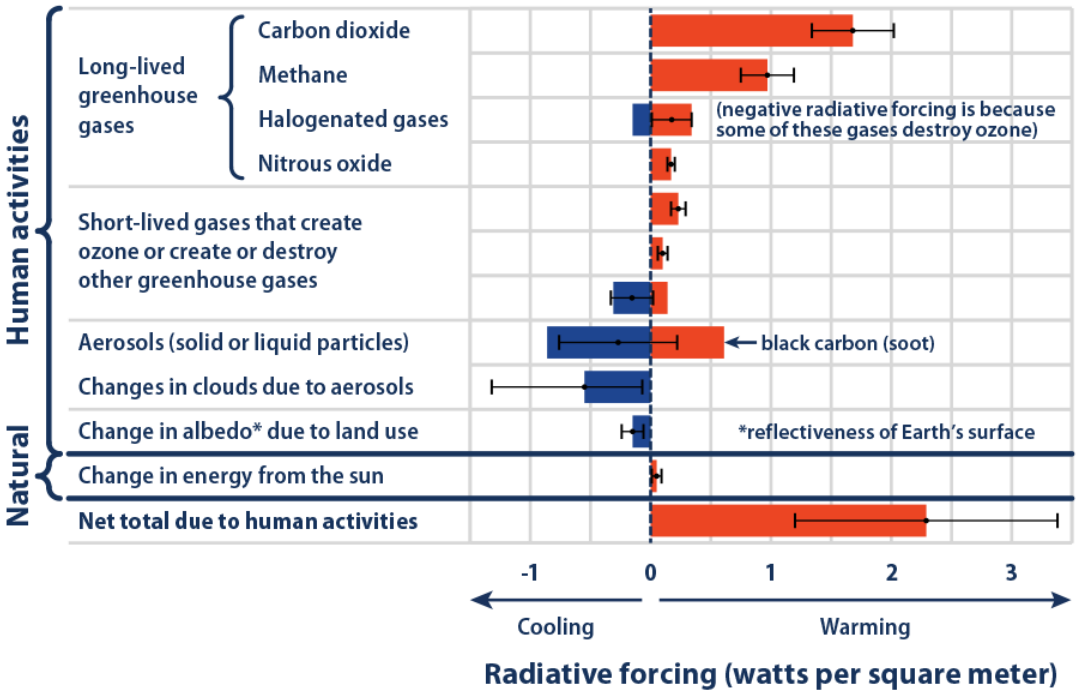


Figure 3.5.1: Natural and anthropogenic climate forcing mechanisms. Any one of these forcing mechanisms can have an amplifying or balancing effect on the Earth's climate, depending upon whether there is an increase or decrease in degree. (EPA graphic)

In our modern environment as in the Earth's past, the climate system is pivotal to life processes. The figure above lists an array of anthropogenic (human) forcing factors that influence the climate system. Changes in any one of these will drive the climate to warm or cool. However, a change in one factor could actually lead to a change in another, independent of any warming or cooling. That is the nature of a system – it is not linear. Interconnections matter. Forcing events never happen in isolation, there are always additional direct and indirect repercussions.

Feedbacks

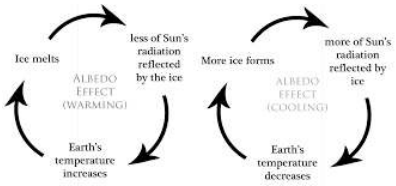


Figure 3.5.2: The ice albedo effect can be a reinforcing feedback for both periods of warming and cooling in the Earth's climate (Used with permission: Climate Interactive)

Feedbacks are the results of forcings. No matter what the forcing event is, there are always feedbacks that occur as a result. Feedbacks can amplify the initial forcing (positive or amplifying feedback). They can also balance that forcing (negative or balancing feedback). A great modern example of an amplifying feedback occurs as a result of the reduction of polar ice due to global warming. Because polar ice is reduced by warming, more dark-colored water is exposed, reducing the albedo that the ice once had in that area of the ocean. Seawater is much darker in color than ice. More energy is then absorbed by the water than was by the ice, further warming the water and melting even more ice. Because polar ice also helps moderate the planet's climate, less ice leads to an amplification of warming.

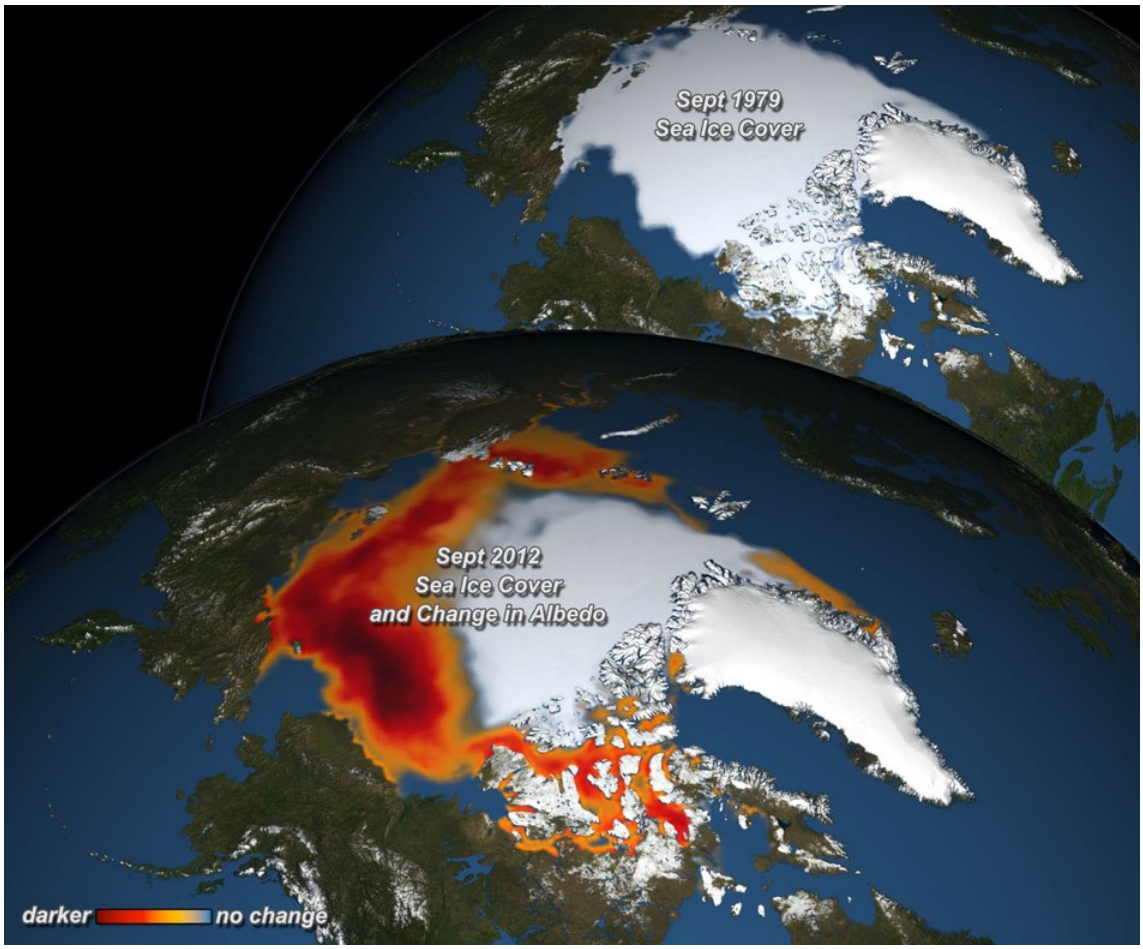


Figure 3.5.3: Decreasing sea ice and its effect on albedo in the Arctic. (NASA graphic)

Amplifying feedback loops tend to move an already off-balance system further away from regaining equilibrium. The image below illustrates the ice-albedo amplifying feedback. Arctic sea ice loss between 1979 and 2012 is shown on the map with areas north of eastern Russia and Alaska shown in false-color to indicate the change in albedo color over time. Dark red areas represent much greater albedo change and, thus, warmer seas. The reduction in white sea ice and snow cover amplifies global warming forced by anthropogenic greenhouse gases.

Balancing (negative) feedback loops provide sustainability in systems. They are the foundations of systems at dynamic equilibrium. In the example to the left, when predator numbers get too high, overkill of prey will lead to a natural decrease in predators over time due to starvation. Likewise, an increase in the birthrate of prey will lead to an increase in the number of predators, until balance is reached once more.

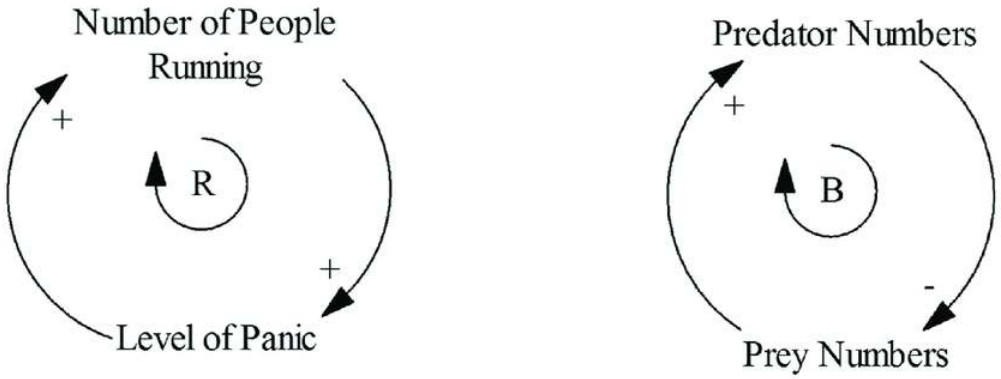


Figure 3.5.4: Examples of amplifying (positive) and balancing (negative) feedback loops. Amplifying feedback loops tend to further force a system that is out of balance further out of balance. Balancing feedback loops represent sustainability, in this case, the predator/prey relationship. (Neely and Walters, 2016)

Eventually, balancing feedbacks will work to bring a system to equilibrium once again. Amplifying feedbacks are not always going to happen, by contrast.

Sinks

Within systems, energy and materials can be stored until it is moved from one part of a system to another. Locations within a system where this occurs are referred to as sinks. Every biogeochemical cycle has sinks. In the carbon cycle, carbon is stored for very long periods of time in the form of limestone or fossil fuels. On the shorter term, it is stored in ocean water, in plants, or in the atmosphere. When a forcing event occurs, carbon may begin flowing from one sink to another. In the case of modern climate change, the carbon is flowing from coal and petroleum products into the atmosphere via combustion.

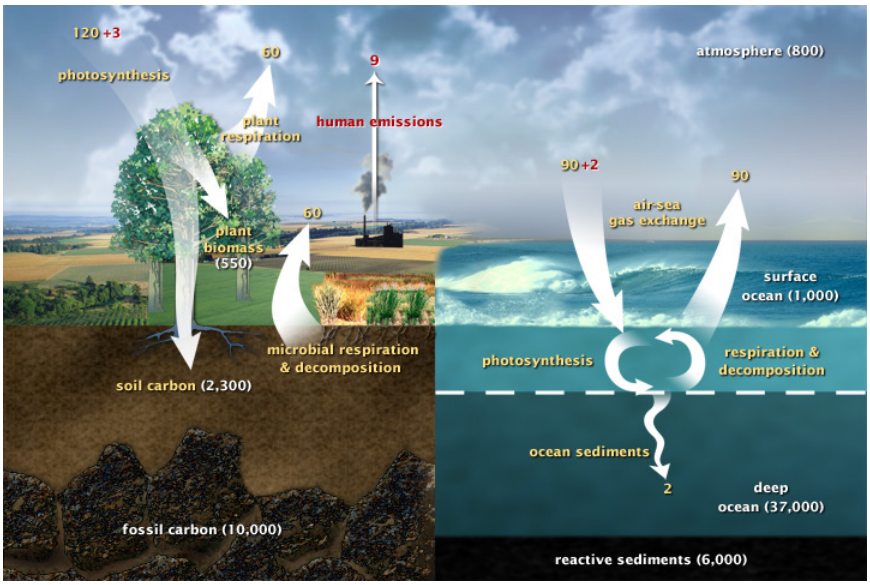


Figure 3.5.5: Short-term and long-term carbon cycles as they currently exist in the Earth System today. (NASA graphic)

In the nitrogen cycle, we remove nitrogen from a sink, the atmosphere (Haber-Bosch Process), to create ammonia-based fertilizers that we then spread on crops. This nitrogen moves into storage in plants (a sink), but excess is washed into waterways where it eventually leads to eutrophic conditions as algae blooms out of control, dies, and is digested by microbes that pull the oxygen out of the water column. This kills everything else. But, that nitrogen then becomes a part of the sediment and, eventually, denitrifies back into the air. Throughout this biogeochemical cycling, a molecule of nitrogen can be stored in a wide range of sinks for varying duration.

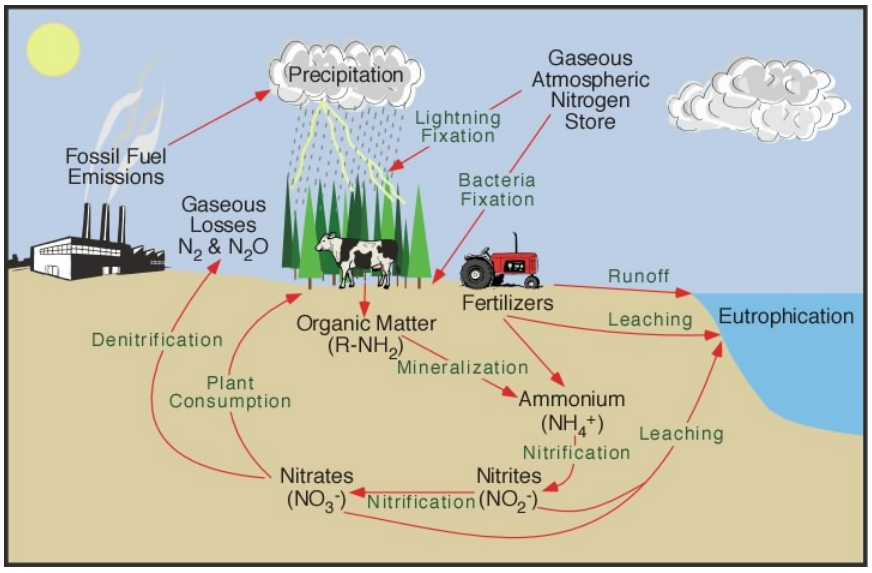


Figure 3.5.6: The nitrogen cycle (Pidwirny and Jones, physicalgeography.net)

Tipping Points

It is possible for a system to move so far out of its prior balance that, once it regains a sustainable equilibrium, it has a very different look and behavior than it did before. Initial forcings, followed by the actions of amplifying feedbacks can eventually cause a system to tip into an entirely new circumstance.

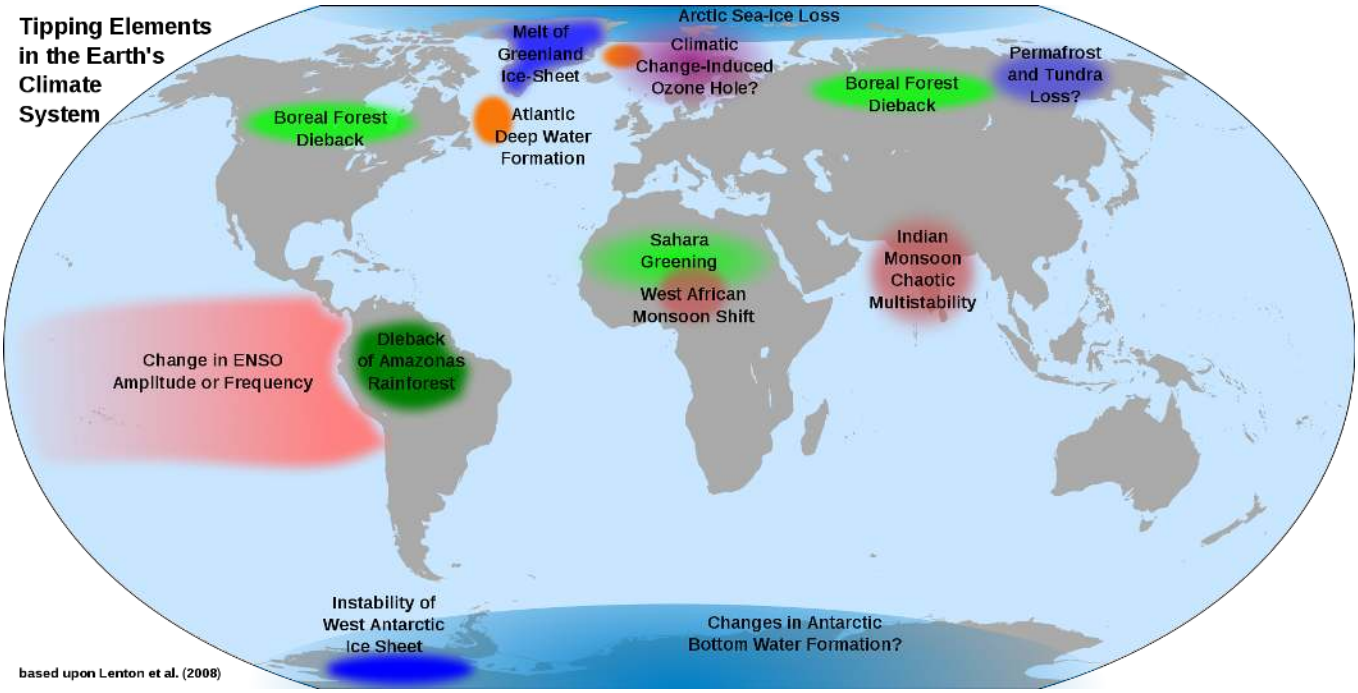


Figure 3.5.7: Global climate tipping points. While these are certainly based upon today's environment, many of these would have also been tipping points at various times in Earth's past. (Wikipedia Commons)

An excellent natural and historical example of this from the Earth's past might be the Great Oxygenation Event. Once photosynthesis evolved, organisms using this new energy pathway could harness an abundant resource, sunshine, and radiate far and wide throughout the nascent biosphere. Evolution took on new pathways and the course of life would lead to eukaryotes and multicellular structures and, eventually, to humans. However, not before the Earth system would find itself reacting violently (geologically) to this reduction in atmospheric carbon dioxide and increase in atmospheric oxygen. The result was a series of massive global glaciations as the Earth system struggled to gain a new equilibrium state.

Eventually, this period of global glaciation would subside. Afterward, the Earth that existed prior to these snowball Earth events was gone and the new Earth was much different. A new dynamic equilibrium was established.

Causal Loops

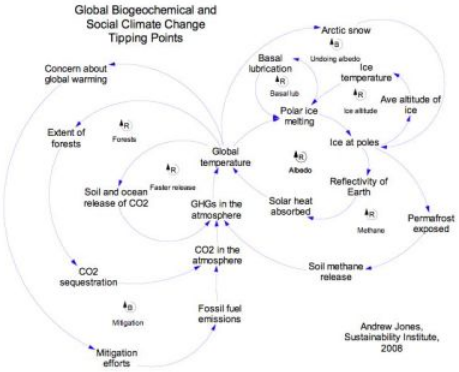


Figure 3.5.8: Causal loop diagram describing climate change tipping points with various feedbacks built in (amplifying and balancing). (Andrew Jones, Climate Interactive)

If we bring all of these systems characteristics together to model a system, we can create causal loop diagrams. These diagrams allow us to do two important things. First, they are effective tools for analyzing a system in as many directions as possible. You can see that this example, focused on climate change and, particularly, global biogeochemical cycles and climate tipping points, is quite complicated. These diagrams also help determine potential areas of focus for research. Finally, in a modern context we can use these to develop mitigation and management strategies for dealing with Earth system forcing events.

Did I Get It? - Quiz

? Exercise 3.5.1

An event in one Earth system will have an impact on other systems. When that impact accentuates that effect, what kind of feedback is that?

a. Balancing Feedback

b. Reinforcing Feedback

Answer

b. Reinforcing Feedback

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3.6: Scales of Measurement- Geography

Earth system forcings (events) happen at different geographic scales. That is, some have global effects while some have more regional or local effects. We will explore some of these below.

Global Effects

There have been many events throughout the Earth's past that have affected change on a global scale. These tend to be the big events, though they may have begun as small events that cascaded into larger-scale forcings. There are many to choose from, but for this section we will discuss the forcing effects caused by the evolution of vascular plants during the Silurian Period.

Evolution of Vascular Plants

Vascular plants entered the scene on land about 420 Ma. One of the challenges plants would encounter as they evolved to colonize land would be how to get water to move up toward the top of a plant against the downward pull of gravity. When you consider the Giant sequoia or Redwood trees in the Pacific northwest of the United States, you can really tell that plants found a way to master this problem. However, it is also true that there are limits to this. Plants overcame gravity in part by taking advantage of some of water's unique properties, such as capillary action. The evolution of specific tissues called xylem and phloem was the key development that allowed plants to harness this.

Xylem is a vascular tissue that transports water upward while phloem is a vascular tissue that transports photosynthetic cells to other parts of the plant for food or storage. The evolution of these tissues made it possible for plants to stand upright and, thus, colonize land.

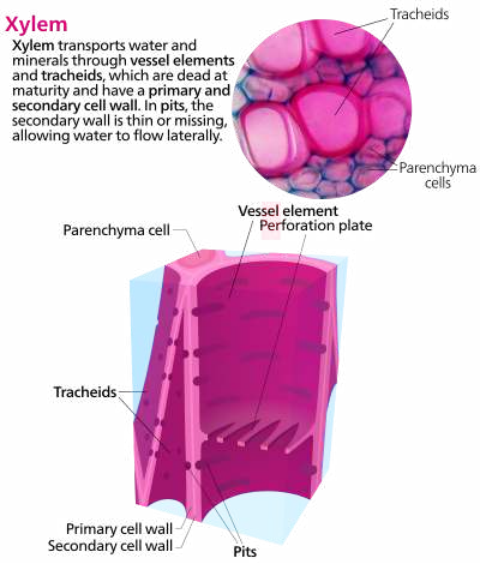


Figure 3.6.1: Xylem transports water and nutrients from roots to stems and leaves. (CC BY-SA 3.0; Kelvinsong – Own work)

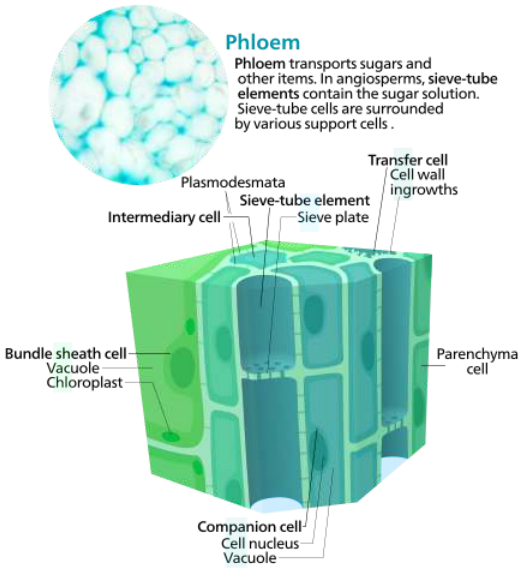


Figure 3.6.2: Phloem is living tissue in plants that transports soluble products of photosynthesis, such as sugars, where needed in plants. (CC BY-SA 3.0; Kelvinsong – Own work)

Why is this "event" a forcing mechanism? Well, consider the effect that photosynthetic vascular plants have on our atmosphere today. They are not only a critical part of the water cycle because of the transpiration of water through their tissues, but also for the removal of carbon dioxide from the atmosphere. Ultimately, plants' ability to convert carbon in carbon dioxide into glucose is a critical process for all life on Earth. Once vascular plants evolved, they spread prolifically, as there was a nearly endless array of niches to fill on land. This proliferation of plant material certainly forced changes in the atmosphere. Plant photosynthesis would remove carbon dioxide and replace it with oxygen, leading to a cooling climate globally. Their effect was so impactful through their drawdown of atmospheric CO_2 that it likely contributed to the Devonian extinction. They would not be outdone by their non-vascular cousins, who likely had the same impact on the end-Ordovician extinction.

Regional Effects

Some Earth system events create interesting regional dynamics. These are still affected by global changes, but the regionalism of impact simply means that global effects manifest themselves in different ways in different locations.

In terms of regional variation within Earth system, there are several examples that are important to understand today and that have applied or may have been different in the Earth's past. In each of the examples used to discuss region effects, the biosphere will be the focus.

Latitudinal Biodiversity Gradients

Species richness is highest in the tropics and lowest at the poles. In a general sense, this gradient is

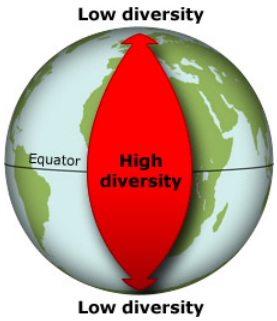


Figure 3.6.3: Modern latitudinal biodiversity gradient.

directly impacted by seasonality and the current, relatively moderate, climate. In Earth's past, particularly during times like the late Paleocene, it is thought that the tropics were so hot that biodiversity was lower than at temperate latitudes. Diversity in polar regions was also lower, but the climate was warm enough to support palm trees and alligators above the Arctic Circle. Biodiversity

gradients, then, were not only very different then, but have changed throughout Earth's past. The Biodiversity gradient we see today from the tropics to the poles has not always looked the way it does today.

Why would latitudinal biodiversity gradients change over time? In the image below, from Mannion et al. (2013), one idea is that diverse tropics have not always been the norm. In cooler periods (blue bands), it may be, but during warmer global climates, the tropics may be much more inhospitable. This would lead to a situation where the biosphere may have had bands of high diversity at mid-latitudes that trended downward north to the poles and south to the equator (in the northern hemisphere).

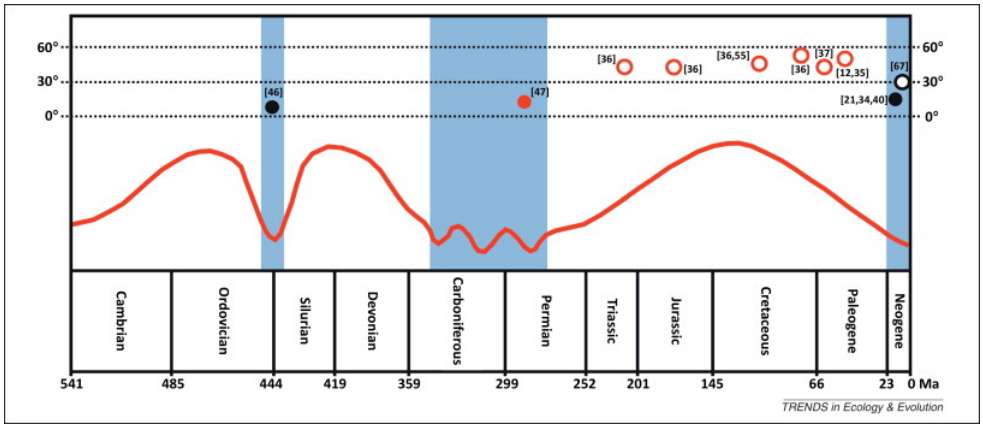


Figure 3.6.4: Summary of changes in latitudinal biodiversity gradients throughout the Phanerozoic Eon, according to Mannion et al. (2013). The smooth red line indicates global temperature changes. Filled circles represent times when the fossil record indicates a biodiversity gradient similar to today (poleward decrease). Open circles occur at times when evidence suggests that greatest diversity existed at mid-latitudes. Red circles are terrestrial data and black are marine. Blue bands indicate global icehouse periods.

Mathematically, latitudinal biodiversity gradients are governed by a simple relationship:

$$S = cA^z$$

Or

$$\log(S) = \log(c) + z\log(A)$$

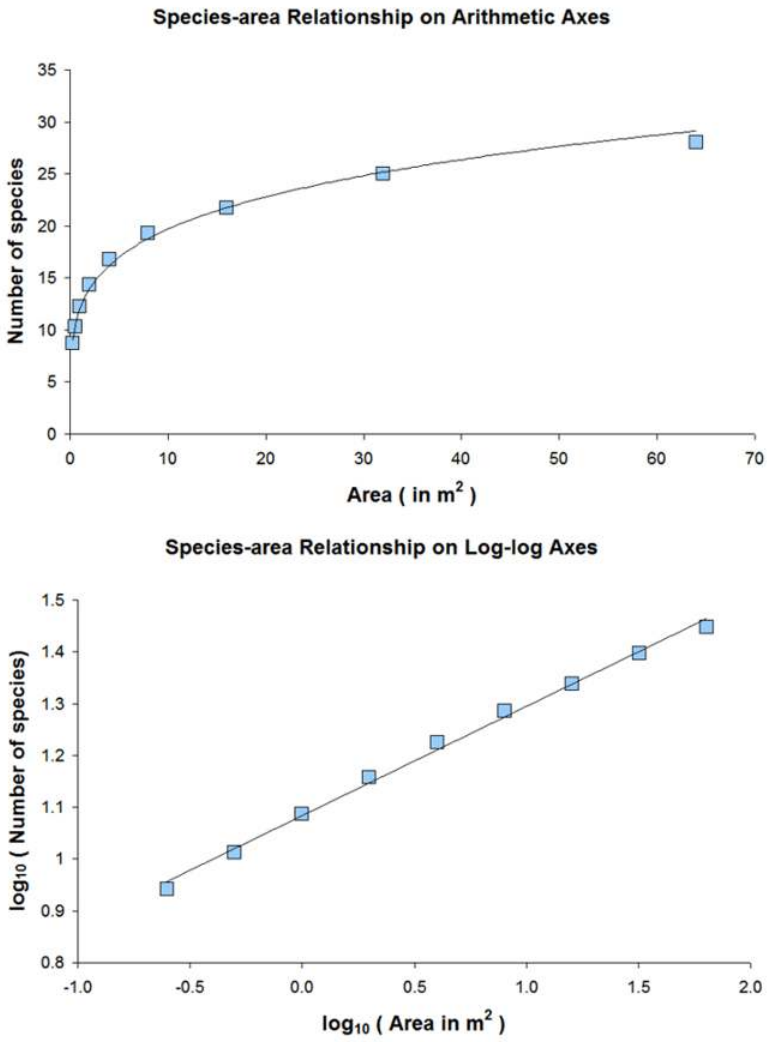


Figure 3.6.5: Species-area curve in arithmetic and log-log space. Z represents the slope of this curve. C is akin to the y-intercept, and A is the independent variable. The top image is the SAR curve with arithmetic axes and the bottom image is the SAR function in log-log space.

For any given area, S is the number of species predicted by the formula, per unit area. A is the area of the region of study, calculated in some unit for area. C and Z are constant values. In the equation above, C is the y-intercept and is typically a value that relates to the Area and defines the number of species that would exist in one square unit of area at that location. Z values represent the slope of the line. Z values are the most important factor in the equation and different environmental characteristics can be assigned different z values for the purposes of modeling. A region's climate is one such characteristic, though other items as wide-ranging as reproductive capacity and life mode can be assigned z values also. The species-area relationship is typically used to describe individual species and not entire ecosystems. However, once you calculate the number of a given species using this formula, knowledge of population ecology and ecosystem dynamics can be used to express a more holistic view of an environment. Generally, as the size of an area increases, so does the number of species.

Because of the climate, a very small area in a tropical region like the Amazon can not only have a dizzying array of species, but a very small area might also serve as the only location where that species can be found! When using the species-area relationship to model species richness as a function of climate, the lack of diversity in polar regions is a predicted result of the modeling. In addition, a single species might cover a very large area, such as spruce forests or Arctic Willow do in portions of the Canadian wilderness.

It is reasonable to assume that mathematical relationships like the species-area relationship not only apply to environments from the Earth's past as it does today. It should then be possible to model not only course global effects of forcings/events, but to examine

regional variations also.

Did I Get It? - Quiz

? Exercise 3.6.1

When you consider today's latitudinal biodiversity gradient and the species-area relationship, where do we find most biodiversity on Earth today?

a. Canadian Arctic

b. Appalachian Mountains

c. Southeastern United States

d. Equatorial Tropical Rainforests

Answer

d. Equatorial Tropical Rainforests

Continental Interiors and Margins

The location on a continent is another important climate regionality. Generally, the interiors of continents have more extreme temperature and moisture differences during cold and warm periods (day/night, summer/winter). Areas along a coastline, by contrast, tend to have much more moderate temperature and moisture variation. This is due to the already discussed elevated specific heat of water and its ability to buffer high temperatures. Continental interiors also tend to have higher elevations. Such locations could be mountainous or high plains. In either case, the average temperature drops as altitude increases. Seasonally, this leads to much greater variation in temperatures between day and night or summer and winter.

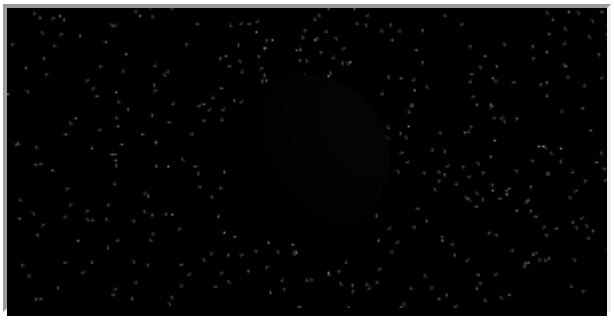


Figure 3.6.6: Earth circa 500 Ma. Note the much different position of the continents and of sea level. (Maps by Christopher Scotese of the PALEOMAP project, with virtual globe by Ian Webster of Dinosaur Pictures.org; reproduced with permission.)

Throughout Earth's past, the paleogeography has changed as plate tectonics forced continents into new positions. These forcings (events), though very long term in nature, led to significant changes in ocean current patterns and regional weather patterns. Short-term Earth system events, such as volcanic eruptions, might affect these patterns for a short time regionally or globally, but the return of the earlier dynamic equilibrium can typically be expected. Under average conditions, continental interior regions will have large range between moist and dry or warm or cold periods than their coastal counterparts.

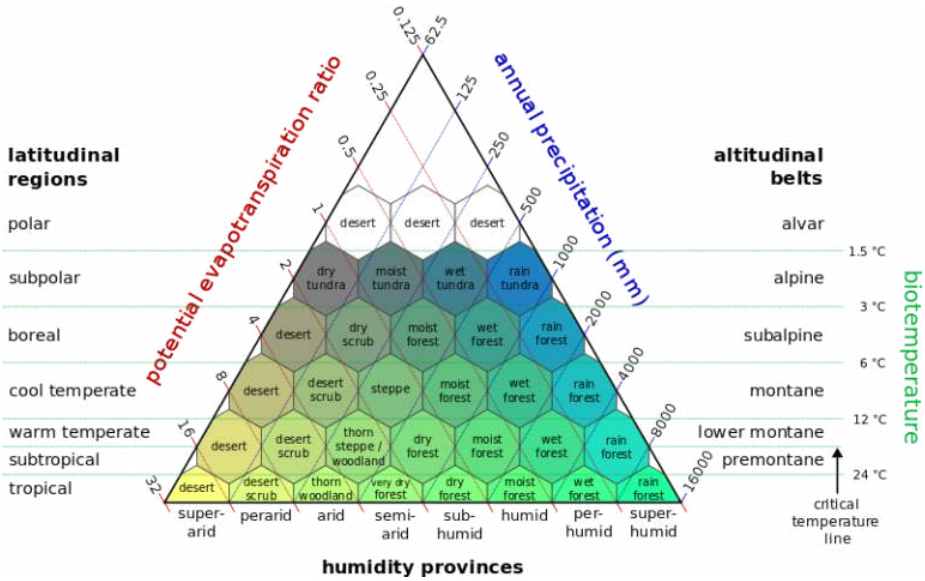


Figure 3.6.7: Regional climate zones are determined by a variety of factors, including whether an environment is within a continental interior or along the margin. Altitude is another important factor. All of these variables lead to variation in precipitation and evapotranspiration. Thus, different locations experience the hydrologic cycle differently, leading to differing climates. (Wikipedia Commons)

Ultimately, all of these geographic variations lead to different areas experiencing the water cycle differently. The key climatic variables, independent of temperature, for any location that determine the kinds of ecosystems that can exist are evapotranspiration and precipitation. There are myriad natural biomes and climatic zones that can be described by these two variables.

Local Effects

Earth forcings (events) also have local effects that vary widely, including as these affect local climate. While detecting local changes due to particular Earth events in the rock record is usually very challenging or even impossible, it is important to note that we can use modern analogs (uniformitarianism) to hypothesize how a locality might have behaved. Global events, like large asteroid impacts, certainly had global impact and detectable regional impact. Localized impact can be harder to see, but even if we do not see evidence for it, we can know that it happened.

Orographic Effects and Slope Aspect

A good example of local climate effects can be found in the impact that mountains have on local (and regional) weather patterns. On a single northern hemisphere hillslope, the north side of a mountain or hill might be covered in tall pines while the south slope might be a grassland. The temperature and moisture effects of the degree of incident sunlight on a hillslope, referred to as slope aspect, can have a profound effect on the type of ecosystem and communities that it supports.

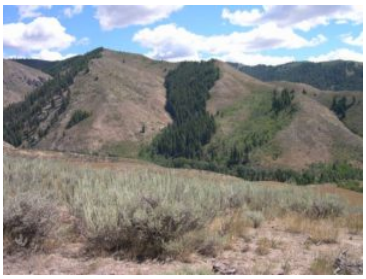


Figure 3.6.8: Slope aspect and its effect on vegetation near Anderson Ranch, Idaho. South-facing slopes get much more direct sunlight and less snow over. They sustain lower growing, desert flora. North-facing slopes hold snow cover and moisture longer and can support healthy forests.

Likewise, in situations where weather is moving from west to east, the eastern slope of a mountain range will typically experience a lower degree of moisture than the western slopes. This rain shadow effect is produced as orographic uplift causes weather systems to drain their energy and precipitation on one side of the mountain, leaving the leftovers (so to speak) to drop a pittance of their

former moisture on the other side. Such local effects may or may not be recorded in the rock record. Once the position of ancient mountain ranges is determined and their latitude defined, it is then possible to craft hypotheses about how weather may have been manifest in a specific area at that time.

Forests

Forests can certainly extend over a very large area, but their effect is also very regional to local in terms of climate. Forests are excellent places for moisture to accumulate and carbon to be sequestered. Transpiration of water through leaves is one way that plants transport soil moisture back to the atmosphere, raising the local relative humidity. This has a cooling effect. Forests can then have a similar effect on land as water has on coastlines, producing a more even temperature and moisture profile and reducing the severity of swings for these two variables.

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3.7: Scales of Measurement - Time

Earth system events and forcing mechanisms also vary across time. Some events are very short in duration but bring with them an intense amount of energy. This leads to massive and rapid change and a completely new dynamic equilibrium, eventually. Some events are much more gradual.

Daily/Diurnal Events

If you stop and think, you can certainly come up with a list of daily environmental "events" that go by without much notice. The Sun rises and sets while the Earth rotates on its axis. Tides come in and go out along coastlines. You sleep, you awaken. Temperatures rise and fall with the Sun. Such events are not always recorded in the rock record, yet we know they played an important role through time.

The nature of the term diurnal itself has variation across Earth's past. The length of a diurnal cycle today is just under 24 hours. However, this has not always been the case and will not remain this way into the future. Using fossil coral and fine layering in some sedimentary deposits that can be attributed to changes in lunar cycles, it is possible to extrapolate the change in diurnal cycles over time. Every 100 years, the length of a day is increased by 0.0024 seconds.

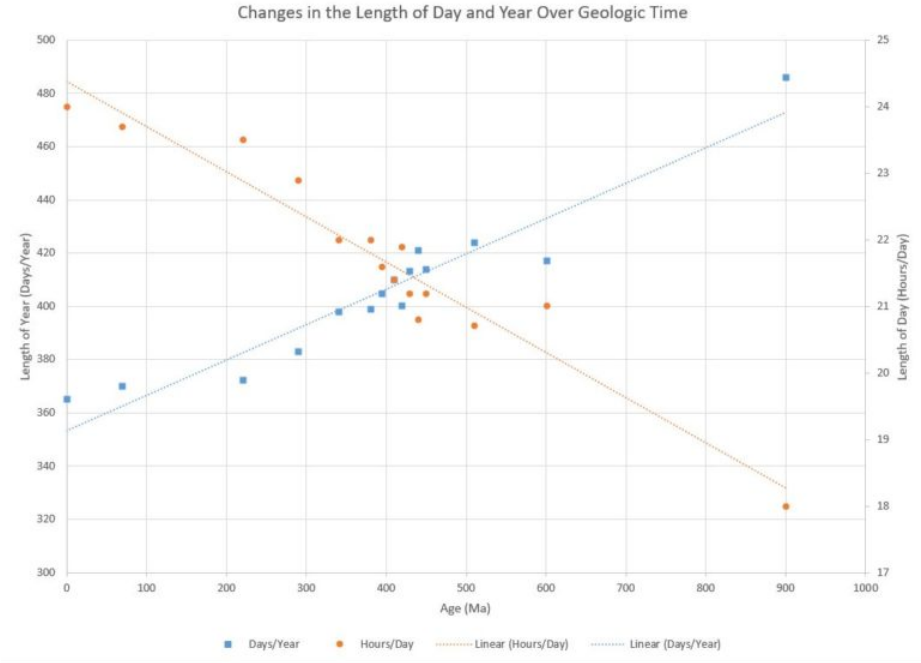


Figure 3.7.1: Changes in the length of a year and day on Earth over time. (NASA graphic)

In the graph above, note that while a day has gotten longer over time, the number of days in a year has decreased. This is because it is assumed that the Earth revolves around the Sun in a fixed amount of time, 365.25 days.

Why does this matter? Many, perhaps most, species today are very much connected to the diurnal cycle. The entire biosphere then is very much linked with this cycle, as it is with seasonal cycles throughout a year.

Annual and longer events

Earth system changes are ideally measured in terms of years. This can be a shorter term event that affects only a given year or two (small volcanic eruption, earthquake, forest fires). Or, important Earth system events can be measured over millions of years. A great example of this is silicate weathering.

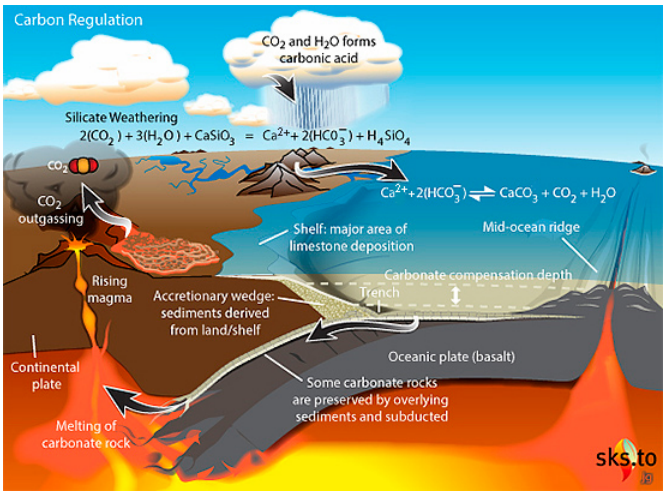


Figure 3.7.2: The silicate weathering cycle.

When two continents collide, mountains are raised up. We call this orogeny. As you drive those silica-rich rocks into the atmosphere they begin to weather, or break down. This is done through a process called hydrolysis, where weakly acidic rain (carbonic acid due to carbon dioxide) breaks down feldspar minerals in the rocks. In the image below, the role of silicate weathering in carbon regulation is highlighted. The long term effect of building up a mountain range is to force the climate to cool over a long period of time. Chemical weathering by hydrolysis breaks up silicate minerals and pulls carbon dioxide out of the air. Two by-products of the process are free calcium ions (aqueous) and aqueous bicarbonate. These new inputs into nearby basins can lead to a flourishing of carbonate life forms, who use these two substances to build their shells. Over time, they die, and their shells and the carbon they contain become limestone, a very effective sink for carbon storage.

Anomalies (Comparison to past trends/normals)

When analyzing data from any particular time period, modern or ancient, how do we know whether or not it is unusual or typical?

In order to define any variable within a system as complex as the Earth system at any given moment, it is useful to define anomalies. Simply put, an anomaly is a departure from normal behavior. The degree of anomaly can help scientists define the intensity of a forcing event and the state of the system at the time.

A very useful modern example, again taken from climate studies, comes from ground-based NASA surface temperature data, also known as GISSTEMP (Goddard Institute for Space Studies Surface Temperature Analysis). These charts, like the one below, are updated monthly. However, in order to define surface temperatures, the models use a baseline set of data, typically several decades in length, as an average for comparison.

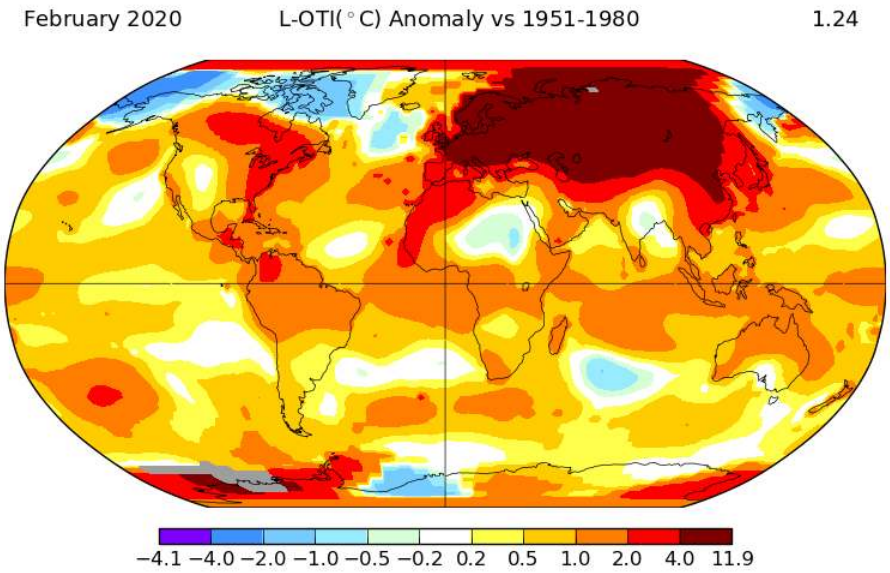


Figure 3.7.3: GISSTEMP surface temperature anomaly analysis for January 2020. Average January 2020 temperatures for various surface temperatures around the globe are measured against a baseline average taken from records spanning 1951-1980. The map shows anomalies, or departures, from the average for this thirty year period. Which areas experienced a particularly warm winter? (NASA GISSTEMP)

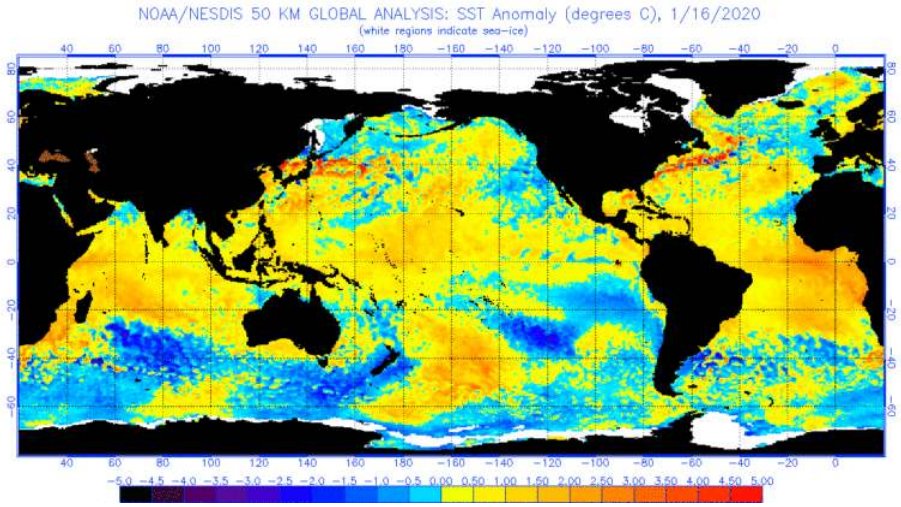


Figure 3.7.4: Sea surface anomaly analysis for January 16, 2020. Which regions of the sea surface are warmer or cooler than normal? (NOAA map)

The sea surface temperature anomaly map above comes from satellite data collected by the National Oceanic and Atmospheric Administration. Using the GOES-15 satellite, is is possible to see which localized regions of the oceans are experiencing warmer temperatures than normal. For analyzing the degradation and bleaching of coral reefs, as one useful example, such data is critically important.

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3.8: Ground-based instrumentation and satellite data

How do we study the Earth system as a whole?

Earth observing systems maintained by NASA and other space agencies around the world still use ground-based data. Such ground-based data measures variables on land and sea, as well as in marginal areas along coastlines. This data is often the highest accuracy data available for areas where it is collected. However, as station distribution tends to be uneven, as does maintenance and reporting, ground-based data has significant resolution challenges. Even still, it is very important for describing and forecasting a wide range of modern environmental variables.

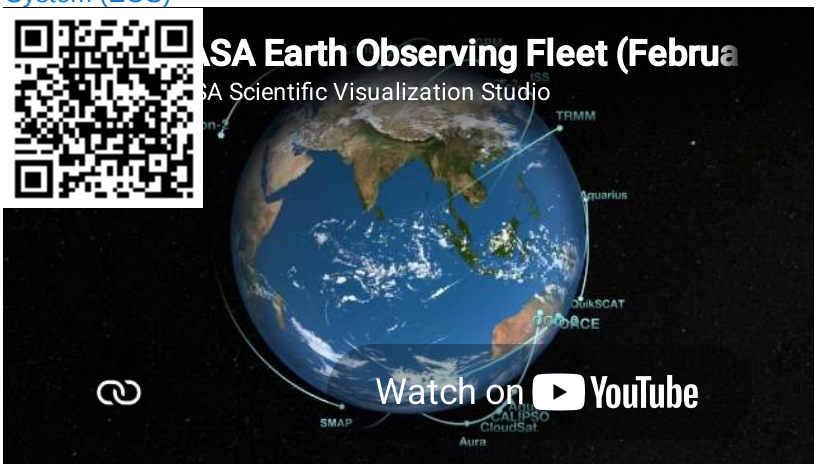
Ground based systems used to observe the Earth system include data collected by a wide spectrum of instrumentation. Some of these are completely automated, such as weather stations or tidal gauges. Others are maintained by a vast army of human volunteers, such as the GLOBE program. Perhaps the most comprehensive system of ground-based observation is maintained and run jointly by NASA, NOAA, EPA, and other governmental agencies.

Some of the challenges of ground-based data are resolved through the use of satellites.

Satellite data

Over the last forty years, a great deal of energy and money has been invested in the development of satellite systems to study the Earth from space. Such systems are very difficult to develop, very expensive, and very difficult to maintain. Despite these and other challenges, satellite observation systems now make up the backbone of data collection on the Earth system.

NASA Earth Observing System (EOS)



The U.S. space agency NASA maintains a vast fleet of satellites that are constantly measuring changes in the Earth system. Called the Earth Observing System (EOS), it consists of a constellation of satellites focused on studying every sphere. The goal of these missions is to collect long-term data about the Earth's spheres. Such data is used for daily forecasting models for your weather. It is also used for various levels of basic scientific research. All of these missions have polar orbits or orbits of low inclination, allowing them to have the best resolution possible with as much coverage of the planet as possible.

Below is an image the outlines the current constellation of satellites. Some of these current satellites are descendants of earlier missions. The LANDSAT program is a great example. LANDSAT 8 is the latest iteration of the program with missions as far out as LANDSAT 10 in planning. LANDSAT 7 is a great example of longevity in an Earth observing mission, as it is still active over 20 years after launch!



Figure 3.8.1: Current NASA Earth observing missions (NASA graphic (2019))

Nearly all of the data produced by these missions is in the public domain. In fact, NASA has developed websites that allow non-professional users to access and explore it. A very useful example is the NEO website. Perhaps the best total clearing house for NASA Earth observing data is one that places it in the context of global climate change. NASA's climate change website provides an excellent dashboard for keeping track of key data related to climate, but also provides links and news about other related missions.

Space Weather Observations

Understanding the dynamics of the Earth system is not complete without exploring space weather – the forcing mechanisms of the exosphere. Dominated by the Sun and its solar output, such phenomena also include monitoring the Earth's magnetic field, forms of cosmic radiation from the Sun and Milky Way, asteroids, and geomagnetic storm potential. Again, NASA has a number of missions exploring these things. More information about them can be found here.

Did I Get It? - Quiz

? Exercise 3.8.1

We study Earth's systems in order to better understand their interconnectedness, our planet's past, and how we interact with it. Base upon what you have read in this chapter, which statement best defines the Earth as a system?

a. A set of things that work together as a part of a mechanism or interconnected work.

b. A collection of several distinct systems that work independently and sometimes interact with one another.

c. A single entity that feeds energy in and through itself.

Answer

a. A set of things that work together as a part of a mechanism or interconnected work.

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3.9: Further Reading

The Climate Leader – Climate Interactive Online Course in Systems Thinking

The International Geosphere-Biosphere Programme – 2015

Kleidon, Axel (2012). How does the Earth system generate and maintain thermodynamic disequilibrium and what does it imply for the future of the planet? *Philosophical Transactions of the Royal Society A*. DOI:10.1098/rsta.2011.0316

Steffen, Will, Johan Rockström, Katherine Richardson, Timothy M. Lenton, Carl Folke, Diana Liverman, Colin P. Summerhayes, Anthony D. Barnosky, Sarah E. Cornell, Michel Crucifix, Jonathan F. Donges, Ingo Fetzer, Steven J. Lade, Marten Scheffer, Ricarda Winkelmann, and Hans Joachim Schellnhuber (2018). Trajectories of the Earth Systems in the Anthropocene. Proceedings of the National Academies of Sciences of the United States of America, 115(33).

Turner, David (2018). *The Green Marble: Earth Systems Science and Global Sustainability*. New York: Columbia University Press.

NASA Earth Science – Main Page

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CHAPTER OVERVIEW

4: Earth Materials - The Rock Forming Minerals

4.1: Introduction

4.2: What is a mineral?

4.3: The Building Blocks of Rock

4.4: The Chemistry of Minerals

4.5: Formation of Minerals

4.6: The Silicate Minerals - Nine of "The Big Ten"

4.7: Other Important Rock Forming Minerals

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4.1: Introduction

Geologists view Earth as an operational system of interacting spheres. These spheres include the Geosphere (the solid body of the Earth), the Atmosphere (the gas envelope surrounding the Earth), the Hydrosphere (water in all its forms at and near the surface of the Earth), the Cryosphere (frozen water part of Earth), and the Biosphere (life on Earth in all its forms and interactions, including humankind).

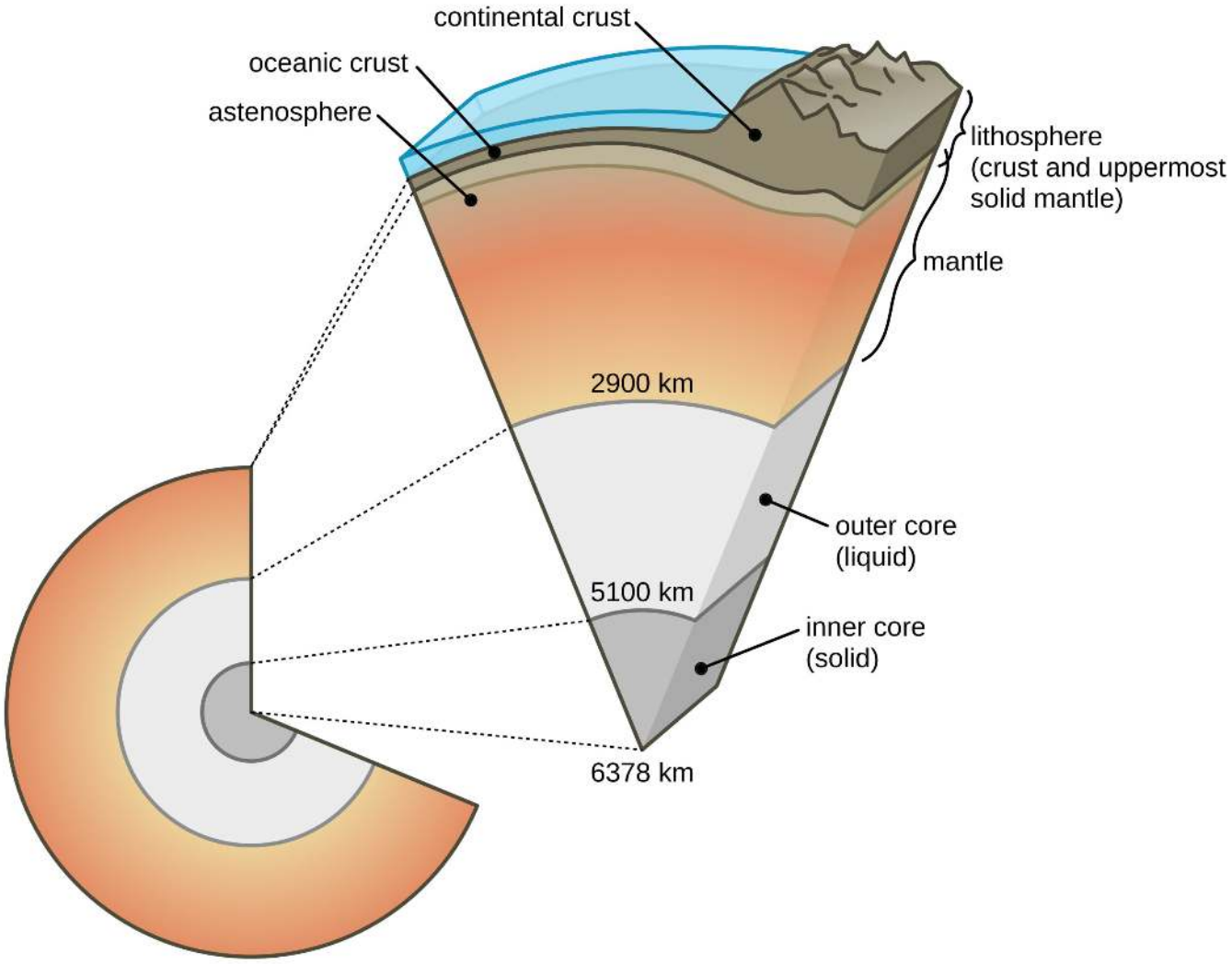


Figure 4.1.1: The layered interior of the Geosphere. Note that Lithosphere includes both the crust and uppermost portion of the solid mantle. This layer composes Earth's tectonic plates. (https://pubs.usgs.gov/gip/dynamic/inside.html United State Public Domain; U.S. Geological Survey, Department of the Interior.)

Earth materials are the foundational building blocks of the Geosphere and are generally thought of as the minerals and rock that compose the solid Earth. However, soil and loose fragments of rocks and minerals that exist at the surface are also considered Earth materials. Fossil fuels and water also exist as Earth materials in solid, liquid and gaseous states. Additionally, the envelope of gases that encircle the Earth is also considered Earth material. Therefore, Earth Materials encompass the entire Earth System where each of Earth's spheres interact. This chapter examines the origin and properties of materials largely found in the outermost layer of the solid Earth, the lithosphere. The lithosphere is the outermost rigid sphere of the Geosphere which includes both the crust and the solid upper mantle. These bound layers compose Earth's tectonic plates.

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4.2: What is a mineral?



Figure 4.2.1: The mineraloid opal filling a void space in a rock sample from Australia. Mineraloids are not actual minerals because they do not contain an orderly and repeating crystalline structure. (CC-BY-SA-3.0; Rob Lavinsky, iRocks.com)

The term “minerals” as used in nutrition labels and pharmaceutical products is not the same as what “mineral” means in a geological sense. In geology, the classic definition of a **mineral** is a substance that is: 1) naturally occurring, 2) inorganic, 3) solid at room temperature, 4) has an orderly and repeating internal crystalline structure, and 5) a chemical composition that can be defined by a chemical formula. Some natural substances technically should not be considered minerals, but are included by exception. For example, water and mercury are liquid at room temperature. Both are considered minerals because they were classified before the room-temperature rule was accepted as part of the definition. Although the mineral calcite, with the chemical formula $CaCO_3$, is quite often formed by organic processes, it is considered a mineral because it is widely found and geologically important. Because of these discrepancies, in 1985, the International Mineralogical Association amended the definition to: “A mineral is an element or chemical compound that is normally crystalline and that has been formed as a result of geological processes.” Typically, substances like amber, pearl, opal, or obsidian do not fit the definition of mineral because they do not have a crystalline structure. They are referred to as “mineraloids.”

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4.3: The Building Blocks of Rock

A **rock** is a solid substance that is made of one or more minerals or mineraloids. As discussed elsewhere, there are three families of rock composed of minerals: igneous (rock crystallizing from molten material), sedimentary (rock composed of the products of mechanical weathering, [sand, gravel, etc.] and/or chemical weathering [minerals and mineraloids precipitated from solution]), and metamorphic (rock produced by the chemical and physical reorganization of other rock under conditions induced by elevated heat and/or pressure).

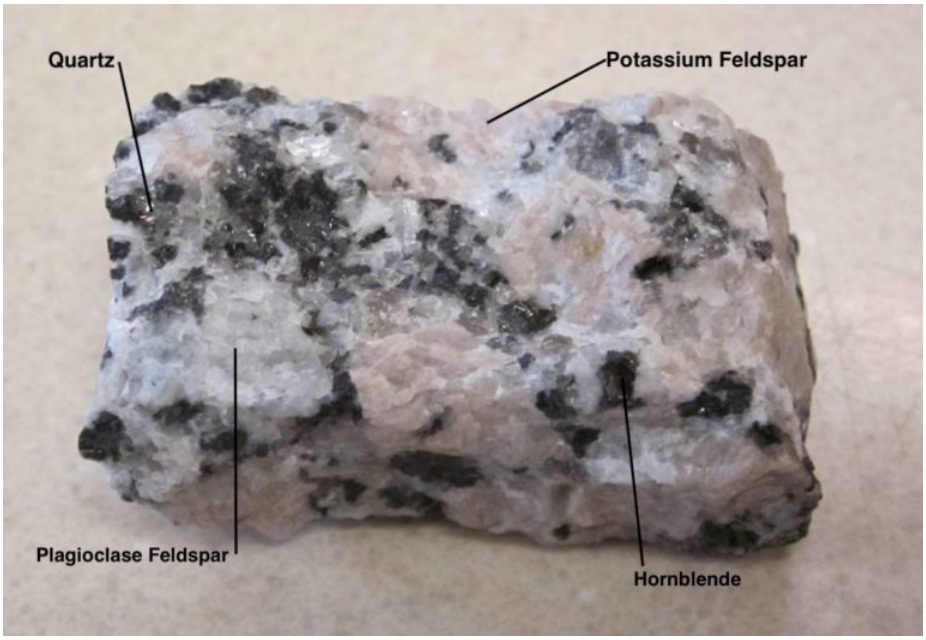


Figure 4.3.1: The common igneous rock, granite. The different minerals that compose the granite are labeled. (Creative Commons Attribution 3.0; Ralph L. Dawes, Ph.D. and Cheryl D. Dawes.)

Mineral identification is the first step in understanding the formation of a rock and its history. Geologists learn to "read the rock" to understand Earth's history at any given location where a rock is found in an outcrop. This allows geologists to understand what the environment was like at the moment the rock formed. Was there a volcano erupting or does the rock tell us that it formed deep inside a magma chamber? Was the rock formed by burial of an ancient beach? Was the rock formed by compressive forces deep within the crust as continents collided and new mountains were forming? The clues to these widely different environments of formation are "written in the rock." The first step in understanding the rock's history is being able to identify, characterize and quantify the minerals that compose the rock. Rocks are fascinating to a geologist because every rock has a story to tell. As we read the rock from one location to the next, it helps us piece together the fascinating story of the Earth.

As of 2018, there were over 5,000 minerals officially recognized by the International Mineralogical Association. Many of these formed under very specific chemical and geological conditions and may only occur in one location on Earth. Fortunately for geology students, only a small subset of these minerals are common and truly necessary for identifying Earth's most common rocks. These minerals have been dubbed "the Big Ten" by the prominent American mineralogist Mickey Gunter (Dyar, M.D., Gunter, M.E. and Tasa D. Mineralogy and Optical Mineralogy: Mineralogical Society of America, Chantilly, Virginia, USA). These ten minerals, plus a handful of others, will be our focus and will recur in many discussions in this text.

"The Big Ten" minerals are: olivine, augite, hornblende, biotite, calcium-rich plagioclase (anorthite), sodium-rich plagioclase (albite), potassium-rich feldspar (commonly orthoclase), muscovite, quartz, and calcite.

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4.4: The Chemistry of Minerals

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

Atoms

Matter is made of atoms. Atoms consist of subatomic particles: protons, neutrons, and electrons. A simple model of the atom has a central nucleus composed of protons, which have a positive charge, and neutrons which have no charge. A cloud of negatively charged electrons surrounds the nucleus, with the number of electrons equaling the number of protons, balancing with the positive charge of the protons for a neutral atom. Protons and neutrons each have a mass number of 1. The mass of an electron is less than 1/1000th that of a proton or neutron, meaning most of the atom's mass is in the nucleus. For our purposes, we can simplify the math by rounding the electron's mass down to zero.

The Periodic Table of Elements

Matter is composed of elements which are atoms that have a specific number of protons in the nucleus. This number of protons is called the atomic number of the element. For example, an oxygen atom has 8 protons and an iron atom has 26 protons. An element cannot be broken down chemically into a simpler form - they are 'elemental' for this reason! (Of course, nuclear reactions can break an atom down to a smaller, simpler form, but nuclear reactions are not chemical reactions). Down to the atom, each element retains unique chemical and physical properties, leading to particular behaviors in nature. This uniqueness led scientists to develop a periodic table of the elements, a tabular arrangement of all known elements listed in order of their atomic number and chemical properties.

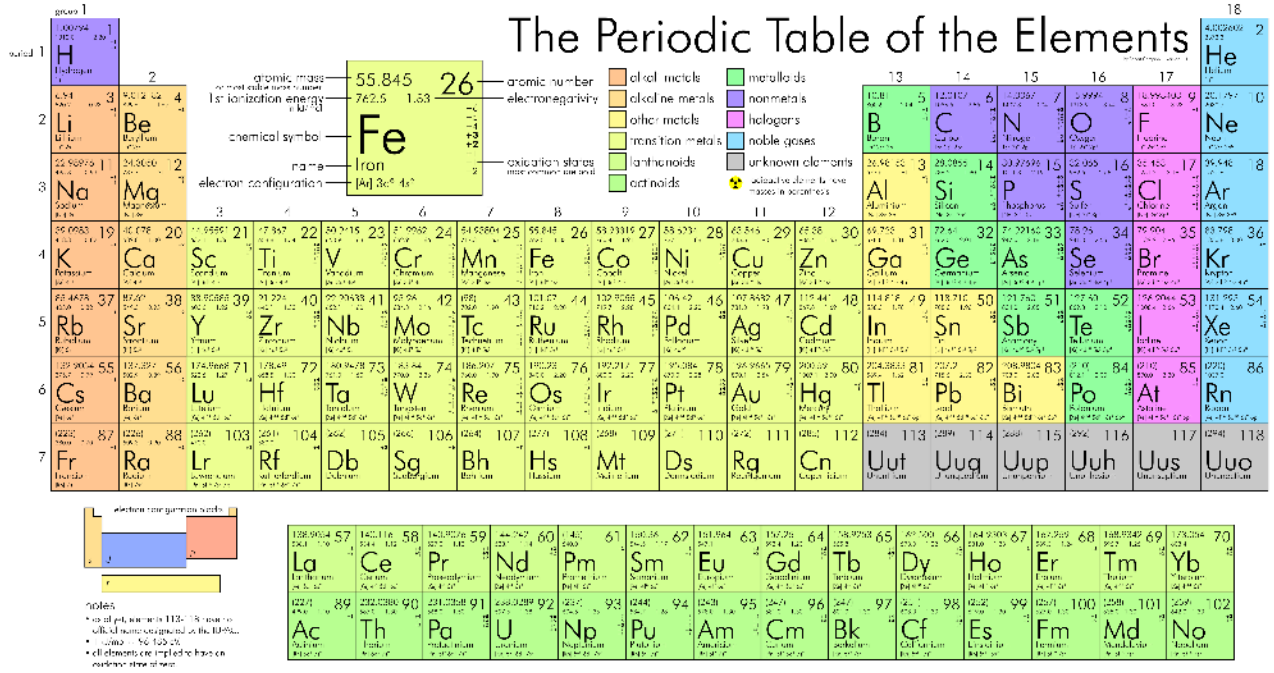


Figure 4.4.1: The periodic table of the elements of the world, showing element names, oxidation states, electron negativities, first ionization energies, and electron configurations. (Public domain, CC0 1.0 Universal; Wikimedia)

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

neutrons, the atomic mass is effectively the number of protons plus the number of neutrons. For example, the element silicon (Si) is atomic number 14 on the periodic table. Silicon has 14 protons and 14 neutrons in its nucleus and 14 electrons located in its electron cloud. The atomic mass of silicon is 28.01 mass units.

Isotopes

The atomic mass of an element represents an average mass of all of the atoms of that element found in nature. The atomic mass, noted on the periodic table, is usually not a whole number. Atoms of the same element can have differing numbers of neutrons in their nucleus. These are referred to as isotopes of that element. Isotopes behave the same chemically however they are slightly different in mass.

Isotopes are very important to Historical Geology. Applications of the use of isotopes in geology is discussed elsewhere in this text.

Common Elements of the Earth's Crust

Among the 118 known elements appearing on the periodic table, the heaviest are fleeting human creations known only in high energy particle accelerators, and they decay rapidly. The naturally occurring elements on the periodic table end with uranium, atomic number 92. Of these 92 elements, only eight are abundant in the Earth's crust and are shown in the pie chart below. Therefore, it is of these eight elements that the most common rock-forming minerals are composed.

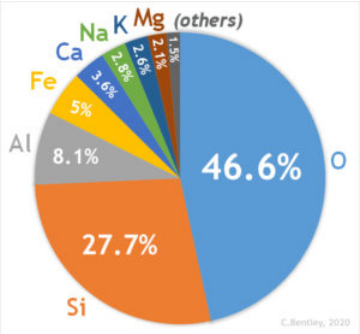


Figure 4.4.2: A pie chart showing the elemental composition of the crust.

Chemical Bonding

Minerals are formed by the chemical bonding that occurs between these elements. Most minerals are compounds containing multiple elements bonded together in a specific arrangement. Chemical bonding describes how these atoms attach with each other to form compounds, such as sodium (Na) and chlorine (Cl) combine to form the mineral halite with the chemical formula: NaCl. The mineral **halite** is common table salt. There are some minerals that are composed of only one element, such as native Copper (Cu), native Gold (Au), or native Silver (Ag). These single elements bond to each other to form molecules, however they are not compounds because they are made of only one element.

Valence, Charge and Ions

The electrons around the atom's nucleus are located in shells representing different energy levels. The outermost shell is called the valence shell. Electrons in the valence shell are involved in chemical bonding. In 1913, Niels Bohr (1885-1962) proposed a simple model of the atom that states atoms are more stable when their outermost shell is full [3, 4]. Atoms of most elements thus tend to gain or lose electrons so the outermost or valence shell is full. Elements that appear on the left side of the periodic table have smaller numbers of electrons in their outermost (valence) shell while elements on the right side have increasingly greater numbers of electrons in the valence shell. It takes a great deal of energy for the nucleus of an atom to hang on to a small number of valence electrons. These elements tend to search out partners that need a few electrons to fill their valence shell. If an element gains or loses electrons in its valence shell, it is then referred to as an ion. Because the number of electrons and protons are no longer equal in an ion, it has a charge. The charge is positive if protons outnumber electrons, and negative if electrons outnumber protons.

Let's use a familiar example to help illustrate. Sodium (Na) is located on the far left side of the Periodic Table. It has one lonely electron in its valence shell that it desperately wants to shed. Remember, one atom of sodium (Na) has the same number of protons in its nucleus as it has electrons in its cloud. Therefore, if it loses any electrons it will have an excess positive charge because the

positive charge of the protons in the nucleus will now outnumber the electrons. As sodium (Na) relinquishes its one valence electron is becomes a positively charged ion called a cation and is designated Na^+. Coincidentally, chlorine (Cl) has one spot left for an electron in its valence shell. If chlorine (Cl) gains an electron, it will have a net negative charge and is referred to as an **anion** and is designated Cl^-.

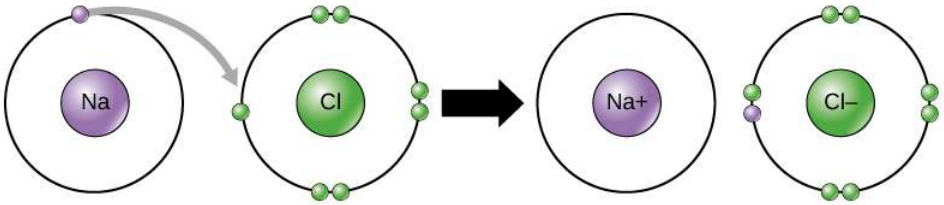


Figure 4.4.3: The formation of an ionic chemical bond by electron transfer. (CC BY 4.0; OpenStax, https://louis.oercommons.org/courseware/module/601/student/?task=9)

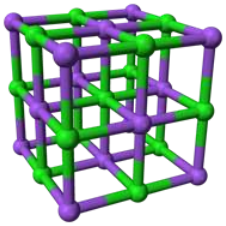


Figure 4.4.4: The unit cell for the mineral halite. (Public domain, CC0 1.0; Benjah-bmm27, Wikimedia)

A chemical bond will form between these oppositely charged ions which creates a molecule. This type of chemical bond is an ionic bond where **electron transfer** and a strong attraction between oppositely charged ions creates the bond. This molecule of NaCl (sodium chloride) is not a mineral. Remember, a mineral must have an "orderly and **repeating** internal crystalline structure." The smallest representation of a mineral is **one unit cell**. The unit cell for the mineral halite (sodium chloride, NaCl) appears in the figure below where the sodium atoms are purple and the chlorine atoms are green.

Explore the following 3D molecular crystal structure of the mineral halite, below. Roll your cursor over to grab and rotate the structure.

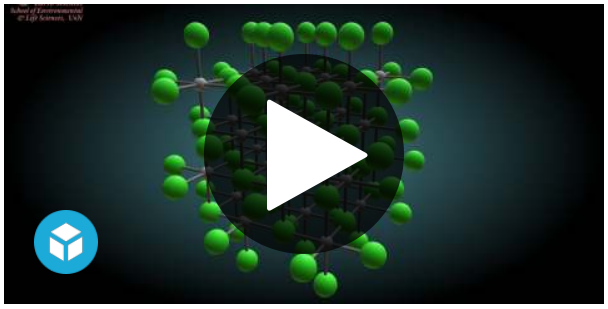


Figure 4.4.5: 3D molecular structure of the mineral halite. (With permission by: Dr. Bill Landenberger, Earth Sciences, University of Newcastle, Australia on SketchFab.)

Now, take a look at a 3D representation of perfect mineral crystals of halite.



Figure 4.4.6: Halite crystals by Malopolska`s Virtual Museums (Public domain; Sketchfab)

Do you notice the similarity of the molecular structure with the outward appearance of the crystals? The outward physical appearance of a mineral will reflect its internal atomic structure. How cool is it that the atomic level structure of a mineral is revealed in its macroscopic (hand-sample sized) shape! We will look at this in more detail in the upcoming section on diagnostic mineral characteristics.

Another type of chemical bonding can occur between elements in the formation of minerals that involves **electron sharing**. This is most easily explained in looking at a simple molecular model of H_2O, water. Oxygen has 6 electrons in its valence shell and hydrogen has only one. By sharing these electrons, the valence shells will contain 8 electrons for oxygen and 2 for hydrogen. Both configurations are very stable.

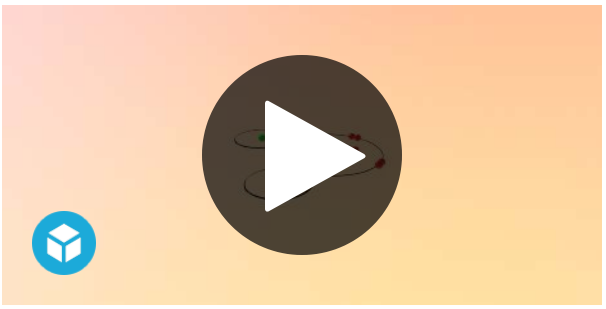


Figure 4.4.7: 3D rendering of the electron sharing through covalent bonding of a water molecule. (CC BY: 4.0; Danarmenta on Sketchfab)

The Silica Tetrahedron

The most common covalent bond in the formation of minerals is the bond that occurs between silicon and oxygen. The pie chart in the previous section shows us that the two most common elements in the Earth's crust are oxygen and silicon. It makes sense that the most common minerals in the crust will therefore contain a lot of these two elements. The first nine of "The Big Ten" minerals are indeed from the largest mineral group, the silicate minerals. The silicates are named for the silicon and oxygen foundation in their composition. The basic framework of the silicate minerals is the silicon - oxygen (or silica) tetrahedron where four oxygen atoms pack very tightly around one silicon atom. The resulting molecular structure is shaped like a tetrahedron.

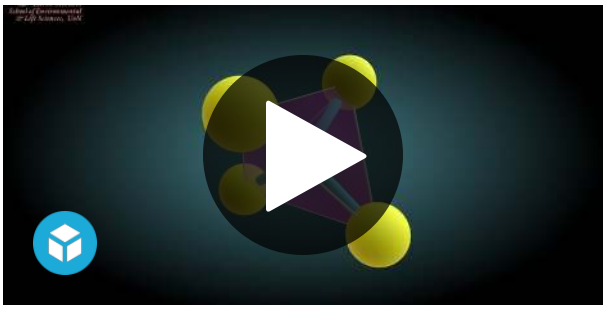


Figure 4.4.8: This model depicts the basic building block of most silicate minerals - the SiO_4 tetrahedron. It is represented by a ball-and-stick portrayal of the molecule with a single cation of silicon (Si^{+4}) in the centre (in red) bonded with four surrounding oxygen anions (each O^{-2}) in a tetrahedral arrangement (yellow). This is combined with a transparent (purple) depiction of the coordination polyhedron. The molecule cannot exist by itself as there is a residual negative charge of -4. This charge is accounted for by sharing oxygen anions with adjacent tetrahedra (e.g. in quartz), or by bonding with other cations in silicate minerals (e.g. Mg^{+2} in olivine). (With permission; Dr. Bill Landenberger, Earth Sciences, University of Newcastle, Australia on SketchFab.)

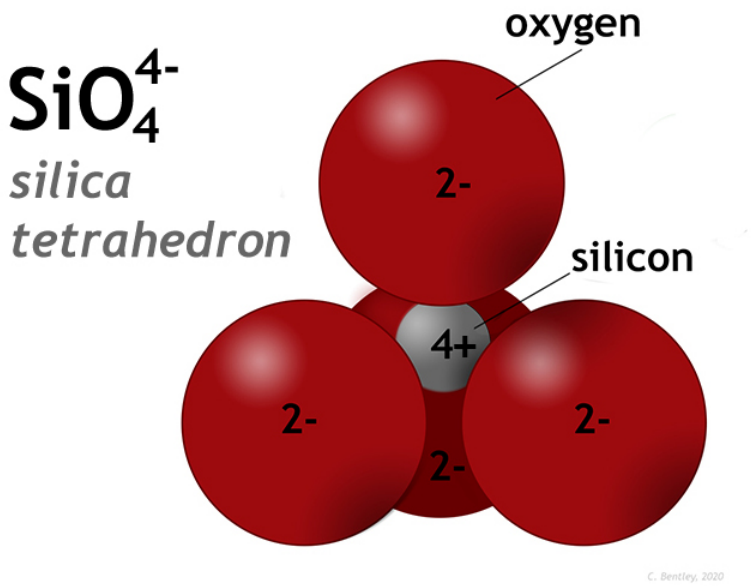


Figure 4.4.9: The atomic structure of the silica tetrahedron, a foundational molecular structure in all silicate minerals. (Callan Bentley)

The silica tetrahedron is referred to as an "complex molecule." This means that the silicon and oxygen atoms are very tightly bonded however the overall charge is negative making this complex molecule a complex ion. The silica molecules acts as an anion. The silicon atom has four valence electrons while the oxygen atoms each have six valence electrons. To fill the outermost electron shell, each atom would need to share a total of eight electrons. This cannot happen since silicon has only four electrons to share. The result is a complex molecule with a net negative charge of -4. As a result, this complex silica molecule is ready to bond with other available elements or complex molecules. This is how all of the silicate minerals form.

Let's look at one of our "Big Ten" silicate minerals, olivine, to see how this works.

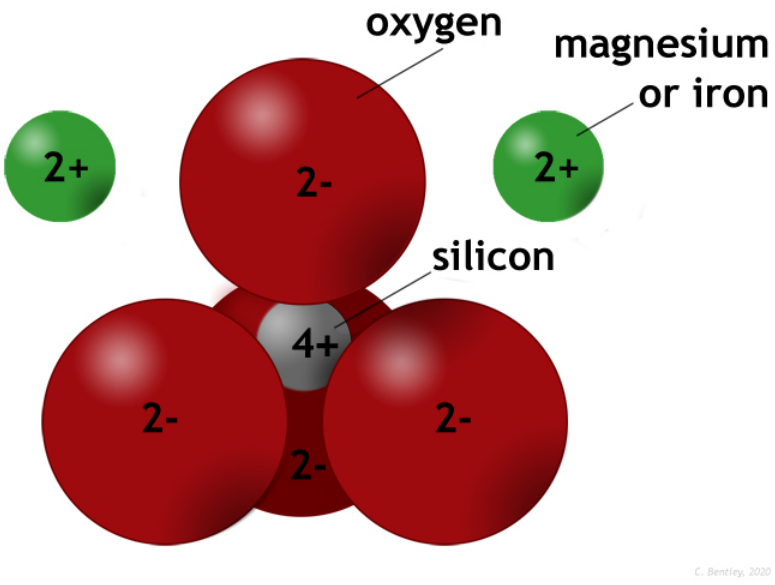


Figure 4.4.10: The unit cell of the mineral olivine features atoms of oxygen, silicon, and iron or magnesium (either will do just fine) in this arrangement. (Callan Bentley.)

Olivine is largely an "igneous mineral" meaning that olivine most commonly forms as a magma cools and minerals begin to crystallize. The chemical formula for the mineral olivine is: $(\mathrm{Mg,Fe})_2\mathrm{SiO}_4$ The elements of Mg (magnesium) and Fe (iron) appear

together in parentheses because they can readily substitute for each other in the mineral's crystalline structure depending on which of the elements are available in the cooling magma.

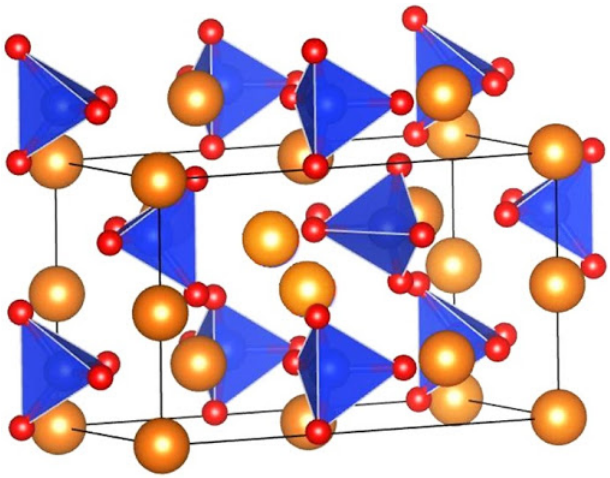


Figure 4.4.11: One unit cell of the mineral olivine. Silica tetrahedra represented in blue with oxygen atoms in red. Iron (Fe^{+2}) and/or magnesium (Mg^{+2}) ions are represented in gold. (Creative Commons Attribution 3.0; S Mahendran et al 2017 Modelling Simul. Mater. Sci. Eng. 25 054002)

Both Mg and Fe have a valence charge of +2; both Mg and Fe have two electrons in their outermost shells, e.g. electrons that they are looking to "give away." They are swimming around in that magma looking for partners! As the magma cools, it will reach a temperature at which bonding can occur. As discussed, the complex silica molecule forms with the overall -4 charge. The available Fe+2 and Mg+2 will form an ionic bond with the complex silica molecule which will balance out the overall charge. However, as indicated above with the example of the mineral halite, one molecule does not constitute a mineral. The smallest representation of a mineral is one unit cell which appears in the diagram (right) for olivine.

[1] Mendeleev, D. The relation between the properties and atomic weights of the elements. *Journal of the Russian Chemical Society* **1**, 60–77 (1869)

[2] Scerri, E. R. *The Periodic Table: Its Story and Its Significance*. (Oxford University Press, USA, 2007)

[3] *Science literacy: concepts, contexts, and consequences*. (National Academies Press (US), 2016).

[4]Bohr, N. On the constitution of atoms and molecules. *Philos. Mag*. **26**, 1–24 (1913)

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4.5: Formation of Minerals

Minerals form when atoms bond together in a crystalline arrangement. In order for a mineral crystal to grow, the **elements** needed to make it must be present in the appropriate proportions, the physical and chemical **conditions** must be favorable, and there must be sufficient **time** for the atoms to become arranged.

Physical and chemical conditions include factors such as temperature, pressure, presence of water, pH, and the amount of oxygen available. Time is one of the most important factors because it takes time for atoms to move from place to place and become ordered. If time is limited, the mineral grains (with crystalline structure) will remain very small. The presence of water enhances the mobility of ions and can lead to the formation of larger crystals over shorter time periods.

Common Processes of Mineral Formation

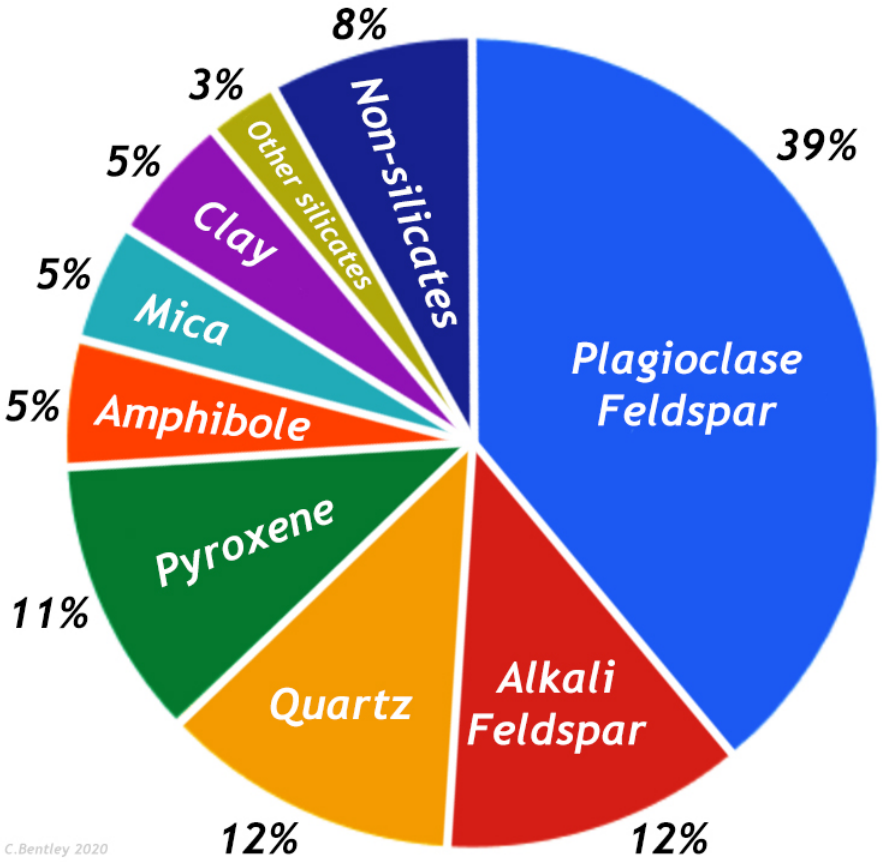


Figure 4.5.1: The most abundant minerals in Earth's crust. The crust is dominantly composed of just a few mineral groups, the vast majority of them silicates. (Callan Bentley.)

- Crystallization from molten rock material (magma or lava). The majority of minerals in the crust have formed this way.
- Organic formation: formation of minerals by organisms within shells (primarily calcite) and teeth and bones (primarily apatite)
- Precipitation from aqueous solution (i.e., from hot water flowing underground, from evaporation of a lake or inland sea, or in some cases, directly from seawater).
- Weathering: during which minerals unstable at Earth's surface (conditions of low temperature, low pressure, high moisture, and high oxygen levels) may be altered to other minerals.
- Metamorphism: formation of new minerals directly from the elements within existing minerals under conditions of elevated temperature and/or pressure.

These processes are responsible for the formation of the various mineral groups that compose the Earth's crust. See the diagram below which illustrates the composition of the Earth's crust by the minerals that compose it.

Crystallization from Magma and the Formation of the Silicate Minerals

Most of the minerals of Earth's crust formed through the cooling of molten rock (magma or lava). Molten rock is very hot, typically on the order of 1000 °C (1800 °F) or more. Heat is energy and temperature is a measure of that energy. Heat causes atoms to vibrate and temperature measures the intensity of the vibration. If vibrations are very strong, chemical bonds will break and the pre-existing minerals in Earth's crust and mantle will melt. Melting releases ions into a pool, forming a magma chamber. Magma is simply molten rock with freely moving ions. If magma is allowed to cool at depth or erupted onto the surface (then called lava), mineral crystals will form as temperature decreases.

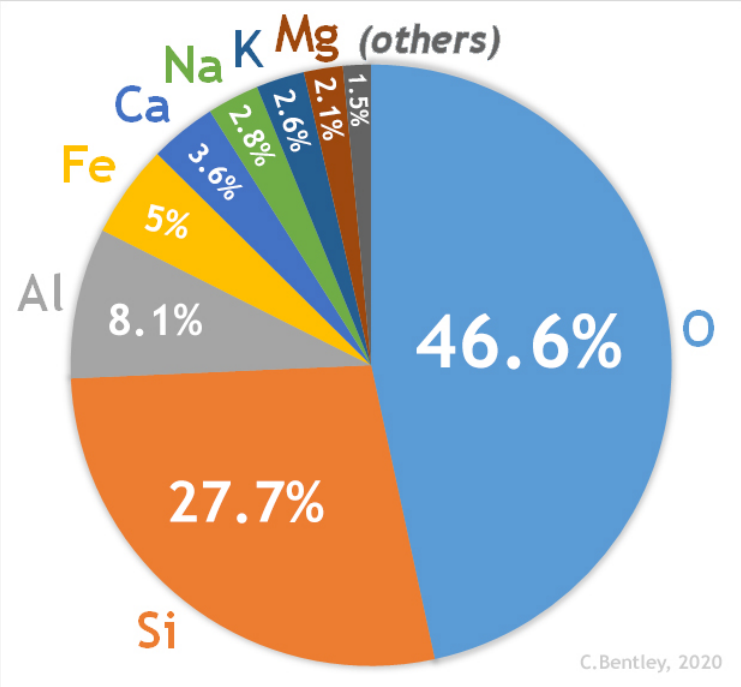


Figure 4.5.2: The elemental composition of the crust. (Callan Bentley.)

The chemical composition of Earth's crust is critical to the discussion of the dominant minerals to form from magma or lava, the silicate minerals. It is important to understand that not every location in the crust will have this composition. All rock is different and virtually no two rocks within the crust will have the exact same composition. The same also goes for magma. No two magma bodies will have the exact same composition. The magma will however, contain these elements in somewhat varying proportions depending on where exactly within the crust (or mantle) the magma was derived. What we can say is that the magma will be largely composed of silicon and oxygen with varying proportions of the remaining six elements plus other elements in trace amounts. The combination of these elements will form the different silicate minerals as the magma cools either deep within the Earth or near/at the surface.

Bowen's Reaction Series

Crystallization of minerals from magma, or cooling lava, follows a very specific sequence. This sequence was first determined experimentally by the scientist Norman L. Bowen (1887-1956) at the Carnegie Institute in Washington, D.C. in the early 1900's. This order is referred to as "Bowen's Reaction Series" and represents the formation of the largest group of minerals, silicate minerals (see diagram below). This specific process of crystallization leads us to the formation of nine of "The Big Ten" minerals: olivine, pyroxene (augite), amphibole (hornblende), biotite, calcium-rich plagioclase (anorthite), sodium-rich plagioclase (albite), potassium-rich feldspar (commonly orthoclase), muscovite, and quartz.

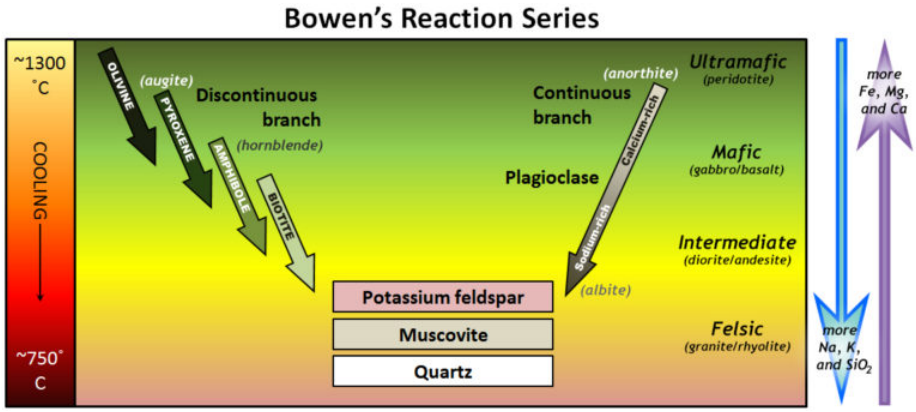


Figure 4.5.3: Bowen's Reaction Series. Developed through laboratory experimentation by Norman L. Bowen in the early 1900's. It establishes the order of crystallization of minerals from a silicate magma. (Creative Commons Attribution 4.0 International License; Modified after Steven Earle, https://opentextbc.ca/physicalgeology2ed/)

The left side of Bowen's Reaction Series Diagram is labeled the "Discontinuous Branch" while the right side is labeled the "Continuous Branch." On the left side, the minerals are most iron- and magnesium-rich at the top. Their crystal structure changes and iron and magnesium content decrease as temperatures drop. On the right side, the plagioclase minerals form in one continuous flow from calcium-rich at the top to sodium-rich at the bottom.

Beginning with the "Discontinuous Branch," olivine forms from single silica tetrahedron ionically bonded to iron and/or magnesium for the crystalline structure. As the magma temperature cools and crystallization proceeds, the silica tetrahedron will link together ("polymerize") to form chains, sheets and frameworks. This allows for the formation of the different minerals in progression along the Discontinuous Branch.

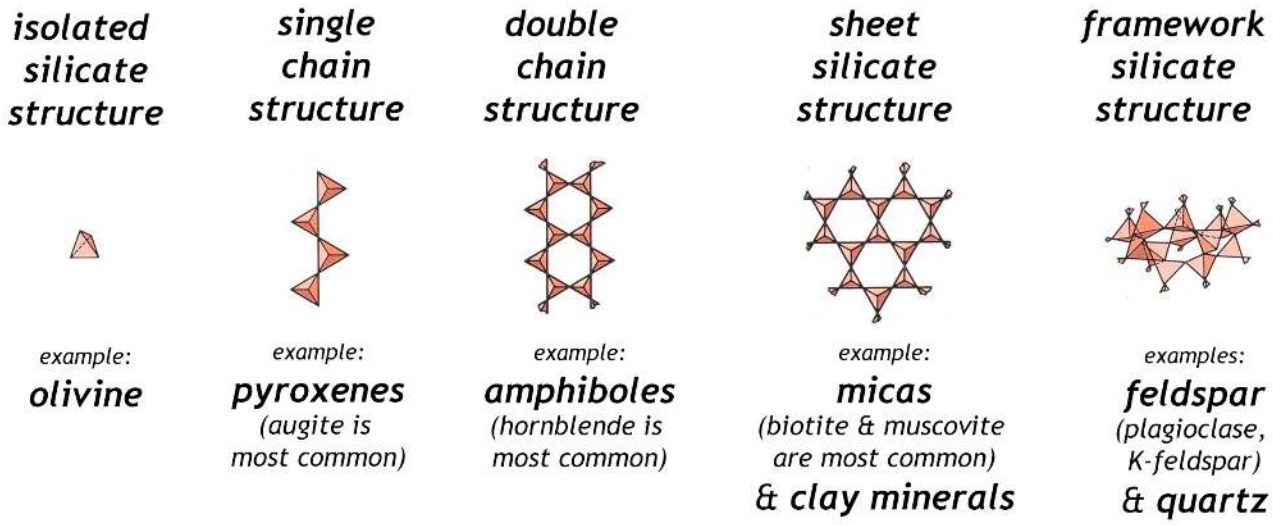


Figure 4.5.4: The various crystal structures of silicate minerals. (Callan Bentley.)

As the temperature drops, and assuming that some silica remains in the magma, the olivine crystals react (combine) with some of the silica in the magma to form pyroxene (commonly augite). As long as there is silica remaining and the rate of cooling is slow, this process continues down the discontinuous branch: olivine to pyroxene, pyroxene to amphibole (commonly hornblende), and amphibole to biotite.

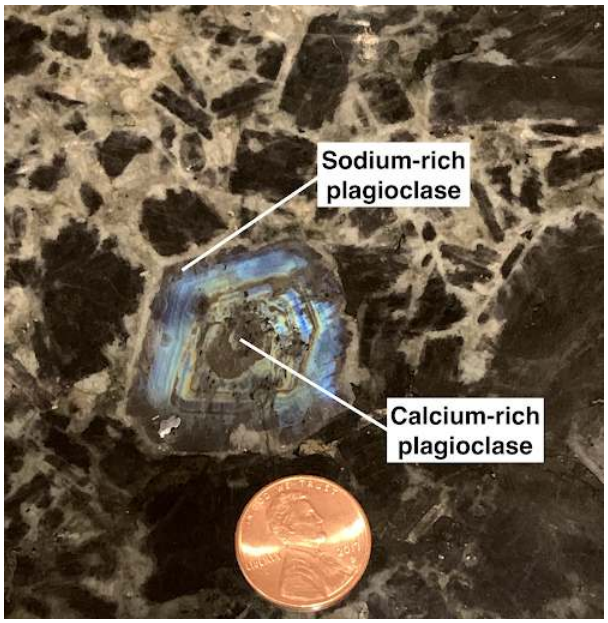


Figure 4.5.5: “Zoning” in a large crystal of plagioclase formed during the slow cooling of magma in an intrusive igneous body. The most calcium-rich concentration is located in the center that formed during the early cooling stages. The plagioclase became progressively more sodium-rich as cooling and crystal growth continued. The blue coloration is called “schillerization” due to microscopic inclusions common to this variety of plagioclase. (Shelley Jaye.)

At about the point where pyroxene (augite) begins to crystallize from the cooling magma, plagioclase feldspar also begins to crystallize on the continuous (right) side of Bowen’s Reaction Series Diagram. As cooling continues, plagioclase minerals retain their original framework silicate crystal structure. The core of the crystal begins calcium-rich and, once crystallization is complete, ends with a rim of sodium-rich plagioclase. See the photo below. Finally, if the magma is quite silica-rich to begin with, there will still be some left at around 750° to 800 °C, and from this last magma, potassium feldspar, quartz, and maybe muscovite mica will form.

Select play on the following animation to see how silicate minerals form from a cooling magma. Use the sliders to change the overall silica content (ultramafic to felsic) and cooling time. This is a very good visual representation of how the ions of the different common elements exist in magma and combine to form the different minerals found on Bowen’s Reaction Series as temperature decreases and the magma crystallizes.

Watch the following video which will introduce you to the silicate minerals:



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4.6: The Silicate Minerals - Nine of "The Big Ten"

Olivine

Chemically, **olivine** is mostly silica, iron, and magnesium and typically green in color. Olivine is the primary mineral component in mantle rock (called peridotite) and ocean floor rock (called basalt). It is characteristically green when not weathered. The chemical formula is \(\ce{(Fe,Mg)2SiO4}\). The comma between iron (Fe) and magnesium (Mg) indicates these two elements occur in a solid solution. Not to be confused with a liquid solution, a solid solution occurs when two or more elements have similar properties and can freely substitute for each other in the same location in the crystal structure. The crystal structure of olivine is built from independent silica tetrahedra.



Figure \(\PageIndex{1}\): The iron- and magnesium-rich silicate mineral olivine (green). (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

The Pyroxene Family

Augite is the most common mineral of the pyroxene family and one of our Big Ten Minerals. Augite is iron and/or magnesium-rich forming a complex structure of elements bonded to polymerized single chains of silica tetrahedra. Augite is typically black or dark green in color. The chemical formula for augite is complex, indicating that different elements may substitute in the structure depending on what is available in the cooling magma: \(\ce{(Ca,Na)(Mg,Fe,Al,Ti)(Si,Al)2O6}\).



Figure \(\PageIndex{2}\): The iron- (Fe) and magnesium- (Mg) rich silicate mineral augite is a member of the pyroxene family. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

The Amphibole Family

As we move down Bowen's Reaction Series, the internal crystal structure of each mineral becomes increasingly more complex. Amphibole minerals are built from polymerized double silica chains. The most common amphibole, hornblende, is usually black. The chemical formula is very complex with several solid solution opportunities and generally written as \(\ce{(Ca,Na)2(Mg,Fe,Al)5(Al,Si)8O22(OH)2}\).

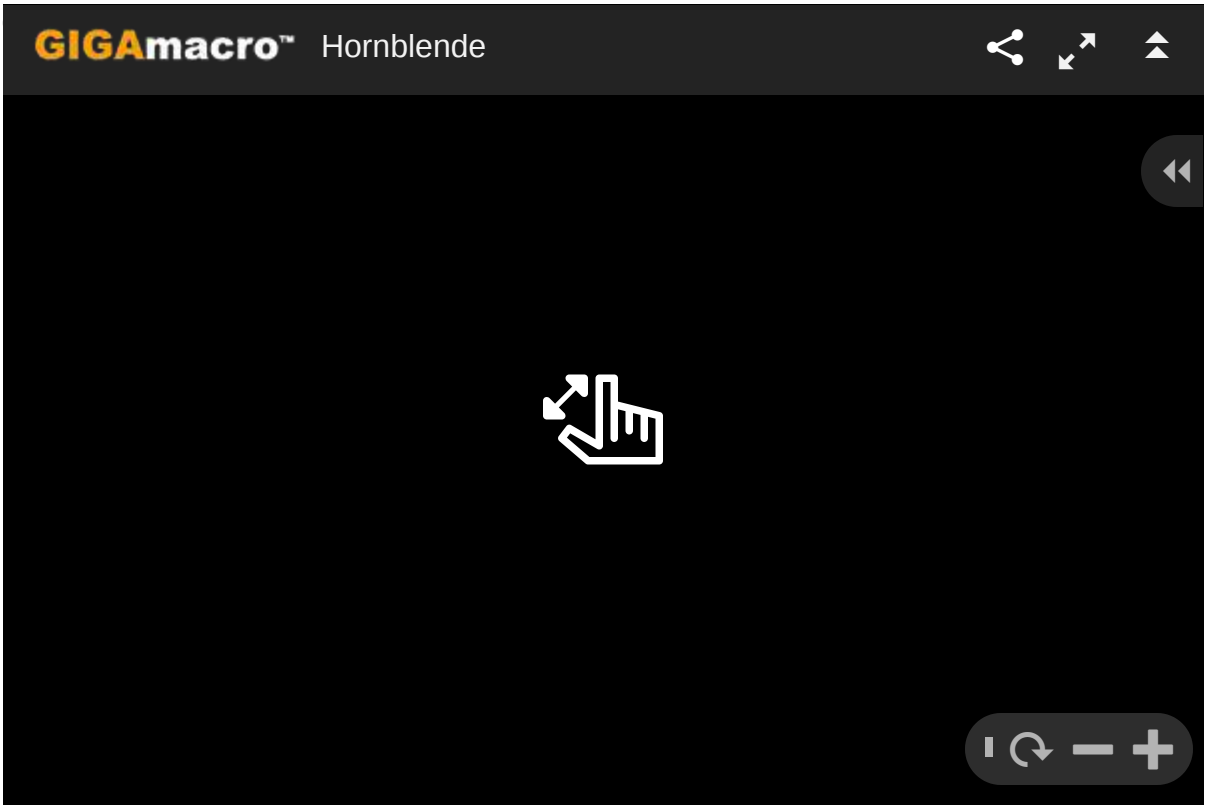


Figure \(\PageIndex{3}\): The iron- (Fe) and magnesium- (Mg) rich silicate mineral hornblende is a member of the pyroxene family. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

The Sheet Silicates

Biotite and muscovite are varieties of mica. The silica tetrahedra in the micas are arranged in continuous sheets. Bonding of elements between sheets is relatively weak which allows these minerals to be split easily along the sheets (this refers to the mineral's characteristic pattern of breaking, or cleavage).



The difference between the two micas is that biotite will contain iron and/or magnesium while muscovite mica, the location of the iron/magnesium is replaced by potassium. Therefore, biotite is dark while muscovite is light in color.



Figure \(\PageIndex{4}\): The sheet silicate mineral biotite. The color indicates that Fe and/or Mg exists in the mineral's chemistry. Note that one thin sheet of dark colored biotite can be virtually colorless. The ability to separate into thin sheets is described as the mineral's cleavage. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

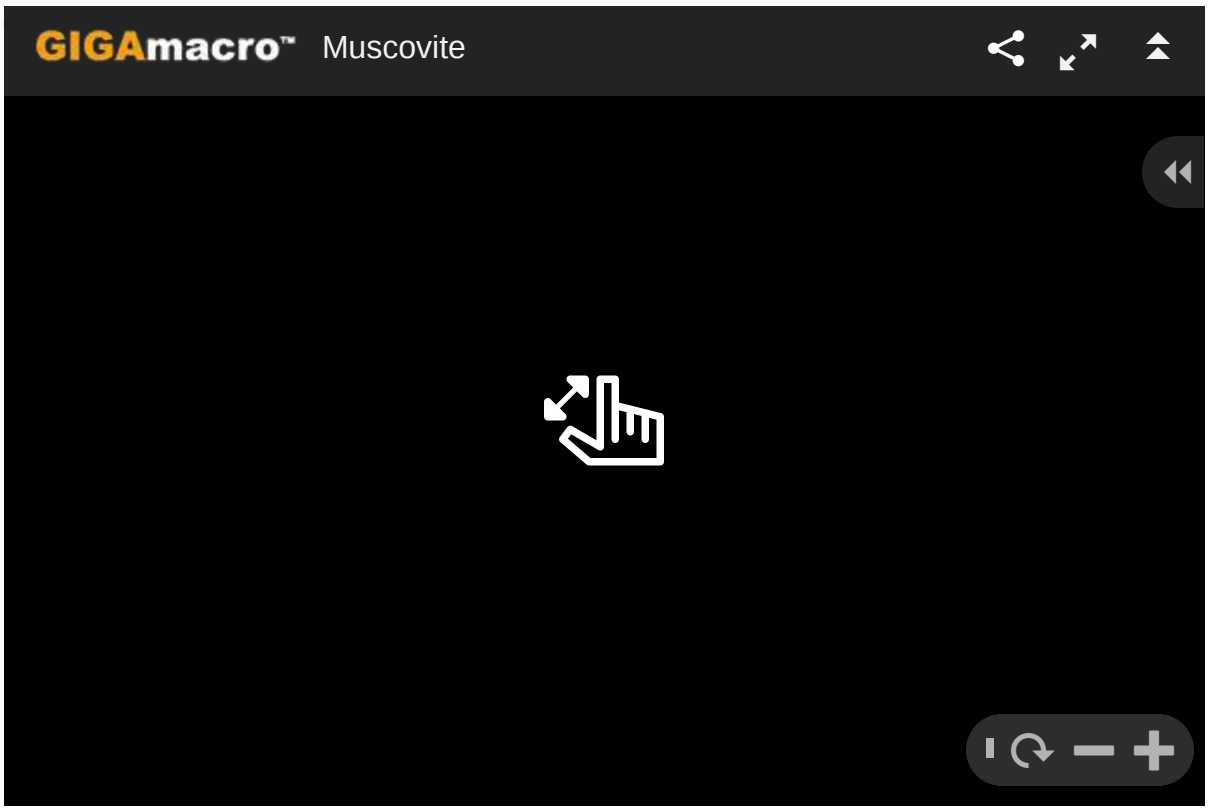


Figure \(\PageIndex{5}\): The sheet silicate mineral muscovite. This mineral belongs to the same family as biotite (mica minerals) however it does not contain any Fe/Mg therefore it is clear in color. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

The Framework Silicates

Quartz and feldspar are the two most abundant minerals in the continental crust. In fact, feldspar itself is the single most abundant mineral in the Earth's crust (see pie chart). The main feldspar minerals are potassium feldspar, (a.k.a. K-feldspar or K-spar) and the continuum of sodium- to calcium-rich plagioclase feldspars where albite is most sodium-rich and anorthite is most calcium-rich. Potassium feldspar and sodium-rich plagioclase feldspar are abundant in the rock of the continental crust while calcium-rich plagioclase feldspar is abundant in the rock of oceanic crust. Together with quartz, these minerals are classified as framework silicates because they are built with a three-dimensional framework of silica tetrahedra. Within the framework of the feldspar minerals are holes and spaces into which potassium, sodium, and calcium can fit giving rise to a variety of compositions.

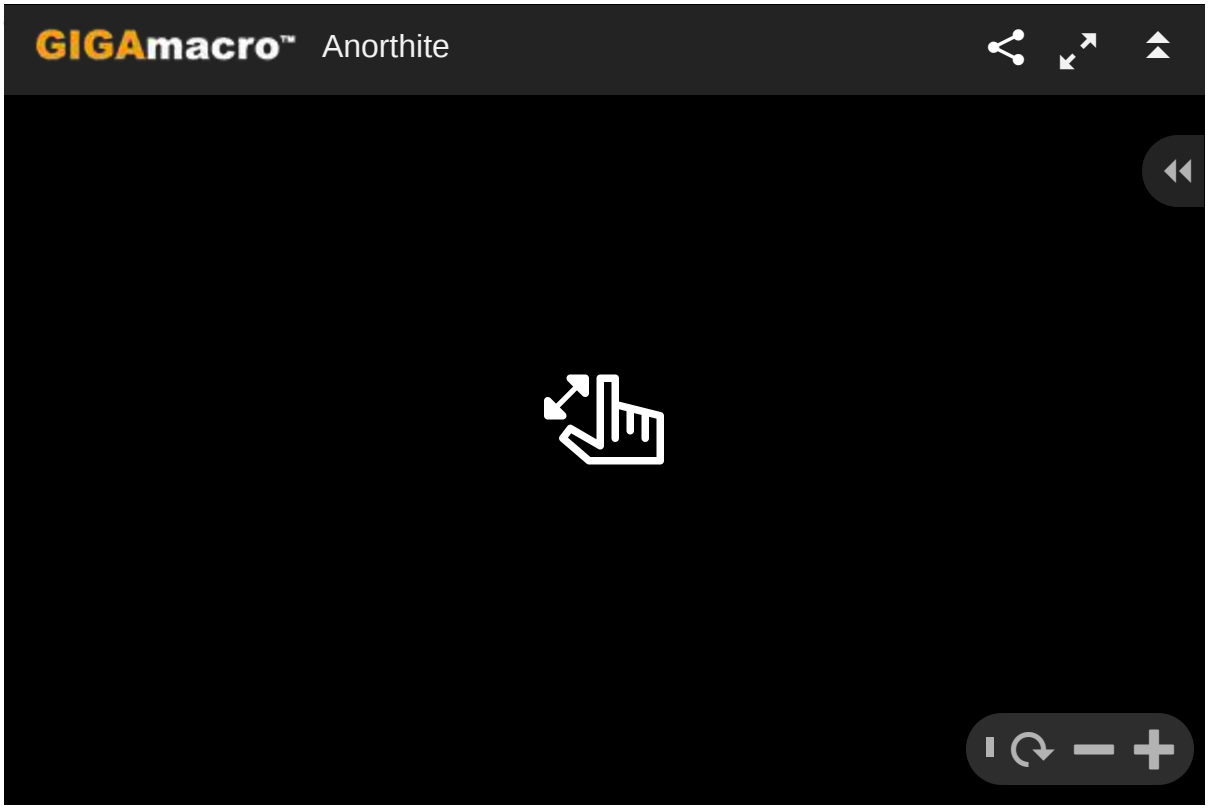


Figure \(\PageIndex{6}\): Calcium-rich plagioclase feldspar, a framework silicate mineral. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)



Figure \(\PageIndex{7}\): Sodium-rich plagioclase feldspar, a framework silicate mineral. Use your trackpad as a zoom. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)



Figure \(\PageIndex{8}\): Potassium-rich feldspar, a framework silicate mineral. Use your trackpad as a zoom. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Quartz is composed of pure silica, \(\ce{SiO2}\) with the tetrahedra arranged in a three dimensional framework. Quartz is the final mineral on Bowen's Reaction Series. It is the last mineral to crystallize from a silica rich magma. In quartz, the silica tetrahedra are bonded in a "perfect" three-dimensional framework. Pure quartz is composed entirely of \(\ce{SiO2}\) however, impurities consisting of atoms within this framework give rise to many varieties of quartz among which are gemstones like amethyst, rose quartz, and citrine.



Figure \(\PageIndex{9}\): Quartz, a framework silicate mineral composed of almost entirely \(\ce{SiO2}\). This is typically how we find quartz in common rocks, both clear and smoky grey in color. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

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4.7: Other Important Rock Forming Minerals

Calcite

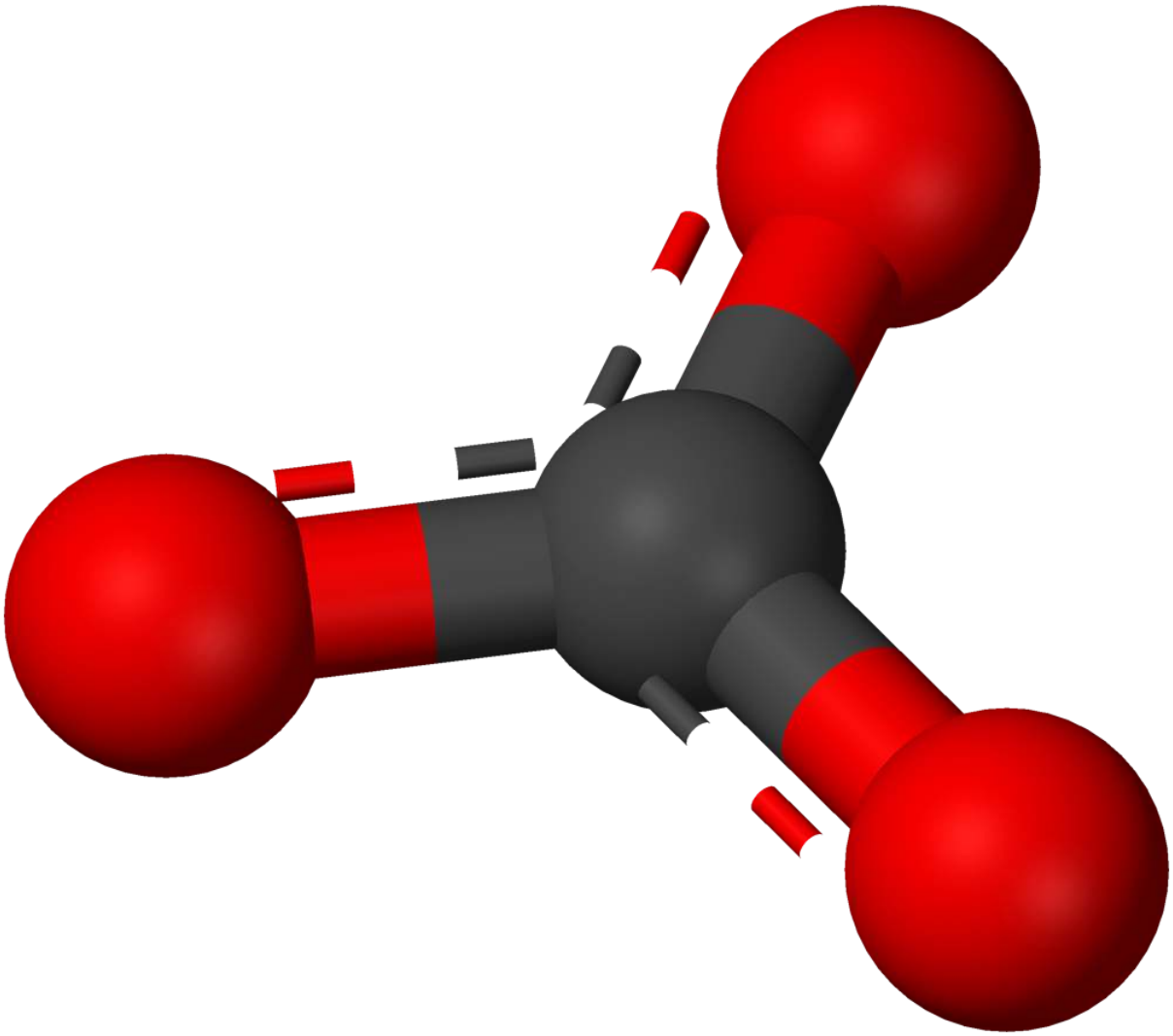


Figure 4.7.1: The complex carbonate ion $(CO_3)^{-2}$ where one carbon cation (brown, center) is surrounded by three oxygen anions (red) in a triangular pattern. This is the basic framework of "The Big Ten" mineral, calcite. (Public domain, CC0 1.0; Benjah-bmm27, Wikimedia)

We've made it through the formation of nine of "The Big Ten" minerals. At this point, you're probably wondering, "what about calcite? As the final 'Big Ten' mineral, how does calcite form?" Calcite formation is most typically associated with sedimentary processes. Let's take a look at calcite and some of the other "non-Big Ten" minerals that are important in Historical Geology.

Calcite is a "non-silicate" mineral (see the pie chart of the minerals of the crust). Calcite belongs to another family, or group of minerals, called the carbonates. The basic chemical and crystalline foundation for calcite and other minerals of the carbonate group is the complex carbonate ion $(CO_3)^{-2}$ where one carbon cation is surrounded by three oxygen anions in a triangular pattern.

Calcite forms by the ionic bonding of the complex carbonate ion with a calcium anion. The diagram above represents one unit cell of the mineral calcite.

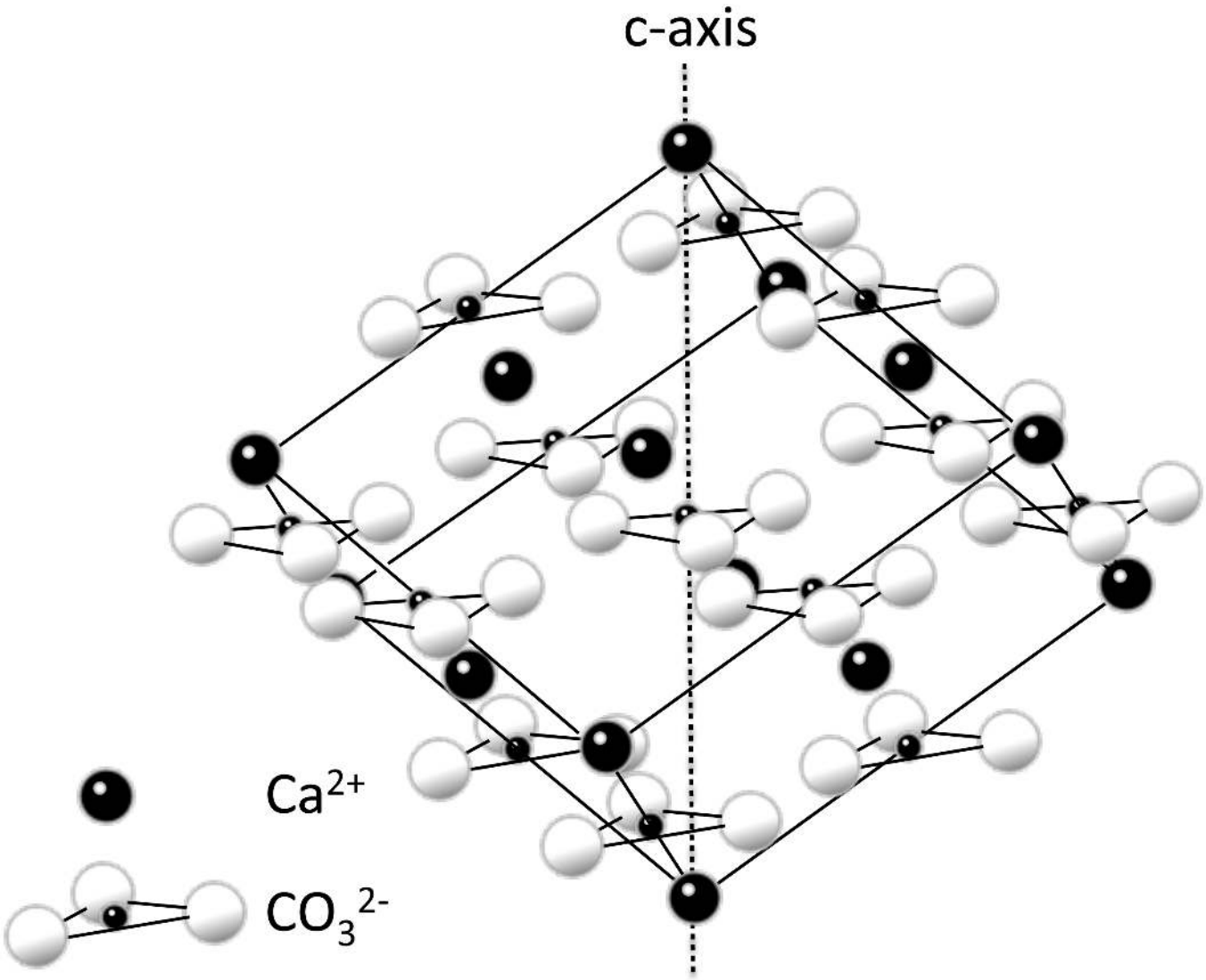


Figure 4.7.2: The unit cell of the mineral calcite, showing a rhombohedral crystal structure. (CC BY Attribution-Share Alike 4.0 International license; Temp4psu.)

Calcite can form by:

1. Biomineralization, where organisms like clams and corals extract calcium and carbonate ions from seawater or fresh water and then combine them inside their tissues to form their skeletal material.

2. Precipitation, where calcium and carbonate ions combine to precipitate micro-crystals of calcite from supersaturated seawater or circulating groundwater.

3. Evaporation, where inland lakes and epicontinental seas become isolated and slowly dry up leaving crystalline calcite plus other minerals.

4. In rare circumstances, calcite can occur through igneous processes.

One of the most common sedimentary rocks found on Earth is limestone. Most limestone is biogenic in origin, meaning, it is formed through the lithification of the skeletal remains of organisms. These organisms, both macroscopic and microscopic, have shells or exoskeletons composed of calcite. When these organisms die, their remains accumulate and their soft body parts are scavenged or decompose. The remaining calcite hard parts become trapped in layers of sediment and eventually are cemented together by additional calcite that precipitates in the pore spaces between the skeletal remains. The resulting sedimentary rock is

called fossil limestone. This accounts for the majority of calcite formation on Earth. Other processes mentioned above require special environmental conditions to allow for the formation of calcite.

Dolomite is another common carbonate mineral closely related to calcite often occurring in similar environments of formation, making it another important mineral in the study of Historical Geology. Dolomite has many similar characteristics to calcite including its chemical formula $CaMg(CO_3)_2$.

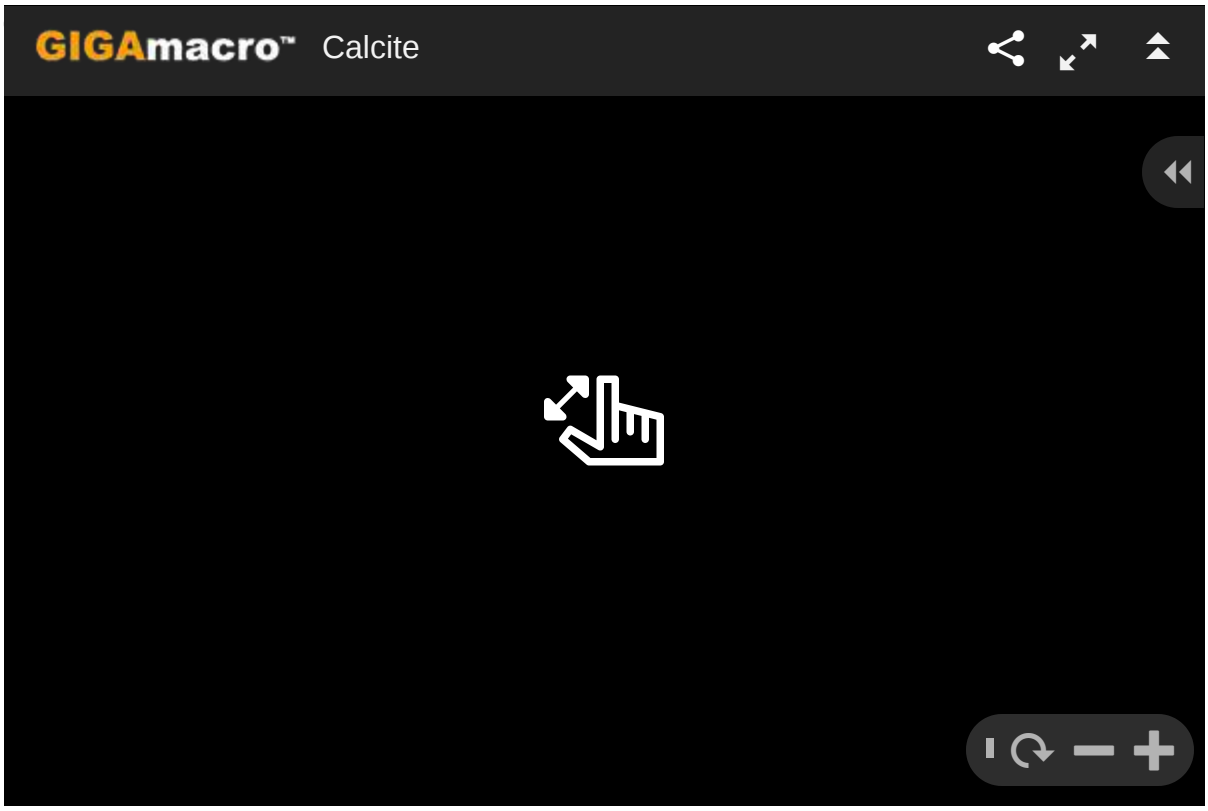


Figure 4.7.3: All of these samples are of the mineral calcite. Calcite is typically clear to white to yellow in color and displays a characteristic rhombohedral cleavage. All calcite will readily effervesce in dilute hydrochloric acid. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Clay Minerals

You might be surprised to learn that clay is a mineral – and not just one mineral – but a whole group of minerals. Clay minerals are also silicate minerals (see previous pie chart) that form through the processes involved with weathering of pre-existing silicate minerals in the presence of weak acid and water. As rain forms in the atmosphere, it combines with carbon dioxide to form a weak acid called carbonic acid with the chemical formula, H_2CO_3. This weak acid rains down on Earth's surface and immediately attacks minerals exposed at the surface. The pre-existing minerals become altered in their chemical composition through a reaction called hydrolysis. Hydrolysis is a type of chemical weathering that forms a wide range of clay minerals.

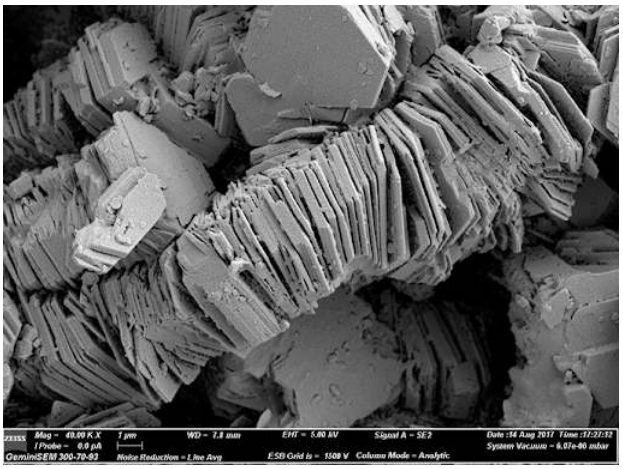


Figure 4.7.4: Scanning electron micrograph of the small sheets of the clay mineral kaolinite. Note the scale is in microns (μm). A micron is 1/1000 of a millimeter! Just to appreciate that scale, a human hair is approximately 75 microns in width! (CC Attribution 3.0; ACEMAC Nano Scale Electron Microscopy and Analysis Facility, University of Aberdeen by GSoil)

Clay minerals are sheet silicates similar to micas. Their chemical formula is dependent upon the minerals from which they were derived however all clay minerals contain a great deal of aluminum and silica plus varying amounts of water. Individual clay crystals tend to be very small, and often cannot be identified without a microscope or other equipment such as a scanning electron microscope (see photo below). Clays are important to the study of Historical Geology because they are the main components of fine-grained sediment (mud) and form another very common sedimentary rock, shale. The following is a hydrolysis reaction that occurs when silicate minerals encounter carbonic acid to produce clay and other ions:

$$\text{Silicate mineral} + \text{carbonic acid } (H_2CO_3) + H_2O \rightarrow \text{clay} + \text{metal cations } (Fe^{+2},\ Mg^{+2},\ Ca^{+2},\ Na^{+},\ etc.) + \text{bicarbonate anions } (HCO_3)^{-1} + \text{silica } (SiO_2)$$

Other Rock-Forming Silicates

The silicate minerals that fall into the group of “other” on the pie chart above (link to it) are largely formed through metamorphic processes. Metamorphism most commonly occurs deep within the Earth’s crust where rocks are buried and experience intense changes in temperature and pressure during periods of mountain building (orogeny). Some of the pre-existing minerals found within these deeply buried rocks can change in physical form and chemical composition as they recrystallize in response to these altered conditions.

One common metamorphic mineral that most students are familiar with is garnet. You might be surprised to learn that garnet is not just one mineral but also a part of a larger group. The most common garnet minerals are dark red in color (like the birthstone) and have the general chemical formula of:

$$(Ca, Mg, Fe)Al_2(SiO_4)_3$$

Garnets are largely formed through the metamorphism of clay minerals which are rich in aluminum and silica. These clay minerals would be found in the common sedimentary rock shale that is subjected to varying degrees of pressure and temperature changes during the collision of tectonic plates and the formation of mountains.

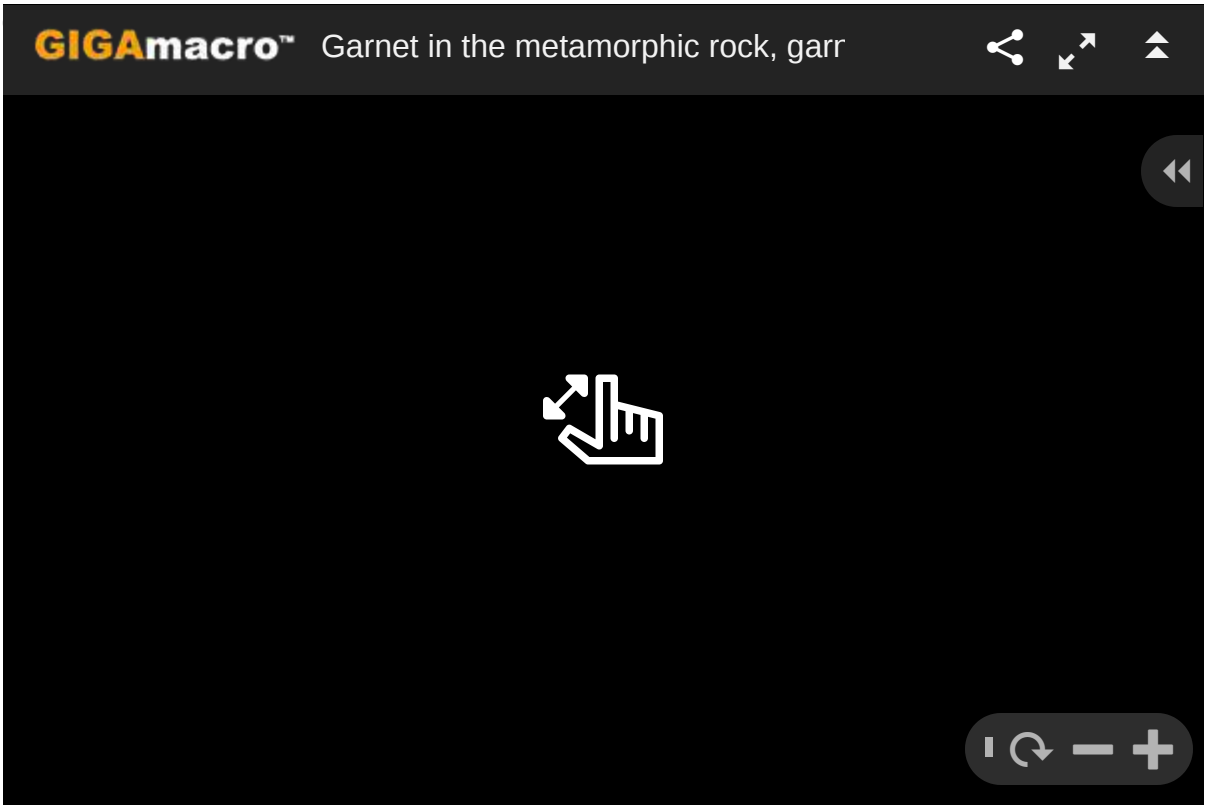


Figure 4.7.5: Very large garnet crystals (porphyroblasts) formed by changes in temperature and pressure deep within the crust during mountain building accompanied by regional metamorphism. From Wrangell, Alaska. (CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Table 4.7.1: Minerals Important to Historical Geology ("Big Ten" in bold)

Mineral Groups	Common Rock Forming Mineral
Silicates	**Olivine** **Pyroxene (Augite)** **Amphibole (Hornblende)** **Biotite** **Muscovite** **Calcium-rich plagioclase feldspar (anorthite)** **Sodium-rich plagioclase feldspar (albite)** **Potassium-rich Feldspar (orthoclase)** **Quartz** Clay Garnet
Carbonates	**Calcite**, Dolomite
Oxides	Hematite, Magnetite
Halides	Halite
Sulfides	Pyrite, Chalcopyrite, Galena, Sphalerite
Sulphates	Gypsum
Phosphates	Apatite
Native Elements	Gold, silver, copper

Oxides, Halides, and Sulfides

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

Oxides

Oxides consist of metal ions covalently bonded with oxygen. Iron oxides are important for producing iron ore deposits from which we extract metallic iron. The formation of Earth's major iron ore deposits are extremely important in studying Earth's history. Prior to about two billion years ago (2 Ga), the early oceans were saturated with soluble iron due to the lack of free oxygen. So much iron existed in the water that our early oceans were thought to have been green in color from the dissolved iron. As oxygen levels slowly increased in the atmosphere and dissolved in the oceans, the oxygen immediately bonded with the iron forming the common ore minerals of hematite (Fe_2O_3) and magnetite (Fe_3O_4). (See the case study on the Great Oxidation Event).

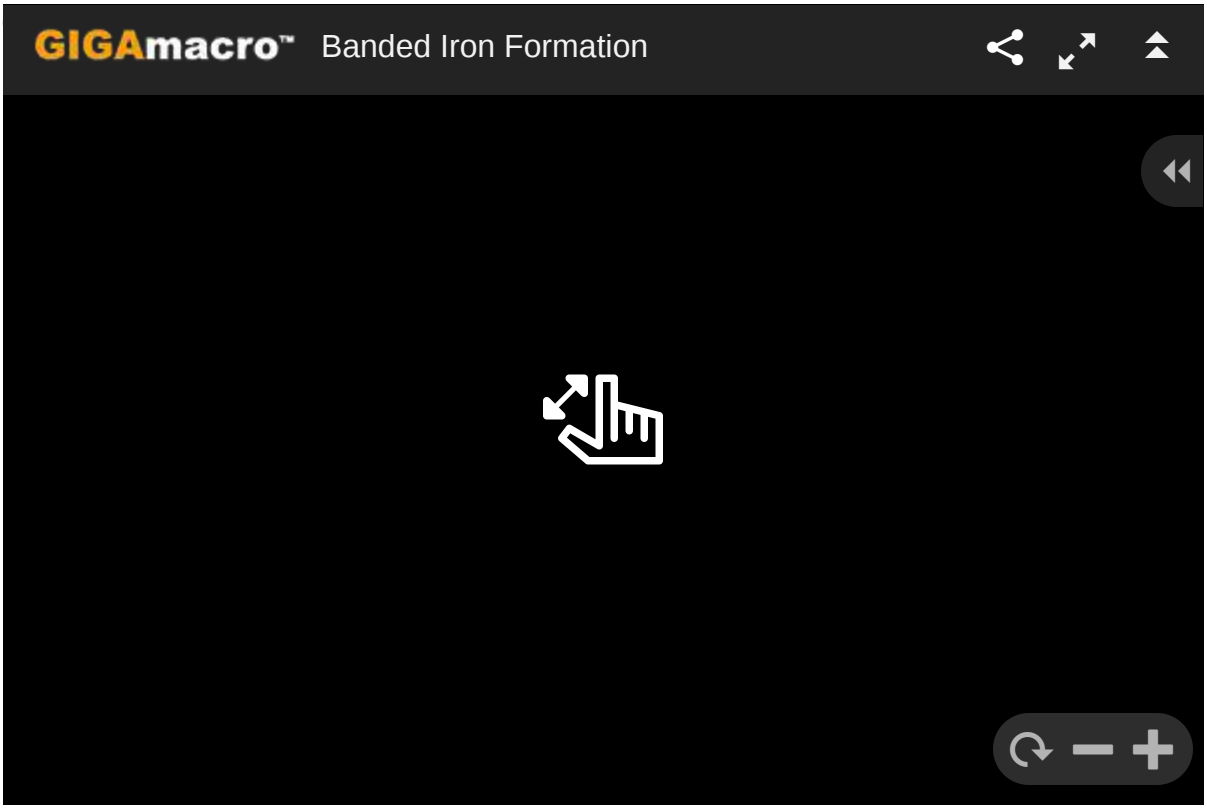


Figure 4.7.6: Banded iron formation. The greyish color is the iron oxide mineral, hematite, the red is the sedimentary rock, chert (when red, it is called jasper), and the gold is tiger's eye. The sample is 26 cm across. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

The most familiar oxide of iron is rust, which is a combination of iron oxides (Fe_2O_3) and hydrated oxides. Hydrated oxides form when iron is exposed to oxygen and water. The red color that we often see in soil and rock is usually due to the presence of iron oxides. For example, the red sandstone cliffs in Zion National Park and throughout Southern Utah consist of white or colorless grains of quartz coated with iron oxide which serves as cement, holding the grains together.



Figure 4.7.7: The "Three Patriarchs" in Zion National Park, Utah. (Public domain, CC0 1.0; Daniel Mayer, Wikimedia)

Halides

The **halide minerals**, also important to the study of historical geology contain the element chlorine ionically bonded with sodium or other cations. These minerals include halite or sodium chloride (NaCl; common table salt) and sylvite or potassium chloride (KCl), commonly used in cooking as a salt substitute.



Figure 4.7.8: The mineral halite which can be processed into common table salt. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Halide minerals are extremely important to the study of Earth's history because their occurrence indicates a period of evaporation of sea water or other isolated bodies of water. A well-known example of modern halide mineral deposits created by evaporation is the Bonneville Salt Flats, located west of the Great Salt Lake in Utah (need a good pic). Watch the following video on extraction of rock salt from the Cayuga Salt Mine in central New York state. The halite mineral of the rock salt represents evaporation of a great inland sea that existed some 385 million years ago in the Great Lakes region.

Watch the following video where at 2300 feet underground, the Cayuga Mine in Lansing, NY, processes approximately 10,000 tons of road salt daily that is shipped to more than 1,500 locations throughout the northeast United States.



Sulfides

Many important metal ores are **sulfides**, in which metals are bonded to sulfur. Significant examples include: pyrite (iron sulfide, sometimes called “fool’s gold”), and chalcopyrite (iron-copper sulfide), galena (lead sulfide), sphalerite (zinc sulfide). Sulfides are well known for being important ore minerals. For example, galena is the main source of lead, sphalerite is the main source of zinc, and chalcopyrite is the main copper ore mineral.

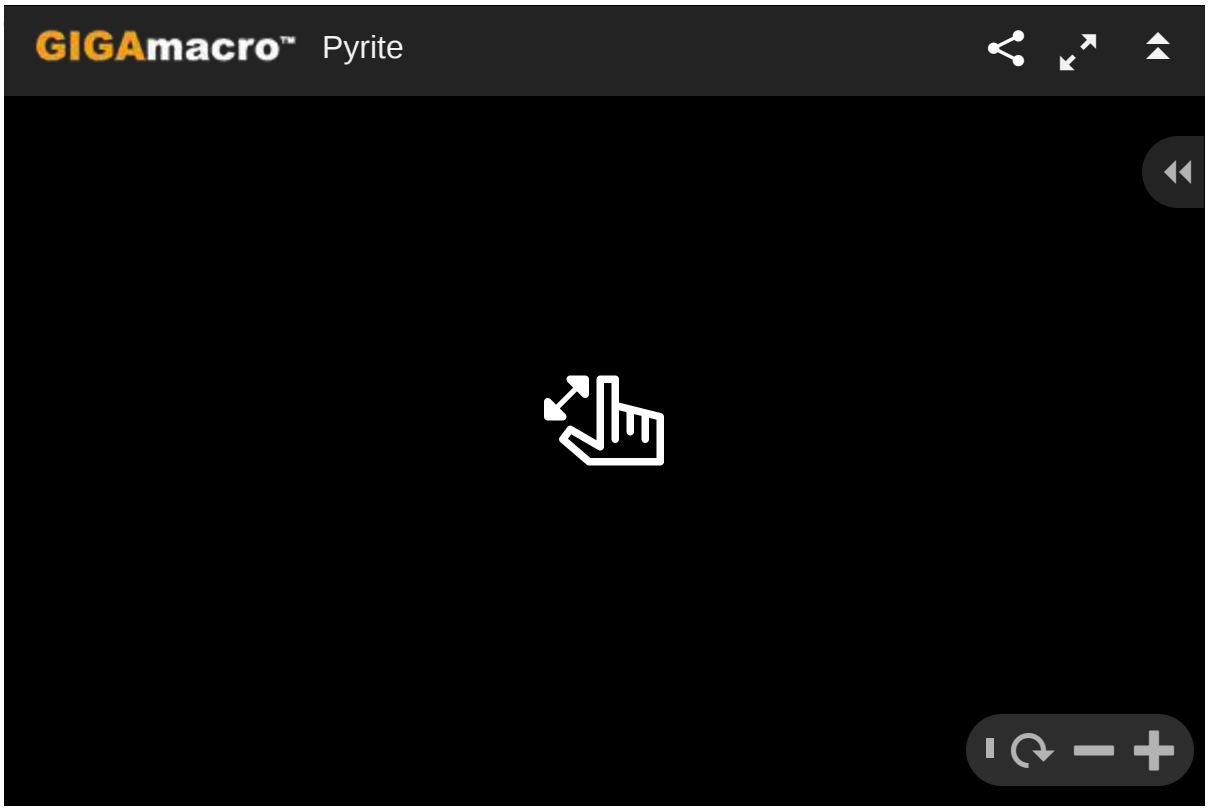


Figure 4.7.9: Common iron sulfide mineral often referred to as “fool’s gold.” (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

The iron sulfide mineral pyrite is of particular importance to Historical Geology. Pyrite is unstable in the presence of oxygen and will readily chemically decompose to form stable iron oxide. However, pyrite occurs in some of Earth’s early (older than 2.0 Ga) sedimentary rocks. Sediments are formed from the weathering of pre-existing rock exposed on Earth’s surface. Pyrite, existing as sedimentary clasts, or pieces, within these ancient rocks, indicates that surface conditions must have been lacking oxygen to allow the pyrite to remain unaltered and ultimately preserved.



Figure 4.7.10: Zoom in close to find the pyrite in this sample of Archean age (greater than 2.5 billion year old) metamorphosed conglomerate. The large grey-white clasts are composed of quartz. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Sulfate Minerals

Sulfate minerals contain a metal ion, such as calcium, bonded to a sulfate ion. The sulfate ion is a combination of sulfur and oxygen $(SO_4)^{-2}$. The sulfate mineral gypsum $(CaSO_4 \cdot 2H_2O)$ is important to Earth's history because it indicates formation from evaporating water and usually contains water molecules in its crystalline structure.



Figure 4.7.11: A variety of the mineral gypsum called alabaster. Gypsum is a common sulfate mineral that is so soft, you can scratch it with your fingernail. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Phosphate Minerals

Phosphate minerals contain the complex cation $(PO_4)^{-2}$ combined with various anions and cations. The best known and important phosphate mineral to the study of historical geology is apatite, $Ca_5(PO_4)_3(F, Cl, OH)$. Variations of this mineral are found in teeth and bones.

Native Elements

Native element minerals, usually metals, occur in nature in a pure or nearly pure state. Gold, silver and copper are examples of common native element minerals. Carbon is often found as a native element in minerals such as graphite and diamond.



Figure 4.7.12: The mineral graphite on the left and diamond on the right. Both minerals are composed of carbon. Graphite is extremely soft and rubs off on your fingers. Diamond is the hardest mineral known to exist. Their difference is in crystal structure. They are known as “polymorphs” of carbon: same composition, different internal atomic arrangement of carbon atoms. (Creative Commons Attribution-Share Alike 3.0 license; Robert Lavinsky, Wikimedia.)

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CHAPTER OVERVIEW

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5.4: Sediments and Sedimentary Rocks

5.5: Classification of Sedimentary Rocks

5.6: Metamorphism and Metamorphic Rocks

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5.1: The Rock Cycle

The Rock Cycle

The rock components of the crust are slowly but constantly being changed from one form to another and the processes involved are summarized in the rock cycle. The rock cycle is driven by two forces: (1) Earth's internal heat engine, which moves material around in the core and the mantle and leads to slow but significant changes within the crust, and (2) the Sun which powers the hydrological cycle, moving water, wind and air along Earth's surface.

The rock cycle is still active on Earth because our core is hot enough to keep the mantle moving, our atmosphere is thick enough to support circulation, and we have liquid water. On some other planets or their satellites, such as the Moon, the rock cycle is virtually dead because the core is no longer hot enough to drive mantle convection and there is no atmosphere or liquid water.

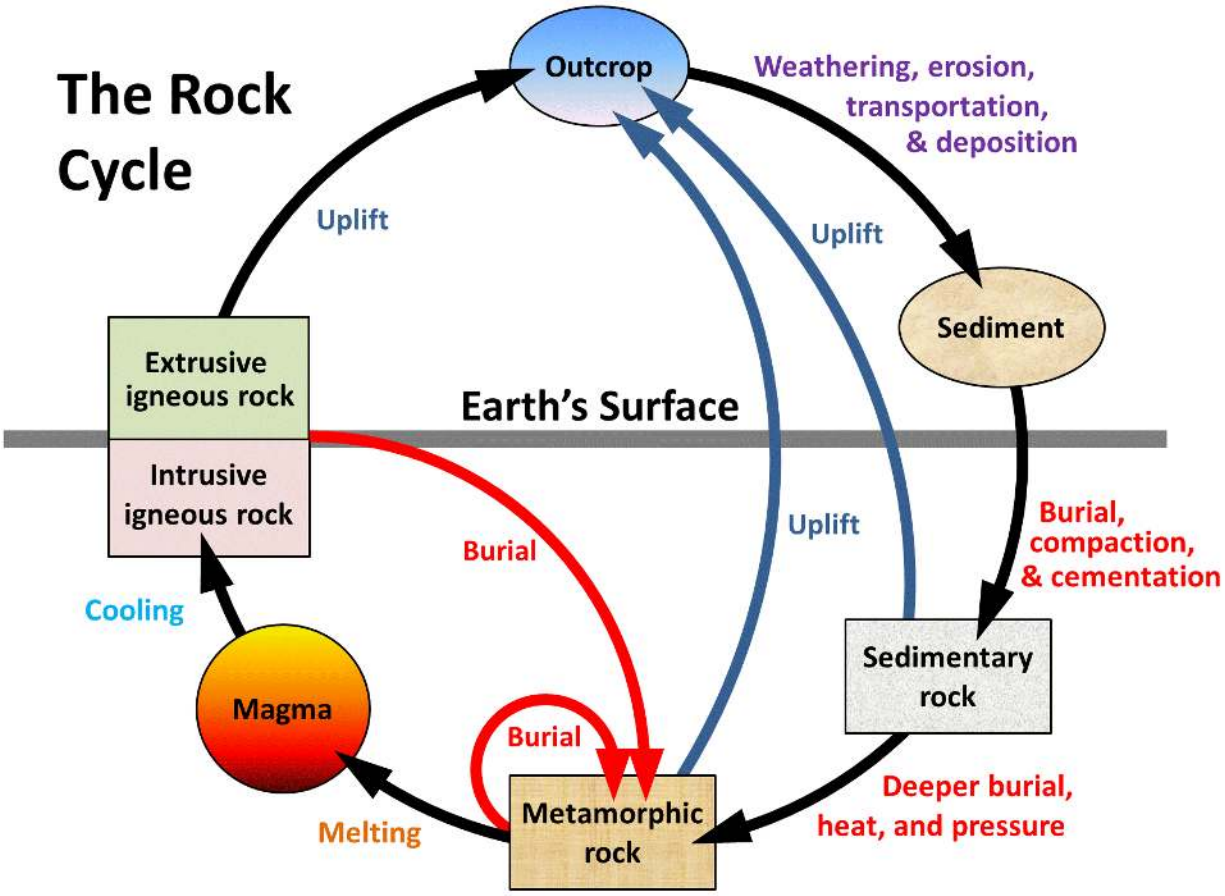


Figure 5.1.1: The Rock Cycle. (Creative Commons Attribution 4.0 International License; Steven Earle)

In describing the rock cycle, we can start anywhere we like, although it's convenient to start with magma because from magma, new minerals and igneous rock are created. Magma is rock that has been heated to the point of being entirely molten. Magma can either cool slowly within the crust (over centuries to millions of years)—forming intrusive igneous rock, or erupt onto the surface, where it is called lava, and cool quickly (within seconds to years)—forming extrusive igneous rock (volcanic rock). Intrusive igneous rock typically crystallizes at depths of hundreds of meters to tens of kilometers below the surface.

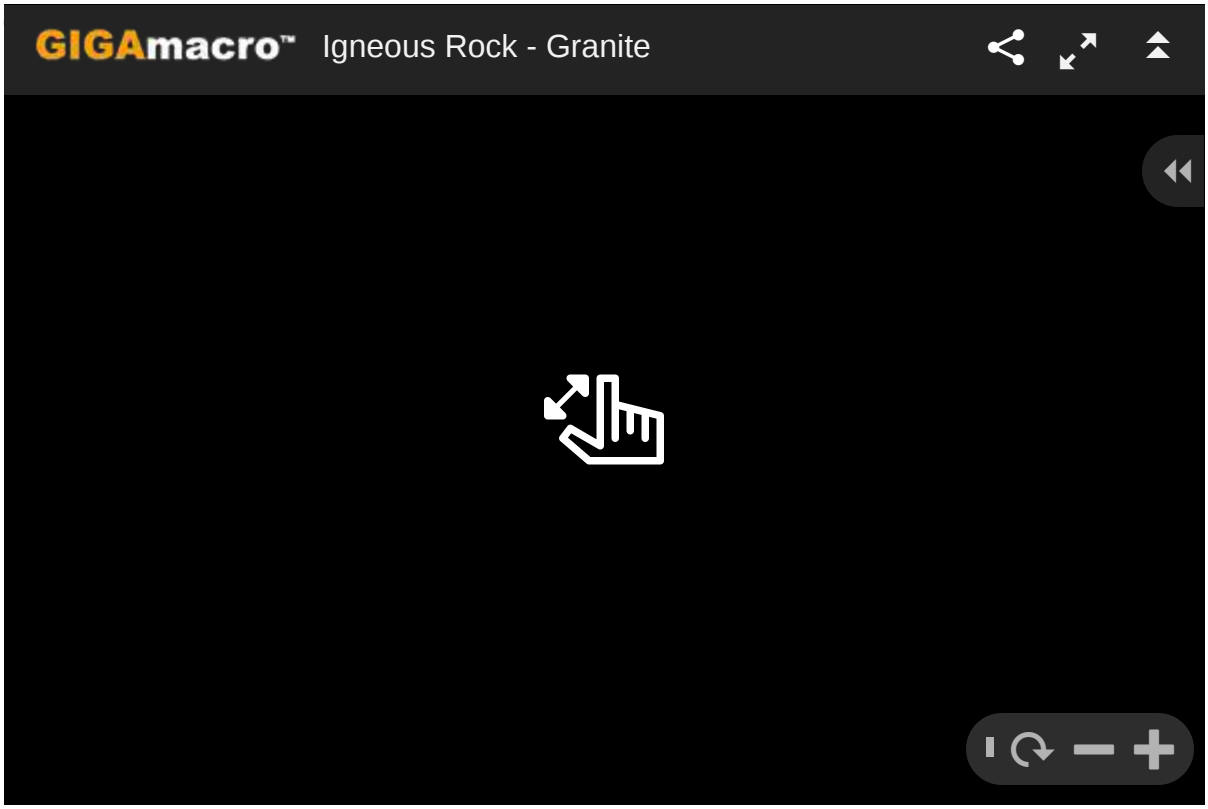


Figure 5.1.2: Granite, an example of an igneous rock. Zoom in close to notice the interlocking fabric of minerals crystals of different size and shape. See if you can identify the common minerals: quartz, sodium-rich plagioclase, hornblende and biotite. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Plate tectonic processes of mountain building can allow for all types of rock to be uplifted and exposed at the surface. These include rock that is buried deep within the continents and even rock that is part of the crust of the deep ocean basins. Once rock of any type is exposed at the surface, it is attacked by the weathering agents of water, wind, ice and gravity. Rock is weathered, both physically (breaking into smaller pieces) and chemically (by alteration of the minerals). These weathering products—mostly small rock and mineral fragments—are eroded, transported, and then deposited as sediments. Transportation and deposition occur through the action of glaciers, streams, waves, wind, and other agents, and sediments are deposited in rivers, lakes, deserts, and the ocean. Sediments will eventually become buried by more sediments in some type of an accumulating basin. At depths of hundreds of meters or more, they become compressed and cemented into sedimentary rock.

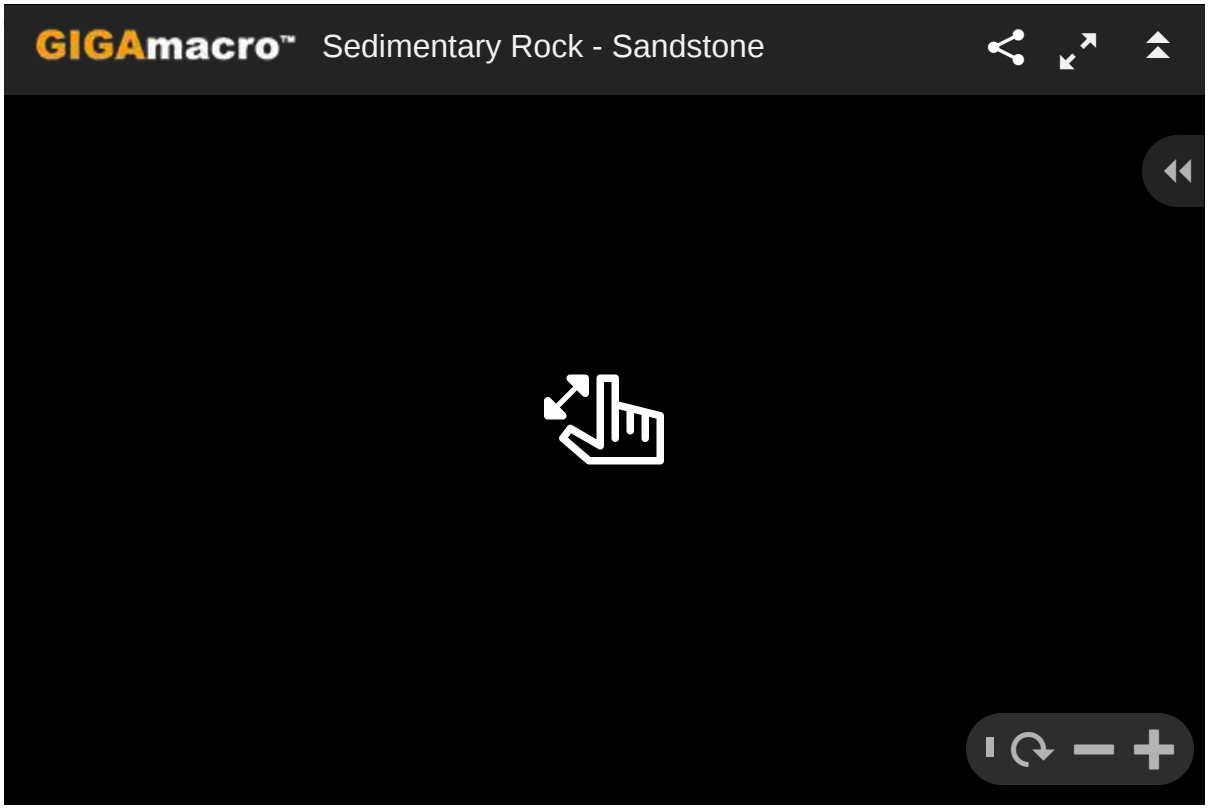


Figure 5.1.3: Quartz sandstone, an example of a sedimentary rock. Zoom in close and notice the rounded quartz grains that are cemented together by white silica cement. The reddish color is the result of chemical weathering of minor iron minerals. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Any rock type that becomes buried deep within the crust is subjected to increasing heat and confining pressure. Tectonic forces of colliding plates can add additional stress to the buried rock. These changing conditions affect the stability of the pre-existing minerals and the fabric, or texture, of the original rock. This results in the formation of metamorphic rock.

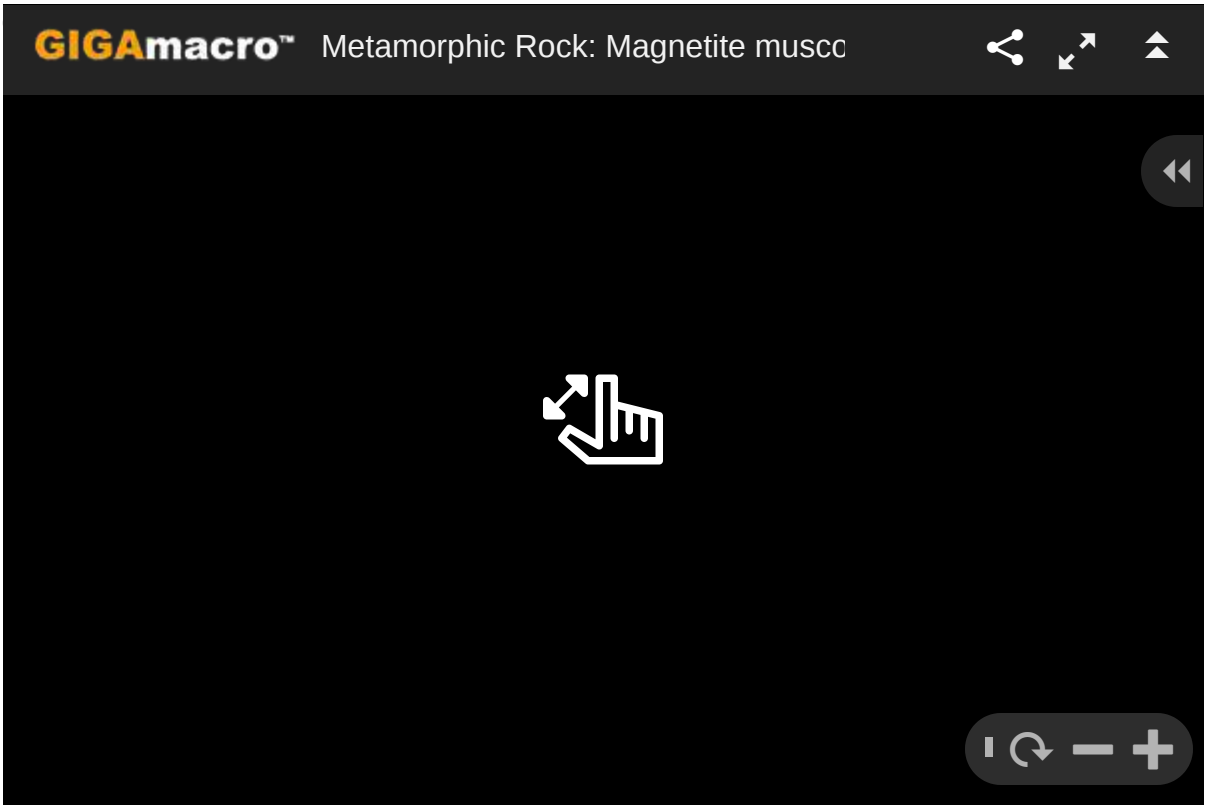


Figure 5.1.4: Mica schist, an example of a metamorphic rock. Tiny muscovite mica minerals give this sample a “glittery” appearance. This metamorphic rock began its life as a clay rich sedimentary rock, like shale. Increased temperature and pressure conditions caused by mountain building deep within the crust caused the metamorphism. The black crystals are the iron-oxide mineral, magnetite. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Did I Get It? - Quiz

Exercise 5.1.1

Which two forces drive the rock cycle?

a. Earth's internal heat and the biological cycle

b. The Sun and gravity

c. Earth's internal heat and the Sun

d. The hydrological cycle and volcanoes

Answer

c. Earth's internal heat and the Sun

Exercise 5.1.2

When describing the processes of the rock cycle, you must begin with igneous rocks.

a. No, the rock cycle processes are continuous and you can begin anywhere.

b. Yes, the rock cycle begins with the formation of igneous rocks.

Answer

a. No, the rock cycle processes are continuous and you can begin anywhere.

Exercise 5.1.3

Agents of the weathering process include

a. Water, temperature, wind and gravity

b. Ice, water, pressure and wind.

c. Gravity, water, wind and ice.

Answer

c. Gravity, water, wind and ice.

Exercise 5.1.4

Metamorphic conditions of change include

a. chemical and physical weathering

b. erosion and transportation

c. pressure and temperature

Answer

c. pressure and temperature

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5.2: Igneous Rock



Figure 5.2.1: Extrusive igneous rock. Basalt lava flow on Old Chain of Craters Road, Hawaii. (With permission by: Garry Hayes and Susan Hayes)

The temperature of magma can range between about 800° and 1600 °C depending on where within the mantle the magma forms. Magma can either cool slowly within the crust (over centuries to millions of years)—forming ***intrusive*** igneous rock (*also called* ***plutonic*** *rock*), or erupt onto the surface and cool quickly (within seconds to years)—forming ***extrusive*** igneous rock (*also called* ***volcanic*** *rock*). Intrusive igneous rock typically crystallizes at depths of hundreds of meters to tens of kilometers below the surface.

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5.3: Igneous Rock Classification

Igneous rocks are simply classified on two characteristics: composition and texture. The composition of igneous rocks is based on the minerals present viewed either with the naked eye or microscopically. For the classification based on composition, we return to Bowen's Reaction Series. As we learned previously, the crystallization sequence of silicate minerals forming from a cooling magma follows the pattern outlined in Bowen's Reaction Series.

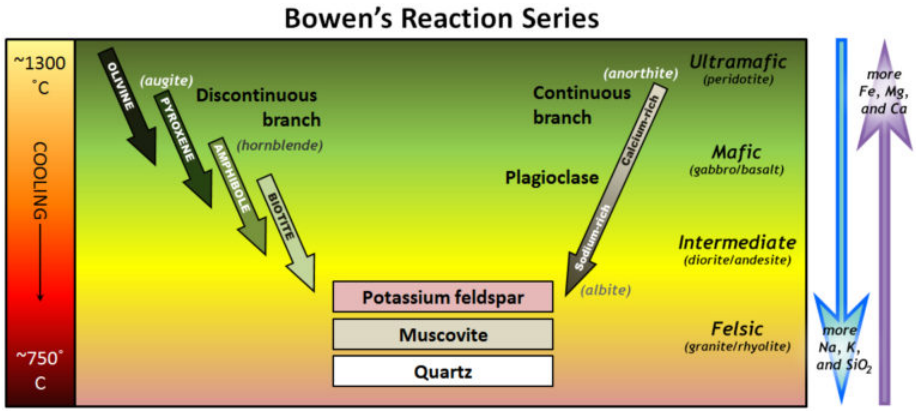


Figure 5.3.1: Bowen's Reaction Series. Developed through laboratory experimentation by Norman L. Bowen in the early 1900's. It establishes the order of crystallization of minerals from a silicate magma. (Creative Commons Attribution 4.0 International License; Modified after an original by Steven Earle)

Igneous Rock Composition

Igneous rock composition is divided into four categories: **ultramafic**, **mafic**, **intermediate** and **felsic**. These categories refer to differing amounts of silica, potassium, sodium, calcium, iron, and magnesium found in the minerals that make up the rocks. Mafic refers to the overall composition being high in concentrations of **ma**gnesium (Mg) and iron (**F**e). The minerals that compose mafic rocks are referred to as the ferromagnesian minerals, meaning, enriched in Fe and Mg. Felsic refers to the overall composition being enriched in the **fel**dspar minerals of sodium-rich plagioclase and potassium feldspar plus **si**lica (SiO_2) in the form of quartz.

It is important to realize these groups do not have sharp boundaries in nature, but rather lie on a continuous spectrum with many transitional compositions and names that refer to specific quantities of minerals. On the Bowen's Reaction Series diagram, we read these igneous rock compositions as horizontal strips that run from right to left. For example, mafic igneous rock will typically contain calcium-rich plagioclase plus varying amounts of the ferromagnesian minerals: olivine, pyroxene (augite) and amphibole (hornblende) minerals. Felsic igneous rock will typically contain sodium-rich plagioclase, biotite, potassium feldspar, muscovite and quartz.

Using the Bowen's Reaction Series diagram alone does not allow us to classify igneous rocks beyond their general composition. To determine their specific classification, we use an enhanced diagram where composition is quantitatively determined by the percentage of each of Bowen's minerals that occur in any given sample. Examine the top portion of the diagram below. In essence, this is Bowen's Reaction Series Diagram rotated 90° to the right. Instead of ultramafic composition appearing along the top, as it does in Bowen's Reaction Series, it exists along the right side of the diagram. The same minerals that we saw in Bowen's Reaction Series that correspond to a rock of ultramafic composition, now exist in the far right column; they include a small percentage of calcium-rich plagioclase, and larger percentages of both pyroxene (augite), and olivine. On this diagram, Bowen's Reaction Series sequence of mineral formation reads from right to left. This diagram helps us understand that rocks of any compositional category (ultramafic, mafic, intermediate and felsic) occur with varying percentages of Bowen's Reaction Series minerals and though the compositional categories are separated by dashed lines, we can see that there are no hard and fast boundaries between each category.

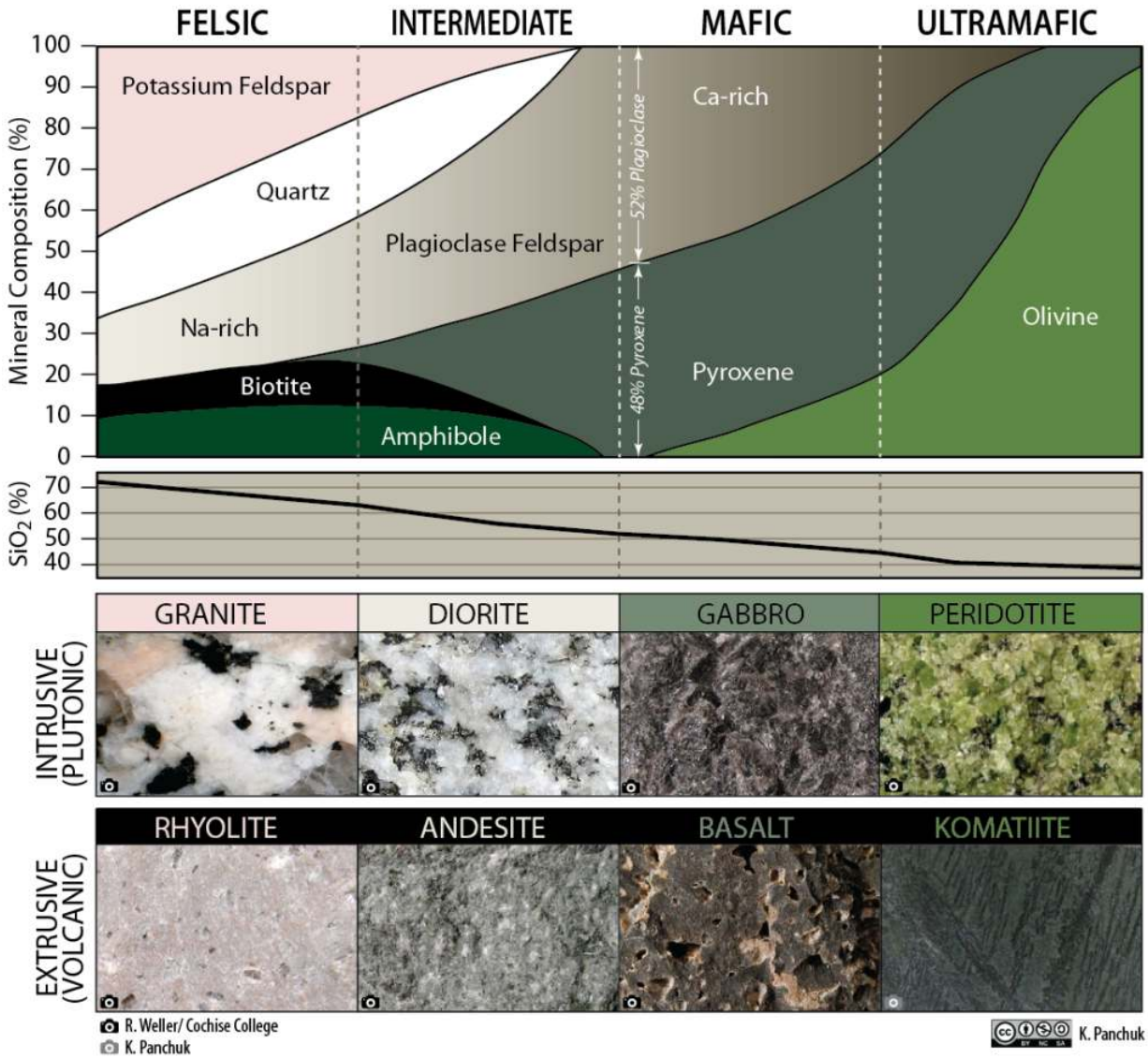


Figure 5.3.2: Igneous Rock Classification Diagram. The diagram is read in columns from the composition at the top to the example textural types at the bottom.

Ultramafic refers to the extremely mafic rocks composed of mostly olivine with lesser percentages of pyroxene and calcium-rich plagioclase. Rock of this composition is rarely found on Earth's surface; Where outcrops do occur, they represent rock derived from the upper mantle. This rock is very high in iron and magnesium and poor in silica, in the 40% or less range.

Mafic refers to igneous compositions with an abundance of dark ferromagnesian minerals of olivine and augite plus gray calcium-rich plagioclase feldspar. The oceanic crust is primarily mafic rock. Mafic rock is low in silica, in the 45-50% range.

Intermediate is a composition between felsic and mafic. This rock type contains roughly equal amounts of light and dark minerals, including light grains of plagioclase feldspar and dark minerals like hornblende. Rock of this composition is commonly associated with convergent plate boundary volcanoes. The silica content in rock of intermediate composition is in the 55-60% range.

Felsic refers to a composition with a majority of the light-colored minerals feldspar and quartz. Minor amounts of dark-colored (ferromagnesian) minerals like amphibole and biotite mica may be present as well. Rock of this composition dominates the continental crust. Felsic igneous rock is rich in silica (in the 65-75% range) and poor in iron and magnesium.

Igneous Rock Texture

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

readily visible to the unaided eye.

When lava is extruded onto the surface, or intruded into shallow fissures near the surface and cools, the resulting igneous rock is called extrusive or volcanic. Extrusive igneous rocks have a ***fine-grained*** or ***aphanitic*** texture, in which the grains are too small to see with the unaided eye. The fine-grained texture indicates the quickly cooling lava did not have time to grow large crystals. These tiny crystals can be viewed using a special petrographic microscope used for viewing and magnifying very thin slices of rock (see photos below). In some cases, extrusive lava cools so rapidly it does not develop crystals at all. This non-crystalline material is volcanic glass which is a common component of volcanic ash and rocks like obsidian.

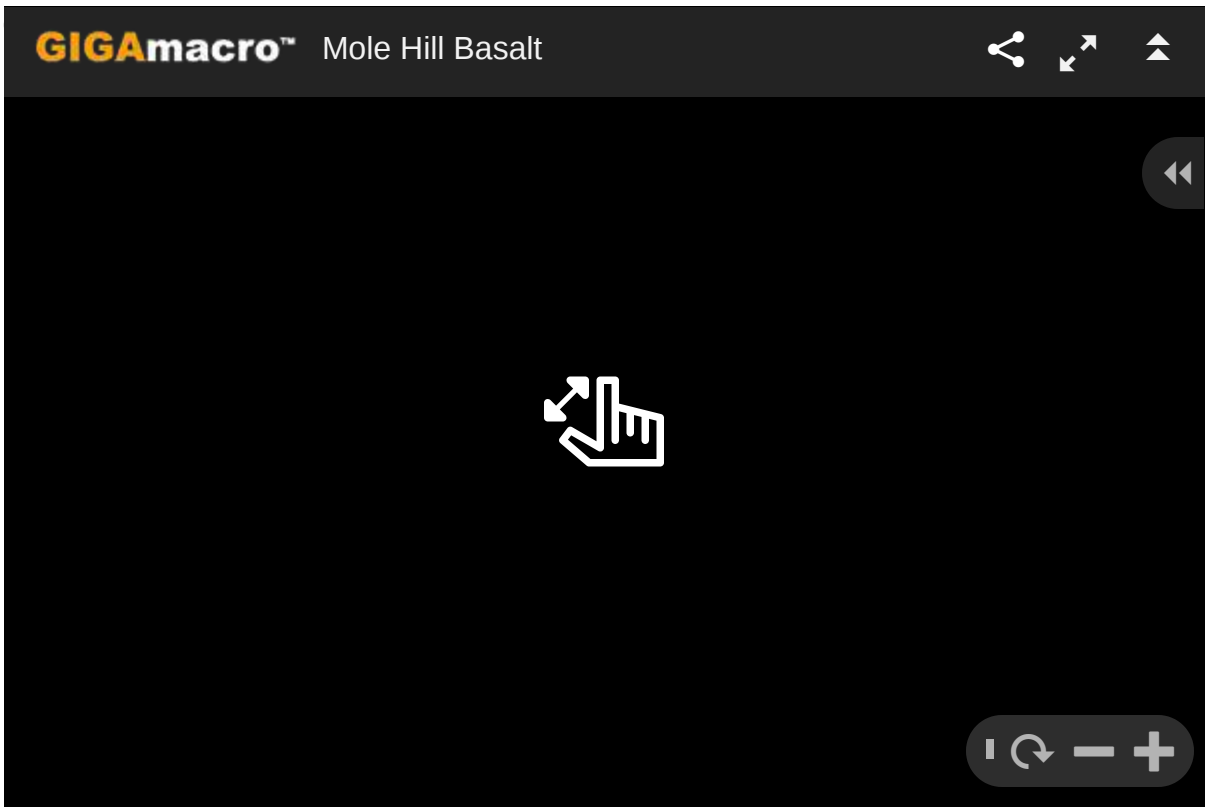
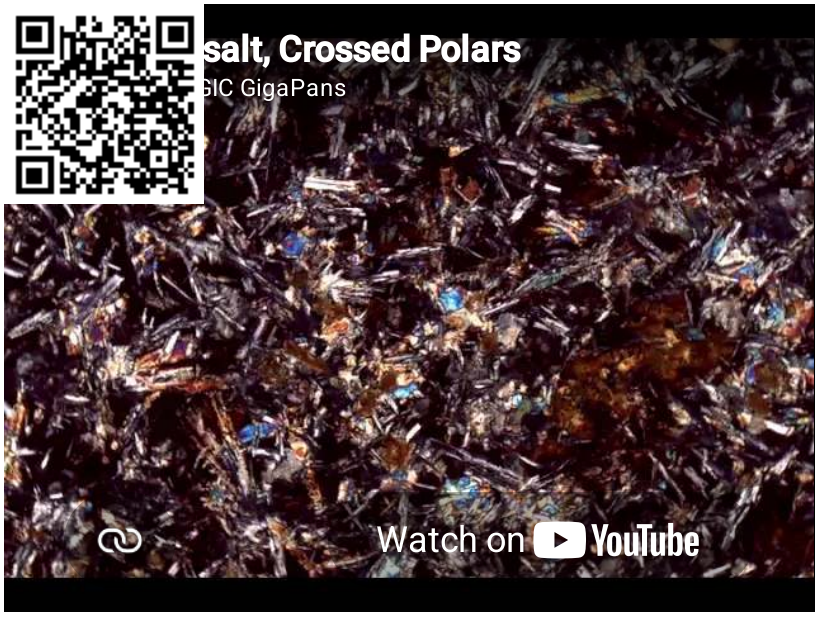


Figure 5.3.3: Hand sample of basalt. Note the aphanitic texture with no visible mineral grains. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)



This is a magnified view of a 30 micron thick slice of a sample of basalt rock using a petrographic microscope. All of the light gray, elongate crystals are plagioclase. The very bright blues and greens are olivine. The golden brown crystals are pyroxene (augite). (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Once composition and texture are determined, the classification diagram can be used to give an igneous rock a name. The Igneous Rock Classification diagram (above) is read in columns. Let's take a look at the "FELSIC" column on the left side of the diagram. As we move down the column we can see that felsic igneous rock will contain the following minerals in varying percentages: potassium feldspar, quartz, sodium-rich plagioclase, biotite and amphibole. The silica content (SiO_2 %) will range between 65-75%. If it is an intrusive rock (plutonic) rock, it will have coarse, visible grains and it is given the name **granite**. If it is an extrusive rock (volcanic) rock with no visible (or few, very small) mineral grains, it is given the name **rhyolite**. **Diorite** and **andesite** likewise refer to intrusive and extrusive intermediate rock. **Gabbro** and **basalt** are the intrusive and extrusive names for mafic igneous rock. **Peridotite** is ultramafic and intrusive and **komatiite** as the fine-grained extrusive equivalent. Komatiite has been a rare rock because for most of Earth history, volcanic material that comes directly from the mantle is not common.



Figure 5.3.4: Left: **Granite** characteristics: Felsic composition, phaneritic texture, intrusive, plutonic. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.) *Right:* **Rhyolite**. Felsic composition, aphanitic texture, extrusive, volcanic. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)



Figure 5.3.5: Left: **Diorite**. Intermediate composition, phaneritic texture, intrusive, plutonic. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.) *Right:* **Andesite**. Intermediate composition, aphanitic texture, extrusive, volcanic. Special note: This texture is more correctly "porphyritic." The large crystals formed early in the magma chamber prior to the volcanic eruption. The remaining rock cooled very quickly, preserving the early formed plagioclase crystals. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)



Figure 5.3.6: Left: **Gabbro**. Mafic composition, phaneritic texture, intrusive, plutonic. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.) *Right:* **Basalt**. Mafic composition, aphanitic texture, extrusive, volcanic. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)



Figure 5.3.7: Left: **Peridotite**. Ultramafic composition, aphanitic texture, intrusive, plutonic. Special note: This sample is called a xenolith. It is a piece of the mantle that was ripped away by magma making its way to the surface through a volcanic eruption. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.) *Right:* **Komatiite**. Ultramafic composition, aphanitic texture, extrusive, volcanic. (CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Did I Get It?

? Exercise 5.3.1

Which statement is true for igneous rocks?

a. intrusive = volcanic

b. extrusive = intrusive

c. extrusive = volcanic

d. volcanic = plutonic

Answer

c. extrusive = volcanic

? Exercise 5.3.2

Igneous rock is classified based on which two characteristics?

a. color and luster

b. hardness and density

c. composition and texture

Answer

c. composition and texture

Exercise 5.3.3

Which of the igneous rock compositions would contain the most silica (silicon and oxygen)?

a. Intermediate

b. Ultramafic

c. Mafic

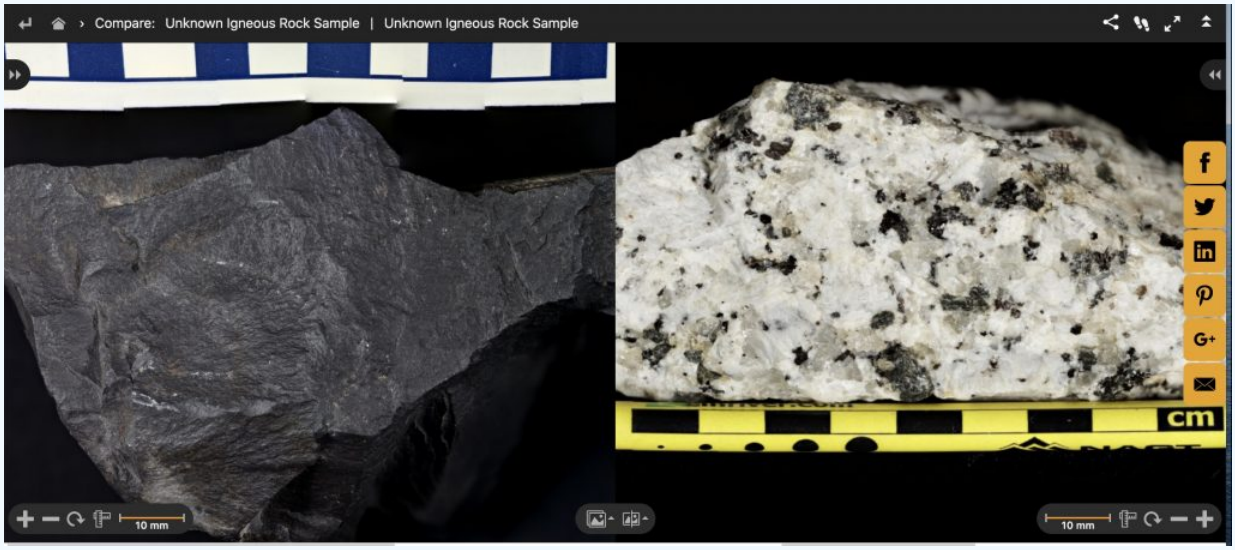
d. Felsic

Answer

d. Felsic

Exercise 5.3.4

Which of the two samples of igneous rock in the photo would be classified as felsic in composition?



a. Neither sample is felsic in composition.

b. Both samples are felsic in composition.

c. The sample on the right is felsic in composition.

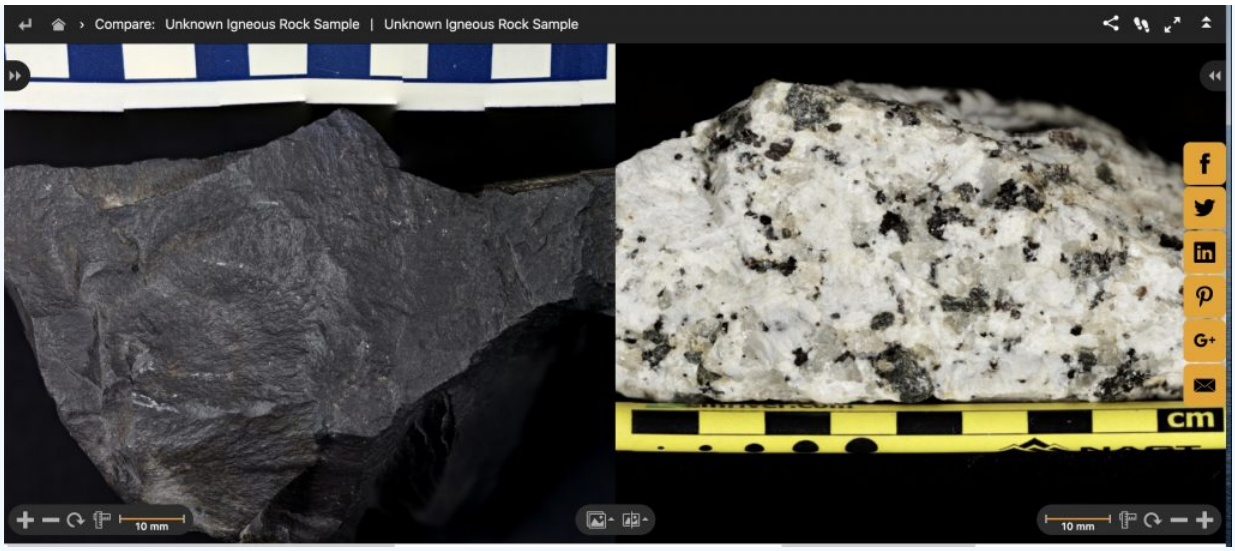
d. The sample on the left is felsic in composition.

Answer

c. The sample on the right is felsic in composition.

Exercise 5.3.5

Which of the two samples of igneous rock in the photo would have formed extrusively?



a. Neither sample formed extrusively.

b. Both samples formed extrusively.

c. The sample on the right formed extrusively.

d. The sample on the left formed extrusively.

Answer

d. The sample on the left formed extrusively.

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5.4: Sediments and Sedimentary Rocks

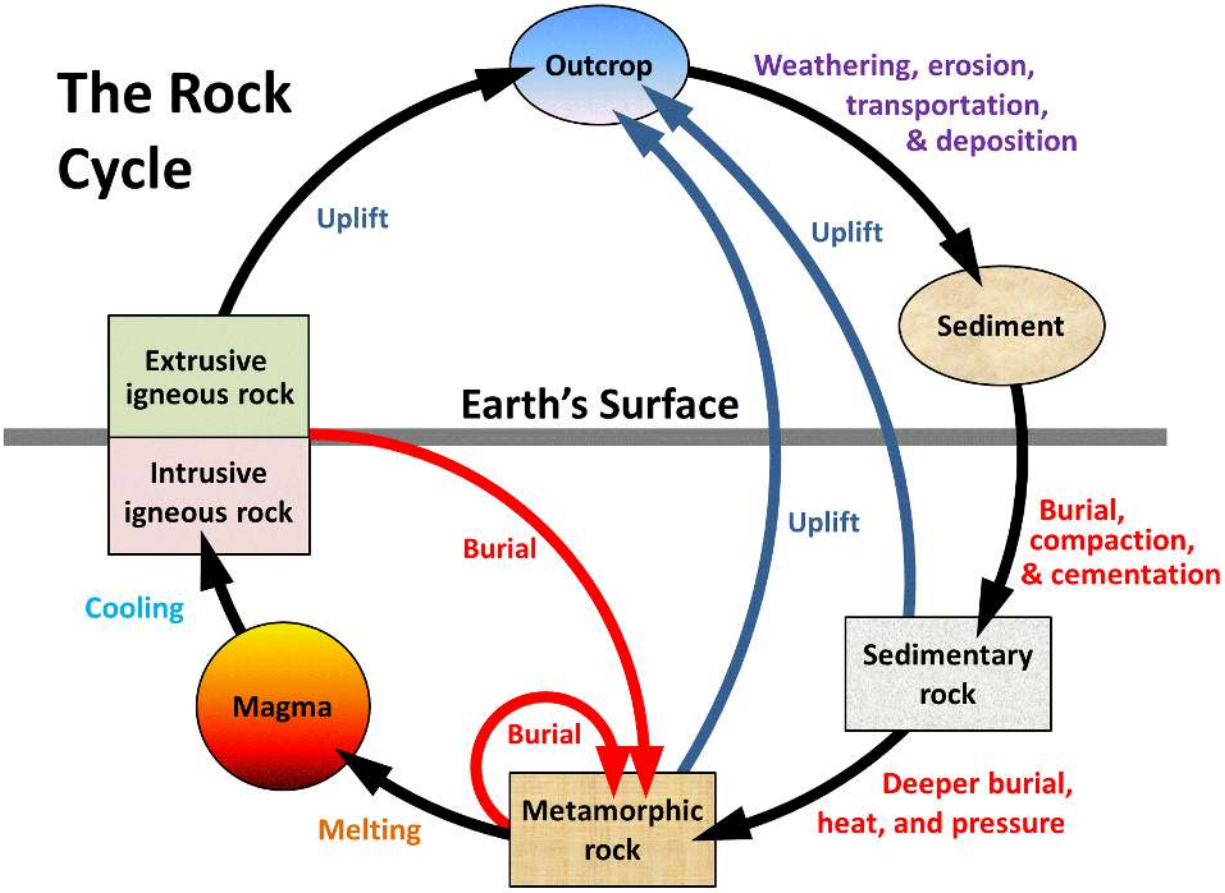


Figure 5.4.1: The Rock Cycle (Creative Commons Attribution 4.0 International License; Steven Earle, https://opentextbc.ca/physicalgeology2ed/.)

Sedimentary rock formation is a cycle unto itself and is discussed in more detail elsewhere in this text. The flow through that cycle appears below.

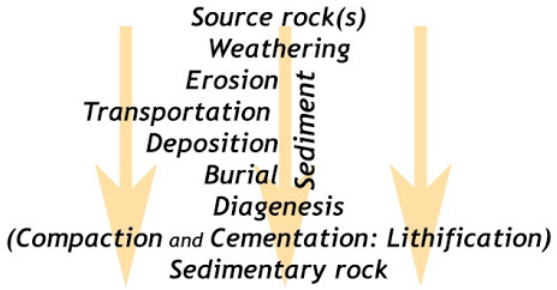


Figure 5.4.2: The steps in the Sedimentary Rock Cycle. (Callan Bentley (2020))



Figure 5.4.3: Boulder of igneous rock from Iceland displaying the effects of physical weathering. (Bridget Wade)

The cycle begins with the exposure of pre-existing rock to the “agents” of weathering and erosion. Weathering involves exposure of pre-existing rock to Earth’s surface weather conditions. These “agents” include wind, water and ice — plus gravity. Erosion is the removal of weathered rock material from its original location. The effects of weathering on pre-existing rock can be both physical and chemical.

Physical Weathering

Physical weathering is simply the process of physical disintegration: the breaking of solid rock into smaller pieces. The “agents” assist in this process which results in the exposure of an ever increasing number of surfaces to additional weathering.

Chemical Weathering



Figure 5.4.4: Chemically weathered surface (red-brown) of an igneous diabase (mafic, shallow intrusive) from Virginia. The chipped surface (black) shows the original color of the unweathered diabase. (Callan Bentley)

Chemical weathering produces changes in the chemistry of minerals in the pre-existing rock. These changes may result in new mineral formation through the addition or subtraction of elements from the original mineral structure or, in the further disintegration of the pre-existing minerals by chemical decomposition. The dominant agent involved in both physical and chemical weathering is water. Water acts as an agent of change, both introducing elements that can interfere with the chemistry of the exposed minerals and also acting as a flux, physically removing weathered rock material and ions lost through decomposition.

Weathering produces particles of sediment (of various sizes) plus ions in water. The sediment is transported to some setting where it gets deposited. This depositional location may be very close to the origin of the sediment (such as a rock fall off a cliff face) or many hundreds (or even several thousand) of kilometers away (for example the beach along the continental shoreline). Following deposition, the sediment is buried, compacted, and cemented together to form solid rock. We call these processes “diagenesis,” and collectively they serve to lithify the sediment into a rock.

The ions in solution (weathered chemical elements) can be:

- used in cementing particles sediment.
- extracted from the water to support life, both biochemical processes (in plants or animals) and in the formation of skeletal material, such as shells.
- precipitated from the water through supersaturation triggered by evaporation or by changing physical conditions of the surrounding water.

The importance of these processes to Historical Geology are discussed in much more detail elsewhere in this text. The simple classification of sedimentary rock is presented below.

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5.5: Classification of Sedimentary Rocks

Clastic Sedimentary Rocks (Detrital)

Clastic sedimentary rock (also called 'detrital' sedimentary rock) consists of sediment pieces (clasts/rock detritus) that come from weathered bedrock, cemented together to make a new rock. The clastic sediment is often referred to as "siliciclastic" because it is derived from the rocks of the crust which, as we have learned, is largely composed of silicate minerals of igneous rock. Clastic (or detrital) rock is classified and named based on the size and composition of the clasts. Their various sizes, from boulders to clay, is referred to as the sediment grain size. The classification and description of the various clastic rocks appears in the top section of the chart below. The size, shape, and composition of the clasts varies widely based upon the source rock and the length of time the sediment has been moving through the system.

The classification and description of the various clastic sedimentary rock types appears in the top section of the chart below. Photos of each rock type appear beneath the chart.

Chemical, Biochemical and Organic Sedimentary Rock

Chemical sedimentary rock is formed by processes that do not directly involve mechanical weathering and erosion. Chemical weathering contributes the dissolved ions in water that ultimately form the various rock types. Inorganic chemical sedimentary rock is made of minerals precipitated from ions dissolved in solution. Inorganic chemical sedimentary rock forms in environments where ion concentration, dissolved gasses, temperatures, or pressures are changing, which causes minerals to crystallize, such as through the process of evaporation.

Biochemical sedimentary rock is formed from shells and bodies of underwater organisms (such as coral). The living organisms extract chemical components from the water and use them to build shells and other body parts. Biochemical sedimentary rocks are commonly composed of calcite, from a wide variety of sea life, or silica, largely from the single-celled microorganisms called radiolarian.

Organic sedimentary rock contains the remains of organic material that has undergone the lithification process. The source materials are plant and animal remains that are transformed through burial and heat, and end up as coal, oil, and methane (natural gas).

The classification and description of the various chemically formed sedimentary rock types appears in the bottom section of the chart below. Photos of each rock type appear beneath the chart.

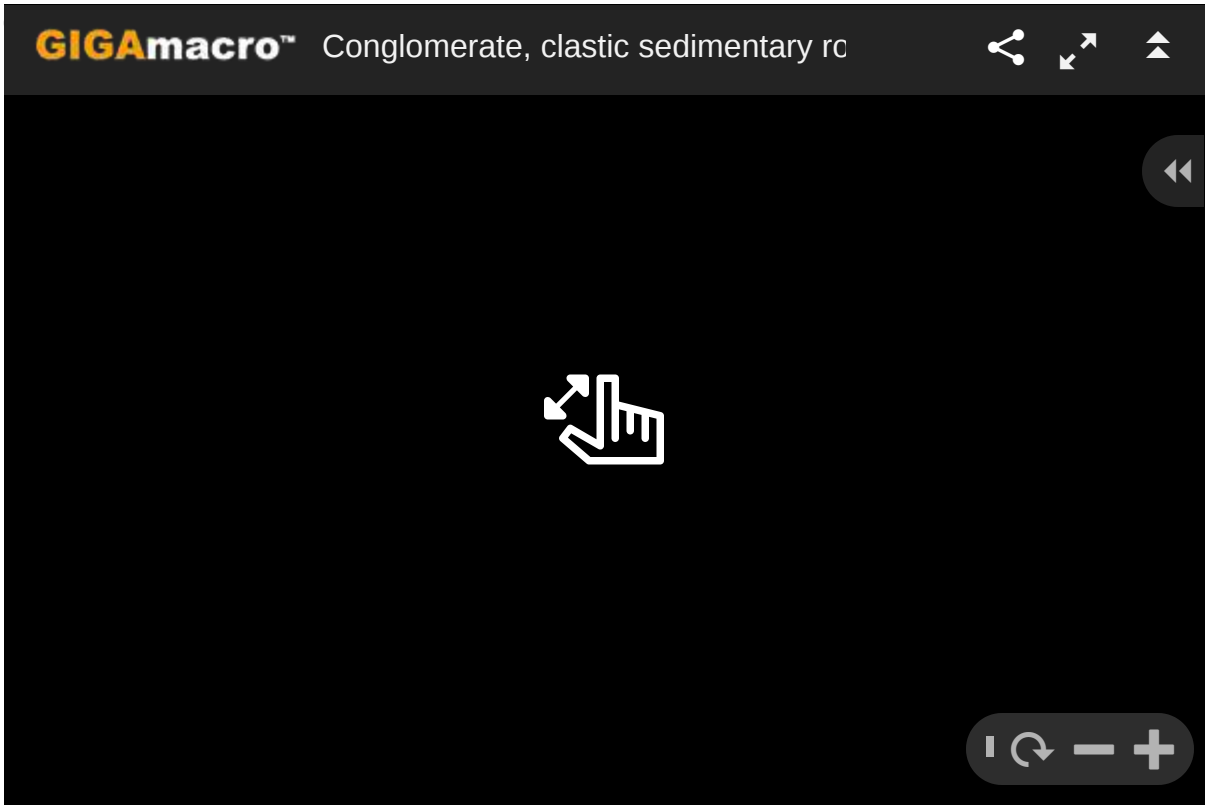
Table 5.5.1: Classification of *Sedimentary Rocks*

Step 1: Determine makeup	Step 2: Determine Grain Size	Step 3: Rock Description	Step 4: Rock Name
CLASTIC: Composed of pieces of rocks and minerals.	Granule (> 2 mm)	Rounded rock or mineral fragments; usually poorly sorted.	Conglomerate
		Angular rock or minerals fragments; usually poorly sorted.	Breccia
	Sand (0.06 mm to 2 mm)	Mostly quartz grains; sorting and rounding variable.	Quartz Sandstone
		Mostly quartz with at least 25% feldspar; rock fragments common; usually poorly sorted with angular grains.	Arkose
		Clay, quartz, feldspars, and rock fragments; usually poorly sorted with angular grains; often dark color.	Graywacke
	Silt (< 0.06 mm)	Silt-sized particles are too small to identify; no layering making it not fissile.	Siltstone
		Silt and clay-sized particles are too small to identify; layering makes rock break in planes making if fissile.	Shale
	Clay (< 0.004 mm)	Clay-sized particles are too small to see.	Claystone
Step 1: Determine makeup	**Step 2: Composition**	**Step 3: Rock Description**	**Step 4: Rock Name**
CHEMICAL, BIOCHEMICAL & ORGANIC: Made of minerals that have crystallized together, or biological fragments of shells of plants.	Calcite - All fizz with acid	Abundant fossils. Possibly in micrite.	Fossilliferous Limestone[1]
		Abundant ooids - coarse, sand. sized spheres with concentric internal layers	Oolitic Limestone
		Microcrystalline[2]; breaks with conchoidal fracture.	Micrite
		Shell fragments loosely cemented with a high porosity.	Coqina
		Microscopic fossil fragments; chalky; white color; soft.	Chalk
		Microcrystalline[2]; color	Travertine

		banding of browns, grays, whites, and blacks.	
		Coarse crystals easily visible.	Crystalline Limestone
	Dolomite	Fizzes with acid only when powdered; fine to coarsely crystalline; may contain fossils.	Dolostone
	Microcrystalline Quartz	Microcrystalline[2]; conchoidal fracture; hardness of 7 so steel nail leaves a metal streak on the chert surface; sharp edges used for pre-historic spearheads and knives; synonymous with flint.	Chert
	Gypsum	Crystalline; very soft with a hardness <2, white, gray, or pink color.	Rock Gypsum
	Halite	Crystalline; salty taste, white or gray color; fairly soft hardness of 2.5.	Rock Salt
	Plant Material	Plant fragments; low density; brown to black color; often crumbly.	Bituminous Coal

[1]Fossiliferous Limestone can be further subdivided into Grainstone, Packstone, Wackestone and Mudstone depending on the amount of mud with the lithified lime mud. Grainstone = less than 10% mud → Mudstone = less than 10% fossil organisms.

[2]Microcrystalline - crystals that are visible only through a high-powered microscope.



Conglomerate. A coarse-grained clastic sedimentary rock that is composed of a substantial fraction of rounded to subangular gravel-size clasts larger than 2 mm (0.079 in) in diameter. Conglomerates form by the consolidation and lithification of gravel. Conglomerates are typically poorly sorted and contain finer grained sediment, e.g., either sand, silt, clay or combination of them, called matrix by geologists, filling their interstices and are often cemented by calcium carbonate, iron oxide, silica, or hardened clay. *Photo credit: CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*



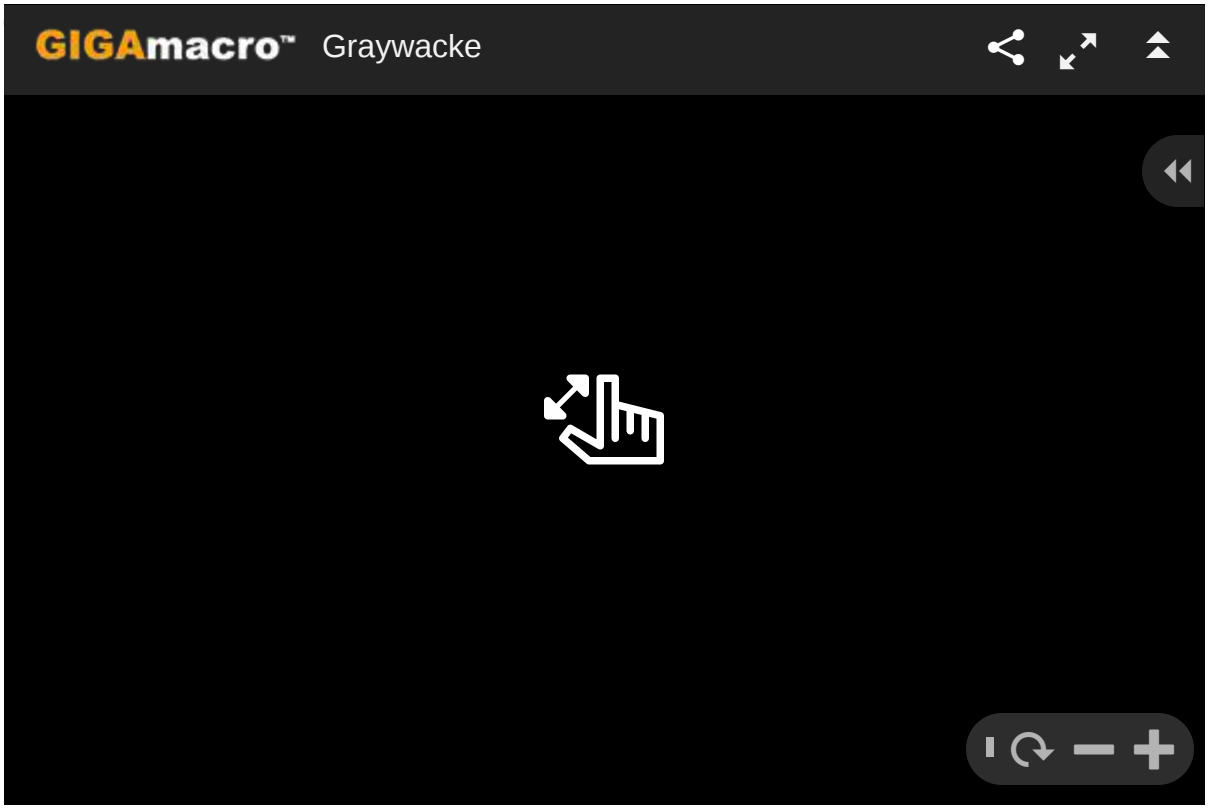
Breccia. A coarse-grained clastic sedimentary rock that is composed of a substantial fraction of angular gravel-size clasts larger than 2 mm (0.079 in) in diameter. Breccias are typically poorly sorted and contain finer grained sediment, e.g., either sand, silt, clay or combination of them, called matrix by geologists, filling their interstices and are often cemented by calcium carbonate, iron oxide, silica, or hardened clay. *Photo credit: CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*



Quartz sandstone. A clastic sedimentary rock composed mainly of sand-sized (0.0625 to 2 mm) quartz mineral grains. White and red (with iron oxide cement) are most common. Variable rounding and sorting. *Photo credit: CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*



Arkose sandstone. A clastic sedimentary rock composed mainly of sand-sized (0.0625 to 2 mm) feldspar mineral grains. Pink (potassium feldspar) and white (plagioclase feldspar) are most common. Commonly angular fragments with variable sorting. *Photo credit: CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.0 Attribution.*



Greywacke. A clastic sedimentary that is a variety of sandstone. It is generally characterized by its dark color, and poorly sorted angular grains of quartz, feldspar, and small rock fragments set in a compact, clay-fine matrix. *Photo credit: CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*



Siltstone. A clastic sedimentary rock. As its name implies, it is primarily composed of silt sized particles, defined as grains < 0.06mm. Siltstones differ significantly from sandstones due to the smaller spaces available between the grains (pore space) and often contain a significant amount of clay. Although often mistaken as a shale, siltstone lacks the fissility (easy separation along flat bedding planes) and laminations which are typical of shale. *Photo credit: CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*



Shale (with shell and plant fossils). A fine-grained, clastic sedimentary rock, composed of mud that is a mix of flakes of clay minerals and tiny fragments (silt-sized particles) of other minerals, especially quartz. Shale is characterized by breaks along parallel layering or bedding less than one centimeter in thickness, called fissility. **It is the most common sedimentary rock.** *Photo credit: CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Claystone



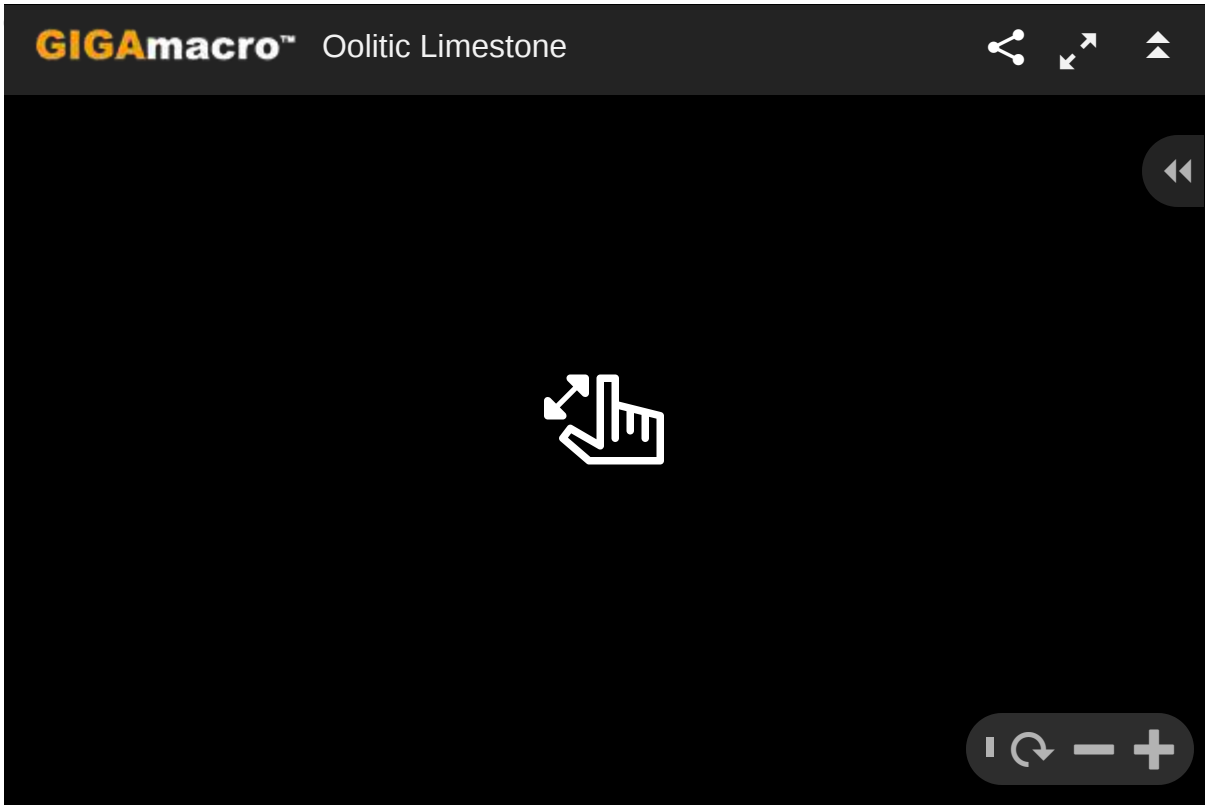
Claystone. A clastic sedimentary rock composed largely of clay, which measures less than 1/256 of a millimeter in particle size (not visible with the naked eye). Claystone is non-fissile, it does not break easily into layers like shale. *Ralph L. Dawes, Ph.D. and Cheryl D. Dawes, licensed under a Creative Commons Attribution 3.0 United States License.*

Chemical, Biochemical and Organic Sedimentary Rock Photos

Presented in order on the chart.



Fossiliferous limestone is any type of limestone, made mostly of calcium carbonate ($CaCO_3$) in the form of the minerals calcite, that contains an abundance of fossils or fossil traces. The fossils in these rocks may be of macroscopic or microscopic size. This particular sample contains an abundance of mollusc and gastropod fossils. *Photo credit: CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*



Oolitic Limestone is a type of limestone formed from ooids, small, spherical grains composed internally of calcium carbonate ($CaCO_3$ – calcite) in concentric layers.
Photo credit: CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.

Micrite



Micrite is a limestone formed of calcareous (composed of calcite) particles ranging in diameter up to four microns. It is formed by the recrystallization of calcareous mud. *"Fossiliferous dismicrite (Nineveh Limestone, Lower Permian; Clark Hill section, Long Ridge, Monroe County, Ohio, USA) 1" by James St. John is licensed under CC BY 2.0 Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Chalk



Chalk is a soft, white, porous, sedimentary carbonate rock, a form of limestone composed of the mineral calcite. It forms under reasonably deep marine conditions from the gradual accumulation of minute calcite shells (coccoliths) shed from micro-organisms

called coccolithophores. *Photo credit: CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Travertine



Figure 5.5.1: **Speleothem** by **Marissa Dudek** on **Sketchfab**

Travertine is a form of limestone deposited by mineral springs, especially hot springs. Travertine is formed by a process of rapid precipitation of calcium carbonate, often at the mouth of a hot spring or in a limestone cave. In the latter, it can form stalactites, stalagmites, and other speleothems (as in the 3D model, above. Grab and rotate the sample to see it in 3D). Travertine can be white, tan, cream-colored, or somewhat rusty in color.

Crystalline Limestone



Crystalline limestone is a carbonate sedimentary rock that is composed of the precipitation of the mineral calcite ($CaCO_3$) from saturated sea water. Its major materials are the minerals calcite and aragonite, which are different crystal forms of calcium carbonate ($CaCO_3$). About 10% of sedimentary rocks are limestones. *Photo credit: Ralph L. Dawes, Ph.D. and Cheryl D. Dawes, licensed under a Creative Commons Attribution 3.0 United States License. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Dolostone



Dolostone is a sedimentary carbonate rock that contains a high percentage of the mineral dolomite, $\mathrm{CaMg(CO_3)_2}$. Dolomite contains nearly equal amounts of magnesium and calcium. Most dolomites formed as a magnesium replacement of limestone or lime mud before lithification. Dolomite is less soluble than limestone and, for identification purposes, must first be scratched into a small power before it will effervesce with dilute hydrochloric acid. *Photo credit: "Dolostone (Silurian; midwestern USA)" by James St. John is licensed under CC BY 2.0. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Chert (and Flint)



Chert is a hard, fine-grained sedimentary rock composed of crystals of quartz (silica SiO_2) that are very small (microcrystalline or cryptocrystalline). It often breaks with a conchoidal fracture and was used to make sharp objects like knives and arrowheads in the prehistoric era. Chert is often of biological origin (organic) but may also occur inorganically as a chemical precipitate or a diagenetic replacement (e.g., petrified wood). Geologists use chert as a generic name for any type of microcrystalline or cryptocrystalline quartz. Varieties dark in color are called **flint**; red varieties are known as **jasper**; multi-colored and banded varieties are known as **agate**. *Photo credit: Ralph L. Dawes, Ph.D. and Cheryl D. Dawes, licensed under a Creative Commons Attribution 3.0 United States License. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Rock Gypsum



Rock gypsum is composed of the very soft mineral gypsum which defining characteristic is that it can be scratched with a fingernail. It is typically white to very pale pink or tan. Gypsum rock forms within layers of sedimentary rock from the slow evaporation of seawater. *James Petts From: Wikimedia is licensed under: Creative Commons Attribution-Share Alike 3.0 license.*



Rock salt is composed of the mineral halite (sodium chloride, NaCl, common table salt) which defining characteristics include that it tastes salty and will dissolve readily in water. It is typically clear to white to pale pink or tan. Rock salt forms within layers of

sedimentary rock from the slow evaporation of seawater. *Photo credit: CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Bituminous Coal



Bituminous coal is an organic sedimentary rock formed by diagenesis of peat bog material. Bituminous coal or black coal is a relatively soft coal containing a tar-like substance called bitumen. Its coloration can be black or sometimes dark brown; hard varieties may shine with an iridescent effect. The carbon content of bituminous coal is around 60–80%; the rest is composed of water, air, hydrogen, sulfur, and heavy metals such as lead, mercury, nickel, tin, cadmium, antimony, and arsenic, as well as radio isotopes of thorium and strontium. Nasty stuff! There is no such thing as "clean coal." *Photo credit: Ralph L. Dawes, Ph.D. and Cheryl D. Dawes, licensed under a Creative Commons Attribution 3.0 United States License. Definition credit: Wikipedia licensed under CC BY: 3.O Attribution.*

Did I Get It? - Quiz

? Exercise 5.5.1

The most important agent involved in chemical weathering processes is

a. ice

b. wind

c. gravity

d. water

Answer

d. water

? Exercise 5.5.2

If you take a hammer and break apart a rock into smaller pieces, this would be a form of which type of weathering?

a. erosional

b. chemical

c. physical

Answer

c. physical

? Exercise 5.5.3

A sedimentary rock that is composed of detrital pieces of pre-existing rock would be classified as

a. chemical

b. biochemical

c. clastic

d. organic

Answer

c. clastic

? Exercise 5.5.4

Under the right conditions, ions of elements dissolved in water can precipitate to form which of the types of sedimentary rock?

a. biochemical

b. clastic

c. organic

d. chemical

Answer

d. chemical

? Exercise 5.5.5

Living organisms that extract ions out water to form their shells produce which type of sedimentary rock?

a. Clastic

b. Chemical

c. Biochemical

d. Organic

Answer

c. Biochemical

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5.6: Metamorphism and Metamorphic Rocks

Metamorphism is the change that takes place within a body of rock as a result of it being subjected to conditions that are different from those in which it formed. In most cases—but not all—this involves the rock being deeply buried beneath other rocks, where it is subjected to higher temperatures and pressures and no longer in equilibrium as it was in its original environment of formation. Metamorphism typically includes the recrystallization of minerals plus the formation of new minerals and different textures though often retaining the same overall chemical composition of the parent, or source rock. This parent rock is referred to as the **protolith**, from the Greek *proto-* meaning first, and *lithos-* meaning rock.

Most metamorphism is the result of tectonic forces during mountain building (orogenic) episodes. Plate collisions, especially continent to continent collisions, introduce tremendous stress and heat, take place over vast areas and last millions of years. These are the primary factors, or conditions needed to allow the dramatic changes to occur from protolith to metamorphic rock product.

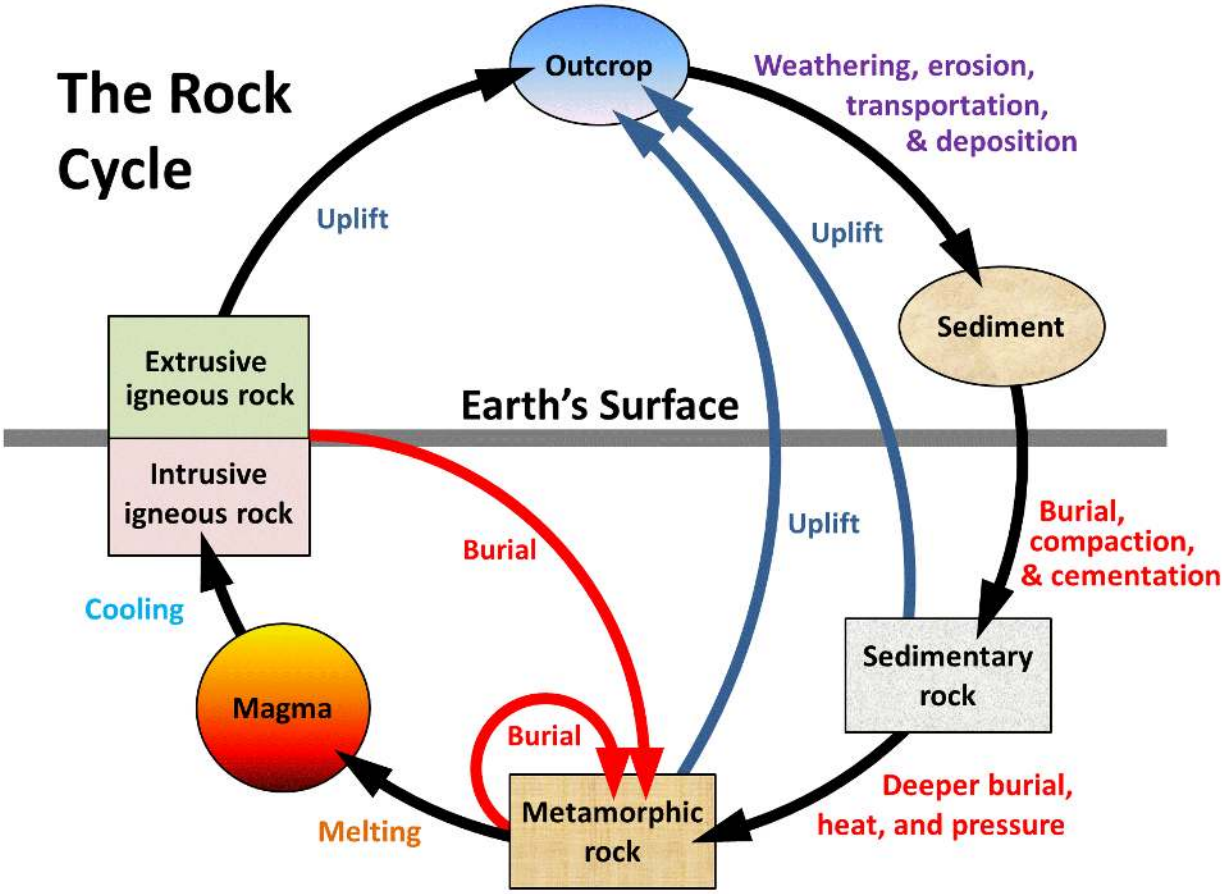


Figure 5.6.1: The rock cycle, showing the processes related to metamorphic rocks at the bottom. (CC BY; Steven Earle from https://opentextbc.ca/physicalgeology2ed/part/chapter-7-metamorphism-and-metamorphic-rocks/)

Types of Metamorphism

There are two primary environments that produce metamorphism that are important to Historical Geology:

Regional metamorphism. This occurs over a wide expanse of the crust, hundreds to more than a thousand kilometers across, resulting from the collision and compression of tectonic plates. Regional metamorphism can affect the pre-existing rock of the continental crust, any oceanic crust caught in the collision, plus the sedimentary rock cover that exists on each. Regional metamorphic processes can also affect the entire thickness of the crust and upper mantle and produce tremendous change in the pre-existing rock occurring tens of kilometers to several hundred kilometers deep.

Contact metamorphism is much simpler and affects pre-existing rock by baking them through contact with molten rock material either by intrusion of magma or extrusion of lava on the surface. Intrusion can occur very deep in the crust and involve large areas affected by heat associated with vast magma chambers. Intrusion can also be much smaller and take place closer to the surface or involve the baking of the surface rock through volcanic eruption of lava.

Metamorphic Processes

The main factors that control metamorphic change of pre-existing rock include:

- the mineral composition of the parent rock,
- the temperature at which metamorphism takes place,
- the amount and type of pressure during metamorphism,
- the types of fluids (mostly water) that are present during metamorphism, and
- the amount of time available for metamorphism.

Parent Rock (Protolith)

The parent rock is the rock that exists before metamorphism starts. As we see in the rock cycle, any type of rock can be a protolith including pre-existing metamorphic rock. The critical component of the parent rock is its mineral composition because it is the stability of each mineral that matters when metamorphism takes place. In other words, when a rock is subjected to increased temperatures, certain minerals may become unstable and begin to recrystallize in size, orientation or into completely new minerals.

Temperature

As we learned in the previous discussions of Bowen's Reaction Series, minerals gain stability as temperature cools. Minerals remain stable over a specific range of temperature. For example, quartz is stable under all surface environmental temperatures will remain stable all the way up to about 1800 °C. On the other hand, most clay minerals are only stable up to about 150° or 200 °C; above that, they transform into micas. Feldspar, the most common mineral of the crust, is stable up to approximately 1200 °C.

Pressure

Pressure is the force exerted on the protolith by burial and/or tectonic stresses. Like heat, pressure can affect the chemical equilibrium of minerals in a rock. The pressure that affects metamorphic rocks can be grouped into confining pressure and directed stress. **Stress** is a scientific term indicating a force. **Strain** is the result of this stress, including metamorphic changes within minerals.

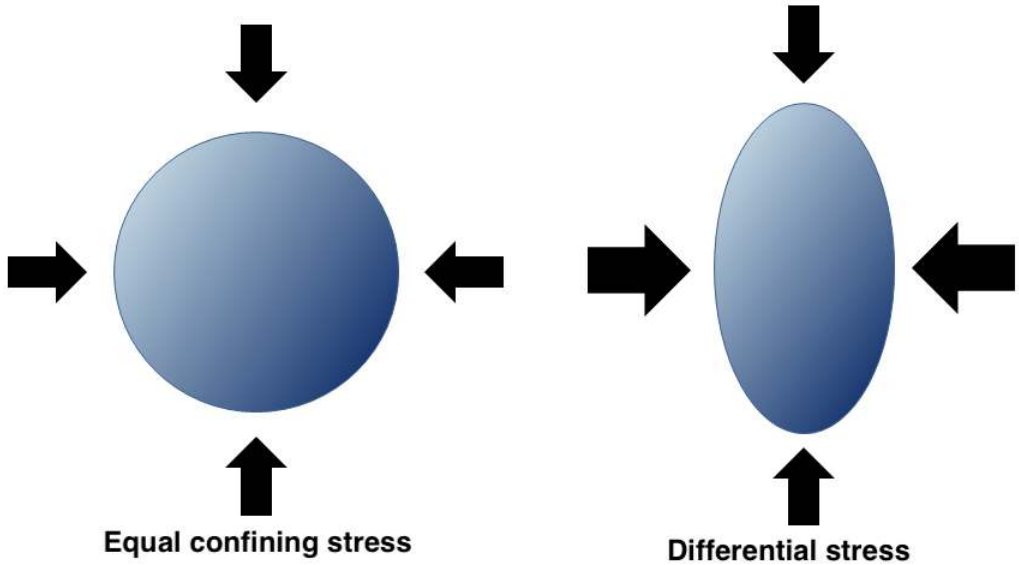


Figure 5.6.2: An illustration of different types of pressure on rocks. Confining pressure, where the pressure is essentially equal in all directions, and differential stress, where the pressure from the sides is greater than that from the top and bottom. In both diagrams, there is also pressure in and out of the page. (CC BY; Steven Earle.)

Confining Pressure

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

Differential Stress

Differential stress is an unequal balance of forces on the protolith in one or more directions (see previous figure). Differential stress is most commonly associated with the tectonic movement of plates during mountain building (orogeny). Differential stress modifies the parent rock at a mechanical level, changing the arrangement, size, and/or shape of the mineral crystals. This creates an identifying texture, known as foliation, which is shown in the figure below.

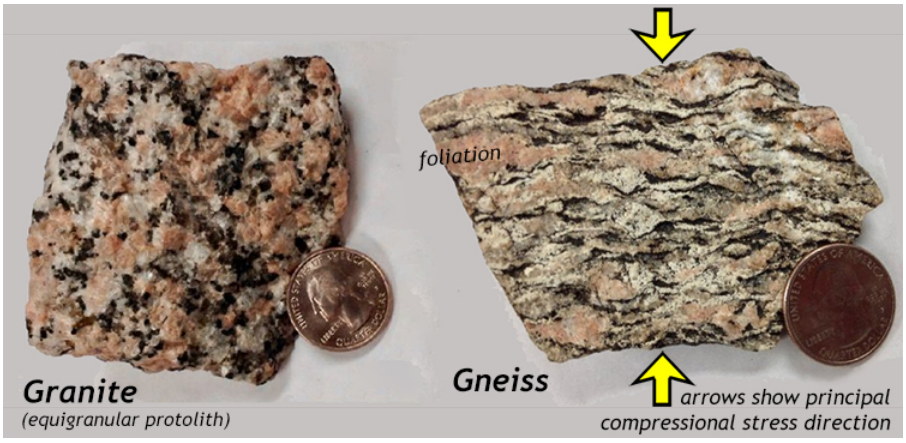


Figure 5.6.3: A comparison of granite and gneiss. The texture of the protolith granite is the same in every direction (equigranular) but the gneiss has a pronounced foliation, which we infer to have developed from squeezing (compressional stress) in a direction that was perpendicular to the resulting foliation. (Peter Davis. Modified by Callan Bentley.)

Differential stress produces rock textures with a visible parallel alignment. In response to the differential stress, mineral crystals can rotate, changing their orientation in space. This is a process of recrystallization in a solid state. The atomic structure of the mineral will respond to the stress and atoms will migrate within the structure in response to the stress. A good example would be the reorientation of the mineral biotite in the samples above. Biotite is a black, sheet silicate that can be easily peeled apart in thin layers. The biotite crystals are randomly oriented in the igneous granite sample on the left. When subjected to the differential stress of tectonic forces, the biotite crystals will realign perpendicular to the directions of stress providing the visually **foliated** metamorphic rock texture.

Recrystallization can also lead to minerals increasing in grain size as atoms migrate from an area of rock experiencing high stress and precipitate or regrow in a location having lower stress. You can visualize this similar to the phenomenon that occurs with small, adjacent soap bubbles coalesce to form larger ones. A good example of this would be the recrystallization of calcite that occurs in the metamorphism of limestone to marble (below). Note the crystalline fabric of the marble on the right versus the very fine mud-like texture of the sedimentary limestone, micrite, on the left.



Figure 5.6.4: Laminated micrite (left) compared to a crystalline banded marble (right). (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Fluids

Water is the main fluid present within rocks of the crust. Water facilitates the movement of ions within the protolith being subjected to metamorphic conditions. Water assists in metamorphic mineral growth and also acts as a catalyst that can increase the rate at which metamorphic reactions take place.

Time

Most metamorphic reactions take place at very slow rates. For example, the growth of new minerals within a rock during metamorphism has been estimated to be about 1 millimetre per million years. For this reason, it is very difficult to study metamorphic processes in a lab.

While the rate of metamorphism is slow, the tectonic processes that lead to metamorphism are also very slow, so in most cases, the chance for metamorphic reactions to be completed is high. A mountain range takes tens of millions of years to form, and tens to hundreds of millions of years to be eroded to the extent that we can see the rocks that were metamorphosed within the deep interior.

Classification of Metamorphic Rocks

Metamorphic rocks are classified based on two characteristics: Texture and Composition. Texture is first characteristic observed in the identification process. There are two main types of metamorphic rocks:

Foliated: formed in an environment subjected to differential stress resulting in a distinct alignment of minerals. Foliated metamorphic rocks are named based on the style of their foliations. Each rock name has a specific texture that defines and distinguishes it.

Non-foliated: either not subjected to directed pressure or, the dominant mineral does not tend to display any alignment.

Table 5.6.1: Metamorphic Rock Identification. (Created by Belinda C. Madsen for Salt Lake Community College)

1. Identify Rock's Foliation	1. Identify Rock's Foliation	2. Textural Features	3. Mineral Composition	4. Rock Name	5. Parent Rock
FOLIATED layered texture	Fine-grained or no visible grains	Flat, slaty cleavage is well developed. Dense, microscopic grains, may exhibit slight sheen (or dull luster). Clanky sound when struck. Breaks into hard, flat sheets.	Fine, microscopic clay or mica.	SLATE	Shale
		Finely crystalline; micas hardly discernible, but impart a sheen or luster. Breaks along wavy surfaces.	Dark silicates and micas.	PHYLLITE	Siltstone or shale
	Medium- to coarse-grained	Schistose texture. Foliation formed by alignment of visible crystals. Rock breaks along scaly foliation surfaces. Medium to fine-grained. Sparkling appearance.	Common minerals include chlorite, biotite, muscovite, garnet and hornblende. Recognizable minerals used as part of rock name. Porphyroblasts common.	MICA SCHIST GARNET SCHIST	Siltstone or shale
		Gneissic banding. Coarse-grained. Foliation present as minerals arranged into alternating light and dark layers giving the rock a banded texture in side view. Crystalline texture. No cleavage.	Light-colored quartz and feldspar; dark ferromagnesian minerals.	GNEISS	Shale or granitic rocks
NON-FOLIATED no layered texture	Fine-grained or no visible grains	Medium- to coarse-grained crystalline texture.	Crystals of amphibole (hornblende) in blade-like crystals.	AMPHIBOLITE	Basalt, gabbro, or ultramafic igneous rocks
		Microcrystalline texture. Glassy black sheen. Conchoidal fracture. Low density.	Fine, tar-like, organic makeup.	ANTHRACITE COAL	Coal
		Dense and dark-	Microscopic dark	HORNFELS	Many rock types

1. Identify Rock's Foliation	1. Identify Rock's Foliation	2. Textural Features	3. Mineral Composition	4. Rock Name	5. Parent Rock
		colored. Fine or microcrystalline texture. Very hard. Color can range from gray, gray-green to black.	silicates.		
		Microcrystalline or no visible grains with smooth, wavy surfaces. May be dull or glossy. Usually shades of green.	Serpentine. May have fibrous asbestos visible.	SERPENTINITE	Ultramafic igneous rocks or peridotite
	Fine- to coarse-grained	Microcrystalline or no visible grains. Can be scratched with a fingernail. Shades of green, gray, brown or white. Soapy feel.	Talc.	SOAPSTONE OR TALC SCHIST	Ultramafic igneous rocks
		Crystalline. Hard (scratches glass). Breaks across grains. Sandy or sugary texture. Color variable; can be white, pink, buff, brown, red, purple.	Quartz grains fused together. Grains will not rub off like sandstone.	QUARTZITE	Quartz sandstone
		Finely crystalline (resembling a sugar cube) to medium or coarse texture. Color variable; white, pink, gray, among others. Fossils in some varieties.	Calcite or dolomite crystals tightly fused together. Calcite effervesces with HCl; dolomite effervesces only when powdered.	MARBLE	Limestone or dolostone
		Texture of conglomerate, but breaks across clasts as easily as around them. Pebbles may be stretched (lineated) or cut by rock cleavage	Granules or pebbles are commonly granitic or jasper, chert, quartz or quartzite.	META-CONGLOMERATE	Conglomerate



Figure 5.6.5: Slate in many different colors, depending on the composition of the protolith. (By C. E. Jones, 2003, from: Geologic Image Archive at the University of Pittsburgh with permission for educational purposes.)



Figure 5.6.6: A slate boulder on the side of Mt. Wapta in the Rockies near Field, BC. Bedding is visible as light and dark bands sloping steeply to the right (yellow arrows). Slaty cleavage is evident from the way the rock has broken (along the flat surface that the person is sitting on) and also from lines of weakness that are parallel to that same trend (red arrows). (CC BY; Steven Earle.)

Slate is a fine-grained metamorphic rock that exhibits a foliation called slaty cleavage that is the flat orientation of the small platy crystals of mica and chlorite recrystallized perpendicular to the direction of stress. The minerals in slate are too small to see with the unaided eye. The thin layers in slate may resemble sedimentary bedding, but they are a result of differential stress and may lie at angles to the original strata (see photo below). In fact, original sedimentary layering may be partially or completely obscured by the foliation. The protolith for slate was a shale-type sedimentary rock composed of clay. Slate represents low-grade metamorphism.

Phyllite is also a fine-grained foliated metamorphic rock in which the platy chlorite and mica minerals have grown larger and the surface of the foliation shows a sheen from light reflecting from the grains. The texture is “phylittic” where we can detect an increase in size of the minerals over slate however the individual minerals are still too small to be differentiated with the naked eye.

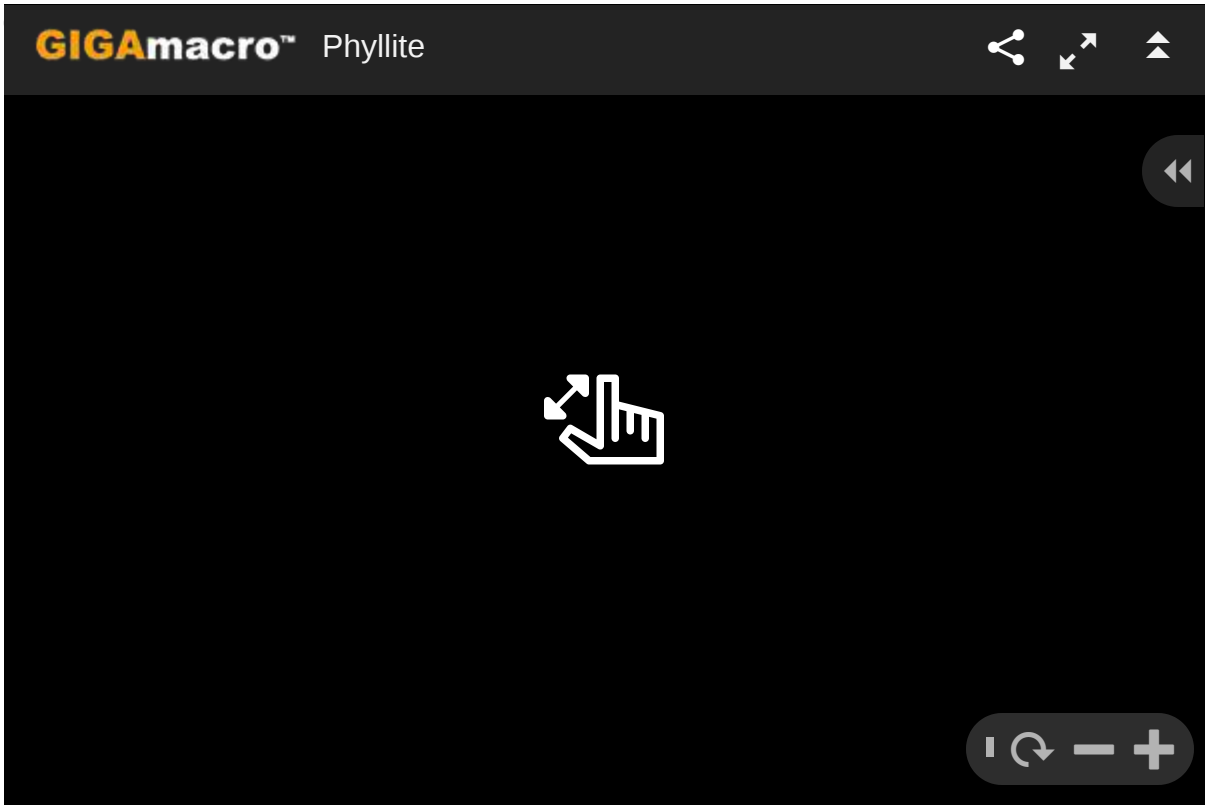


Figure 5.6.7: Phyllite displaying “phylittic” texture. Note displaying cross-bedding of the sedimentary protolith. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Schist. The metamorphic rock schist displays a “shistose” foliation in which the minerals are now visible as individual crystals. Common minerals are muscovite, biotite, and **porphyroblasts** of garnets. A porphyroblast is a large crystal of a particular mineral surrounded by small grains. Varieties of schist are named for their dominant minerals such as mica schist (mostly micas) or garnet mica schist (mica schist with garnets).



Figure 5.6.8: Left: **Biotite schist** displaying coarse-grained "schistose" texture. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.) *Right:* Garnet Mica Schist (both biotite and muscovite). Also displaying the coarse-grained "schistose" texture with large porphyroblasts of garnet. (CC BY Attribution 3.0; Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Gneiss and Migmatite. Mineral banding in a metamorphic rock produces a "gneissic" foliation in which visible silicate minerals separate into dark and light bands. The mineral grains tend to be coarse and the bands are often folded due to the extreme pressure and temperature conditions at which this type of metamorphic rock forms. A rock with this texture is called gneiss. Some gneisses may be subjected to temperatures at which partial melting can occur. This partially melted rock is a transition between metamorphic and igneous rocks called a migmatite. The melting will affect the minerals in the gneiss formed at the lowest temperatures on Bowen's Reaction Series (quartz and potassium feldspar) and will collect in small lenses within the migmatite.

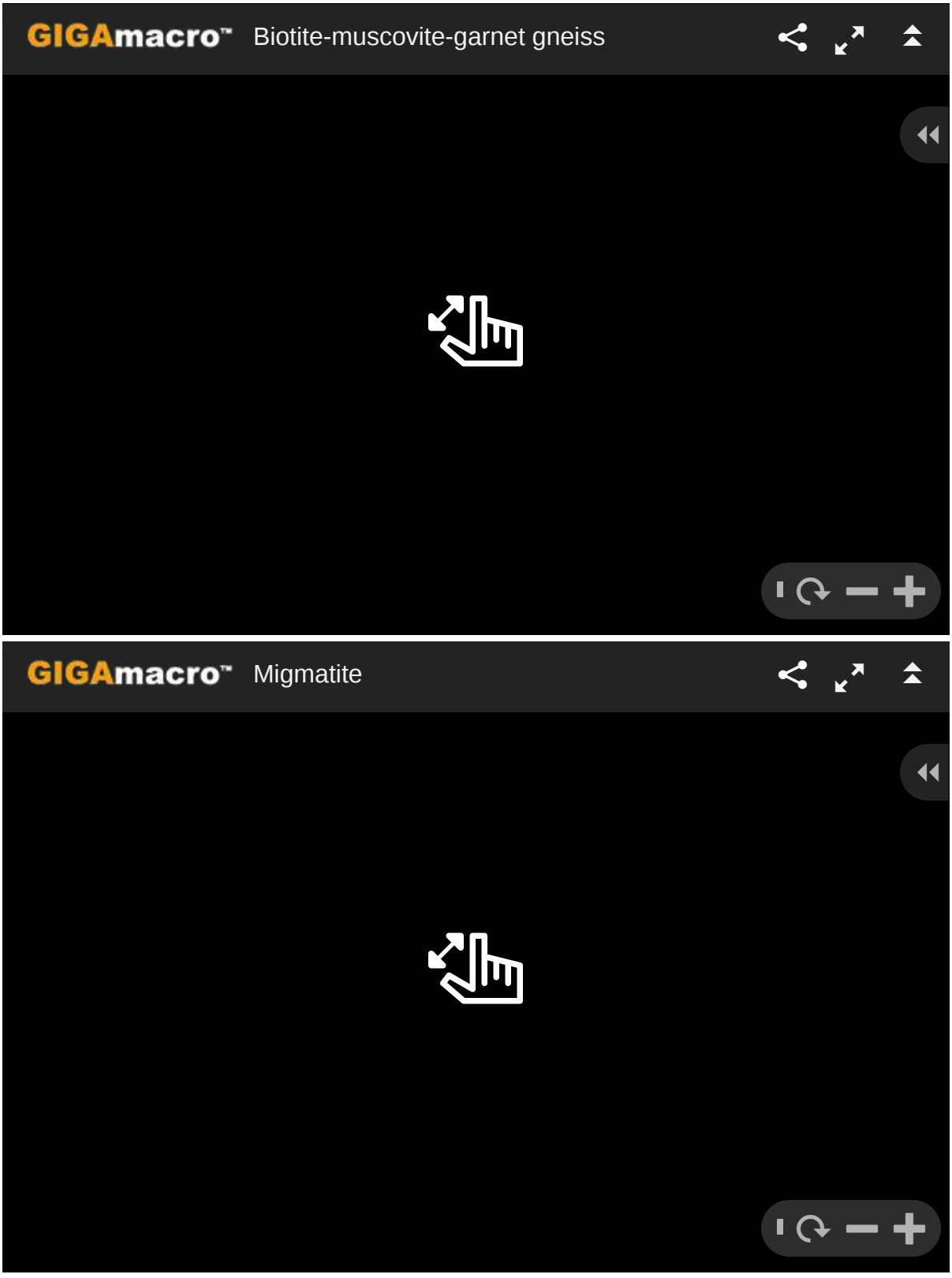
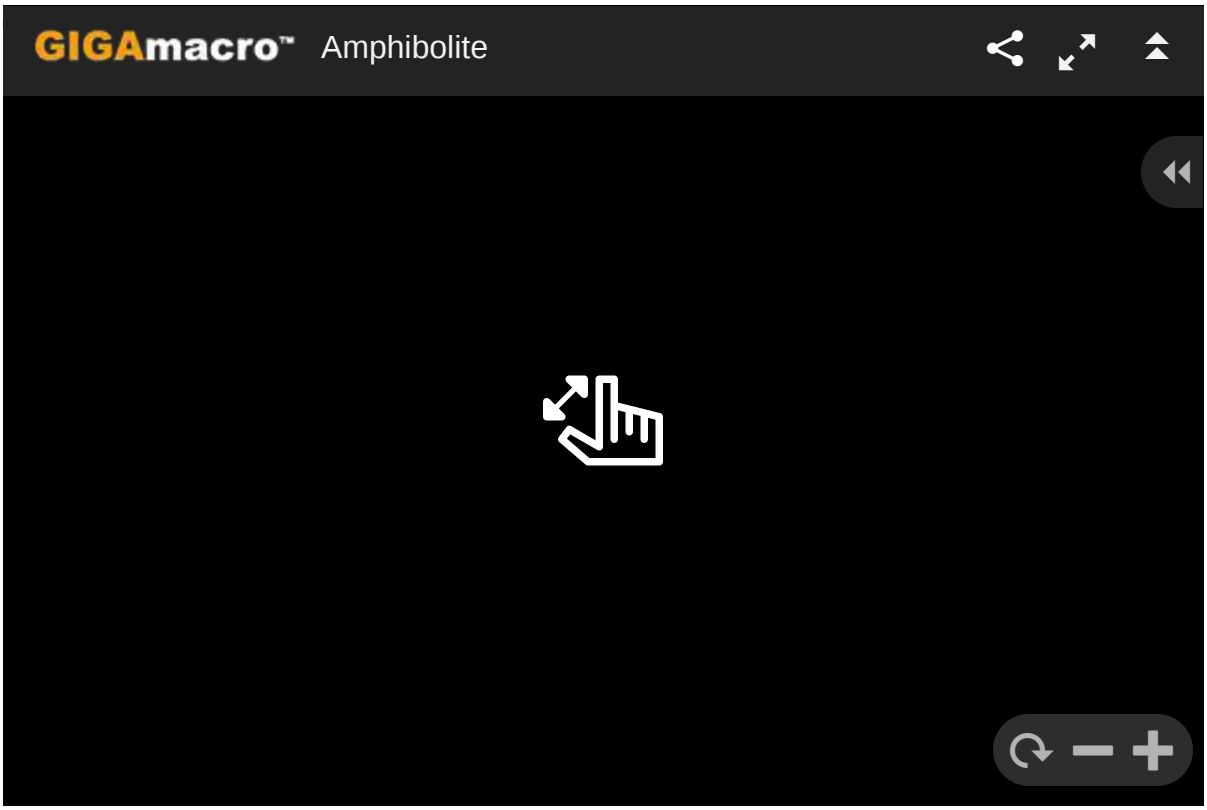


Figure 5.6.9: (Top) Muscovite, biotite, garnet gneiss displaying “gneissic” foliation. (Bottom) Migmatite. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Non-foliated Metamorphic Rocks

Non-foliated textures do not display alignment of mineral grains. Non-foliated metamorphic rocks are typically composed of just one mineral, and therefore, usually show the effects of metamorphism with recrystallization in which crystals grow together, but with no preferred direction. The two most common examples of non-foliated rocks are quartzite and marble.

Amphibolite is the result of metamorphism of a mafic igneous protolith. This may have originally been basalt or gabbro with olivine, pyroxene (augite) and calcium-rich plagioclase. The dominant mineral now is amphibole (hornblende). Zoom in on this sample and you will also see the green clay mineral chlorite.



Greenstone is a fine grained metamorphic rock that is largely composed of the green chlorite mineral plus other typically green metamorphic minerals, epidote and serpentine. The protolith of a greenstone is most commonly the mafic igneous rock, basalt. The original fine grained, iron and magnesium rich, minerals of olivine and pyroxene have recrystallized through metamorphic processes to form the greenstone rock. Greenstone is very important to Historical Geology, preserving some of the oldest slivers of early ocean crust. This is discussed in much more detail elsewhere in this text. The photo on the left, below is a hand sample of greenstone taken from the outcrop location on the right. Zoom into the sample on the left to see the fine-grained green minerals. Zoom into the photo on the left to get a feel for the metamorphic fabric and overall green color of the rock.



Figure 5.6.10: Left: Greenstone. *Right:* Outcrop location. (CC BY Attribution 3.0; Left photo: Robin Rohrback; right photo: Callan Bentley, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Quartzite and Marble



Figure 5.6.11: Left: Quartzite is a metamorphic rock with the protolith being quartz sandstone. In quartzite, the quartz grains and silica cement of the original sandstone become fused and interlocking by the recrystallization process. Zoom in and note the cohesive fabric produced by the interlocking crystals of quartz. (CC BY Attribution 3.0; Callan Bentely, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.) *Right:* Marble is metamorphosed limestone (or dolostone) composed of calcite (or dolomite). Recrystallization typically generates visible interlocking crystals of calcite or dolomite. (CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

Marble and quartzite often look similar however there are simple tests that can distinguish the two. First, the quartz minerals that compose the quartzite are considerably harder than the calcite that composes the marble. A simple hardness test on a piece of glass will show that quartz will scratch the glass, however calcite will not. Another way to easily distinguish quartzite from marble is with a drop of dilute hydrochloric acid. Marble will effervesce (fizz) because it is made of calcite. Remember to scratch the sample to make a little powder and then use a drop of hydrochloric acid to test. It will slowly effervesce if it is a dolomite marble.

Did I Get It? - Quiz

? Exercise 5.6.1

Metamorphism caused by tectonic plate movement that takes place over a great expanse of the crust and upper mantle is called

a. confining metamorphism

b. contact metamorphism

c. differential metamorphism

d. regional metamorphism

Answer

d. regional metamorphism

Exercise 5.6.2

The "protolith" is

a. the resulting metamorphic rock.

b. the parent rock that exists before metamorphism starts.

c.the final rock at the end of the rock cycle.

Answer

b. the parent rock that exists before metamorphism starts.

Exercise 5.6.3

Differential stress will create a metamorphic rock texture called

a. crystallization

b. non-foliation

c. foliation

d. mineral organization

Answer

c. foliation

Exercise 5.6.4

The rate of mineral growth during the recrystallization process of metamorphism is estimated to be

a. 1 millimeter per million years

b. 10 millimeters per 10 thousand years

c. 1 millimeter per billion years

d. 1 meter per million years

Answer

a. 1 millimeter per million years

Exercise 5.6.5

The mineral calcite is the dominant composition of limestone, both chemical and biochemical limestone. When limestone is buried deeply and subjected changes in pressure and temperature the calcite will recrystallize to form the metamorphic rock, marble. Which of these would be considered the protolith?

a. marble

b. limestone

c. calcite

Answer

b. limestone

Exercise 5.6.6

Metamorphism that occurs adjacent to a magma chamber would be of which type?

a. regional

b. temporary

c. contact

Answer

c. contact

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CHAPTER OVERVIEW

6: Plate Tectonics

In the late 1960s and early 1970s, geoscientific consensus gelled around an intriguing idea: that many different observations about our planet could be explained with a single model of Earth dynamics. The idea of “plate tectonics” put together old ideas about continental drift with new data showing seafloor spreading. The new theory was a revolutionary idea that made the distribution of volcanoes, earthquakes, continents, and topography make sense in a way that no idea had achieved before. Fifty years of additional data and testing have confirmed plate tectonics as ***the*** key idea in modern geology.

6.1: Plates
6.2: Tectonics
6.3: Boundaries
6.4: The historical record of plate interactions
6.5: Geophysical phenomena
6.6: Driving forces
6.7: Wilson Cycles
6.8: References, citations, and further reading

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6.1: Plates

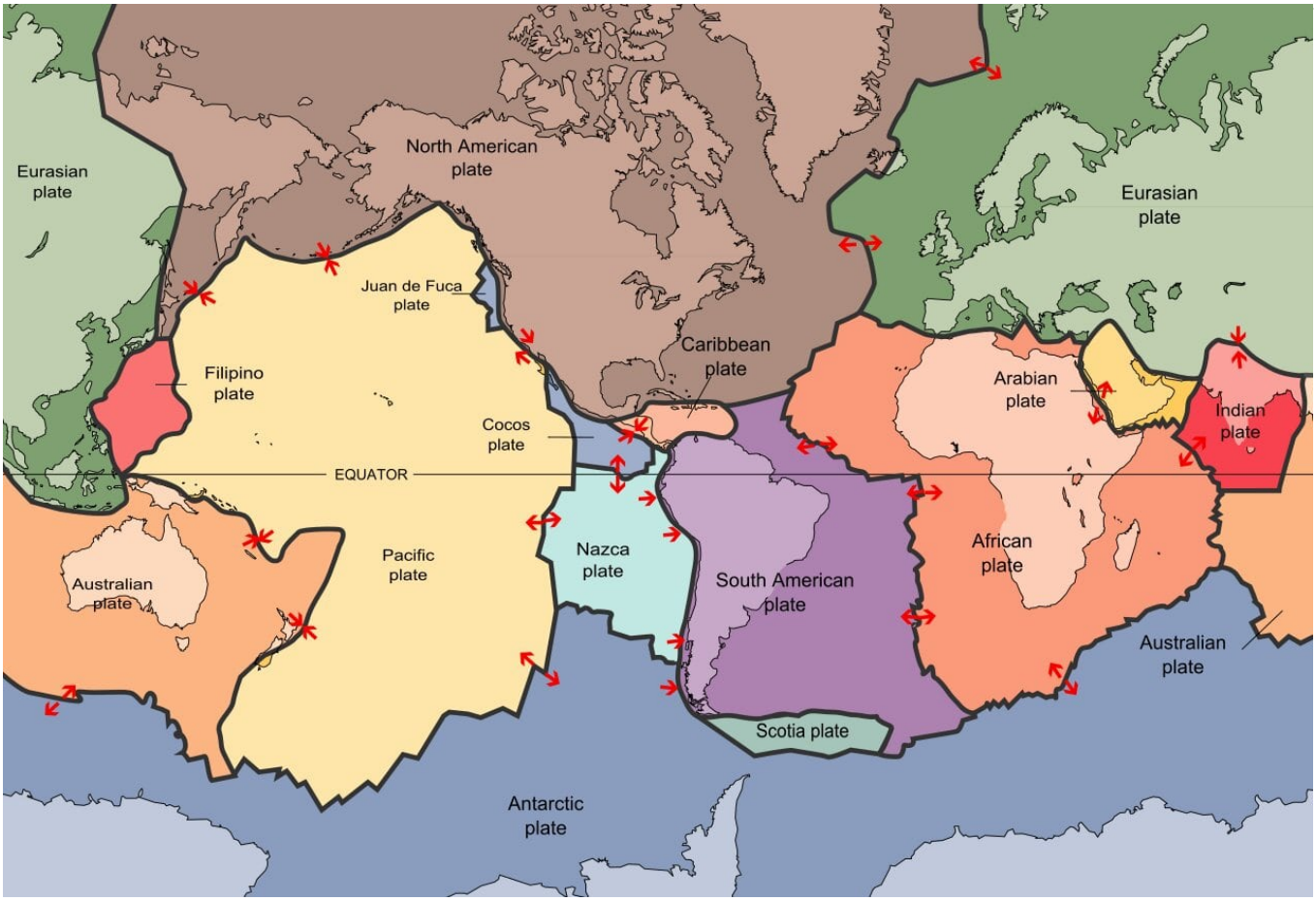


Figure 6.1.1: Plate boundary map. Arrows at the plate boundaries indicate the direction of movement along that particular boundary. (Public domain; Wikipedia.)

Earth's outermost rocky surface, the lithosphere, is broken into a series of big curvitabular pieces that move about relative to each other. Depending on how you count them, there are about 15 of these slabby pieces. We call them **plates**, and they range from about 50 km to 280 km thick and are mostly stiff upper mantle, but the upper part of each consists of crust. As humans, this is the part we can see and explore. The crust comes in one of two varieties: **continental crust** or **oceanic crust**. Together, the crust and the upper mantle comprise the lithosphere.

Oceanic lithosphere is a lot thinner and denser than the continental lithosphere. On average, the continental crust tends to be about 30 km thick, but it can get to be close to 100 km thick under new **mountain belts**, and is thinner beneath **rift zones**, weak spots where the crust is stretching. The oceanic crust is much thinner, around 10 km thick on average. It's thinnest beneath oceanic ridges, prominent features that rise from the seafloor and wrap around the planet through the ocean basins like seams on a baseball. The mantle portion of the lithosphere is the same composition beneath the continents and oceanic crust. It differs from the rest of the mantle not so much in its composition but in the way that it behaves: it rides around with the overlying crust as a more or less stiff unit.

Let us first consider the **position** of the plates in terms of the planet's layered structure. The lithosphere sits atop the asthenosphere, which contrasts in its behavior. The asthenosphere is a weak layer in the mantle that is mostly solid but perhaps 0.5% molten. This tiny amount of melt (99.5% solid) dramatically lowers the strength of the asthenosphere. The prefix *asthenos* means "weak" in Greek. Movement in the asthenosphere contributes to the motion of the lithospheric plates.

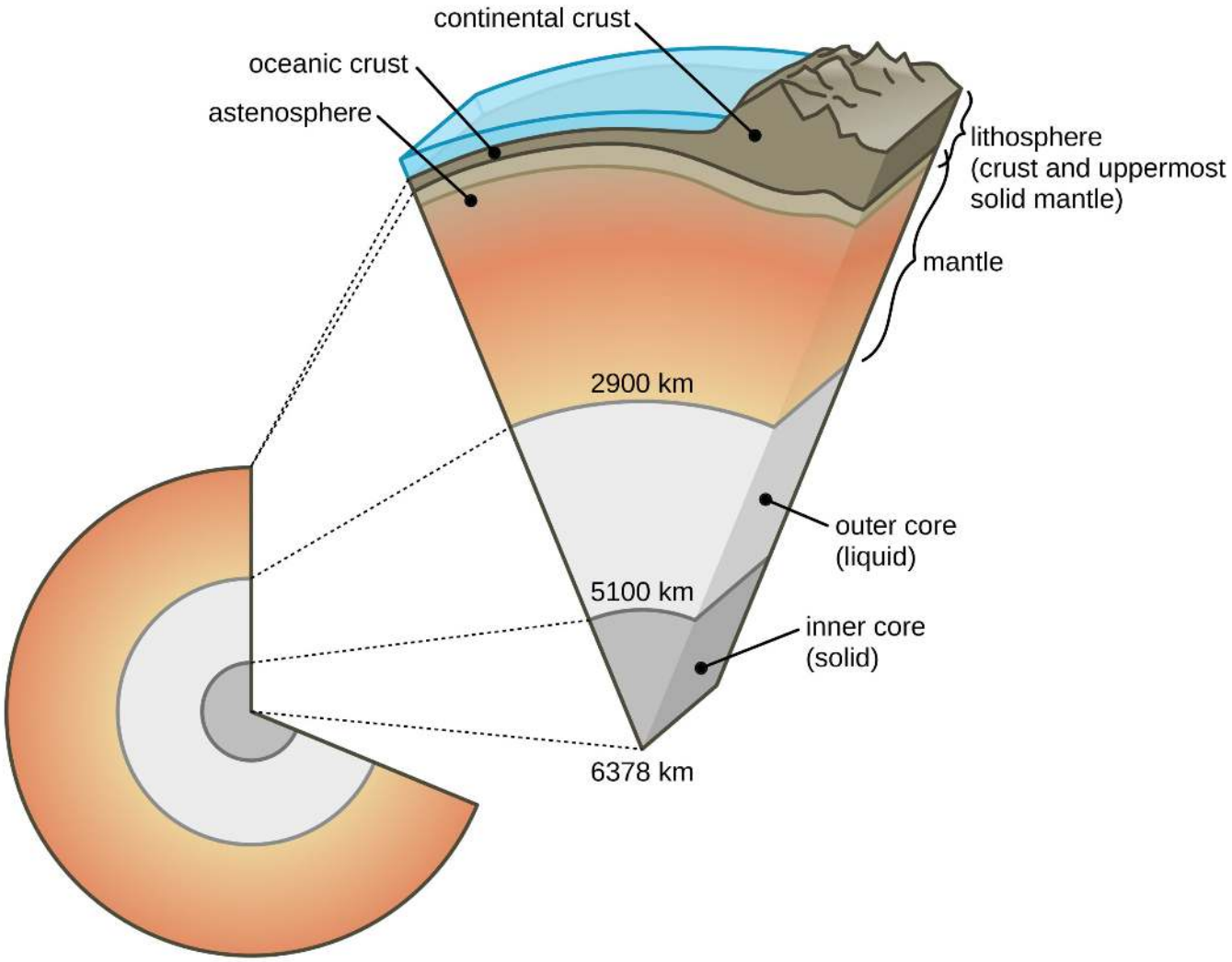


Figure 6.1.2: The layered interior of the Geosphere. Note that Lithosphere includes both the crust and uppermost portion of the solid mantle. This layer comprises Earth's tectonic plates. (U.S. Geological Survey.)

Beneath the asthenosphere is the lower mantle, and beneath that is the core. The core consists of an inner solid portion and an outer liquid portion. Both are made of an alloy that is mostly iron, plus ~5% nickel and some sulfur. The outer core convects, and in so doing generates a powerful magnetic field that extends through the surrounding shell of rock and out into space.

Above the lithosphere are the oceans and/or atmosphere. These sit higher than the rocky plates, since they are less dense. Though they sit at the same level, continental lithosphere is more buoyant than oceanic lithosphere since the continental crust is made of less dense rock than the oceanic crust. It is a situation akin to a Zodiac raft and a surfboard both floating on the surface of a lake. The ordered layering of atmosphere, ocean, crust, mantle, and core is a reflection of a mature planet, that has differentiated into distinct horizons of varying density:

Atmosphere	**0.001** g/cm^3
Ocean water	**1.0** g/cm^3
Continental crust or oceanic crust	**2.7** g/cm^3 *or* **2.9** g/cm^3
Mantle	**3.3** g/cm^3
Core	**11** g/cm^3

As far as their **size** goes, the plates can be large or small. The largest plates are the Eurasian Plate and the Pacific Plate. The smallest is harder to define. The Juan de Fuca Plate is only 250,000 km^2, a relatively tiny slab in the eastern Pacific Ocean, but there are even smaller "microplates" of oceanic lithosphere between the Pacific and Nazca Plates, too. The smallest plate consisting primarily of continental lithosphere is the Arabian Plate, at 5,000,000 km^2. Some consider the crust beneath California's Sierra Nevada and Great Central Valley to be a microplate, one that has foundered a bit to the west.

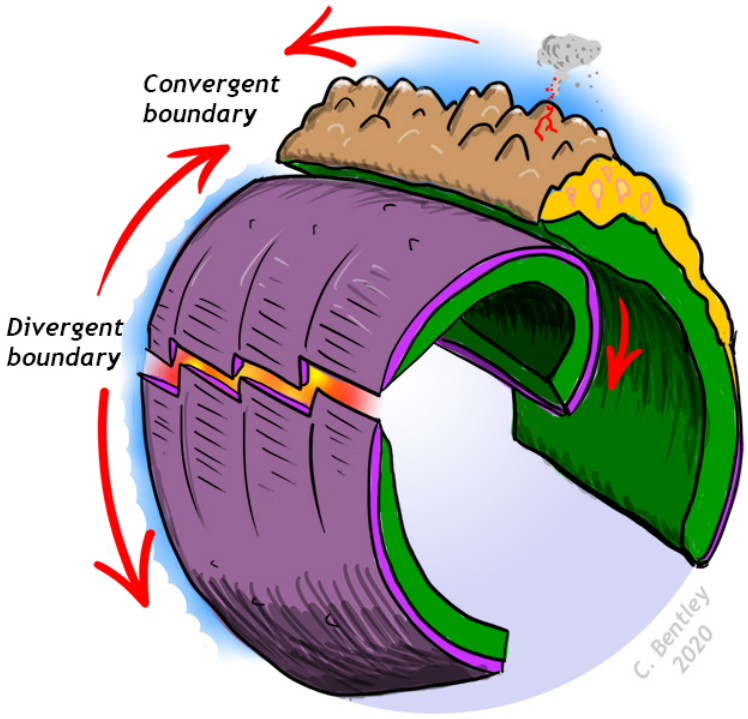


Figure 6.1.3: Three curvitabular plates interact along two boundaries. The middle plate, of oceanic lithosphere, diverges from the left plate, and converges with the plate to the right. (Drawing by Callan Bentley.)

The **shape** of the plates is kind of like fragments of an eggshell, or pieces of orange peel: they are both "slab like" or tabular in some senses, but also curved, as fragmental pieces of an overall spheroidal shell. There's no perfect word to describe such a shape, but perhaps "curvitabular" conveys a sense of it.

In map view, the shape of these curvitabular plates varies tremendously. Their edges can be more or less straight lines, arc-like curves, or zigzag-like patterns resembling irregular stair steps. They can even be broad, diffuse zones rather than crisp, well defined boundaries. And each of these boundary shapes is shared with another plate. Each of these shapes conveys information about the relative sense of motion of the plate movement.

Did I Get It? - Quiz

? Exercise 6.1.1

Plates are slabs of what layer of the Earth?

a. Crust only

b. Lithosphere (crust plus uppermost mantle)

c. Continental crust only

d. Lower mantle and outer core

e. Mantle only

f. Asthenosphere

Answer

b. Lithosphere (crust plus uppermost mantle)

Exercise 6.1.2

How much of the asthenosphere is molten?

a. A substantial amount, around 30%

b. All of it is molten (100%)

c. A tiny amount, perhaps 0.5%

Answer

c. A tiny amount, perhaps 0.5%

Exercise 6.1.3

Which of the following is least dense?

a. Core

b. Ocean water

c. Oceanic crust

d. Mantle

e. Atmosphere

f. Continental crust

Answer

e. Atmosphere

Exercise 6.1.4

What shape are plates?

a. "Curvitabular" (even though that isn't a real word)

b. Linear (pencil shaped)

c. Toroidal (doughnut shaped)

d. Spherical (ball shaped)

e. Tabular (flat slabs)

Answer

a. "Curvitabular" (even though that isn't a real word)

Exercise 6.1.5

Which is thicker (and less dense)?

a. Oceanic crust

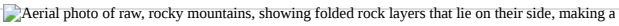
b. Continental crust

Answer

b. Continental crust

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6.2: Tectonics



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Figure 6.2.1: The huge fold in the flank of the 2969 m high Dent de Morcles is the most impressive witness to the tectonic collision between Africa and Europe. The darker layers are Tertiary in age and are younger than the greyish and yellowish limestone of the Cretaceous period. As the Alps were shoved upward, these marine layers were inverted on the upper limb of the fold. (Photograph by Bernhard Edmaier; reproduced with permission.)

Much of the attention we devote to lithospheric plates is focused on their boundaries, where much of their dynamic action takes place. That this dynamic action is located at the plate boundaries was the insight that led to the second part of the name: tectonics. This word means ***action*** in the sense of "building" or "constructing" things – specifically building **belts of mountains**. The root word is again Greek, and *tektonikos* means "building" or "making."

So in literal translation, the phrase plate tectonics means "building [mountain belts] using plates" but in current usage by scientists, the theory of plate tectonics covers much more than just mountains. It also encompasses our understanding of volcanic island arcs, ocean basins, earthquake locations, and disparate lines of evidence that document the long-term motion of lithospheric plates.

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6.3: Boundaries

By and large, the action of plate tectonics happens at the ***edges*** of plates, where they meet their neighbors. These plate **boundaries** come in three principal varieties, determined by the relative motion between the neighboring plates. Are they coming together, moving apart, or simply grinding past one another in opposite directions? We call these situations **convergent, divergent,** and **transform.** The dynamics of their relative motion end up determining which suite of geological processes and phenomena will result.

Did I Get It? - Quiz

? Exercise 6.3.1

What are the three principal varieties of plate boundaries?

a. Left-lateral, right-lateral, and convergent

b. Transform, divergent, and convergent

c. Tectonic, lithospheric, and transform

d. Convergent, volcanic, and lithospheric

Answer

b. Transform, divergent, and convergent

? Exercise 6.3.2

The word "tectonics" refers to what, exactly?

a. The degree of volcanic influence on a sedimentary "deposit."

b. The amount of shaking "induced" by an earthquake.

c. The "boundaries" of lithospheric plates.

d. The act of "building" mountain ranges.

Answer

d. The act of "building" mountain ranges.

Transform boundaries

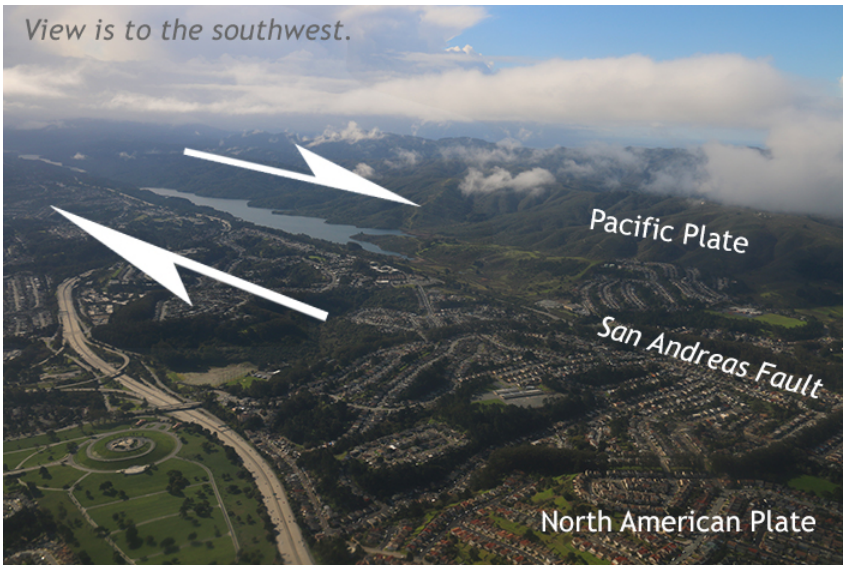


Figure 6.3.1: The modern plate boundary in the San Francisco region is a right-lateral transform plate boundary, seen here south of San Francisco along the trace of the San Andreas Fault. (Callan Bentley photo.)

Transform boundaries are the simplest kind of plate boundary. At a transform boundary, new lithosphere is neither created nor destroyed. Instead, the two plates slide past one another in opposite directions, grinding slowly along a fault zone. Friction acts to restrain the plates' motion. Moments when the friction is overcome — and the plates slip rapidly and violently — are earthquakes. Earthquakes are the main geological phenomenon to be aware of at transform boundaries, but the earthquakes there are distinct in that they are relatively shallow (20 km deep or less), and range in magnitude between small and large (~M7.5 at the most), but they are never huge, like the largest quakes that can occur (~M9) at convergent boundaries. Rock is weaker under shear than under compression, so less stress builds up at transform boundaries.

Transform boundaries can be recognized by offset features that cross the plate boundary fault zone such as bodies of rock, landscape features such as stream valleys, or human-built structures such as roads and fences. The faults that form on transform boundaries are either right-lateral or left-lateral. The San Andreas Fault in California is right-lateral: from the perspective of the Pacific Plate, the North American Plate is moving to the right. From the perspective of the North American Plate, the Pacific Plate is moving to the right.

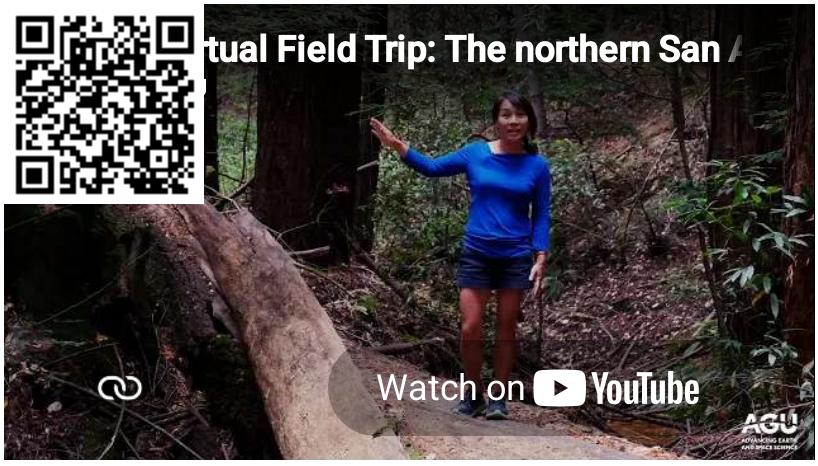




Figure 6.3.2: Crushing of bedrock on the San Andreas Fault in California has encouraged differential erosion of the landscape, producing the Olema Valley and Tomales Bay. (Landsat image from NASA.)

It doesn't matter which side of the fault you stand on; the apparent motion of the opposite side is to the right. The North Anatolian Fault in Turkey is another example of a right lateral fault, with the Anatolian Plate moving west relative to the Eurasian Plate. The North American Plate's southern margin is a left-lateral transform boundary with the Caribbean Plate. This boundary is distributed across two major faults, the Enriquillo-Plaintain Fault and the Septentrional/Oriente Fault.

Faults are are zones where rock is broken. Because crushed-up, pulverized rock weathers and erodes more rapidly than unbroken rock, the landscape of transform boundaries frequently shows linear features parallel to the trace of the fault zone: straight valleys and coastlines, elongated peninsulas, and bays.

The deeper record of this crushing can be exposed much later at Earth's surface. When historical geologists find a zone of **fault breccia**, they can deduce that they are looking at a shallow fault. Deeper in the crust, the rocks are hotter and under more confining pressure, and they deform in a different fashion as a result. Instead of fracturing, the minerals flow in the solid state, producing the smeared-out rock called **mylonite**. Compare the wispy, foliated texture of a mylonite with the chunky texture of a fault breccia here:





Figure 6.3.3: A sag pond developed along the North Anatolian Fault in Turkey. (Callan Bentley photo.)

In places where the trace of the faults on a transform boundary are not parallel to the plates' motion, there can be localized areas of compression or tension. When the orientation of the fault trace weaves a bit to the left or right, we call it a stepover. A right step in a right-lateral transform fault will create a zone of extension. This could manifest as a sag pond if it is small, or if bigger, it will make a larger trough like the Salton Sea or the Sea of Marmara. A left step in a right-lateral fault will generate a zone of compression, forcing rock upward into a pressure ridge. Pressure ridges can be small enough to see in a parking lot pavement, or the size of mountains.

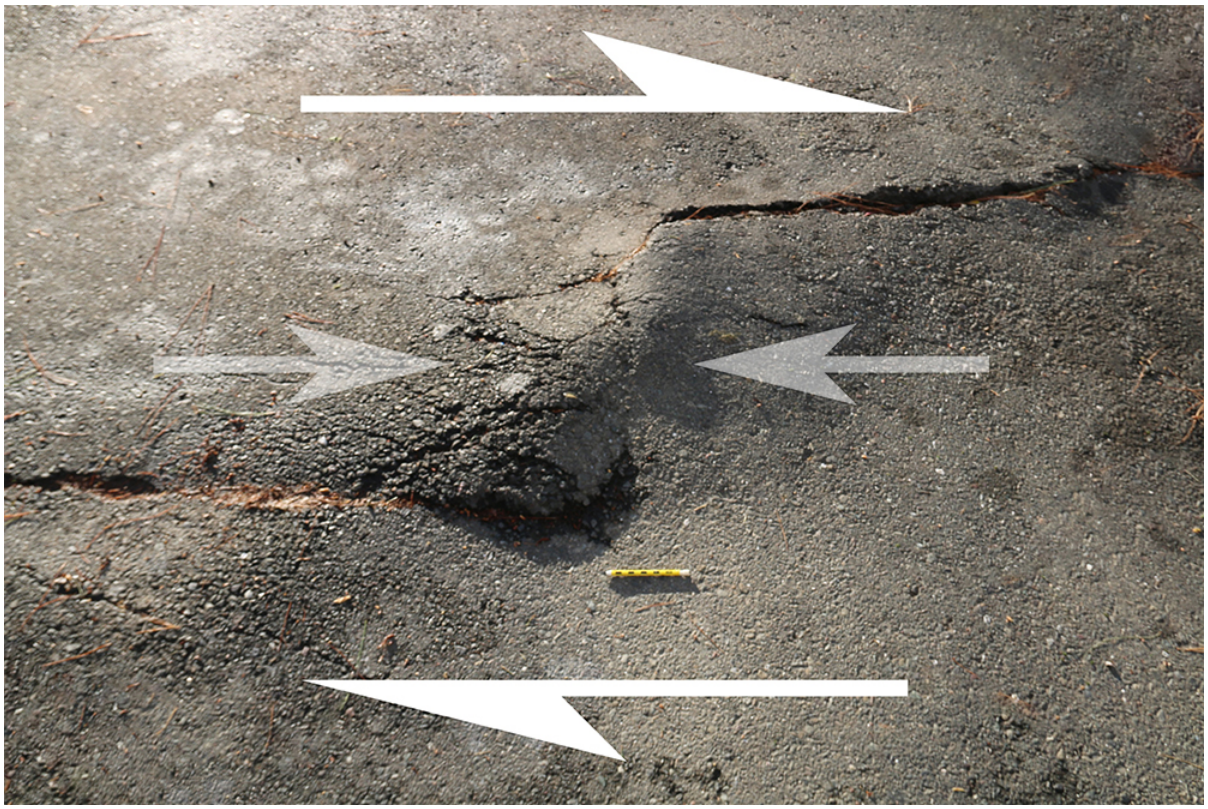


Figure 6.3.4: A small pressure ridge developed in parking lot asphalt at Central Park, Fremont, California. This is a left step in the right-lateral Hayward Fault. Perspective is looking to the southwest across the fault. (Callan Bentley photo.)

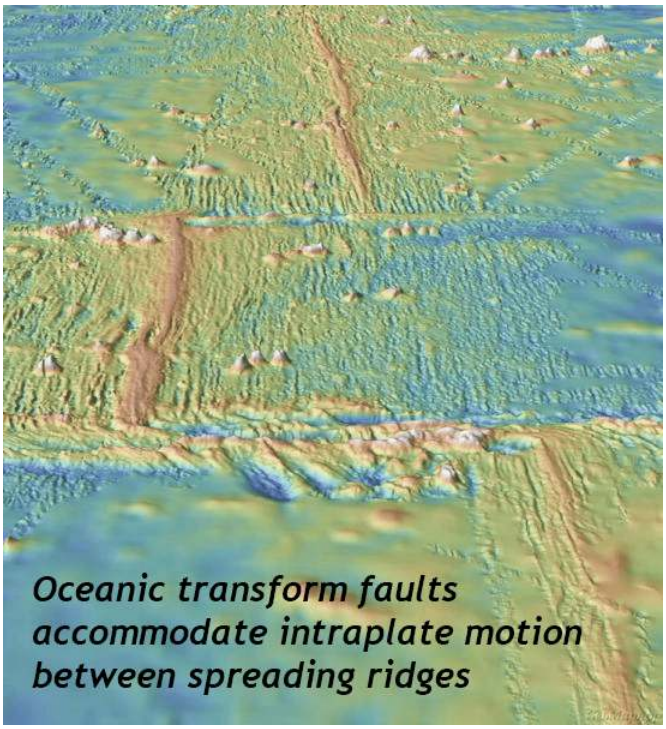


Figure 6.3.5: The oceanic ridge system comes in many segments, offset along perpendicular fracture sets.

Continental transform boundaries are striking, but far more common are examples of transform faults in the oceanic lithosphere. Between segments of oceanic ridge on the seafloor are zones where the plates slide past one another. These are oceanic transform faults, and they are a key feature of overall divergent boundaries in our planet's ocean basins. Together with the oceanic ridges you'll learn about below, the create the distinctive zigzag shape to most oceanic divergent plate boundaries.

Did I Get It? - Quiz

? Exercise 6.3.1

The San Andreas Fault is _____.

a. a left-lateral transform plate boundary

b. a convergent plate boundary

c. a right-lateral transform plate boundary

d. a divergent plate boundary in continental crust

e. a divergent plate boundary in oceanic crust

Answer

c. a right-lateral transform plate boundary

? Exercise 6.3.2

If the orientation of a transform fault veers a little bit, what can happen?

a. Partial melting, producing granite.

b. Deposition of limestone rich in stromatolites, ooids, and hurricane rip-up breccias.

c. Either a sag pond or a pressure ridge, depending on which way the fault jogs.

d. An earthquake!

Answer

c. Either a sag pond or a pressure ridge, depending on which way the fault jogs.

? Exercise 6.3.3

Faulting breaks up rock, making it easy to weather. What is the result of this?

a. Linear mountain ridges and peninsulas.

b. Rectangular valleys and ridges.

c. Triangular mountains and bays.

d. Linear valleys and bays.

Answer

d. Linear valleys and bays.

Divergent boundaries

Divergent boundaries are sites where two plates move away from one another. In this extensional setting, normal faulting dominates, thinning the crust vertically while stretching it out laterally. Divergent boundaries come in two principal varieties that are relevant to historical geology: rift valleys in continental crust and oceanic ridges in oceanic crust.

Continental rifting

When a divergent boundary develops in continental crust, it has two consequences, one sedimentary and one igneous.



Figure 6.3.6: A fresh divergent boundary developing in continental lithosphere, east Africa. (NASA photo, annotated by Callan Bentley.)

The sedimentary consequence is most apparent where divergent boundaries occur in continental lithosphere. The normal faulting down-drops a block of crust, producing a graben or "rift valley." The classic modern example is the Great Rift Valley of east Africa (pictured here).

The edge of the rift valley (the escarpment) is a site of high topographic relief, which encourages clastic sediment to drop downhill from the surrounding highlands into the rift valley. This sediment does not travel far, or for very long. As a consequence, it tends to be compositionally and texturally immature. Alluvial fans of poorly-sorted gravel flank the escarpment, smoothing out the highest relief areas. Finer grained sediment gets carried a bit further, and creates deposits of arkosic sand. Lakes accumulate in the lowest regions, furthest from the rift escarpment, growing salty as they evaporate.

A very similar situation can be seen in older, now-inactive rift valleys. For instance, the Culpeper Basin of central Virginia was an active site of rifting during the Triassic and Jurassic, as the supercontinent Pangaea was breaking apart and the Atlantic Ocean was first 'being born.'

As far as igneous consequences of crustal thinning, consider the decompression of the underlying mantle. Already hot, the lowering of the confining pressure allows melting to occur without the addition of any new heat. Partial melting of ultramafic peridotite generates a mafic magma, which may rise to the surface to erupt as "floods" of low-viscosity basalt. This is plain as a half dozen large black patches in the satellite image here, showing the Afar Triangle region of northeastern Africa.



Figure 6.3.7: Volcanic ash from rift valley volcanism preserves trace fossils of hominids that walked upright.

If the basaltic magma interacts with the crust on its ascent, it can transfer some of its heat and produce intermediate magmas, resulting in composite volcanoes such as Mt. Kenya, Mt. Kilimanjaro, or Mt. Meru. It was felsic ash erupted in the East Africa rift that preserved the footprints of two Australopithecines waking side by side at Laetoli, Tanzania, 3.2 million years ago.

Continental rifting is only the beginning of a divergent boundary however. Once the crust is sufficiently thinned, seafloor spreading takes over. This can be seen just north of East Africa, where a narrow body of oceanic crust separates Africa from Arabia. The Red Sea and the Gulf of Aden are just a bit more 'tectonically mature' than the Great Rift Valley. Here, the continental crust has fully broken into two discrete separate segments, diverging as Arabia moves to the northeast. New oceanic crust is filling the gap between them, a process called **seafloor spreading**.

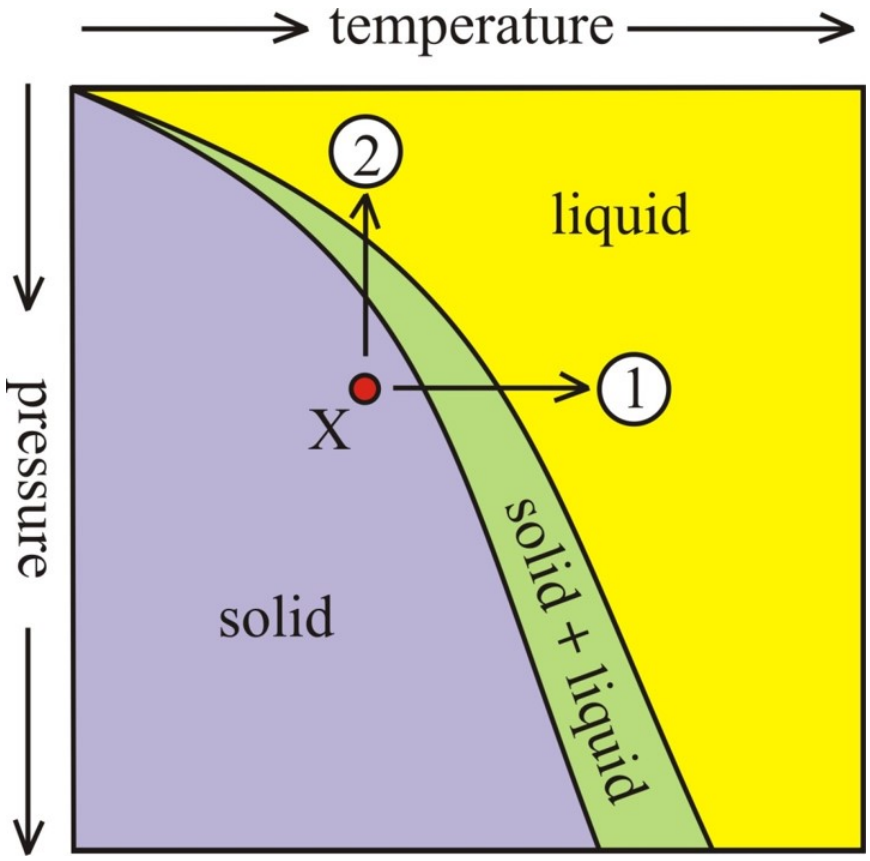


Figure 6.3.8: Two potential pathways for melting mantle peridotite: “1” shows the path of increasing temperature at constant pressure, probably the most familiar to human experience. But pathway “2” is what happens beneath oceanic ridge systems: Mantle rock rises beneath oceanic ridges, experiencing lower and lower pressures, but not changing its temperature at all. Despite no heat being added, it still melts. (Callan Bentley diagram.)

In seafloor spreading, **decompression melting** of the mantle produces mafic magma. This magma cools at various depths to produce new oceanic crust. When freshly minted, this oceanic crust makes a bathymetrically high feature on the seafloor, an **oceanic ridge** (also called a “mid-ocean ridge,” since many of them are in the middle of their host ocean basins).

No new heat needs to be added to mantle peridotite in order for it to melt via decompression: hot peridotite only needs to rise and experience lower pressures. Mantle convection brings warm rock upward, due to the warm rock’s low density. As it rises, it partially (not completely) melts beneath the oceanic ridge. This partial melting produces a magma that is mafic in composition from ultramafic source rocks.

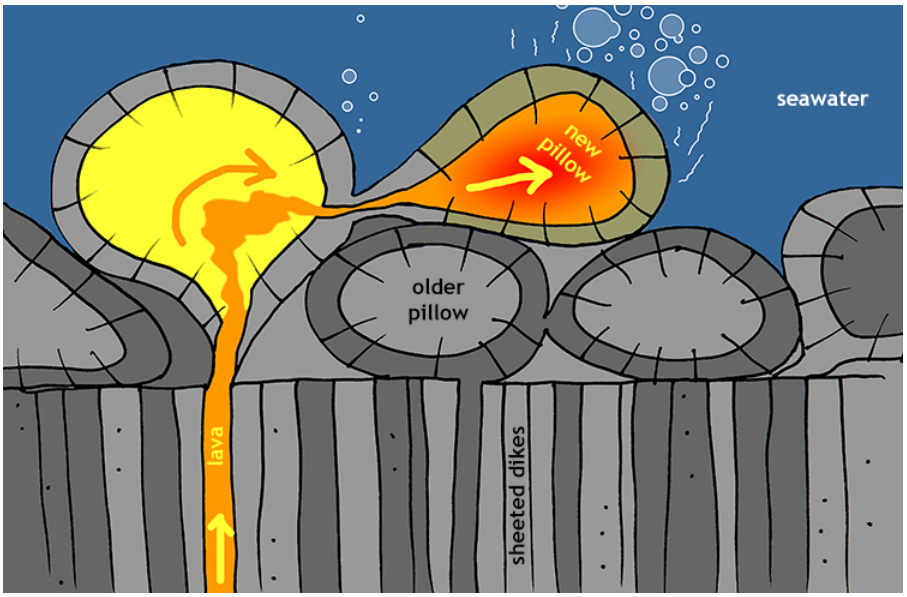


Figure 6.3.9: The formation of pillow structures occurs when basaltic lava erupts underwater. (Callan Bentley cartoon.)

As the plates diverge “overhead,” this mafic magma squirts into the crack between them. Some of the molten rock may make it all the way into the water and the bottom of the ocean, erupting to produce pillow basalts, while some magma cools rapidly and solidifies in the feeder crack itself, making a dike of basalt. Still more mafic magma cools slowly, deeper in the new oceanic crust, producing bodies of gabbro. This sequence of rock types is an ophiolite sequence – an important feature that we will return to again when we discuss convergent boundaries.

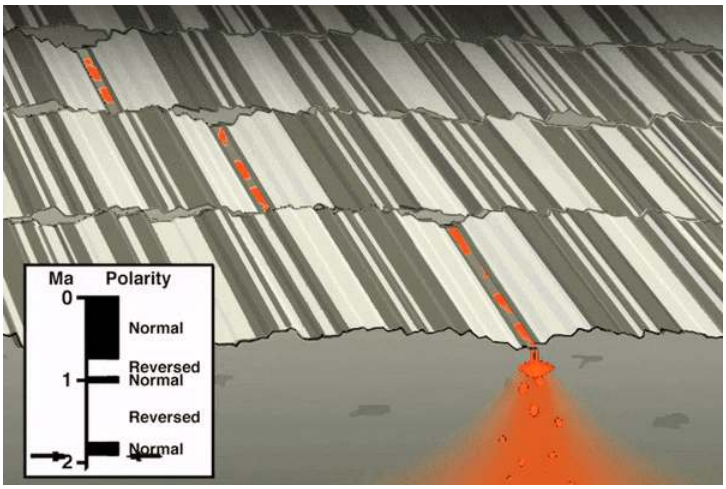


Figure 6.3.10: Spreading along several mid-ocean ridges, showing magnetic striping symmetry. (Animation by Tanya Atwater, UCSB.)

Seafloor spreading happens at different rates in different ocean basins. The East Pacific Rise generates roughly twice as much new seafloor in a given amount of time as the Mid-Atlantic Ridge. At both oceanic ridges, sections of active divergences are offset along perpendicular segments of transform faulting, giving the plates’ edge a shape similar to a staircase. Because the mafic magma that cools to make oceanic crust cools to make magnetically-susceptible minerals such as magnetite, which will rotate into alignment with Earth’s magnetic field before the lava or magma around them solidifies. The generation of new oceanic crust through seafloor spreading therefore records changes in the polarity of Earth’s magnetic field (which happen due to changes in the core, and have nothing to do with the surface processes of plate tectonics).

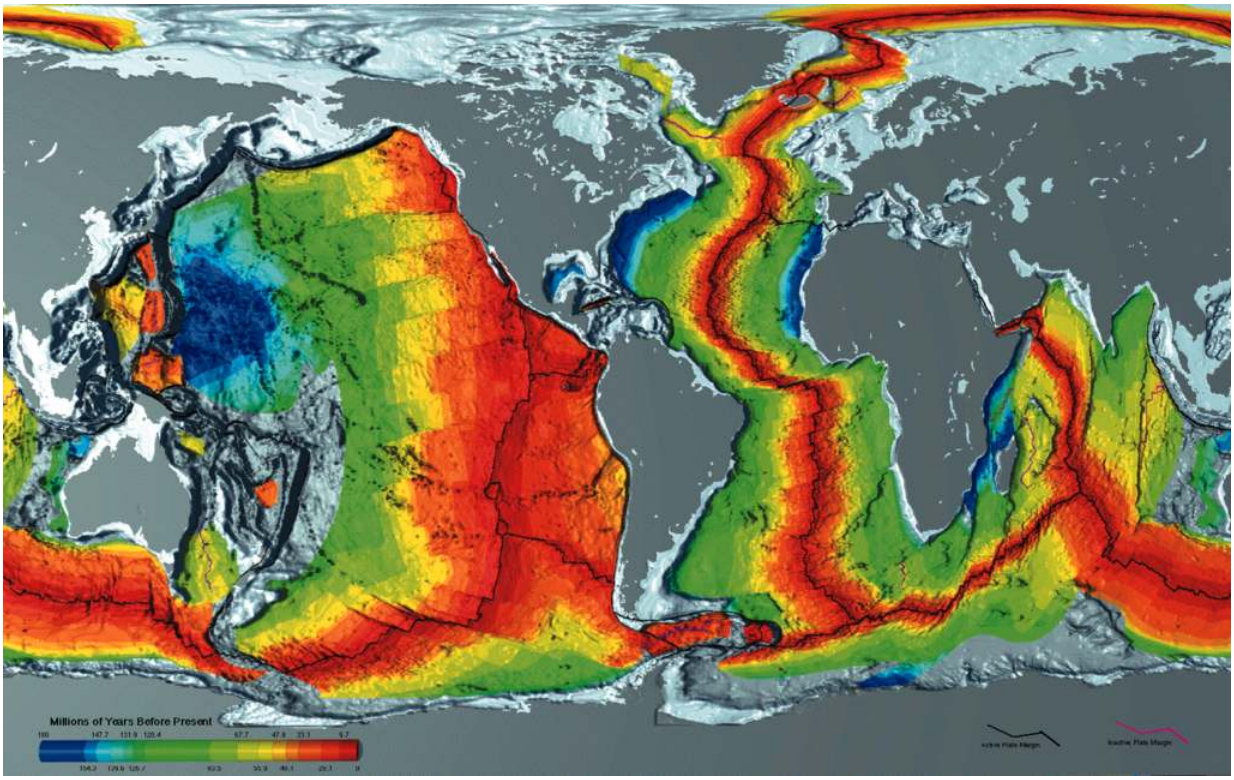


Figure 6.3.11: Age of the oceanic crust on Earth. (NOAA map.)

Over time, the oceanic crust cools. As it cools, it becomes more dense. More dense rock sinks. The lithospheric mantle also builds out below, thickening through time to a maximum of around 100 km by the time it's about 80 million years old. This adds substantial weight to the plate. This density-driven subsidence is combined with sedimentation atop the oceanic crust (adding more mass) to create the relatively smooth abyssal plains that floor most of the ocean basins around the world.

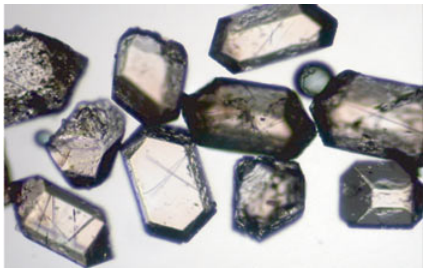


Figure 6.3.12: Tiny crystals called zircons are relatively common in gabbro of the oceanic crust. (Photo: Michael John Cheadle, University of Wyoming, via NSF)

We can determine the age of the seafloor by several methods: the first is direct dating of the crust via radioactive isotope decay systems in certain minerals. For instance, the gabbro of the deeper oceanic crust contains zircon crystals that host uranium→lead timekeepers. Another approach is analyzing the fossil content of sediments deposited on top of it. Index fossils in the deepest (oldest) sediment on a given patch of oceanic crust represent a time immediately after the new crust had formed due to volcanism. Finally, magnetostraigraphy of a stretch of oceanic crust can be measured and compared to the record of Earth's magnetic polarity reversals from well-dated locations on the continents. Depending on the characteristics of the particular location, some of these methods may be more applicable than others. In those lucky locations where multiple methods can be applied, the different techniques can be used to corroborate each other's age determinations.

Did I Get It? - Quiz

Exercise 6.3.1

What kind of faulting is found at divergent plate boundaries?

a. Right-lateral transform faulting

b. Left-lateral transform faulting

c. Mylonite zones up to several kilometers across

d. Thrust faulting

e. Normal faulting

Answer

e. Normal faulting

Exercise 6.3.2

What sort of volcanism dominates in divergent settings?

a. Intermediate

b. Porphyritic

c. Ultramafic

d. Mafic

e. Felsic

Answer

d. Mafic

Exercise 6.3.3

What is the character of sediment in rift valleys (at least during initial stages)?

a. Mature clastic sediment (quartz sandstone, shale)

b. Biochemical sediment: chert, chalk

c. Immature clastic sediment (conglomerate, arkose, shale)

d. Carbonate sediment (wackestone, packstone) + shale

Answer

c. Immature clastic sediment (conglomerate, arkose, shale)

Exercise 6.3.4

Given continued divergence, what do rift valleys turn into?

a. Volcanic island arcs

b. Mountain belts

c. Shear zones

d. Ocean basins

e. Pressure ridges

f. Transform faults

Answer

d. Ocean basins

? Exercise 6.3.5

Why is the Great Rift Valley of East Africa a good place to find fossils of human ancestors?

a. That's where the human ancestors were.

b. That's where the human ancestors were, *AND* it was a site of sedimentary deposition over the relevant time period.

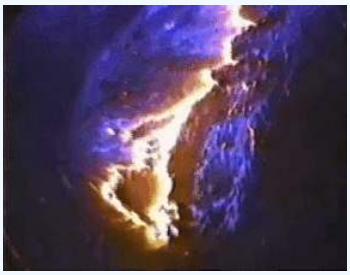
c. It was a site of sedimentary deposition over the relevant time period.

Answer

b. That's where the human ancestors were, *AND* it was a site of sedimentary deposition over the relevant time period.

? Exercise 6.3.6

Pillow basalts form at what sort of plate boundary?



a. Divergent

b. Transform

c. Convergent

Answer

a. Divergent

? Exercise 6.3.7

How does the mantle melt beneath oceanic ridge systems?

a. Adding heat to cold rock

b. Adding water to dry rock

c. Decompression of already hot rock

Answer

c. Decompression of already hot rock

Convergent boundaries

When two plates are headed toward each other, we describe the situation as **convergent**. Depending on whether the leading edge of these two plates consists of oceanic lithosphere or continental lithosphere, several different situations can result.

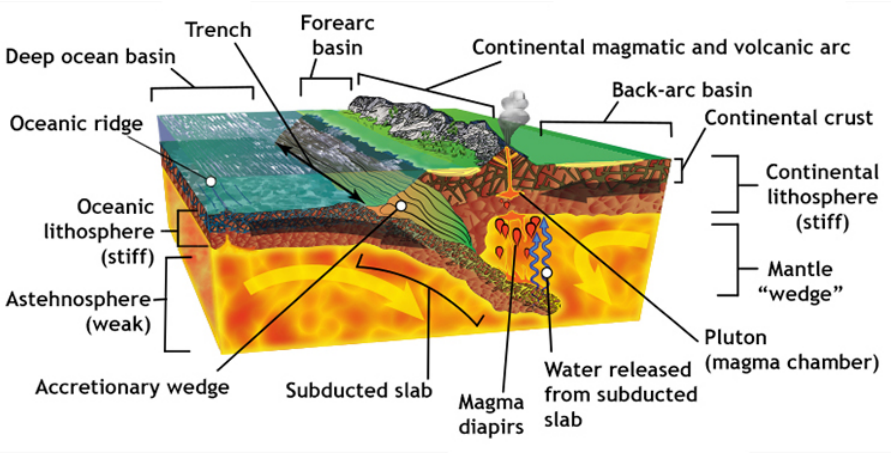


Figure 6.3.13: A cartoon showing key features in an oceanic/continental subduction zone. (Public domain via Wikimedia.)

When oceanic lithosphere converges with a continent, **subduction** happens. Subduction is the situation when the plate of oceanic lithosphere descends down into the mantle beneath the plate bearing the continent. This subduction is marked by many phenomena: an oceanic trench marks the place where subduction begins, but earthquakes are generated all along the subduction zone to great depths beneath the overriding plate. The earthquakes at this megathrust fault can be among the largest known. At the trench, any sediment atop the plate may be scraped off, building up in a thick jumbled pile called an **accretionary wedge**. The accretionary wedge can also include volcanic rocks from seamounts and intrusions, as well as slices of the oceanic lithosphere itself.



Figure 6.3.14: Osorno (left) and Calbuco (right) are classic Andean volcanoes in Chile. (Callan Bentley photo.)

As it descends, the subducted plate releases water into the surrounding mantle. This water enables the melting of the ultramafic mantle rock, generating mafic magma. This magma rises and pools beneath the base of the continental crust, transferring its heat. The minerals of the continental crust tend to have lower melting temperatures, and so they melt, resulting in intermediate composition magmas, which continue to ascend through the crust. They may lodge at depth, crystallizing to make plutons and batholiths, or they may find their way to the surface, where they erupt as volcanoes. The development of a **continental volcanic arc** parallel to the trench is a sure indication of oceanic/continental convergence.

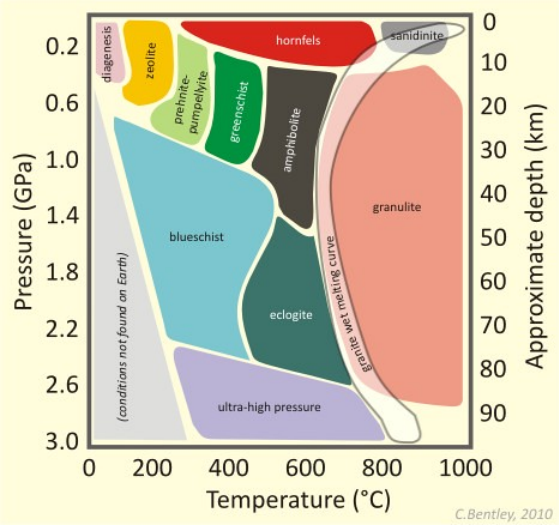


Figure 6.3.15: A plot showing the various metamorphic facies that result from different combinations of temperature and pressure. Note that blueschist and eclogite result from relatively low-*temperature* but relatively high-*pressure* metamorphism: conditions we would expect in a subduction zone. (Callan Bentley diagram)

As the subducted plate descends, its rock experiences higher pressures and temperatures. These trigger **metamorphosis**. Specifically, subduction zones see rocks descending faster than they can warm up, and so they experience **high pressures but relatively low temperatures**. This results in blueschist and eclogite metamorphic facies.

Several places in the modern world are examples of this kind of plate boundary, including the Cascadia volcanic arc in the Pacific Northwest of the United States of America, but the classic example is the western edge of South America, where subduction of the Nazca Plate has resulted in the Andes volcanic arc as well as some of the largest earthquakes ever recorded.

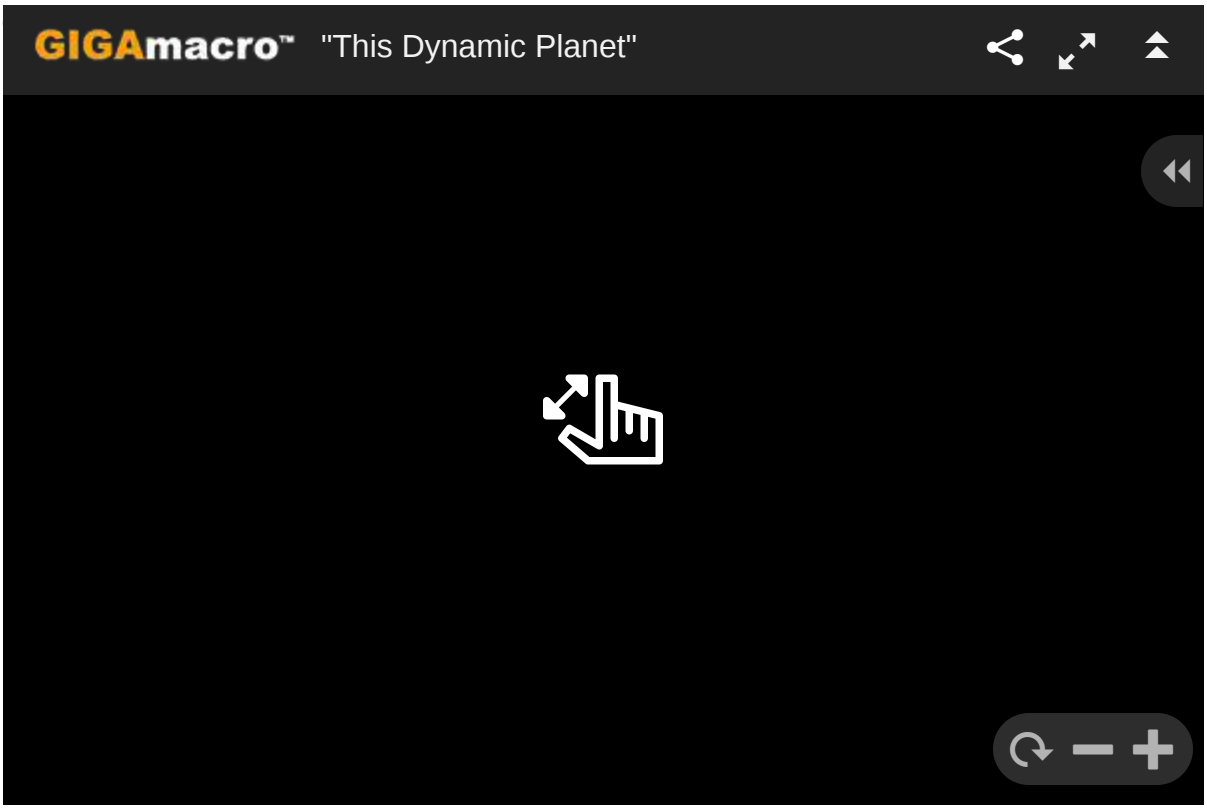


Figure 6.3.16: (Simkin, et al. (2006), USGS.)

Ancient subduction zone complexes can be found in in many places that are no longer experiencing subduction: coastal California, central Turkey, western France, and the Cyclades of Greece are a few examples.

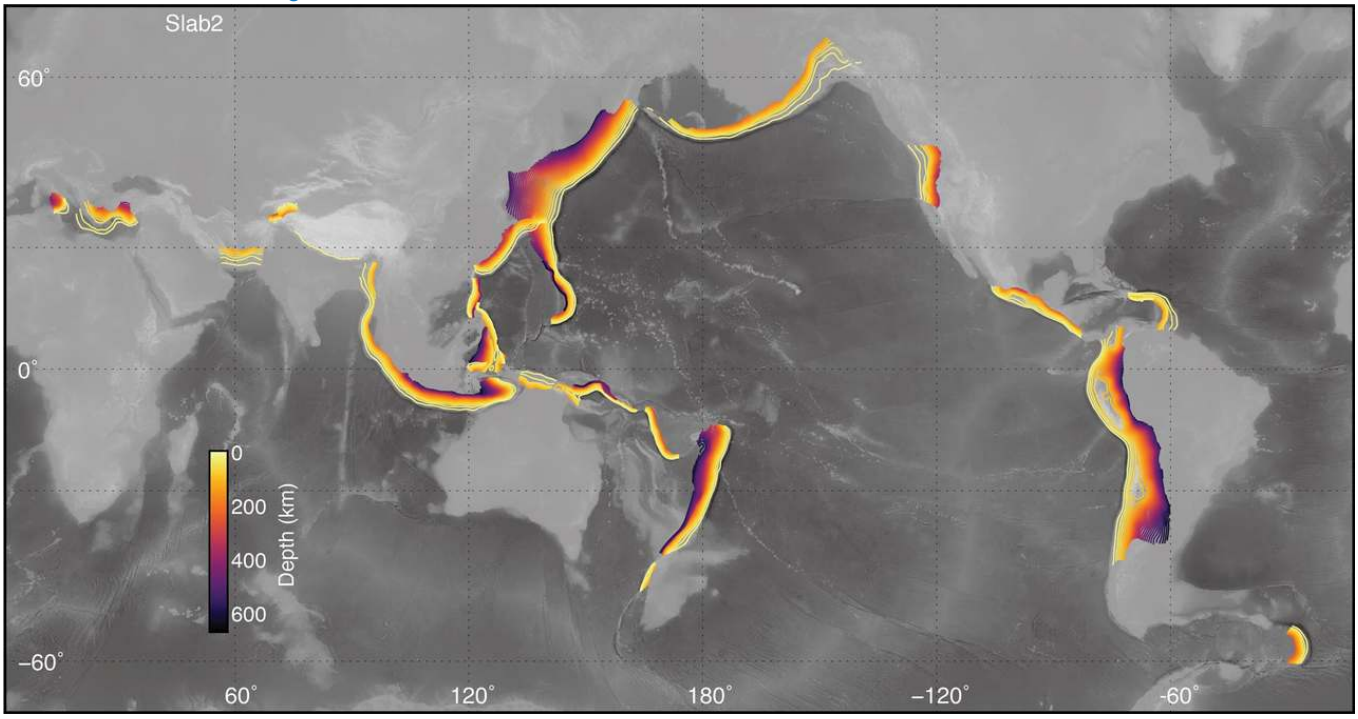


Figure 6.3.17: Global map of subduction zones, with subducted slabs contoured by depth. (Gavin P. Hayes (USGS) map.)

Volcanic island arcs result when two plates of oceanic lithosphere converge. As with the oceanic/continental situation, subduction results when one of the two plates descends beneath the other. Because they are both composed of the same rock types, the key variable is density based on age. Older oceanic lithosphere is colder oceanic lithosphere, and colder rock tends to be denser. For instance, in the western Pacific Ocean, ~200 Ma oceanic lithosphere of the Pacific Plate is subducting beneath ~20 Ma oceanic lithosphere of the Philippine Plate. In a match-up between old and young oceanic lithosphere, the young wins, and the old subducts. Because convergence with both continental lithosphere and young oceanic lithosphere "selects against" old oceanic lithosphere, there is very little of it still around. The oldest oceanic lithosphere left on this planet is Permian in age (~250 Ma), in the Mediterranean Sea.

So old oceanic lithosphere subducts, producing a trench and an accretionary wedge. Again the rock of the subducted slab undergoes dehydration reactions, releasing water into the mantle. This triggers melting of the mantle. The magma rises up to pierce through the overlying crust as a chain of volcanic islands. We call this a **volcanic island arc**.

In this overlay of the "Discovering Plate Boundaries" volcanology and topography maps, you can see this relationship plainly: in each location where we see a deep sea trench, it is paralleled by a volcanic island arc (line of red dots):



Figure 6.3.18: There is a strong coincidence of deep sea trenches and parallel belts of volcanoes. (Dale Sawyer's "Discovering Plate Boundaries" maps.)

Continental/Continental convergence



Figure 6.3.19: Aerial view over the raw mountains formed as India and Asia collide. (Photograph by Susan Hough; reproduced with permission.)

The final circumstance that can happen at a convergent plate boundary is when all the oceanic lithosphere separating two continents has been subducted, and the two continents (formerly separated by an ocean basin) finally meet. Because of their buoyancy, the continents cannot subduct very far. Instead, they crumple upward into mountains. Deep below the plate boundary, the plate crumples too, thickening into a “keel” that extends downward. Extending all the way through the thickest lithosphere on the planet, this is a young collisional **mountain belt**. The act of forming a mountain belt is an **orogeny** (from the ancient Greek for “mountain making”).

The classic modern example of a collisional mountain belt is the Himalaya, which have been forming for the past ~40 million years where the Indo-Australian Plate is converging with the Eurasian Plate. Frequent large, shallow earthquakes mark this collisional zone, but there is no volcanism. (There is neither decompression melting nor slab dehydration occurring at a continent-continent collision.) There is probably **partial melting** occurring within the continental crust at the base of the mountains, making migmatite, but insufficient volumes of felsic magma are produced to generate volcanism.



Figure 6.3.20: Gravel from a stream in the Himalayas, showing porphyroblasts of kyanite and garnet. (Photograph by Stephanie Sparks; reproduced with permission.)

One consequence of this collisional mountain belt is regional metamorphism of the rock there. Where clay-rich protoliths encounter elevated temperatures and pressures at the roots of these mountains, metamorphic rocks such as phyllite, schist, and gneiss are generated. Porphyroblasts of metamorphic minerals such as chlorite, garnet, kyanite, staurolite, and sillimanite grow. Each of these minerals is a chemical manifestation of how atoms in Earth materials react to changing conditions. They represent a physical "snapshot" of pressure/temperature conditions when the rocks equilibrated to novel conditions within a mountain belt. Dating metamorphic minerals using mineral/isotope systems is one way we can determine when an episode of mountain-building happened.

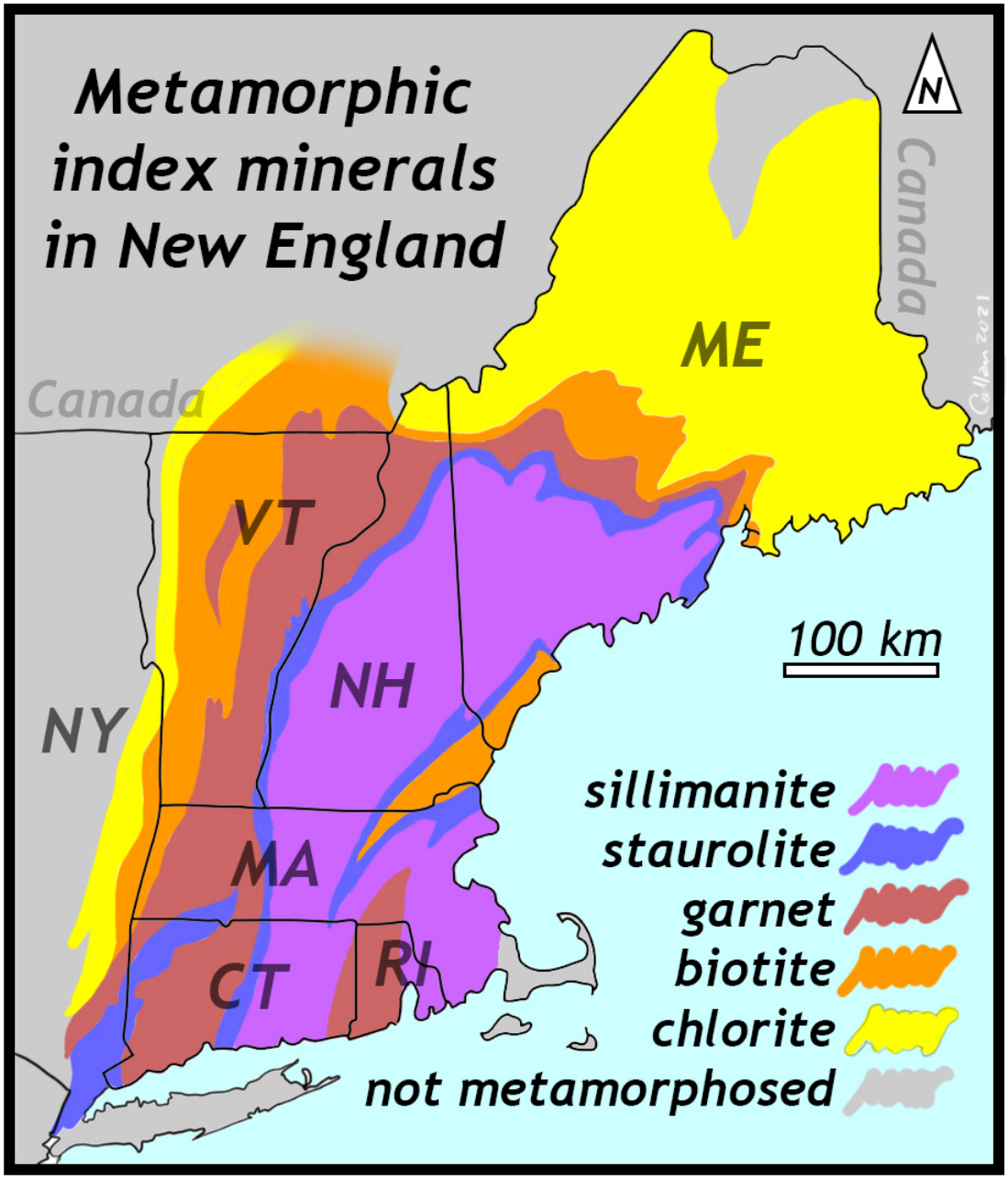


Figure 6.3.21: Metamorphic index minerals show where the most intense temperatures and pressures were in New England during the Acadian Orogeny. (Callan Bentley graphic.)

Even after the mountains' topographic relief has been subdued by erosion, these metamorphic minerals serve as indications of the high temperatures and pressures at the roots of the mountain belt. For instance, in New England (northeastern United States), mountain building cooked and squished the rocks during the late Devonian Acadian Orogeny. A belt of high-grade metamorphic

rocks stretches from south-central Connecticut north through central Massachusetts, where it splits. One of these sillimanite-bearing belts wraps around through eastern Massachusetts and south into Rhode Island, while the other widens at it trends north through New Hampshire and into southern Maine. The sillimanite zone is surrounded by zones of medium- and low-grade metamorphic rocks, as indicated by the minerals they bear as porphyroblasts. The staurolite didn't get as metamorphosed as the sillimanite, but the garnet was even less metamorphosed. Biotite and chlorite mark lower grades of metamorphism yet. The sillimanite-bearing belt marks the rough position of the highest peaks when the mountains were freshly built.



Figure 6.3.22: Zerrissene turbidites, folded by the Damara Orogeny, Namibia. (Photograph by Jay Kaufman; reproduced with permission.)

If the compressional stresses are strong, but the confining pressure and temperature are relatively low, rocks on the flanks of the mountain belt will be **deformed** through folding and thrust faulting. This zone of compressional deformation is distinct and pervasive in mountain belts across time and space. **Fold and thrust belts** form in pre-orogenic layered sedimentary and volcanic strata. Examples include the Pyrenees in Spain and France, the Cape Fold Belt in South Africa, the Valley & Ridge province of the Appalachians, the Ouachita Mountains in Arkansas, the Marathon Mountains in Texas, the Caledonides in Scandinavia, the Sevier Fold and Thrust Belt in the Rocky Mountains, the Damara Belt in Namibia, the Atlas and Anti-Atlas in Morocco, and the Alps in Europe.

Here are a few images highlighting fold and thrust belts around the world:



Figure 6.3.23: Folded sedimentary layers are a signature of compressional forces associated with mountain-building.

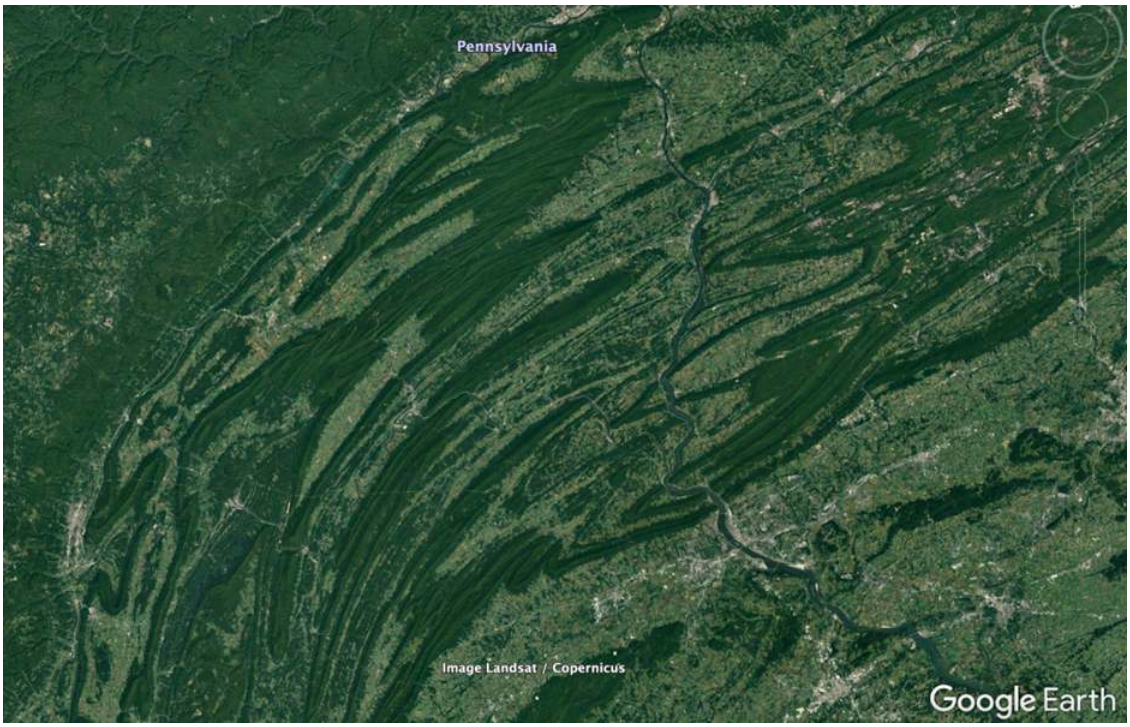


Figure 6.3.24: Google Earth Image of the Appalachian Mountain fold belt in western Pennsylvania at 200 km eye altitude.



Figure 6.3.25: Recumbent Fold from Recumbent fold, Port au Port Peninsula, Newfoundland. (CC BY-NC 4.0; from: https://openeducationalberta.ca/introductorystructuralgeology/chapter/e-folded-surfaces/)



Figure 6.3.26: Asymmetric anticline-syncline pair in marble and quartzite. (By: Marli Miller Photography.)

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Figure 6.3.27: Swift Dam's spillway showcases a 3D view of the contact between folded Mississippian carbonate and Cretaceous siliciclastic strata. This is the steep, short, eastern limb of the fold.



Figure 6.3.28: Angular unconformity, Telheiro Beach, Portugal, showing Triassic sandstone atop folded, eroded, Carboniferous graywacke and shale. (Photo by Joao Duarte, 2019.)



Figure 6.3.29: Neoproterozoic strata that are folded on the southeast coast of Newfoundland.



Figure 6.3.30: Cape Liptrap (southeast of Melbourne in Victoria, Australia). (Photo by Sandra McLaren.)

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Figure 6.3.31: Folded quartzite layers in the Cape Fold Belt, near Gansbaai, South Africa. (Callan Bentley photo.)

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Figure 6.3.32: Folded quartzite layers in the Cape Fold Belt, near Gansbaai, South Africa. (Callan Bentley photo.)

Did I Get It? - Quiz

? Exercise 6.3.1

Subduction occurs at what sort of plate boundary?

a. Divergent

b. Convergent

c. Transform

Answer

b. Convergent

? Exercise 6.3.2

Of the list of following features, which is *NOT* associated with a subduction zone?

a. Deposition of limestone

b. Deep sea trench

c. Accretionary wedge

d. Volcanic arc (either continental or island)

e. Earthquakes

Answer

a. Deposition of limestone

Exercise 6.3.3

If you explore an area, and find (1) foliated metamorphic rocks, (2) granite plutons, and (3) thrust faults + folded strata, what have you discovered?

a. A right-lateral transform fault

b. A rift valley

c. A mountain belt

d. A deep-sea trench

e. An oceanic ridge

Answer

c. A mountain belt

Exercise 6.3.4

If two plates converge, and their leading edges are both composed of oceanic lithosphere, which one will subduct?

a. The younger one, since it will be warmer and less dense.

b. The younger one, since it will be colder and denser.

c. The older one, since it will be colder and denser.

d. The older one, since it will be warmer and less dense.

Answer

c. The older one, since it will be colder and denser.

Exercise 6.3.5

Which type of plate boundary produces the largest earthquakes?

a. Divergent

b. Convergent

c. Transform

Answer

b. Convergent

Plate interiors



Figure 6.3.33: Assateague Beach, Virginia. Note the extremely subdued topography and sand (accumulating mature sediment). (Photo by Doug Kerr via Wikimedia.)

Though most of the geological "action" happens at plate boundaries (edges), the vast interiors of plates merit some attention, too. Largely they move as a coherent block, translating in a common direction or rotating around some axis of motion. **Passive margins** are the "trailing" edges of continents within a plate. For instance, consider the North American continent as it is today. As it moves from east toward the west, the western edge is the active margin (with Cascadia subduction and two giant transform fault zones), and the eastern edge is the passive margin, trailing along with no active tectonics. A place like Virginia is the edge of the *continent*, but the middle of the *plate*. No mountains are being actively built in modern Virginia, but old mountains are being eroded away. Particles of sand and mud with origins in the Appalachian Mountains are tumbled downstream and deposited along the western edge of the the Atlantic Ocean.

A passive margin is a site of tectonic calm. Without plate tectonics to rough up the landscape, the topographic relief is minimal. Cool crust is denser, and subsides. This allows the accumulation of deposits of mature sedimentary strata. These sediments (quartz sand, mud, shelly carbonate material) have been laid down along the modern mid-Atlantic margin in the thick layers of the Coastal Plain and continental shelf.

Other modern examples of passive margins include eastern South America, Western Africa, Western Australia, and southern India.

Did I Get It? - Quiz

? Exercise 6.3.1

Which of the following characteristics is consistent with forming along the passive margin of a continent?

a. Regional metamorphism.

b. Deposition of mature sediment such as quartz-rich sand.

c. Eruption of mafic lava.

d. Contact metamorphism.

e. Deposition of immature sediment such as feldspar-rich sand.

f. Eruption of felsic lava.

Answer

b. Deposition of mature sediment such as quartz-rich sand.

? Exercise 6.3.2

What defines a passive margin?

a. It is the edge of the continent, but not the edge of the plate.

b. It is neither the edge of the continent, nor the plate.

c. It is the edge of the plate, but not the edge of the continent.

d. It is the edge of both the continent and the plate.

Answer

a. It is the edge of the continent, but not the edge of the plate.

Variations and mashups

While it is easiest to conceptualize plate boundary types with a neat triumvirate of possibilities, (1) transform, (2) divergent, and (3) convergent, the real world is frequently more complex. Some plate boundary settings "mix" elements of convergent and transform motion. Others mix divergent and transform motion. These situations arise when the direction of plate motion is neither perpendicular nor parallel to the orientation of the plate boundary. Instead, the motion is ***oblique*** to the boundary.

Transpression

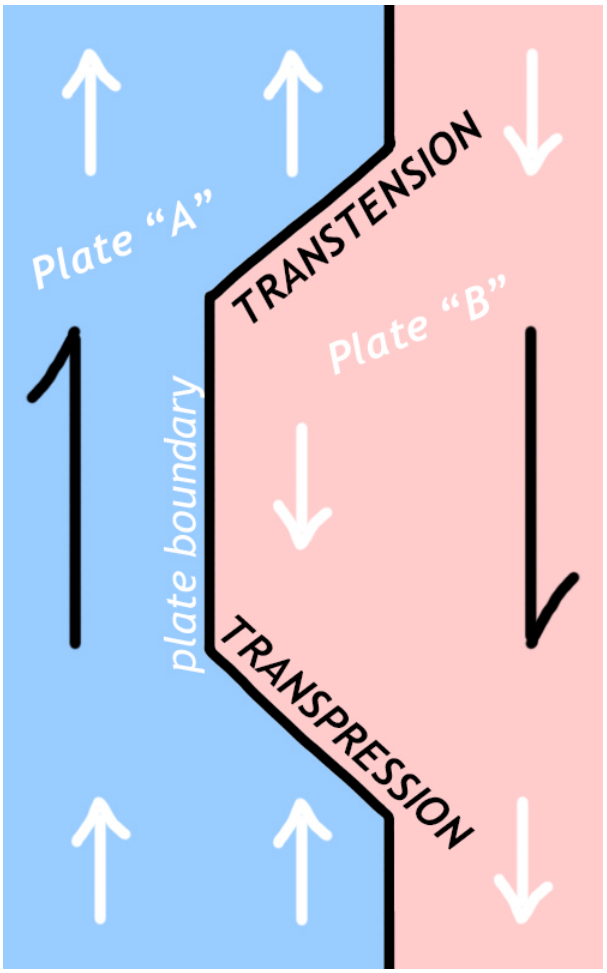


Figure 6.3.34: Transpression and transtension occur along plate boundaries that are oriented obliquely to the direction of plate motion. (Drawing by Callan Bentley)

Transpression is oblique convergence. Two plates converge with one another, but the boundary is oblique to the convergence direction. Rocks caught in a zone of compression are squished and sheared.

If this occurs at sufficient depth for dynamic metamorphism, they can develop an internal **S-C fabric**. S-C fabrics are characterized by a wavy, lenticular "grain" to the rock. The waviness comes from the intersection of two planar fabrics: the foliation ("S") that is typical of so many metamorphic rocks, plus small shear bands ("C") that cut across and merge with the foliation. Conveniently, the S fabric tends to be wavy, much like the letter S. The C surfaces tend to be more planar, **c**utting across the S surfaces.

As an example of transpression that might be relevant to a geologist deciphering the tectonic history of a region, consider the GIGAmacro image below. It shows a slab of deformed granodiorite from a Mesozoic-aged transpressional shear zone in the Sierra Nevada of California. Examine the orientation of the light and dark minerals. You can see a diagonal foliation (the "S" surface) as

well as a left-to-right series of small shear bands (the “C” surface). This particular pattern indicates that the rock was being compressed “top to bottom” as well as being sheared right-laterally (top to the right):



Figure 6.3.35: Cross-sectional view of S-C fabric in Rush Creek Granodiorite, near Gem Lake, California. (GIGAmacro by Robin Rohrback.)

S-C fabrics are the deep record of transpression, and are thus reasonably likely to be preserved in the geological record over the long term. Eventually (through uplift, erosion, and exhumation), historical geologists can find S-C fabrics in transpressionally sheared rocks, and extract from them information about which way the plates were moving when those rocks were deformed.

Closer to the surface, transpression may cause the extrusion of soil and rock material vertically. Rock moves upward through ductile flow at depth, but through brittle offsets at shallower depths. This net upward movement of Earth materials makes topographically “high” features, called pressure ridges. We mentioned them briefly above in the context of transform faults where the fault’s orientation is oblique to the relative motion along the fault.

Landscape features are useful to the historical geologist for understanding recent tectonic activity, but the shape of the land is ephemeral, and unlikely to be preserved as useful information over the geological long-term.

Transtension

Transtension is oblique divergence. Two plates diverge from one another, but the boundary is oblique to the divergence direction. Unlike transpression, transtension tends not to last very long as a discrete entity unto itself. Instead, it will rapidly organize itself into a typical “zigzag” pattern of divergent rift valleys and transform faults, as we see along the global oceanic ridge system. For instance, a recent example can be found on the seafloor of the Gulf of California (a.k.a. the “Sea of Cortez”). These, we can see a distinct stairstep pattern to the bathymetry. The Baja California peninsula has been recently ripped off the coast of mainland Mexico and transported to the northwest along a zone of transtension. This transtension has been accommodated through the development of a series of very ***short*** segments of oceanic ridge (spreading centers), interspersed along relatively ***long*** stretches of transform faults. The transform faults are oriented precisely perpendicular to the ridge axes:

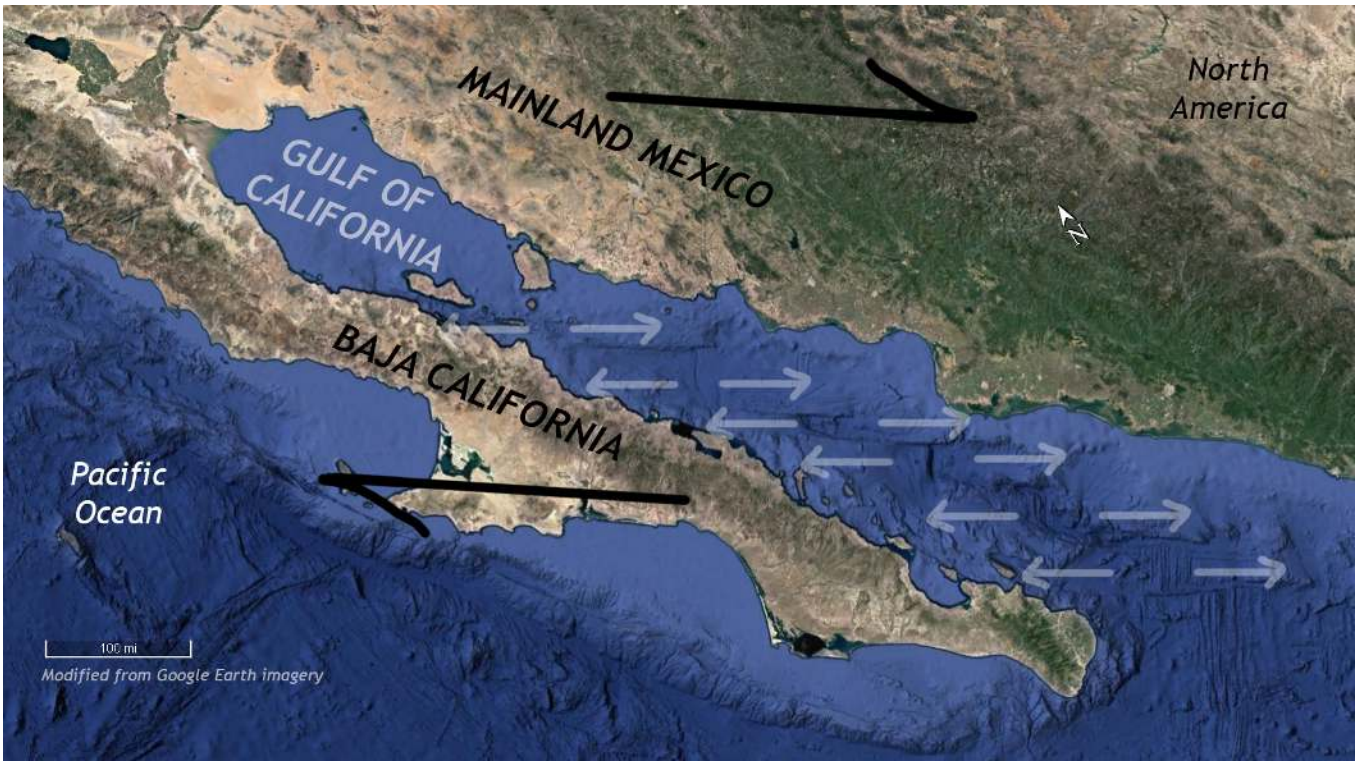
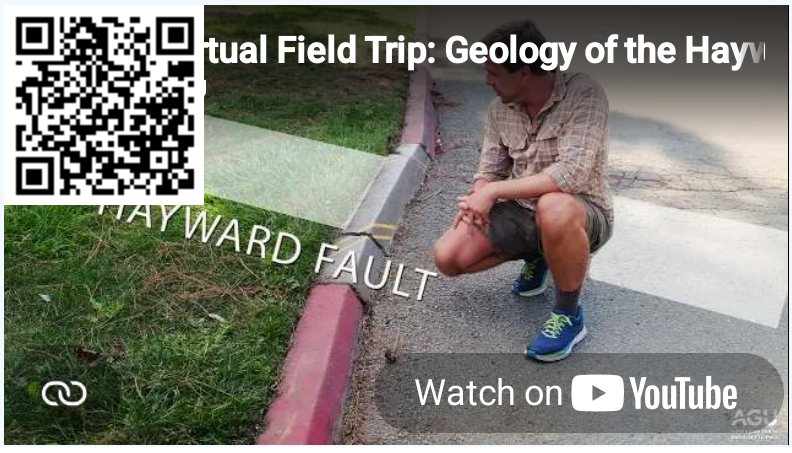


Figure 6.3.36: Baja California has separated from mainland Mexico along a transtensional zone, the Gulf of California. (Modified from a Google Earth image.)

Example 6.3.1: Along the Hayward Fault

Let us now summarize the manifestations of transpression and transtension with a short video looking at the small-scale landforms in Fremont, California's Central Park, along the trace of the (transform) Hayward Fault:



Did I Get It? - Quiz

Exercise 6.3.1

Oblique convergence is also known as _____.

a. transpression

b. transtension

Answer

a. transpression

Exercise 6.3.2

Land surface evidence of ongoing transpression includes _____.

a. migratory birds

b. S-C fabrics

c. granite magmatism

d. pressure ridges

e. sag ponds

Answer

d. pressure ridges

Exercise 6.3.3

A sag pond is evidence of _____.

a. ductile flow

b. transtension

c. transpression

Answer

b. transtension

Exercise 6.3.4

What does an S-C fabric indicate?

a. Ductile flow and recrystallization of rock experiencing transpression deep in the crust.

b. Violent volcanic eruptions.

c. Brittle deformation of rock along a fault zone.

d. Subduction.

Answer

a. Ductile flow and recrystallization of rock experiencing transpression deep in the crust.

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6.4: The historical record of plate interactions

When plates interact in different ways (convergent, divergent, *etc.*), various phenomena result. Some of these phenomena are short-lived and thus unlikely to end up preserved in the "deep time" geologic record (offset streams at transform boundaries, for instance). Others create more durable signatures that can be identified in rock millions and billions of years later (such as metamorphic belts of a consistent age, or sedimentary basins that record rifting). When a region experiences divergent tectonics, it forms one set of geological signatures, while a completely different set of rocks, structures, and topography forms as a result of subduction tectonics.

Mountain belts

Convergent plate boundaries between blocks of continental lithosphere are recorded in the geologic record through the construction of **mountain belts**. When fresh, these mountain belts are mountainous: that is, they show high relief landscapes that occur in linear belts, like the modern Alps or Himalayas. However, intense topographic relief is unlikely to persist in the geologic record. Over time, mountains are worn down by erosion: between landslides, glaciers, and rivers, once-lofty rock is taken downhill, reducing the overall topographic relief.

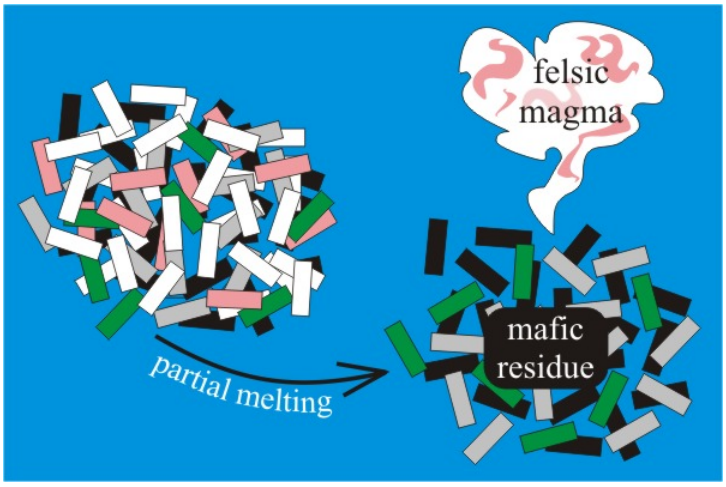


Figure 6.4.1: Partial melting makes magmas that are more felsic than their source rocks, and leaves behind solid rocks that are more mafic. (Cartoon by Callan Bentley.)

However, the signatures of mountain-building at the **roots** of those same mountains may be preserved. These include (1) folded and thrust-faulted sedimentary strata on the flanks of the mountain-belt and (2) foliated metamorphic rocks in the core of the mountain belt. If orogenic temperatures were high enough, some of the minerals in some of the metamorphic rock may melt, forming migmatite. Sufficient partial melting will generate (3) bodies of felsic magma that can cool to form granite plutons. These durable geologic signatures show us where the mountains used to be, even after the mountains themselves are gone.

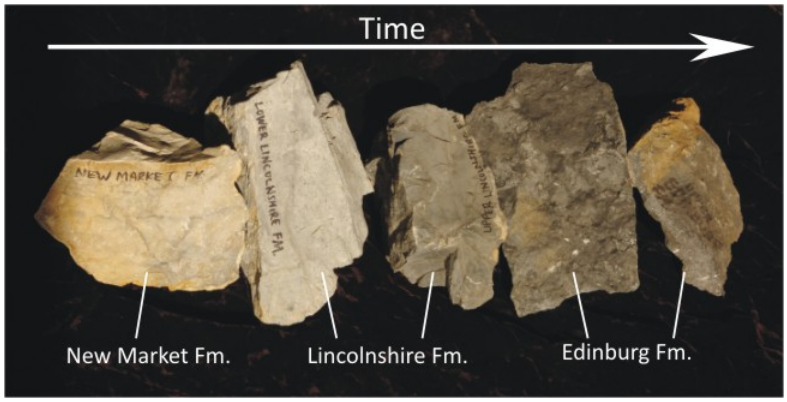


Figure 6.4.2: Over the course of the late Ordovician, light-colored shallow limestones gave way to increasingly dark deep-water limestones and shales. This "dirtiness" signals the onset of nearby mountain-building. (Callan Bentley photo.)

Additionally, consider the situation from the perspective of the stuff that's removed: the sediment ***generated*** by the erosion of the mountains has to go ***someplace***. It goes downhill, and ends up deposited somewhere else, perhaps nearby or perhaps far away, but it adds clastic sediment to the geologic record somewhere adjacent to the mountain belt. A shift in the sedimentary character of a given stratigraphic column from carbonate to clastic can signal an orogeny turning on. Because these clastic strata can contain fossils, we can use their age to determine the timing of mountain-building. A subsequent reversion to carbonate sedimentation thereafter will be a signal that the mountains have been removed as a source of sediment.

Marine sediment shed off a neighboring mountain belt is collectively dubbed ***flysch***. It consists of black shale, bentonite, graywacke, and conglomerate. Terrestrial sediment shed off a neighboring mountain belt is collectively dubbed ***molasse***. Typically it consists of red shale, submature river sandstones, and conglomerate. Before early geologists had come to terms with the non-intuitive ways that metamorphic and plutonic rocks formed, these clastic strata were the first recognized signals of ancient mountain building. The key recognition here is that the bajillions of particles that make up clastic strata represent a tremendous amount of eroded source rock, a huge volume equivalent to vanished mountain ranges.

Cratons and the accretionary growth of continents

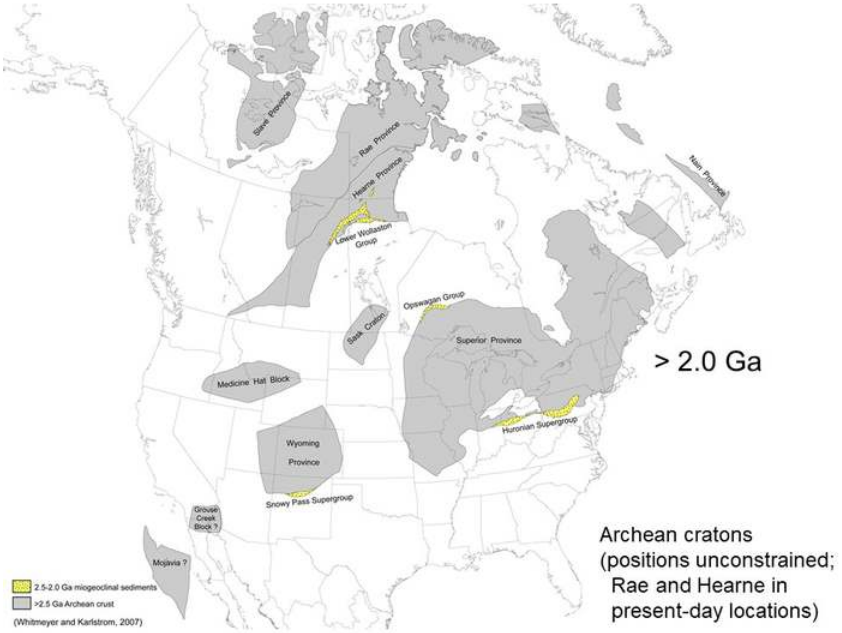


Figure 6.4.3: A quick summary of the evolution of North America over the Archean & Proterozoic eons from Whitmeyer & Karlstrom (2007). Explore this dataset at your own pace with this Google Earth Pro file. (Animation derived from Whitmeyer & Karlstrom (2007).)

Over time, collisional events between masses of continental crust cause small blobs of crust to glom together and create larger, composite blobs of crust. This map shows the age of the various blocks of crust that make up the North American continent. If the crustal components are old, we tend to call them "cratons," but if they are more recently tacked on, the preferred term is "terrane." Whatever the name, the pattern is plain: the oldest bits of crust are surrounded by younger bits of crust. Continents start off relatively small, as resistant blobs of low-density material that survive subsequent collisions with neighboring blobs. Through convergence, they are sutured together and stay together. Seafloor sediment, volcanic island arcs, and other small masses of continental crust merge and stick: continents grow through the accretion of new terranes on their margins.

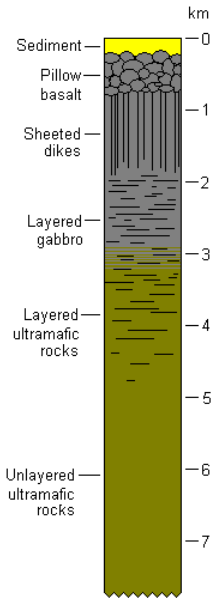


Figure 6.4.4: Rock units in an ophiolite sequence reflect the structure of the oceanic lithosphere. (Wikimedia image.)

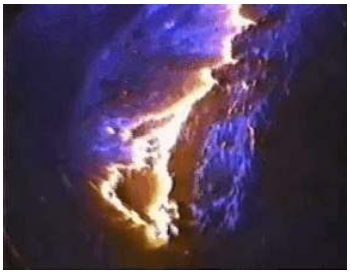


Figure 6.4.5: Modern pillow lava forming underwater.

Earlier, we described the structure of the oceanic lithosphere as it forms through seafloor spreading at a divergent boundary. From bottom to top, it consists of mantle peridotite, gabbro, sheeted dikes of basalt, pillow basalt, and that may be topped with deep sea sediments, such as chert or shale. This sequence is both the structure of the oceanic crust as well as the structure of small slivers of rock we often find between accreted terranes. In other words, on dry land, in the heart of mountain belts, we often find what appear to be scraps of oceanic crust. These are called ophiolites. Before the advent of plate tectonics, it was difficult to explain ophiolites, but now we can recognize them as remnant shreds of the oceanic lithosphere that used to separate the two terranes – the scant remains of an ocean basin that was otherwise mostly destroyed through the process of subduction.

For instance, at Gros Morne National Park in Newfoundland, Canada, there is a huge area of exposed peridotite. Though the fresh peridotite would be green because of its olivine content, the wet climate of Newfoundland has rusted it all at Earth's surface, making the landscape surface orange. Adjacent to the peridotite is layered gabbro, and adjacent to that are sheeted dikes and pillow basalts.

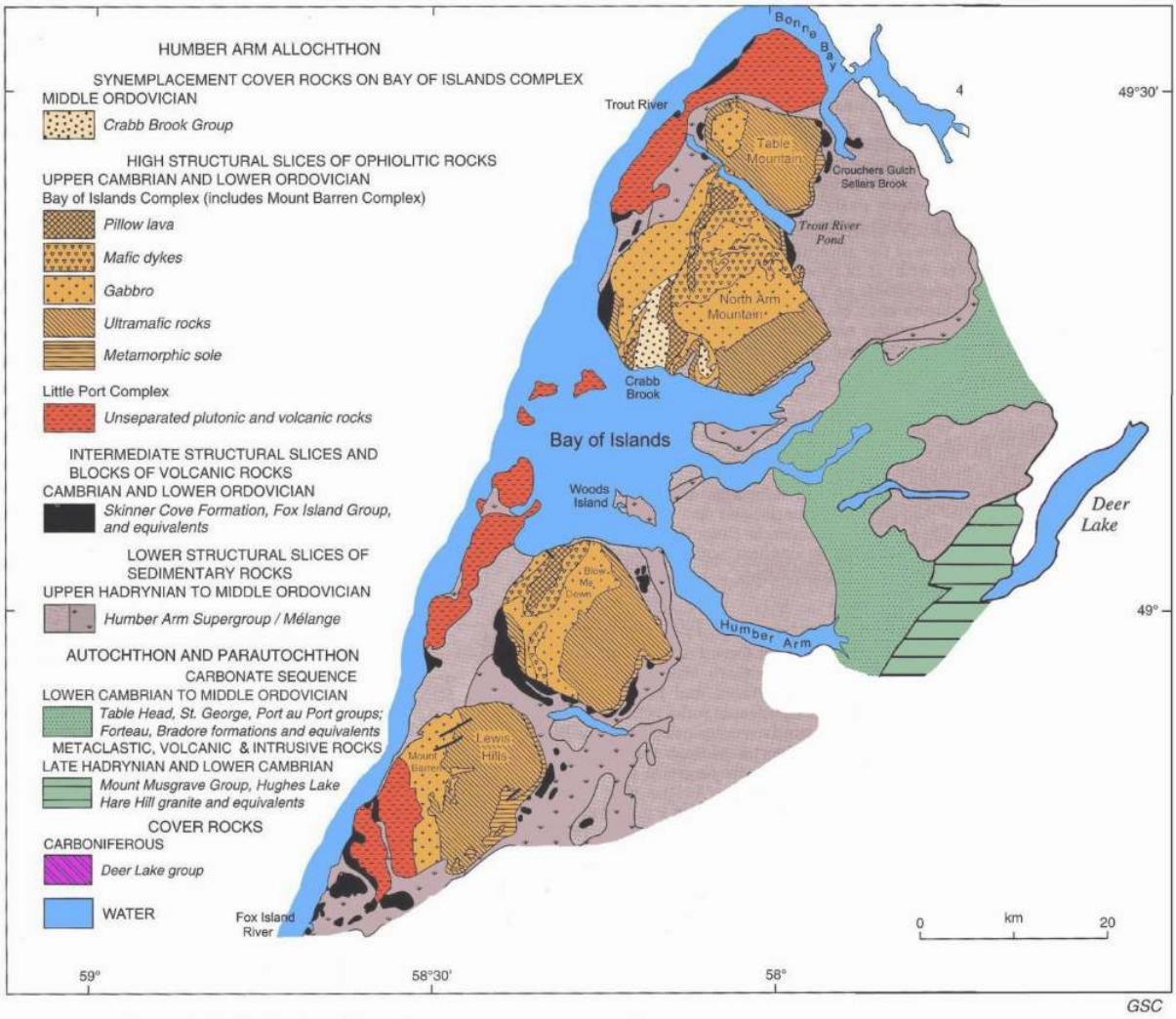


Figure 6.4.6: Distribution of discrete volcanic blocks and slices within the Humber Arm Allochthon and internal lithic units of the Bay of Islands Complex. Geologic map of the area immediately southwest of southwest Gros Morne National Park, Newfoundland. (Geological Survey of Canada map.)

The record of subduction



Figure 6.4.7: Blueschist is evidence of a high-pressure & low-temperature metamorphic regime. (Callan Bentley photo.)

Subduction is a key component of plate tectonics. To recognize places where subduction has occurred in the geologic past, we look for metamorphic, igneous, and structural signatures. One of the basic aspects of modern subduction is the parallelism of oceanic trench and volcanic arc. These two settings have different blends of temperature and pressure, resulting in different "flavors" of metamorphic rocks. The trench and Wadati-Benioff zone will result in **high pressure + low temperature** metamorphic rocks, while the neighboring magmatic arc will produce **moderate pressure + high temperature** metamorphic rocks. These "paired metamorphic belts" allow us to distinguish the earliest records of subduction from the geologic record, and the rather different tectonic situation that preceded subduction during the Archean.

Few places preserve this better than coastal northern California. The rock there formed (and deformed) during the Mesozoic subduction of the Farallon Plate beneath the western North American Plate.

Did I Get It? - Quiz

? Exercise 6.4.1

Interpret the following geological characteristics as a plate boundary setting: blueschist, accretionary wedge, thrust faults.

a. Transform

b. Convergent (mountain-building)

c. Divergent (continental rifting)

d. Convergent (subduction)

e. Divergent (seafloor spreading)

Answer

d. Convergent (subduction)

? Exercise 6.4.2

Interpret the following geological characteristics as a plate boundary setting: mafic volcanism, immature clastic sediments, normal faults.

a. Convergent (mountain-building)

b. Convergent (subduction)

c. Divergent (rift valley in continental crust)

d. Transform

e. Divergent (seafloor spreading)

Answer

c. Divergent (rift valley in continental crust)

? Exercise 6.4.3

Interpret the following geological characteristics as a plate boundary setting: foliated metamorphic rocks, granite plutons, folding, ophiolites.

a. Divergent (seafloor spreading)

b. Convergent (subduction)

c. Divergent (continental rifting)

d. Convergent (mountain-building)

e. Transform

Answer

d. Convergent (mountain-building)

? Exercise 6.4.4

Continents get larger through time due to _____.

a. docking of cratons or accretion of terranes

b. rifting

c. eruption of pillow basalts

d. erosion

Answer

a. docking of cratons or accretion of terranes

? Exercise 6.4.5

Ophiolites represent what process(es)?

a. Seafloor spreading followed by mountain-building

b. Left-lateral transform faulting

c. Subduction

d. Seafloor spreading

e. Mountain-building

Answer

a. Seafloor spreading followed by mountain-building

Sedimentary basins

An important consequence of plate tectonics for Historical Geology is how the different plate boundary settings give rise to circumstances where sediments can accumulate and be preserved over the geological long term. Substantial loci of sedimentary deposition are referred to as **basins**.

Transform

Because transform plate boundaries largely conserve the land surface as it is, merely offsetting it in opposite directions, they do not tend to be exceptionally good basin formers. However, where there are small “jogs” in the orientation of the fault trace, transtension can open up relatively small wrench basins.

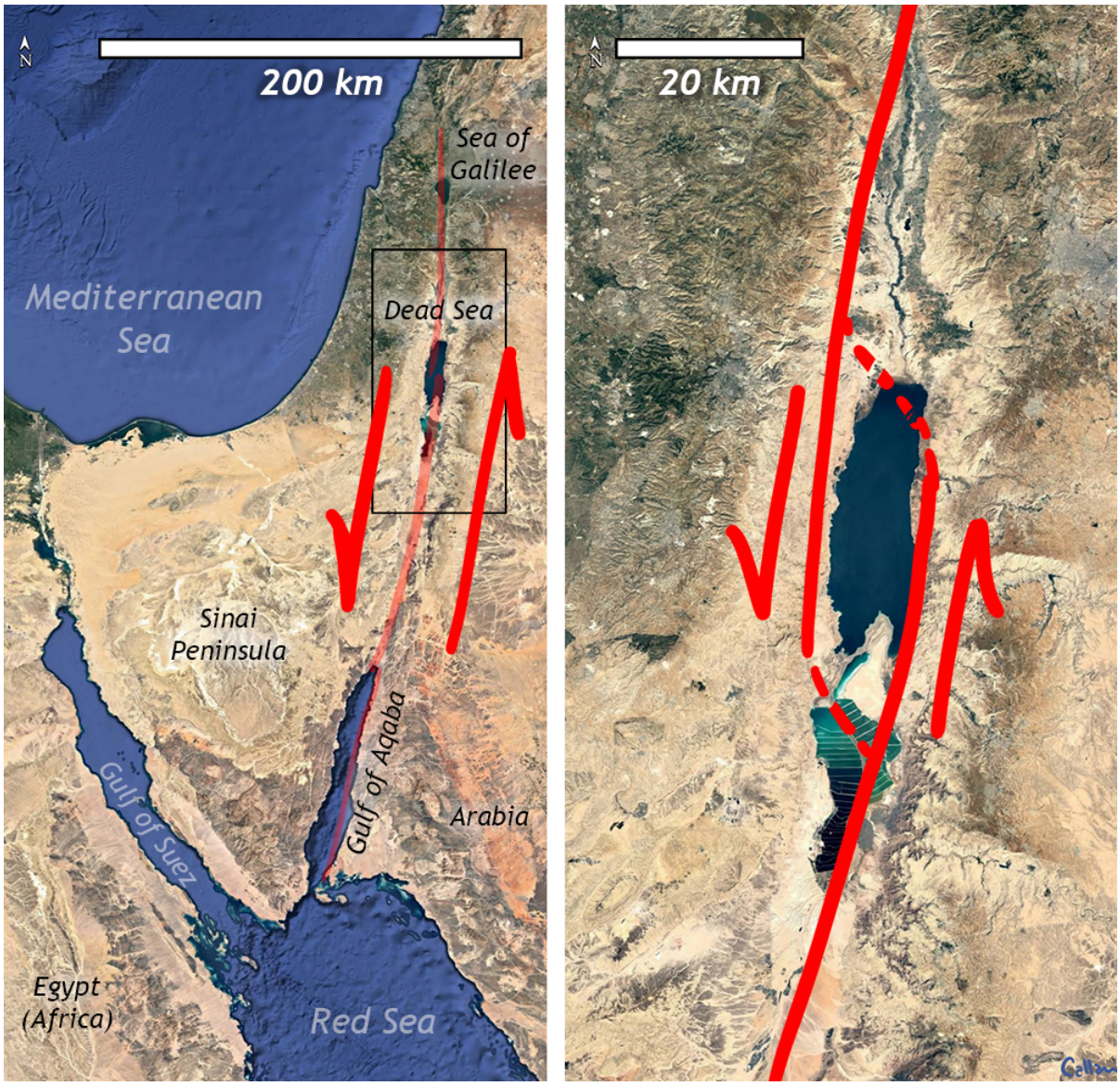


Figure 6.4.8: Left steps in a left lateral transform fault between Sinai and Arabia have opened a series of wrench basins including the Dead Sea. (Modified from Google Earth imagery.)

Wrench basins form when a right-lateral fault steps over to the right, or a left-lateral fault steps over to the left. The land between the end of one fault segment and the beginning of the next will sag. As we saw above, on a small scale this can generate a sag pond (or a low spot in a parking lot!), but on a larger scale it can produce substantial basins: these catchments dilate and create a topographic low spot into which sediment can pour from the surrounding highlands.

The Dead Sea, on the boundary between Israel, the West Bank, and Jordan, is an excellent modern example. Here, the Dead Sea Transform Fault, a left-lateral series of faults between the Sinai Peninsula and the Arabia Plate. As at Aqaba, and the Sea of Galilee, left steps in the fault system have stretched open low spots in the landscape, filled by water. At Aqaba, it's an offshoot of the Red Sea. The Sea of Galilee is landlocked, and filled with freshwater.



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Figure 6.4.9: Halite gravel from a beach on the western Dead Sea. (Photo by Daniel Garcia-Castellanos; reproduced with permission.)

The Dead Sea is the lowest spot on land in the entire world. *(There are of course, deeper spots in the ocean basins, such as the Marianas Trench.)* The hot, dry climate and lack of replenishing water (especially in recent years) have resulted in an extremely salty body of water occupying the basin, currently at 430.5 meters below sea level (-1,412 feet in elevation), but dropping further every year.

Other modern examples include the Salton Sea (on the San Andreas Fault) and the Sea of Marmara (south of İstanbul on the North Anatolian Fault).

Wrench basins are not just a modern phenomenon, however. In true uniformitarian spirit, we find many preserved in numerous contexts in the geologic record. They can serve as excellent repositories of local lithologic diversity in situations where the source rocks have been completely eroded away.



Figure 6.4.10: Strained metaconglomerate of the Archean-aged Seine Group, southern Ontario. (Callan Bentley photo.)

For instance, in the heart of the Superior Craton of North American, there are a series of wrench basins that captured cobbles derived from the surrounding highlands: mountains that haven't seen topographic relief since Archean times. Thanks to the preservation of these wrench basin conglomerates, we can see not only which areas were made low by tectonics, but what composition rocks made up the surrounding high country.



Figure 6.4.11: Diabase is a mafic igneous rock found in the Culpeper Basin of Virginia. Note the difference between the rusty weathering rind and the fresh surface, which is dark. (Callan Bentley photo.)

The principal variety of basin that forms at a divergent setting is a rift basin. When rift basins are fresh, they accumulate immature clastic sediment. Over time, especially in large rift basins, the sediments may mature towards a higher proportion of quartz and clay. Some even see limestones at positions far removed from the basin margins. Volcanism is important too: as the crust stretches in a rift, it lowers the pressure on the underlying mantle. Decompression melting follows, which produces mafic magma. Eruption of basalt and intrusion of gabbro and diabase are common igneous features of rift basins.

RIFT BASINS

As we've already mentioned, the Great Rift Valley system of east Africa offers an outstanding modern example of a rift basin. Because of its location and timing, this is ***the*** place to look for fossils of human ancestors. Other modern examples of rift basins include Lake Baikal in Siberia (and a smaller equivalent to the south in Mongolia, Lake Khovsgol). The valley of Þingvellir *("Thingvellir")* in southwest Iceland, is celebrated by Icelanders as the seat of the world's oldest functioning democracy, but geologists see it in a different light: it is a place where the Mid-Atlantic Ridge can be seen above sea level. All of these basins are bound by normal faults.

A fine example of an ancient rift basin can be found in central Virginia, where the Culpeper Basin is the largest of a dozen basins that preserve terrestrial sediments (and mafic igneous rocks) related to the break-up of the supercontinent Pangaea and the initial opening of the Atlantic Ocean.

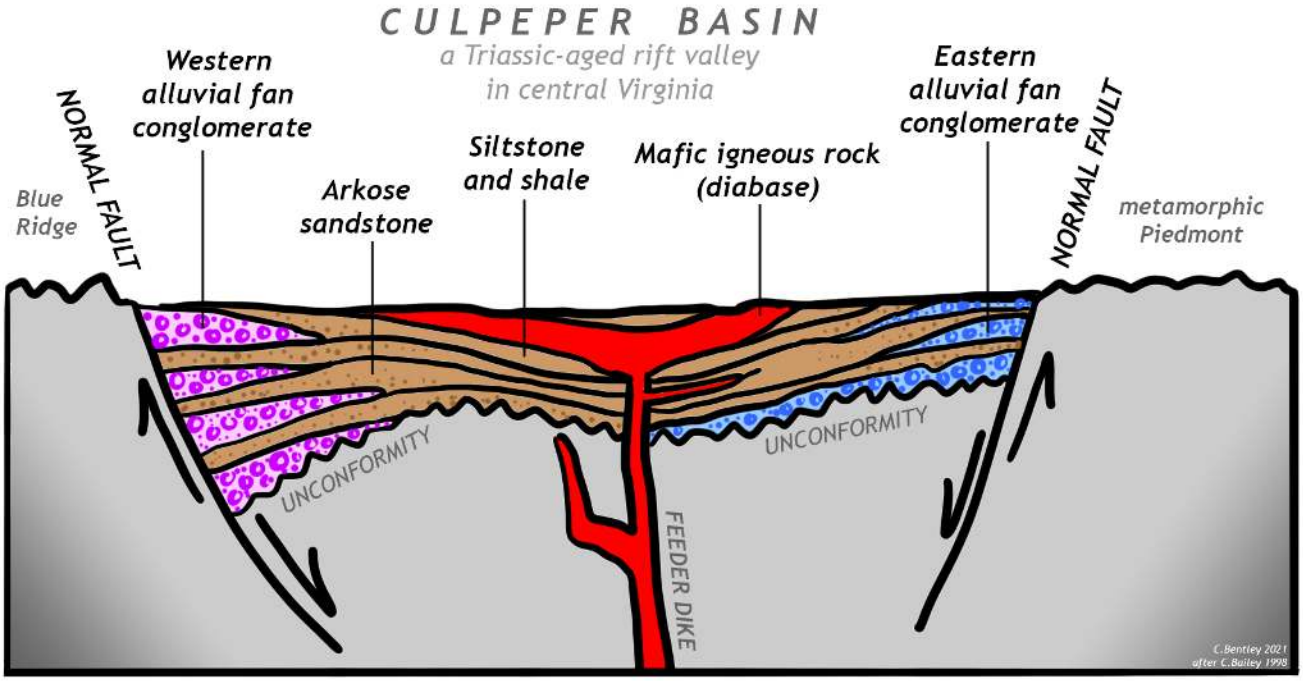


Figure 6.4.12: Conceptual cross-section through the Culpeper Basin. (Callan Bentley cartoon.)

The Culpeper Basin is filled with clastic sediment and mafic igneous rock. It shows coarsest-grained sediment at the margin basins. These alluvial fan deposits are different in composition in the west versus the east, as the alluvial fans in the two locations were draining highlands of very different compositions. In the west, limestone and basalt clasts predominate. *(One of these western fanglomerates, the Leesburg Conglomerate, makes up the distinctive columns of the Hall of Statuary in the United States Capitol building.)* In the east, the large clasts are mainly foliated metamorphic rocks derived from the neighboring Piedmont geologic province. Further from the basin margins, the sediments get finer. They include arkosic sands and mud deposits.

Lake deposits from the central Culpeper Basin have been contact metamorphosed to yield layers of multicolored hornfels that produce surprisingly durable cliffs:



Figure 6.4.13: Contact-metamorphosed meta-mudstones (rift valley lake sediments) of the Triassic Balls Bluff Formation exposed in road cuts along the Dulles Greenway between Reston and Leesburg, Virginia. (GigaPan by Chris Johnson.)

The Culpeper Basin is one of many basins that formed in the heart of Pangaea during its breakup. Some of these basins connected up — a line of "weakest links" — and ripped open to make the Atlantic Ocean. Basins east of that line are now in Africa and Europe. Basins west of that line, including the Culpeper Basin, are called the "Newark Supergroup" basins. Other examples in this collection include the Gettysburg Basin, the eponymous Newark Basin, the Hartford Basin of Connecticut, and Canada's Bay of Fundy. Collectively, the Newark Supergroup basins are quite extensive.



Figure 6.4.14: Pavement outcrop of stromatolites in limestone the Helena Formation, Belt Supergroup. (Callan Bentley photo.)

Another huge rift basin can be found in northwestern Montana and adjacent states and territories: the Belt Supergroup was deposited in the Belt Basin in Mesoproterozoic sedimentary time. Like the Culpeper Basin, it's coarsest at the edges and at the

bottom. It's also similar in having intercalated mafic igneous rocks. It's different in being much larger, and lasting long enough to develop more mature sediments, including thick carbonate packages within its stratigraphic sequence.

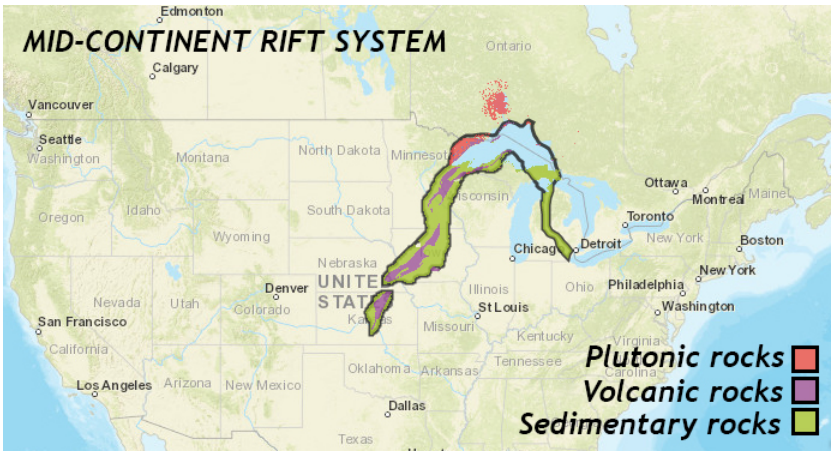


Figure 6.4.15: Extent of the Mesoproterozoic Mid-Continent Rift System in central North America. (Modified from a USGS map; explore more here.)

Another ancient rift basin worth noting is the Mid-Continent Rift of North America (also called the Keweenawan), which extends from Lake Superior in Minnesota to the southwest, diving beneath sedimentary cover for much its length (but still detectable via its distinctive positive gravity anomaly signature). This gigantic rift system developed in the Mesoproterozoic immediately before the Grenville Orogeny.

Convergent

Subduction at convergent plate boundaries can also induce sedimentary basins to form. There are three principal varieties to consider: foreland basins, forearc basins, and backarc basins.

FORELAND BASINS

Foreland basins form on the continental crust as subduction loads the crust with a heavy mountain range. Thrust faulting thickens the crust adjacent to a subduction zone, and this additional rock mass causes the adjacent crust to sag. The sagging creates accommodation space for sediment.



Figure 6.4.16: The Persian Gulf is a foreland basin formed in association with the Zagros Mountains.

A modern foreland basin can be seen in the Persian Gulf. The Persian Gulf is a foreland basin between the Arabian Peninsula and the southern edge of Eurasia: Iran's Zagros Mountains. As the Arabian Plate and the Eurasian Plate converge, thrust-loading of the crust in the Zagros has caused the adjacent crust to sag and make the low spot of the Persian Gulf.

An Ordovician example can be found in the mid-Atlantic region: the Queenston clastic wedge is a thick package of sediment including both marine flysch and terrestrial molasse that formed in response to downward flexure of the crust induced by the Taconian Orogeny. The same pattern repeated during the Devonian Acadian Orogeny: the Catskill clastic wedge is the name given to this subsequent foreland basin, partially overlapping in space the region occupied by its Ordovician predecessor.

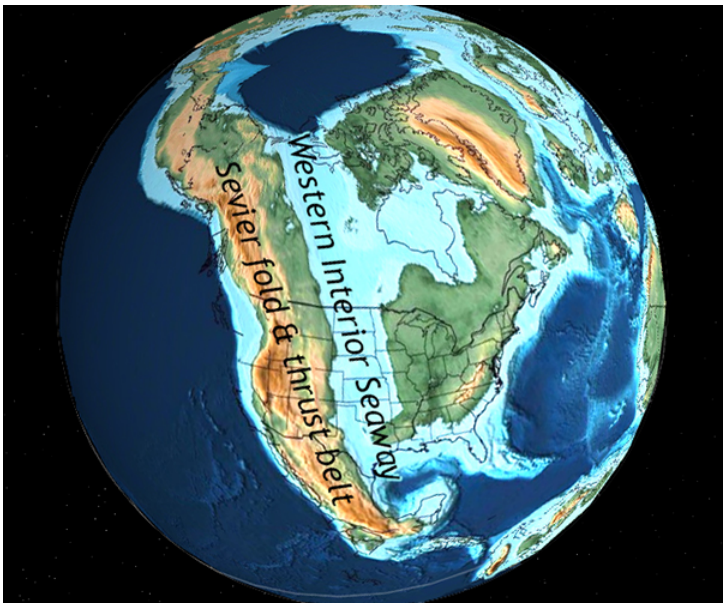


Figure 6.4.17: The Western Interior Seaway was an extensive foreland basin that developed during Cretaceous mountain-building. (Modified from a paleogeographic map by Christopher Scotese.)

In Cretaceous western North America, another foreland basin formed: the Western Interior Seaway stretched from the Gulf of Mexico to the Arctic Ocean. Seawater flooded a vast trough of land that sagged downward as subduction of the Farallon Plate caused the Sevier Orogeny. This episode of thin-skinned deformation thrust thick slices of sedimentary rock into towering peaks.

FOREARC BASINS

Forearc basins form on the trench side of volcanic arcs. Subduction at the trench keeps the crust low, and the neighboring volcanic arc sheds copious sediment that can fill the basin.



Figure 6.4.18: Half Dome is a prominent granite peak in Yosemite. All the rock you see here used to be felsic magma beneath a continental volcanic arc. (Photo by Diliff, via Wikimedia Commons.)

A classic example is the Great Central Valley of California, which is a forearc basin of Jurassic/Cretaceous age. At that time, subduction of the Farallon Plate generated a thick accretionary wedge of seafloor rocks. This is preserved in the modern day as the bulk of California's Coast Ranges. Inland, this same subduction manifested as a vibrant volcanic arc, the roots of which are today preserved as the Sierra Nevada batholith, the voluminous wad of granites that can be found in places such as Yosemite National Park. During the Cretaceous, the Sierra Nevada would have looked much like the modern Andes Mountains of South America. Though the volcanoes themselves have since been eroded away, the extent of their source magma chambers is gloriously revealed in the high country of that excellent mountain range.

Between the Coast Ranges' accretionary wedge and the Sierra's magmatic arc is the vast, hyperflat Great Central Valley. Though the valley is a modern topographic low, it corresponds very closely in areal extent with the Cretaceous submarine forearc basin:

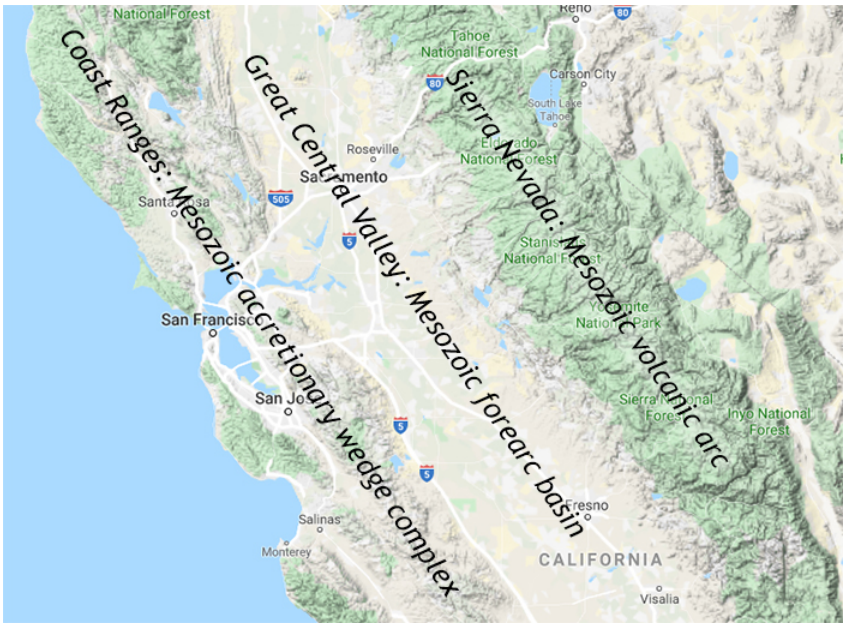


Figure 6.4.19: North-central California shows a three-part physiography that corresponds to the Mesozoic origins of this part of North America as a subduction zone at a convergent plate boundary.



Figure 6.4.20: Turbidites in the Venado Sandstone (Great Valley Sequence) at Lake Berryessa, California. (Photo by Mike S. Clark via Wikimedia.)

The Great Central Valley is 60 to 100 km (40 to 60 miles) wide. It stretches approximately 720 km (450 miles) along a north-northwest to south-southeast trend, paralleling the volcanic arc to the east and the accretionary wedge to the west. Clastic sediments deposited in this basin make up the 12 km-thick (40,000') Great Valley Sequence. Mostly shale, it also includes substantial bodies of sandstone and conglomerate that probably represent Mesozoic abyssal fan systems, sourced from the Sierra Nevada volcanic arc.

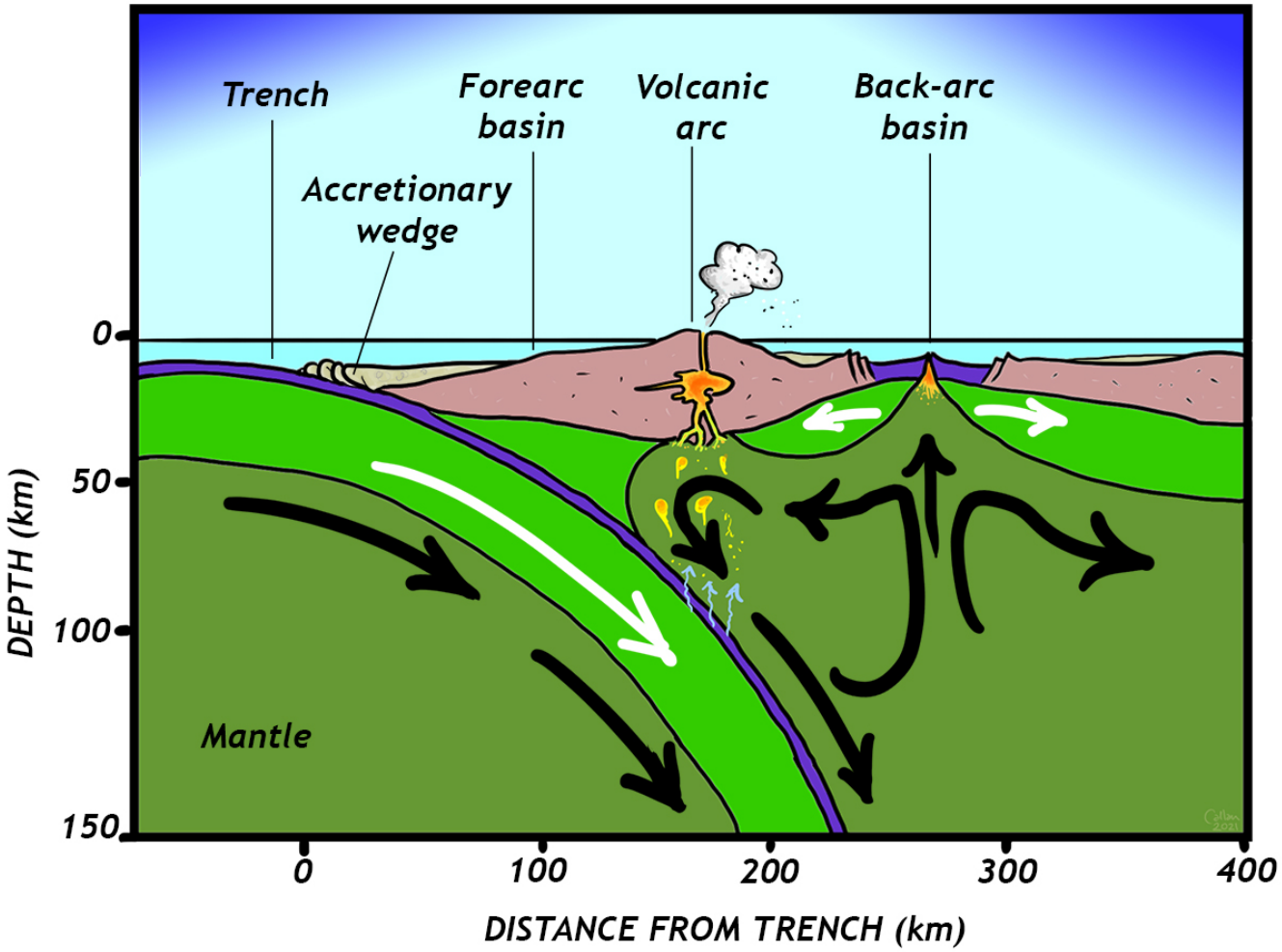


Figure 6.4.21: An “eddy” of convection in the mantle wedge above a subducted slab makes a back-arc basin. (Callan Bentley cartoon.)

Back-arc basins are the final category associated with convergent boundaries. There’s a plot twist though: ironically, they actually require divergence to form! They result from tensional forces caused by rising mantle material coupled with oceanic trench rollback (the oceanic trench shifts over time in the direction of the source of the subducting plate). The arc crust is under extension or rifting as a result of the sinking of the subducting slab. A good modern example of back-arc spreading is the Sea of Japan, a modest ocean basin that has opened between the islands of Japan (a volcanic arc) and the Asian mainland.

Did I Get It? - Quiz

? Exercise 6.4.1

Transform boundaries can produce what kind of basin?

a. Foreland basin

b. Wrench basin

c. Rift basin

d. Back-arc basin

Answer

b. Wrench basin

Exercise 6.4.2

What kind of basin forms as a result of "loading" the crust with a thick mountain belt due to thrust faulting?

a. Foreland basin

b. Forearc basin

c. Wrench basin

d. Back-arc basin

Answer

b. Foreland basin

Exercise 6.4.3

The Persian Gulf is what kind of basin?



a. Wrench basin

b. Foreland basin

c. Rift basin

d. Forearc basin

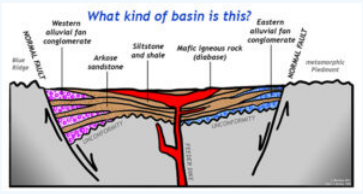
e. Back-arc basin

Answer

b. Foreland basin

Exercise 6.4.4

What kind of basin is shown?



a. Wrench basin

b. Forearc basin

https://geo.libretexts.org/@go/page/22635

c. Back-arc basin

d. Rift basin

e. Foreland basin

Answer

d. Rift basin

? Exercise 6.4.5

A modern-day example of a back-arc basin is _____.

a. central Virginia

b. the Wasatch front of Utah

c. east Africa

d. the Mid-Atlantic Ridge

e. west of Japan

Answer

e. west of Japan

? Exercise 6.4.6

What sort of basin formed in the middle of the ancestral North American continent during the Mesoproterozoic Eon? (near the western Great Lakes region)

a. Wrench basin

b. Foreland basin

c. Rift basin

d. Forearc basin

Answer

c. Rift basin

? Exercise 6.4.7

What sort of basin formed in the middle of North America during the Cretaceous, running from the Arctic Ocean south to the Gulf of Mexico?



a. Wrench basin

b. Foreland basin

c. Back-arc basin

Answer

b. Foreland basin

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6.5: Geophysical phenomena

Several manifestations of plate tectonics are best examined through the lens of geophysics.

Earthquakes

The vast, vast majority of earthquakes on Earth are associated with plate boundaries. Examine this map showing more than a century of epicenters to get an immediate, visceral sense of this pattern:



Earthquakes are caused by the sudden slip of two bodies of rock past one another, after tectonic stresses have accumulated to the point that they overwhelm frictional inertia. Most (95%) of the places where rock is forced into the uncomfortable position of scraping over other rock are located at plate boundaries. The exceptions (5%) are at sites where the crust is being stressed for other reasons: excessive loading due to sedimentation, intrusion of magma, isostatic adjustments, or artificially induced seismicity due to elevated pore pressure.



Figure 6.5.1: Mount Augustine, a composite volcano in Cook Inlet, Alaska, erupting in 2006. (Photo by Game McGimsey, USGS.)

Similarly, most volcanoes on this planet are associated in some way with plate boundaries [map]. There are four major options here:

1. Subduction of oceanic lithosphere under continental lithosphere (example: **Andes**)
2. Subduction of oceanic lithosphere under oceanic lithosphere (example: the **Philippines**)
3. Divergence (rifting) of continental lithosphere (example: **East African Rift**)
4. Divergence (seafloor spreading) of oceanic lithosphere (example: **Mid-Atlantic Ridge**)

But volcanoes can also form in ***non***-plate boundary settings. When we find eruptions piercing plates in locations far from their edges, many geologists invoke hot spots or mantle plumes. The idea here is that, independent of plate motions, the mantle has isolated point loci where rising warm material convects upward, and in so doing partially melts. The magma makes its way through the overlying lithosphere and erupts. This can occur in a couple of contexts:

1. Hot spot volcanism through oceanic lithosphere (example: **Hawaii**)
2. Hot spot volcanism through continental lithosphere (example: **Yellowstone**)



Figure 6.5.2: GPS receivers can measure the movement of Earth's landmasses very precisely. (UNAVCO photo.)

One of the greatest innovations of the modern era is the advent of global navigation satellite systems (GNSS), including the American global positioning system (GPS). With dozens of satellites orbiting the Earth, a given receiver on the ground can trilaterate its position exceptionally precisely. This is not only useful for Waze getting you to your job interview on time, but can also be used for documenting the subtle motions of Earth's crust.

GNSS measurements show the crust moving relative to other parts of the crust. To analyze GNSS information, a geodesist must choose a frame of reference. Since no spot on Earth's surface is stationary over the long term, it is important to realize that all plate motion is relative to other plates. For instance, if you choose North America as your reference frame, but examine the motions of various spots west of the Wasatch Front, you'll find an interesting set of patterns:

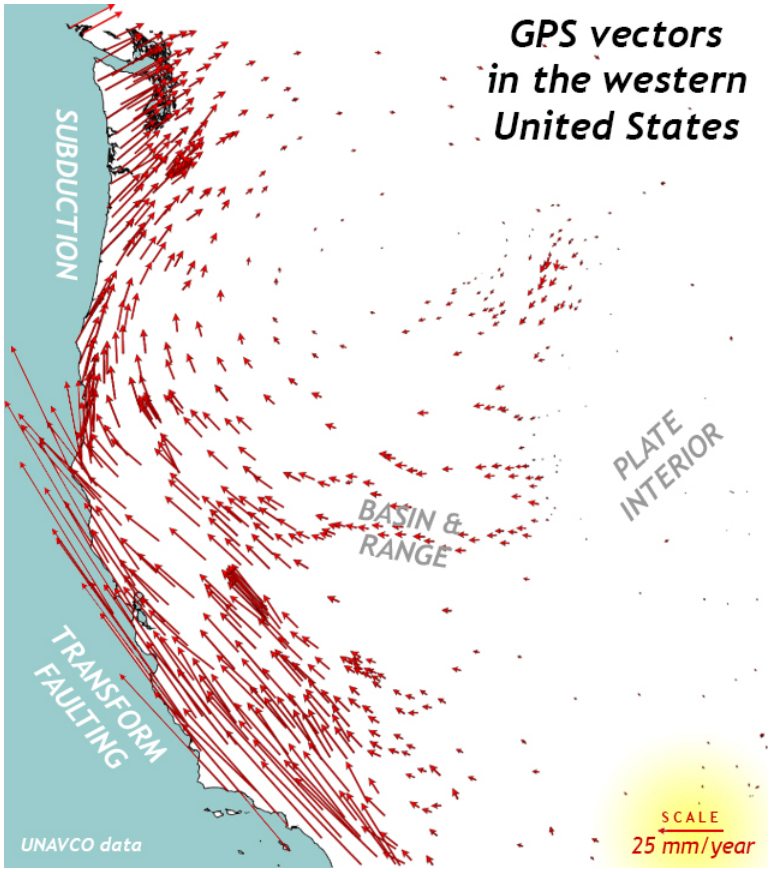


Figure 6.5.3: GPS vectors in the western United States reflect different tectonic settings. (UNAVCO data.)

Each arrow shows the vector of motion* of that location. Vectors are ways of describing motion in terms of both magnitude and direction. The longer the arrow, the faster that bit of the crust is moving. The arrow points in the direction that crust is moving

relative to the North American frame of reference.

*Technically, what is shown in that map is the ***horizontal*** component of the velocity. There can also be a ***vertical*** component of the velocity (*i.e.*, up or down motion), especially at divergent or convergent boundaries, or sites of transpression or transtension. Displaying this vertical motion is an additional option on the UNAVCO GPS Velocity Viewer application.

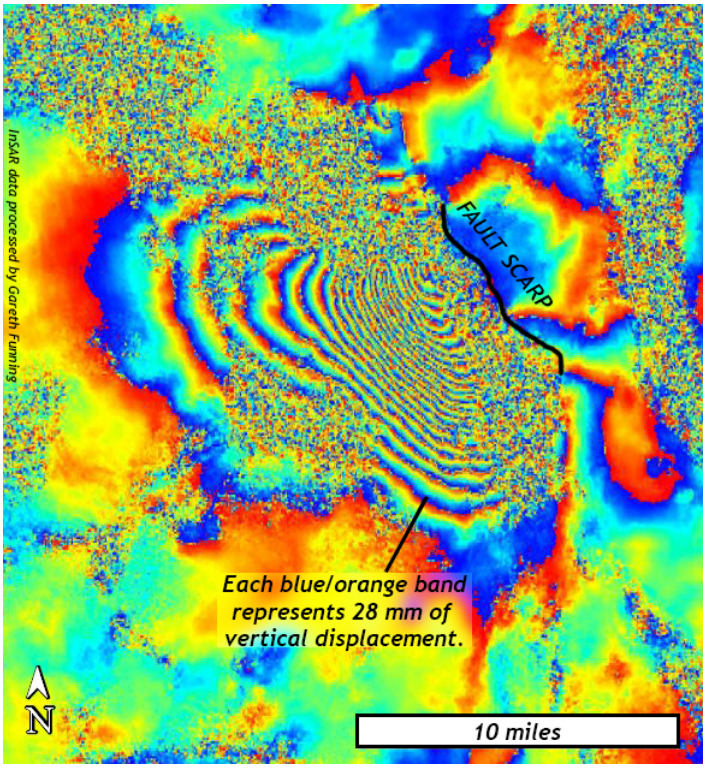


Figure 6.5.4: InSAR-derived interferogram for an earthquake in Dinar, Turkey in 1995. (Modified from an original interferogram by Gareth Funning.)

Another aspect of satellite measurement can be utilized for examining land positions before and after an earthquake. InSAR uses microwaves to repeatedly measure the shape of Earth's surface, and if there is a change (sudden or gradual), InSAR can detect the difference between one radar mapping "pass" by a satellite and its successor. What's literally being measured here is whether the land surface has moved toward or away from the satellite. InSAR data is displayed in rainbow-scale maps called "interferograms." Each psychedelic blue/orange band in such an image represents a vertical displacement of 28 mm. Counting the number of bands and multiplying by 28 will give a fair estimate for the vertical offset of a fault in a given earthquake, as measured in millimeters.

Did I Get It? - Quiz

? Exercise 6.5.1

Where can we find 95% of all earthquakes?

a. Along plate boundaries

b. In plate interiors

c. In the lower mantle

d. Along the edges of continents

Answer

a. Along plate boundaries

Exercise 6.5.2

GPS stands for _____.

a. granite posts selfies

b. geostationary positron sycophancy

c. global positioning system

d. gopher performs soliloquy

Answer

c. global positioning system

Exercise 6.5.3

What is the overall pattern of the continental crust in the western part of the contiguous United States of America?

a. a uniform southward march, but faster in the north than in the south

b. a clockwise spin, with highest velocities west of the San Andreas Fault

c. a counterclockwise spin, with highest velocities east of the Wasatch Fault

Answer

b. a clockwise spin, with highest velocities west of the San Andreas Fault

Exercise 6.5.4

What is InSAR good for?

a. Quantifying annual plate motion relative to other plates

b. Measuring surface deformation after earthquakes

c. Quantifying subsurface movement of magma

d. Evaluating risk due to future earthquakes

Answer

b. Measuring surface deformation after earthquakes

Exercise 6.5.5

Which of the following locations is thought to be a modern example of a hot spot?

a. Hawaii

b. Virginia

c. Mid-Atlantic Ridge

d. Andes

e. Philippines

Answer

a. Hawaii

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6.6: Driving forces

The magnitude and direction of lithospheric plate movement is the result of the various forces acting on the plate, and the plate's resistance to those same forces. What could make such a massive slab of rock overcome inertia and move though geologic time? Convection in Earth's mantle is relevant, but probably greater momentum comes from the forces of ridge push and slab pull.



IRIS video.

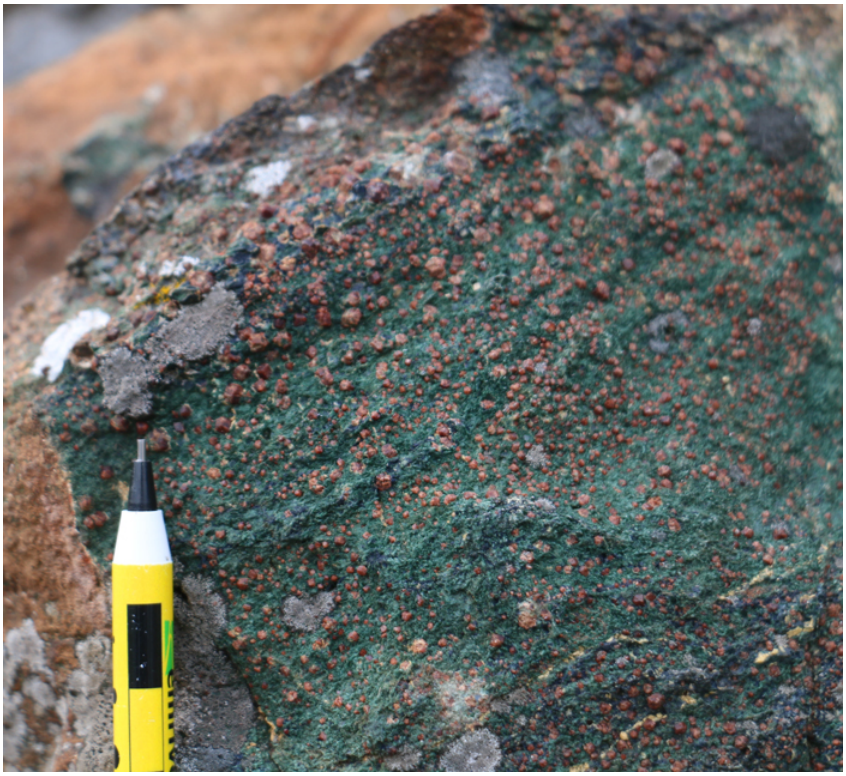


Figure 6.6.1: Eclogite consists of green omphacite (a variety of pyroxene) and prominent red pyrope (a variety of garnet). (Callan Bentley photo.)

When oceanic lithosphere subducts to a sufficient depth, the basalt is metamorphosed to blueschist and then **eclogite**. Eclogite is particularly dense (3.45 to 3.75 g/cm^3), and that extra dense material on the leading edge of the subducting slab may yank on the rest of the plate with an urgency that cannot be resisted. In other words, the metamorphic reactions that make eclogite help contribute to the slab pull force.

Did I Get It? - Quiz

Exercise 6.6.1

Is mantle convection alone sufficient to explain the motion of lithospheric plates?

a. No; the forces of "ridge push" and "slab pull" must also be considered.

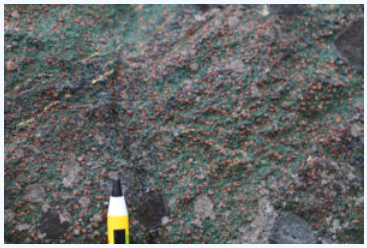
b. Yes; no need for any additional causes.

Answer

a. No; the forces of "ridge push" and "slab pull" must also be considered.

Exercise 6.6.2

What role does eclogite potentially play in driving plate tectonics?



a. When eclogite is erupted from oceanic ridges, it pushes down strongly on the adjacent crust.

b. It slows down subduction because of all its gritty garnet crystals.

c. The metamorphic reactions that turn ocean floor basalt and gabbro to eclogite make it much denser, and that extra weight can pull on subducted plates.

Answer

c. The metamorphic reactions that turn ocean floor basalt and gabbro to eclogite make it much denser, and that extra weight can pull on subducted plates.

Exercise 6.6.3

How old does oceanic lithosphere need to be before it reaches its maximum thickess?

a. 1800 million years

b. 180 million years

c. 10 million years

d. 80 million years

Answer

d. 80 million years

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6.7: Wilson Cycles

It is critical for historical geologists to recognize that the dynamics of a given plate boundary (divergent, convergent, etc.) are not eternal. A given spot on Earth's surface may be divergent for some time, then convergent, then divergent again. Just because a given place is tectonically inactive today, that doesn't mean it was not active in the past. Things change!

From the perspective of continents, another way of putting this is to say: **supercontinents form, and then break up.** A given spot on a continent may be a site of divergence and rifting (opening up a new ocean basin next door), and later a site of convergence and mountain-building (as that ocean basin is closed and the continents on either side collide). Tuzo Wilson recognized these signatures and described them in a classic 1966 paper about the opening and closing of ocean basins. Because he was first to publish the idea that the change of tectonic setting could be cyclical, we often refer to supercontinent cycles as "Wilson cycles" today.

Mid-Atlantic Wilson Cycles

The eastern margin of the United States offers a classic example of how a given place can shift from convergent to divergent to passive to convergent to divergent to passive again. These sequential shifts accompanied the construction of two subsequent supercontinents (Rodinia and Pangaea) and their subsequent break-up. These repetitive "cycles" of supercontinent formation and destruction are particularly well recorded in the Appalachian mountain belt.

At locations up and down the east coast of North America, there are four major batches of active tectonism, interspersed with times of passive margin conditions:

1. From 1.2 to 1.0 Ga, **mountain-building** dominated, marked by granites, metamorphism, and deformation. This made the supercontinent Rodinia.

2. From 700 Ma to 565 Ma, **rifting** dominated, marked by mafic volcanism and immature sedimentary deposits. This broke Rodinia up into fragments, and opened the Iapetus Ocean between them.

3. Subsidence and **passive margin sedimentation** occurred through the Cambrian and early Ordovician (~550 to ~480 Ma).

4. From 300 to 250 Ma, **mountain-building** dominated, marked by granites, metamorphism, and deformation. This made the supercontinent Pangaea.

5. From 200 to 180 Ma, **rifting** dominated, marked by mafic volcanism and immature sedimentary deposits. This broke Pangaea up into fragments, and opened the Atlantic Ocean between them.

6. **Passive margin sedimentation** resumed, from the Cretaceous (~100 Ma) until today.

The Grenville Orogeny



Figure 6.7.1: The Old Rag Granite, a 1.04 Ga felsic intrusion in Shenandoah National Park, Virginia. (Callan Bentley photo.)

The oldest rock on the east coast of North America is intrusive. Much of it is felsic in composition, reflecting the partial melting of older source rocks. Dating zircons from these granites and granite gneisses of the “basement complex” gives ages of 1.2 to 1.0 Ga. In general, the older units are more foliated than the younger units, suggesting they were around for a longer period of time during mountain building, with more time to build up a pervasive compressional fabric. Basement rocks of this age are found from Texas to Newfoundland, including the anorthosites making up the Adirondack Mountains in New York, and the granites of the Blue Ridge province in North Carolina and Virginia.

Iapetan rifting



Figure 6.7.2: Basaltic cooling columns of Neoproterozoic-aged Catoctin Formation, exposed atop Compton Peak in Shenandoah National Park, Virginia. (Callan Bentley photo.)

When Rodinia broke up, it was marked by the intrusion of mafic dikes and the eruption of large volumes of mafic lava. The resulting flood basalts bear vesicles, xenoliths, and cooling columns. Normal faulting led to rift valleys that filled with immature clastic sediment, including gravel, arkosic sand, and mud. Some of these rift basins connected up, becoming “the weakest link” between ancestral North America (Laurentia) and the various Proterozoic continental fragments that drifted away from it. Seafloor spreading took over, filling in the gap between continents with fresh injections of basalt. The Iapetus Ocean was born. Inch by inch, this ocean basin widened for hundreds of millions of years, adding a “skirt” of oceanic crust to the edge of the ancestral North American plate.



Figure 6.7.3: Petersburg Granite, a ~320 Ma pluton exposed at the Falls of the James River in Richmond, Virginia. (Callan Bentley photo.)

The Appalachian Mountains mark the suture between ancestral North America (Laurentia) and ancestral Africa (the leading edge of Gondwana). They represent the collision between these two large continental landmasses, formerly separated by the Iapetus Ocean. Over the course of the Paleozoic, the Iapetus began to close (resulting in two earlier, smaller orogenies), and the temporary "skirt" of oceanic crust was recycled into the mantle via subduction. This brought Africa and North America ever closer, until eventually they merged in a collisional mountain belt. When the Appalachians were young, they ran through the heart of the Pangaean supercontinent on a scale that may have dwarfed the modern Himalayas. Based on the age of the rocks deformed by compressional stresses and the dating of granites and metamorphic rocks, it appears that the final orogeny occurred in the late Paleozoic, starting around 300 Ma (in the Pennsylvanian) and wrapping up by about 250 Ma (in the Permian). This compressional event resulted in the folding and faulting of older strata in what is today the Valley & Ridge province, metamorphism of the rock in the Blue Ridge and Piedmont province, intrusion of granite plutons, and suturing of African crust to the North American crust.



Figure 6.7.4: Outcrop of Waterfall Formation rift-valley alluvial fan conglomerate, near Haymarket, Virginia. (Callan Bentley photo.)

When Pangaea began to break up in the Triassic, the tectonic extension was first marked by the intrusion of mafic dikes. Normal faulting resulted in the opening of numerous parallel rift basins up and down the east coast. Some of these rift basins connected up, becoming "the weakest link" between North America and Africa/Europe. They stretched wide enough apart that seafloor spreading took over, filling in the gap between continental fragments with fresh injections of basalt. Inch by inch, the Atlantic basin has been widening ever since. Numerous "failed" rift basins remain in the Piedmont geologic province, sort of like "stretch marks" in the crust. They are dominated by two kinds of rocks: immature clastic sediments, and mafic volcanics. The immature clastic sediments include conglomerates and arkose sandstone near the margin of the basins, but also lake deposits in the basins' centers.

Did I Get It? - Quiz

? Exercise 6.7.1

Which of the following happened during the Mesoproterozoic on the east coast of North America?

a. Rifting of Pangaea

b. Grenville Orogeny

c. Appalachian Orogenies

d. Rifting of Rodinia

Answer

b. Grenville Orogeny

Exercise 6.7.2

Which of the following happened during the Triassic on the east coast of North America?

a. Rifting of Rodinia

b. Passive margin sedimentation

c. Appalachian Orogenies

d. Grenville Orogeny

e. Rifting of Pangaea

Answer

a. Rifting of Pangaea

Exercise 6.7.3

Which of the following was a signature of the rifting of Rodinia during the Neoproterozoic on the east coast of North America?

a. Intrusion of granite

b. Eruption of floods of basalt

c. Thrust faulting

d. Deposition of limestone and quartz sandstone

Answer

b. Eruption of floods of basalt

Exercise 6.7.4

Which of the following was a signature of Appalachian mountain-building during the Paleozoic on the east coast of North America?

a. Deposition of limestone

b. Eruption of floods of basalt

c. Intrusion of granite

d. Normal faulting

Answer

c. Intrusion of granite

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CHAPTER OVERVIEW

7: Geologic Time

The Earth is 4.566 billion years old. [1]

Knowing this is not simply some esoteric piece of trivia. Instead, the discovery of Earth's great antiquity is the crowning achievement of the geosciences.

Why does it matter? If Earth were only a mere 10,000 years old, then human civilization would represent a significantly large portion of Earth's history. Instead, in the context of an Earth that is 4.566 billion years old, modern human civilization hardly registers at all, amounting to a trivially small sliver (~0.0003%) of Earth's history. Understanding geological time is central to properly understanding our place in nature and history.

If you think about large numbers and try to comprehend the difference in the amounts, say the difference between 50,000 and 50,000,000, you have no frame of reference to visualize the difference. This is the difficulty in trying to understand the vastness of geological time—which is measured by millions and billions of years. Essayist John McPhee coined the phrase "deep time" in his 1981 book, *Basin and Range*. Deep time is truly staggering, a view captured eloquently in a recent essay in *the New York Times* by Peter Brannen.

Have a look at the image below (or download this PDF file), which illustrates 5,000 individual zeros.

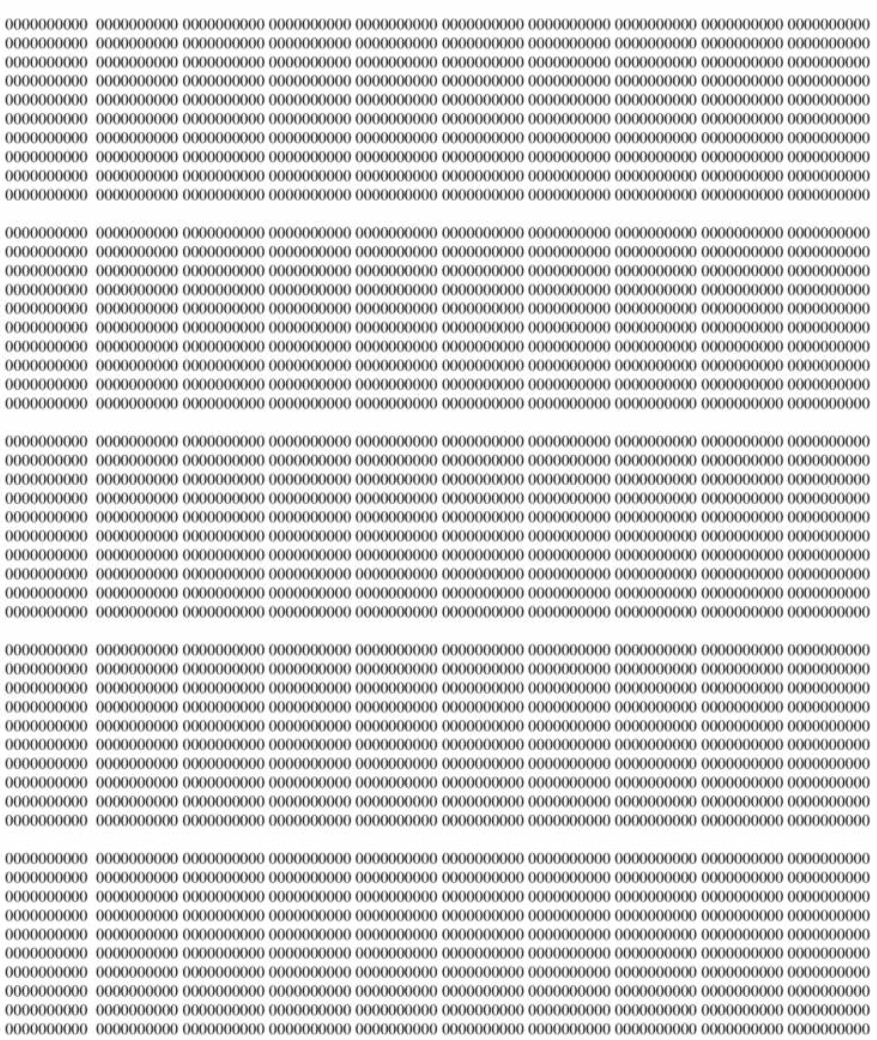


Figure 7.1: Image by Jonathan R. Hendricks. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License.)

How many copies of this image (or sheets of paper) would you need to print to reach a total of a million zeros?

$$1,000,000/5,000 = 200 \text{ sheets of paper}$$

Now, how many copies would you need to print to reach one-billion zeros?

$$1,000,000,000/5,000 = 200,000 \text{ sheets of paper}$$

The Earth is 4, 566, 000, 000 years old. How many copies would you need to print to cover the age of the Earth with each year represented by 1 zero?

$$4,566,000,000/5000 = 913,200 \text{ sheets of paper}$$

Copy paper is bundled by 500 sheets. This is called a ream of paper. How many reams of paper would you need to replicate the age of the Earth?

$$913,200/500 = 1,826 \text{ reams of paper}$$

Typically there are 6 reams of paper per box, that makes 304 boxes of copy paper just to print out the number of zeros in the age of the Earth.

That is a long time...

7.1: Abbreviating geological time

7.2: Geology as a historical science
7.3: Foundational Concepts of Historical Geology
7.4: The Geological Time Scale
7.5: References
7.6: Further Reading

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7.1: Abbreviating geological time

Geologists measure events in Earth's history in years before the present date. They use certain conventions for abbreviating intervals of time. One thousand years is represented by the abbreviation "ka," which means "kilo-annum." The prefix "kilo-" should be familiar to you: a kilogram equals one thousand grams and a kilometer equals one thousand meters; "annum" means year. Similarly, one million years is represented by "Ma," meaning "mega annum." One billion years is abbreviated as "Ga" for "giga annum." These prefixes are also used to describe computer memory: kilobyte, megabyte, and gigabytes of storage. Sometimes geologists use "kya" and "mya" to characterize events that occurred, respectively, thousands or millions of *years ago*. For example, most dinosaur species went extinct 66 mya. The process of assigning actual ages –in years before the present date– is referred to as numerical (absolute) dating, which is covered in a later section.

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7.2: Geology as a historical science

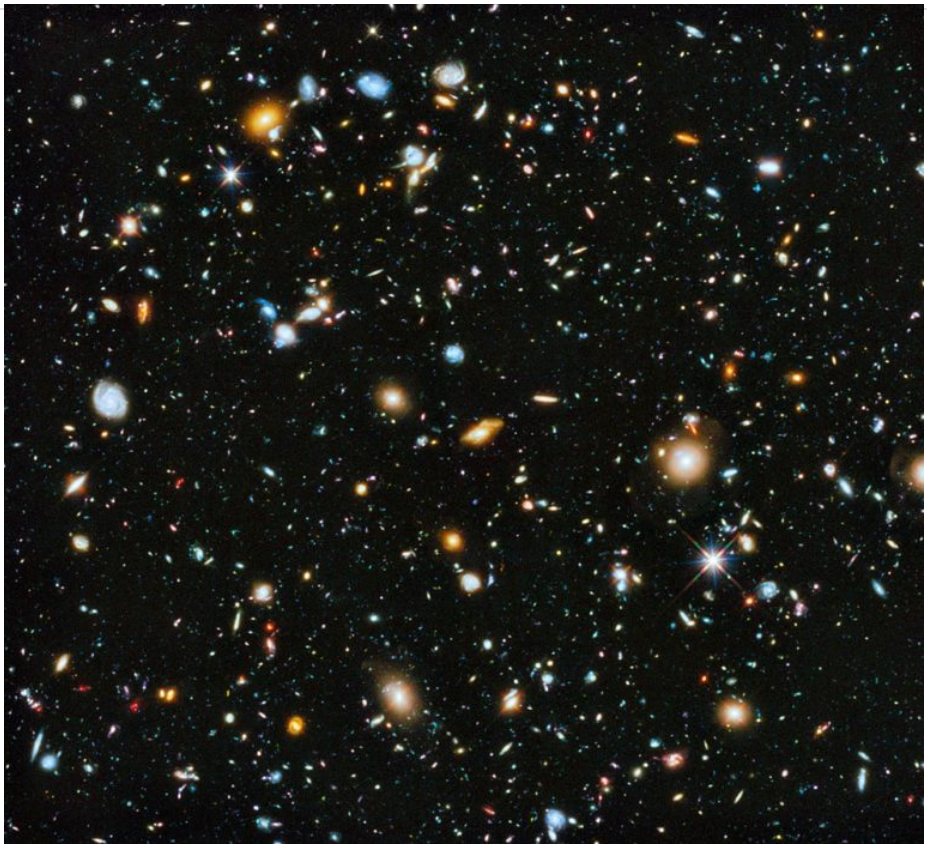


Figure 7.2.1: The Hubble Ultra-Deep Field image, showing many different far, far away galaxies; in most cases, it took billions of years for the light from these galaxies to reach Earth. Each galaxy shown may be composed of many billions of individual stars, each of which may have its own solar system of planets. (Image by NASA, ESA, H. Teplitz and M. Rafelski (IPAC/Caltech), A. Koekemoer (STScI), R. Windhorst (Arizona State University), and Z. Levay (STScI) – http://hubblesite.org/newscenter/arc...14/27/image/a/ (image link), Public Domain, https://commons.wikimedia.org/w/inde...urid=33189266.)

Geology is the study of the Earth and its history. While some geologists study modern geological phenomena such as earthquakes and volcanic eruptions, much of the science is devoted to discovering the past events that shaped and created the world we live upon today. In this respect, geology is considered a historical science because it deals with events that happened long ago–typically long before humans had evolved. Just as a historian or archeologist uses available records, artifacts, and other clues to paint a picture of past human lives, a geologist reconstructs the past using preserved evidence such as clues from layers of rock and the minerals and fossils found within those rocks. Geology shares good company with other historical sciences, including much of evolutionary biology, climatology, and archaeology. Astronomy is another example of a historical science: it has taken billions of years for light leaving far away galaxies (that is, galaxies that are billions of light years away) to reach the Earth, so when viewing such galaxies using powerful telescopes like the Hubble Space Telescope, we are looking backwards in time. If some of these galaxies went dark right now, observers on Earth would not know it until billions of years from now.

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Uniformitarianism

Our ability to use evidence left behind in rocks to interpret and reconstruct events that happened thousands, millions, or billions of years ago is dependent upon geology's most fundamental assumption: **uniformitarianism**. Uniformitarianism is the assumption that the chemical and physical laws of nature have not changed over the course of Earth's long history. The laws have remained constant, or "uniform" in their operation.This means, for example, if geoscientists determine rates of plate motion are a few centimeters per year, there has likely never been a time that pieces of the lithosphere were zooming around the planet at 50 kilometers per hour! While geoscientists cannot have direct observations of Earth processes occurring billions of years ago, there is no evidence to suggest the basic principles of biology, chemistry, and physics should be different than today.



Figure 7.3.1: Portrait of James Hutton by Sir Henry Raeburn (public domain; Wikimedia Commons).

The beauty of uniformitarianism is that it allows us to use observations of the modern world to understand Earth's history. We credit James Hutton (1726-1797) with development of the assumption of uniformitarianism. Hutton was a careful observer of the landscapes around him and he paid close attention to the slow rates at which rocks naturally erode, and the similarly, slow rates at which new layers of sediment are formed. Given such slow rates, what were the implications for interpreting the time it must have taken to form the countless thick layers of rocks exposed at Earth's surface? Hutton's observations led him to conclude that geological time must be unimaginably vast. He asked, "What more can we require? Nothing but time." He viewed Earth as constantly changing and renewing itself, a view radically different from the orthodoxy of the day, which commanded that Earth had been created a few thousand years ago just as we see it today. Referring to Earth's history, Hutton famously quipped that it shows "no vestige of a beginning, no prospect of an end."

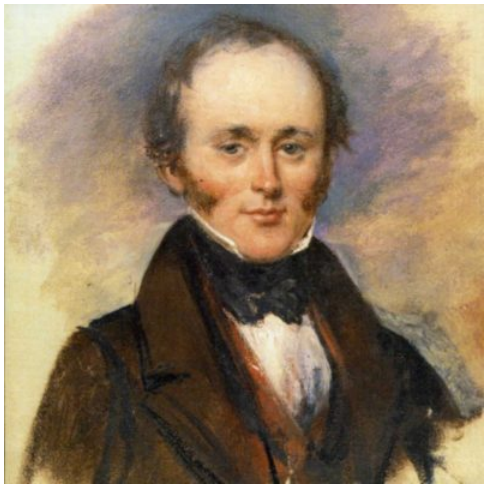


Figure 7.3.2: Portrait of Charles Lyell by Alexander Craig (1840) (public domain; Wikipedia).

While Hutton developed the concept of uniformitarianism, Charles Lyell (1797-1875) made the idea famous in his influential book *Principles of Geology*, first published in 1830. Based on many observations and examples, he convinced many people that

geological processes act slowly and continuously. We now know that this is not strictly true: the intensities and rates of some processes *have* changed over Earth's history (for example, the interior of the Earth used to be much hotter than it is today, as it once contained much more heat-generating radioactive material) and Earth's history has been occasionally punctuated by catastrophic events (such as the asteroid impact that sealed the fate of most dinosaurs 66 Ma). But, we still accept the uniformitarian view that the chemical and physical laws of nature have nevertheless remained constant over time.

Fundamentally, the work of Hutton and Lyell solidified the uniformitarian view that **"the present is the key to the past."** This means that by studying how Earth processes work today, we can also make sense of how Earth operated in the past, as well as predict its future. Because of this, uniformitarianism is the single most important and fundamental idea in all of geology.

Principles of Relative Dating

It may surprise you to learn that geologists were able to determine much of the history of the Earth and its life without knowing anything about the actual ages of the rocks that they studied. Through use of numerical (absolute) dating techniques (which were developed during the 20th century; discussed later), they were able to later assign dates in years before present to important events in Earth's history. But, before that, they relied upon a different approach to first determine the sequence of important events in Earth's past: **relative dating**.

Relative Dating



Figure 7.3.3: Grand Canyon, Arizona. (CC BY-SA (https://creativecommons.org/licenses/by-sa/4.0); By Niagara66, https://commons.wikimedia.org/wiki/File:Grand_Canyon_National_Park_06.jpg.)

The principles of relative dating are largely applied to unraveling the geologic history of the thin blanket of sedimentary rock that covers much of continental land surfaces. Most of what we know about the development, diversity, and evolution of complex life is preserved in this thin blanket. The thickness of the sedimentary rock layers, or strata, varies quite a bit. In the northern Canadian Shield areas, which have been stripped clean of sedimentary cover during the past several million years of ice age glaciations, the cover is essentially zero. In locations uninterrupted by recent mountain building or the effects of glaciation, like the Gulf Coast of the United States, this cover can be substantial, up to 20 km thick (~12 miles)! The average cover is about 1800m thick (5900ft or just over 1 mile). The thickness of the sedimentary rock layers exposed in the Grand Canyon is close to the average at 1850m. This extent of visual exposure in the canyon is not found elsewhere in the U.S. which is what makes it so spectacular.

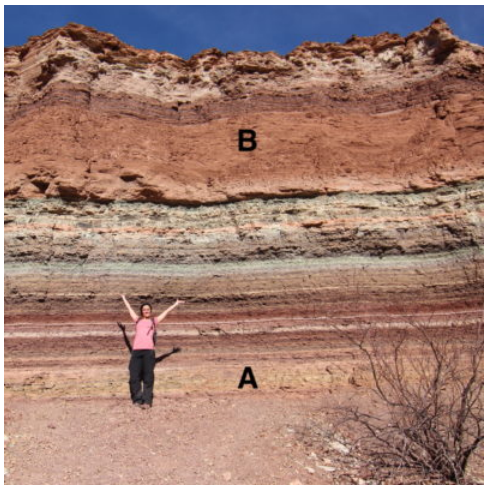


Figure 7.3.4: Sedimentary rock layers exposed at Quebrada de Cafayate, Salta (Argentina). (CC BY-SA 2.0, https://commons.wikimedia.org/w/inde...curid=24779211; by travelwayoflife https://www.flickr.com/photos/travelwayoflife/6164348161/in/photostream/)

The following relative dating principles apply to this covering of strata which is largely sedimentary in origin. However, igneous rock may also be included in the layers when it forms horizontally, in layers parallel to the Earth's surface, as in lava flows, pyroclastic flows and ash falls.

Very simply, relative dating has to do with determining whether one geological or paleontological event happened before or after a second event. For example:

Did rock layer A form before or after rock layer B?

Did the crinoid from Indiana (below, left) live before or after the trilobite from Oklahoma (below, right)?



Figure 7.3.5: Fossil Trilobite. Cravat Member, Bois d'Arc formation, Lower Devonian (Held.) age, Clarita, Coal Co., Oklahoma, U.S.A. (CC BY-SA 3.0; Moussa Direct Ltd., https://commons.wikimedia.org/w/inde...?curid=4461171)



Figure 7.3.6: Fossil Crinoid. Agaricocrinus americanus, Carboniferous, Indiana. (CC BY-SA 3.0; https://commons.wikimedia.org/wiki/User:Vassil.)

We can answer these questions with a handful of principles that help guide our thinking when establishing how old one geologic feature is relative to another. Relative dating has to do with determining the ordering of events in Earth's past. Geologists employ six simple principles in relative dating; the most important of which is the **principle of superposition**

Principle of superposition



Figure 7.3.7: Portrait of Nicholas Steno (public domain; Wikimedia Commons).

Nicholas Steno (1638-1686) was a well traveled Danish anatomist and one of the earliest scientists associated with studying paleontology and geology. He was the first to study relationships between layers of rock and the association of sediment, minerals and fossils to the surrounding and containing rock. He recognized the stories that rock could tell through their history of formation read as clues found within each rock. Steno's **principle of superposition** is simple, intuitive, and is the basis for relative dating of geological formations. It states that rocks positioned below other rocks are older than the rocks above.



Figure 7.3.8: Taughannock Falls near Trumansburg, New York, illustrating the Principle of Superposition. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

The image to the right demonstrates this principle as it shows a sequence of Devonian-aged (~380 Ma) rocks exposed at the magnificent waterfall at Taughannock Falls State Park in central New York. The rocks near the bottom of the waterfall were deposited first and the rocks above are subsequently younger and younger. Superposition is observed not only in rocks, but also in our daily lives. Consider the trash in your kitchen garbage can. The trash at the bottom was thrown out earlier than the trash that lies above it; the trash at the bottom is therefore older (and likely smellier!).

The photograph below was captured at Volcano National Park on the Big Island of Hawaii. Use superposition to determine which is older: the road or the lava flow? How do you know?



Figure 7.3.9: Volcano National Park, Big Island of Hawaii. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

Principle of faunal succession

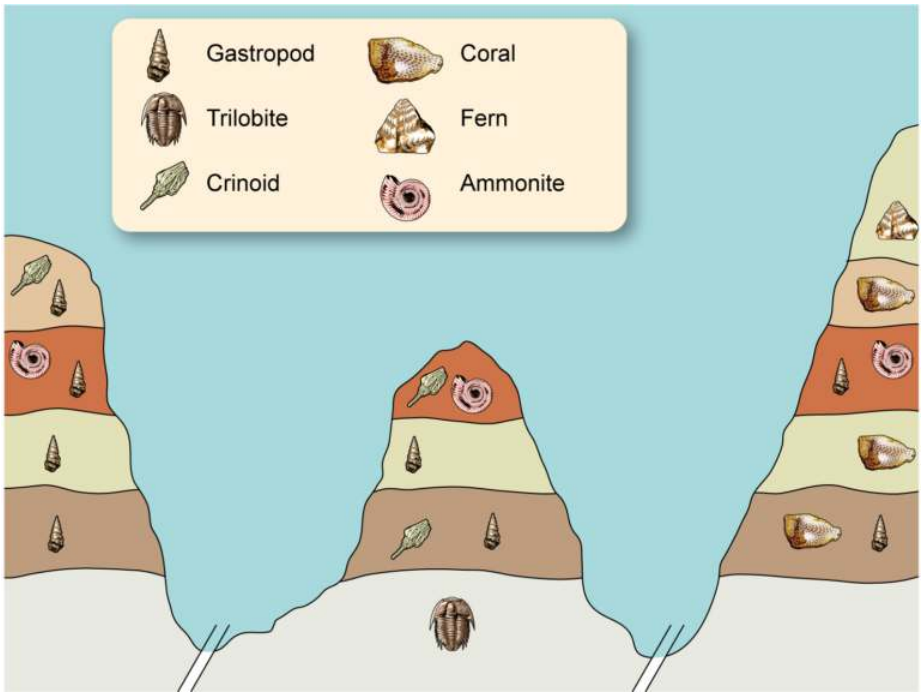


Figure 7.3.10: Sequence and correlation of fossils from different locations. (CC BY-SA; https://commons.wikimedia.org/wiki/File:Fossils.png by: Kurt Rosenkrantz)

William Smith (1769 – 1839) worked as a surveyor in the coal-mining and canal-building industries in southwestern England in the late 1700s and early 1800s. While doing his work, he had many opportunities to look at the Paleozoic and Mesozoic sedimentary rocks of the region, and he did so in a way that few had done before. Smith noticed the textural similarities and differences between rocks in different locations, and more importantly, he showed that fossils could be used to correlate rocks of the same age. Smith is credited with formulating a fundamental concept and basis for the development of the geologic time scale – the **principle of faunal succession**. Smith found the same ordering of fossil species from place to place; Fossil A was always found below Fossil B, which in turn was always found below Fossil C, and so on. By documenting these sequences of fossils, Smith was able to correlate rock layers (strata) from place to place. He established that rock layers in two different places were the same age based upon the fact that they include the same distinctive types of fossils.

Principle of original horizontality

The principle of original horizontality states that sediments are deposited by gravity in layers parallel to the Earth's surface. This applies also to volcanic igneous rocks that have been formed by lava flows, pyroclastic flows and ashfall deposits. Layers that accumulate along the edges of basins may be slightly sloped in toward the basin. This principle is particularly useful when rock

layers have been deformed by the forces of tectonic activity. Students must learn to “unfold the folds” or reconstruct faulted and contorted layers that were originally horizontal to the Earth’s surface to understand their geologic history. Check out the contorted folds in the sedimentary rock layers in the photo below. From the picture it is difficult to put all the rock layers back in their originally horizontal position. A geologist would walk this area, and any adjacent areas of exposed rock layers, to make sketches, take samples and pictures and use their imagination to attempt to conceptually “unravel” the deformed layers to understand the complete story of what happened from the original deposition of sediments to get them to their current appearance.



Figure 7.3.11: Folds exposed in a roadcut along Highway 14 in Palmdale, California, adjacent to San Andreas fault. (With permission by: Garry Hayes and Susan Hayes from: http://geotripperimages.com/Tectonic_Processes/Folds.htm)

Principle of lateral continuity

The principle of lateral continuity states that layers of sediment initially extend outward 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 (for instance, Grand Canyon), can be assumed to be originally continuous.

Layers of sediment do not extend indefinitely or infinitely. Instead, limits are controlled by the amount and type of sediment available, and the size and shape of the sedimentary basin. As sediment is transported to an area, thickest deposits are found closest to the source, and gradually thin as the distance increases.

In the panoramic photo below, the originally laterally continuous layers of sedimentary rock have been cut into by the Colorado River at Dead Horse Point State Park in southern Utah.



Figure 7.3.12: By Jean-Christophe BENOIST (Own work, CC BY 3.0, https://commons.wikimedia.org/w/index.php?curid=10152833)

Principle of cross-cutting relationships

The principle of cross-cutting relationships states that any geological feature that cuts across, or disrupts, another feature must be younger than the feature that is disrupted. In the photo below, a fault disrupts the once continuous layers of sedimentary rock . The dark red “marker bed” can be used to see the displacement that has occurred on either side of the fault. The fault “cross-cuts” the layers of sedimentary rock. This indicates that the sedimentary layers pre-date the fault; the fault is the younger feature.



Figure 7.3.13: Normal fault offsetting layers in the Noonday Dolomite, Mosaic Canyon, Death Valley, California. (With permission by: Garry Hayes and Susan Hayes from: http://geotripperimages.com/images/DSC01776%20Normal%20fault%20in%20Mosaic%20Canyon.jpg)


Intrusive Dike
Contact
Metamorphism
(Baking)

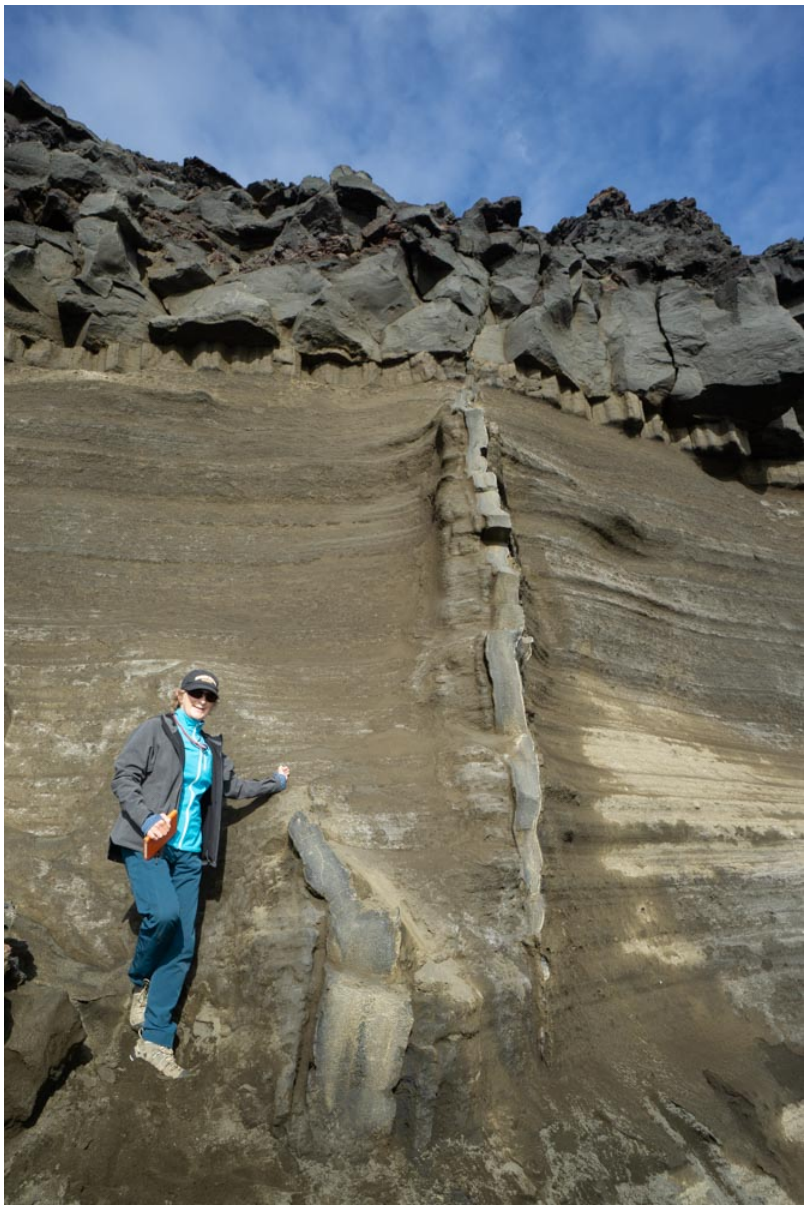


Figure 7.3.14: Above: A thin basaltic dike cross-cutting volcanic sediments of Paleogene age at Lagganulva, on the Isle of Mull, Scotland. (By permission, John Faithful, University of Glasgow, Scotland.) *Below:* Feeder dike intruding tephra with basaltic lava flow above, Iceland. (By: Marli Miller from https://geologypics.com/geological_entries/gallery. For educational purposes.)

Another example of a **cross cutting relationship** includes an igneous intrusive feature known as a dike. In the picture to the left, hot magma forced its way into a fracture, or weakness, in pre-existing rock of the crust. The hot magma also "baked" the pre-existing rock through the process of contact metamorphism. These two clues from this exposure help us understand that the dike is younger than the rock it cross-cuts.

A final possibility occurs when a new pattern, such as tectonic cleavage or folding overprints a pre-existing structure, such as sedimentary bedding or metamorphic foliation. The pattern of folding cuts across (reconfigures) the old pattern. Here is an example from Connecticut, showing both this "overprinting" variety of cross-cutting, as well as a traditional igneous dike:



Figure 7.3.15: What sequence of steps were needed to make this Connecticut outcrop? (Photograph by Gregory Simones; reproduced with permission)



Figure 7.3.16: The sequence of steps needed to make the Connecticut outcrop. (Original photograph by Gregory Simones; reproduced with permission)

Principle of inclusions

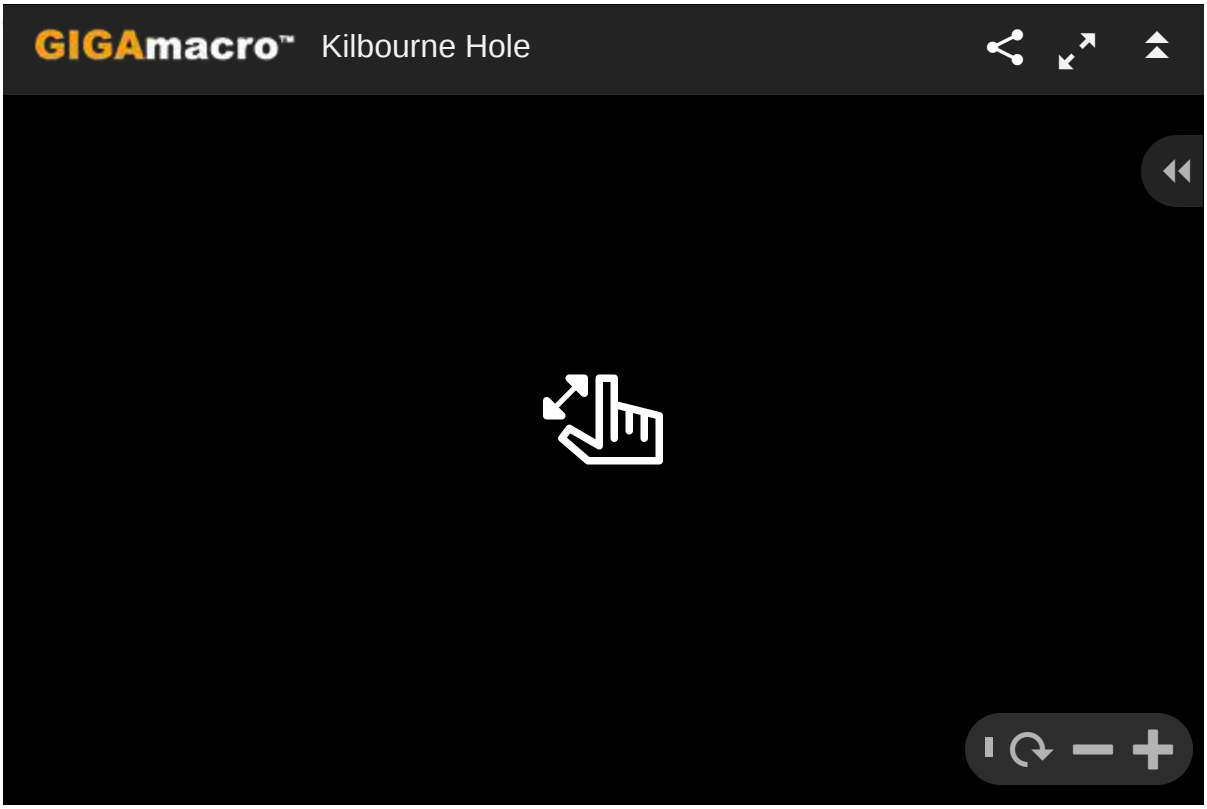
The principle of relative dating by inclusions states that any rock fragments that are included in another body of rock must be older than the rock in which they are included. The photo below provides a good example. Both are igneous rock. The pink rock is granite while the black rock is basalt. A **xenolith** of the granite is included in the mass of basalt rock. How did this happen? Basalt intruded into the pre-existing pink granite. The injection of mafic magma caused a piece of the solid granite to break off, rotate a bit, and then get locked in place as an “inclusion” within the basalt as the dark magma cooled.



Figure 7.3.17: Included fragment (xenolith) of granite within basalt near Georgeville, Nova Scotia. (CC BY-SA 3.0; By M.C. Rygel)

Another interesting feature is apparent in this photo. The pink xenolith displays a much deeper, more rosey color which is also seen along most of the contact between the granite and basalt. This deeper, rosier color is due to the baking of the granite by the basalt. This is referred to as contact metamorphism and is also a relative dating principle. The rock that has been “baked” must have been in place prior to the intrusion of the hot magma/lava which caused the baking. You can’t cook something that doesn’t yet exist.

The following gigapixel panorama is a fascinating example of an inclusion embedded into pyroclastic ash flow deposits. The photo is from Kilbourne Hole in southeast New Mexico. The black basalt volcanic bomb was ejected during a volcanic eruption and landed in a recent pyroclastic ash flow deposit that had yet to solidify. You can easily see how this volcanic bomb deformed the layers of ash as it landed and became embedded. The lower ash layers obviously had to be there prior to the volcanic bomb landing and embedding. Later ash layers buried the bomb.



The Grand Canyon example

The Grand Canyon of the Colorado river in the U.S. state of Arizona illustrates many of the stratigraphic principles discussed thus far. The photo below shows layers of rock on top of one another in order, from the oldest at the bottom to the youngest at the top, an illustration of the principle of superposition. The predominant white layer just below the canyon rim is the Kaibab Limestone. This layer is laterally continuous with the upper white layers of strata in the far distance, even though the intervening canyon separates them. The rock layers exhibit the principle of lateral continuity, as they are found on both sides of the Grand Canyon, ten miles apart.



Figure 7.3.18: Many people consider Grand Canyon National Park the world's premiere geologic landscape and a "geologic wonder". The eroded cliffs reveal 1.7 billion years of fossils, volcanic activity, and geologic history. (United States Public Domain; National Park Service from: https://www.nps.gov/articles/age-of-rocks-in-grand-canyon.htm.)

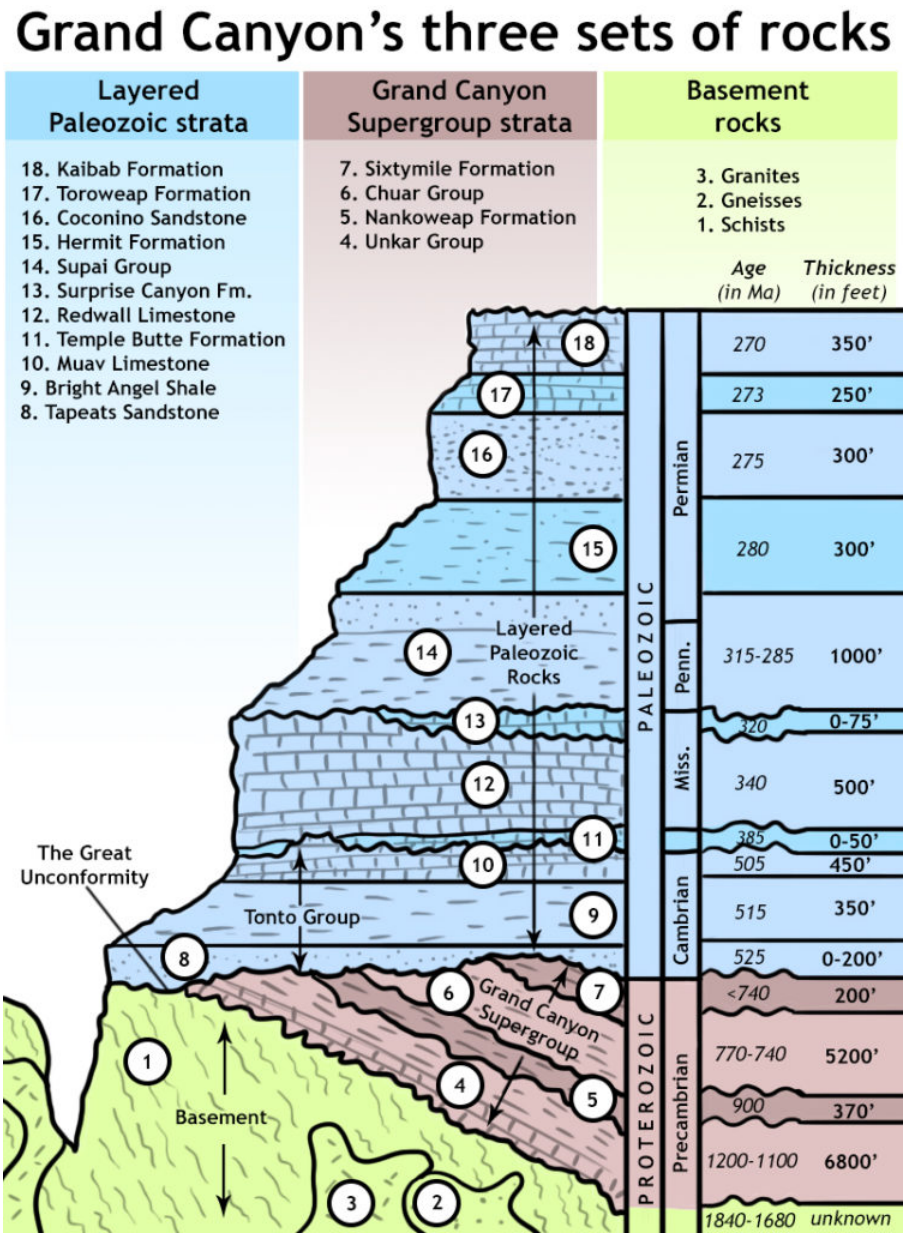


Figure 7.3.19: A schematic cross-section of the strata exposed in the Grand Canyon. (United States Public Domain; Callan Bentley after National Park Service from: https://www.nps.gov/articles/age-of-rocks-in-grand-canyon.htm.)

The diagram (left) shows a cross-section of the rocks exposed on the walls of the Grand Canyon, illustrating the principle of cross-cutting relationships, superposition, and original horizontality. In the lower parts of the Grand Canyon are the oldest sedimentary formations, with igneous and metamorphic rocks at the very bottom. (For this reason, these crystalline rocks are sometimes called "the basement.") The principle of cross-cutting relationships shows the sequence of these events. The metamorphic schist (#1) is the oldest rock formation and the cross-cutting granite intrusion (#3) is younger. As seen in the figure, the other layers on the walls of the Grand Canyon are numbered with #18 being the youngest, of all of the rocks exposed in the canyon, #18 was the last to form. This illustrates the principle of superposition. The Colorado River carves through the Colorado Plateau, exposing the horizontal strata, that follows the principle of original horizontality. These rock strata have been barely disturbed from their original deposition, except by a broad regional uplift.

Unconformities – Missing Time

An unconformity represents an interruption in the process of depositing sedimentary layers. Recognizing unconformities is important for understanding time relationships in sedimentary sequences. We can use the rock exposed in the Grand Canyon to

illustrate examples of all of the different types of unconformities. Locate these on the diagram of Grand Canyon stratigraphy above as each is discussed.

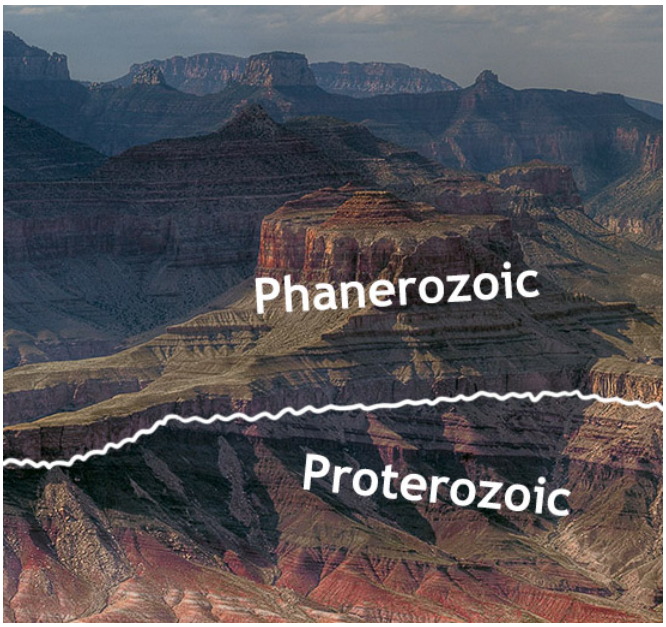


Figure 7.3.20: The great angular unconformity in the Grand Canyon, Arizona. The tilted rocks at the bottom are part of the Proterozoic Grand Canyon Supergroup (aged 825 to 1,250 Ma). The flat-lying rocks at the top are Paleozoic (540 to 250 Ma). The boundary between the two represents a time gap of nearly 300 million years. (Credit: Doug Dolde / Doug Dolde at en.Wikipedia / Public domain. Annotated by Callan Bentley.)

An example of one of the most iconic unconformities is shown in the lower center portion of the photo below from the Grand Canyon. It is known as “The Great Unconformity.” Proterozoic rocks of the sedimentary sequence known as the Grand Canyon Supergroup, have been tilted and then eroded to a flat surface prior to deposition of the younger Paleozoic rocks. The specific type is an **angular unconformity**. The difference in time between the youngest of the Proterozoic rocks and the oldest of the Paleozoic rocks is close to 300 million years. Tilting and erosion of the older rocks took place during this time, and if there was any deposition going on in this area, the evidence of it is now gone.

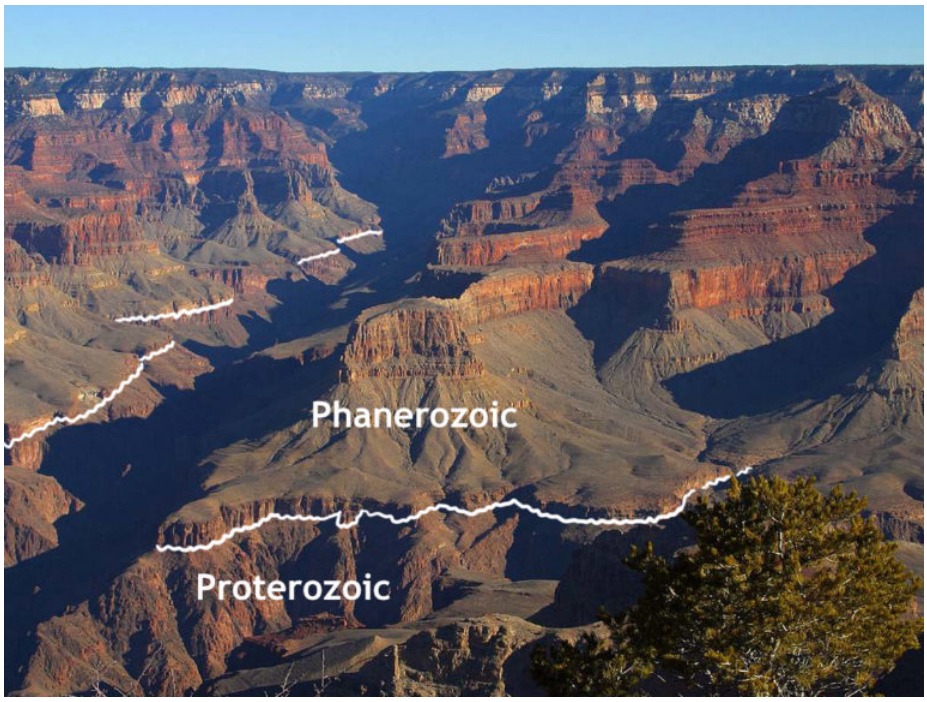


Figure 7.3.21: The view is centered on Sumner Butte, at southeast terminus of Bright Angel Canyon, intersecting with Granite Gorge. (The original uploader was Szumyk at Polish Wikipedia. – http://pdphoto.org/PictureDetail.php?mat=pdef&pg=5506, Public Domain, https://commons.wikimedia.org/w/inde...?curid=5883465. Annotated by Callan Bentley.)

A **nonconformity** represents an erosional surface separating rocks of different types. The erosional surface will separate igneous and/or metamorphic rock from overlying sedimentary rock such as in the photo below, from a different spot in the Grand Canyon. In the photo, right, a nonconformity representing the erosional surface between 1.75 Ga Proterozoic age rock at the base of the Grand Canyon and ~545 Ma Phanerozoic age rock. Over 1 billion years separate the basement rock from the first overlying layer.

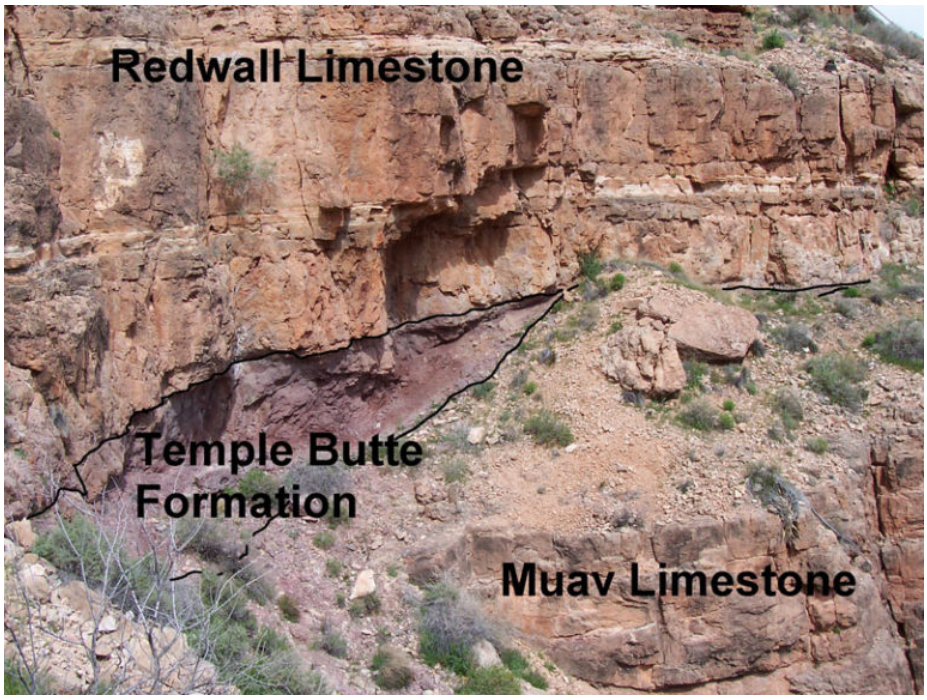


Figure 7.3.22: Redwall, Temple Butte and Muav formations in the Grand Canyon. An erosional surface (disconformity) exists between both the Muav Limestone and the Temple Butte formations as well as between the Temple Butte and the Redwall Limestone formations. (United States Public Domain; National Park Service from: https://upload.wikimedia.org/Wikipedia/commons/e/eb/Redwall%2C_Temple_Butte_and_Muav_formations_in_Grand_Canyon.jpg.)

A disconformity is a little harder to discern: it is not as obvious as the previous examples. Disconformities occur between parallel sedimentary layers. Several exist in the horizontal layers of the Grand Canyon. One occurs between the Muav Limestone formation which is approximately 515 Ma and the overlying Temple Butte and Redwall Limestone formations which are approximately 350 Ma and 335 Ma, respectively. This erosional surface, representing over 150 million years is well displayed in the photo at left.

Example 7.3.1: Unconformity Examples

Let's take a look at some additional examples of the different types of unconformities.

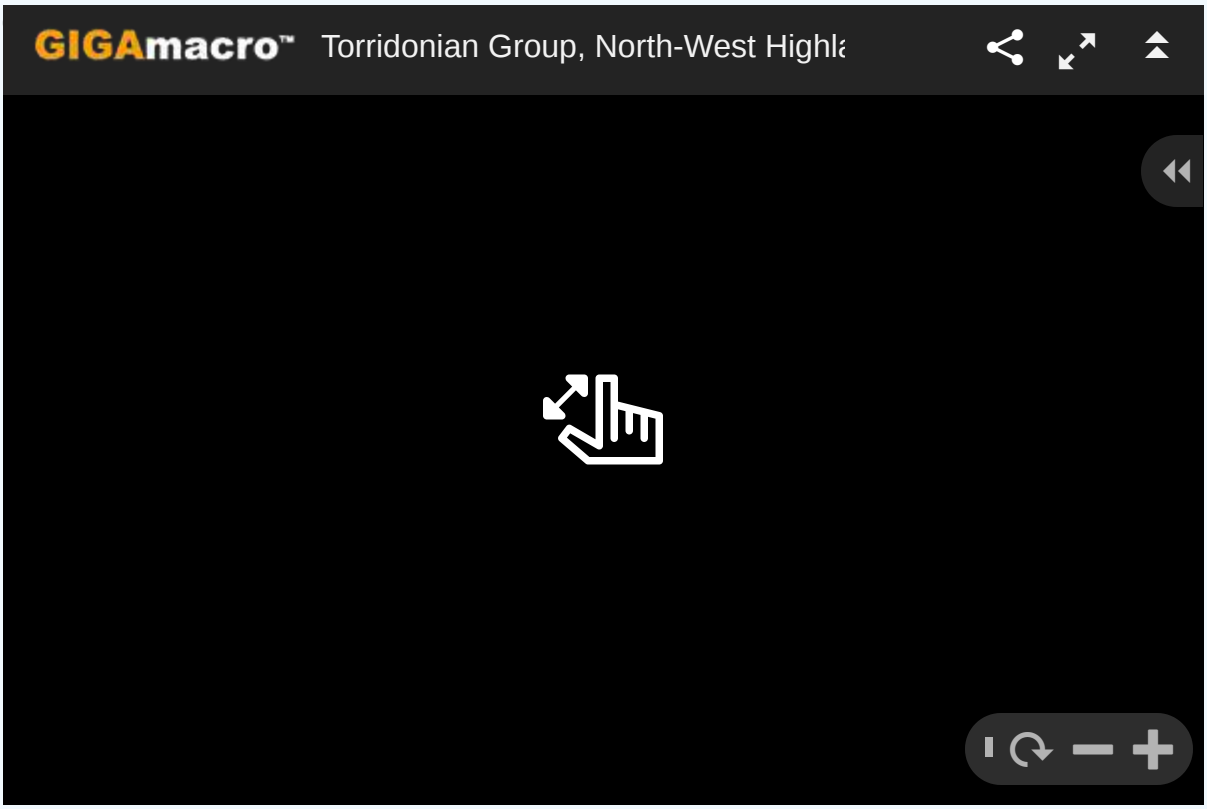
Disconformity:



Angular unconformity:



Nonconformity:



More unconformities:



Figure 7.3.23: Angular unconformity, Telheiro Beach, Portugal, showing Triassic sandstone atop folded, eroded, Carboniferous graywacke and shale. (Photo by Joao Duarte, 2019.)



Figure 7.3.24: Angular unconformity in Loughshinny, Ireland, showing flat-lying Pleistocene till overlying folded limestone and shale. (Photo by Jason Loxton, 2018.)



Figure 7.3.25: Angular unconformity in Loughshinny, Ireland, showing flat-lying Pleistocene till overlying folded limestone and shale of Carboniferous age. (Photo by Jason Loxton, 2018.)



Figure 7.3.26: Carboniferous / Cretaceous angular unconformity at Ense-Bremen, North Rhine-Westphalia, Germany. (Photo by Henning Bergermann; reproduced with permission.)

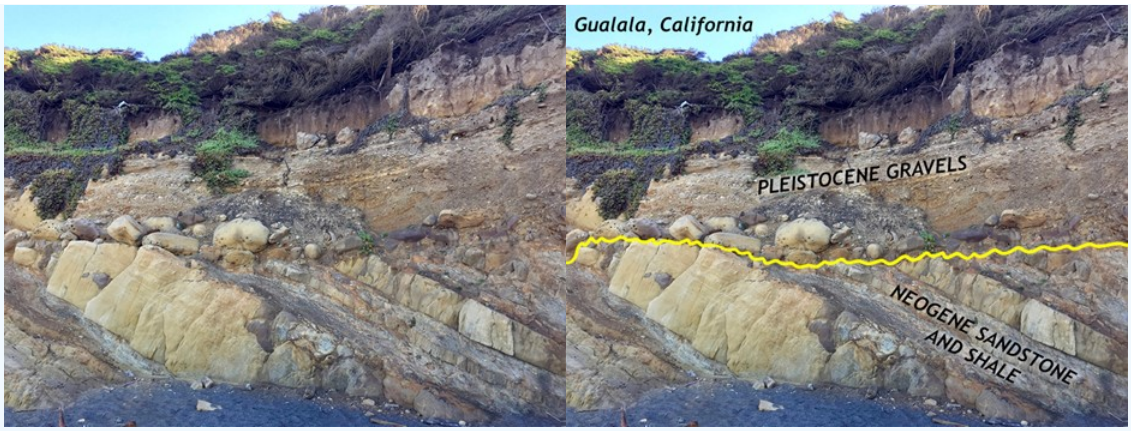


Figure 7.3.27: Tilted layers of Neogene-aged sandstone and shale overlain by Pleistocene gravels (with clam-bored boulder bed just above the unconformity surface). Bonus: a layer of recent colluvium overlies the gravels! (Photo by Seth Finnegan; reproduced with permission.)



Figure 7.3.28: Jurassic-aged Morrison Formation (shale and sandstone) folded and eroded, overlain by Quaternary sandstone and conglomerate. (Photo by Elizabeth Bollen; reproduced with permission.)



Figure 7.3.29: Unconformity between Neoproterozoic andesite (~570 Ma) and overlying Triassic sandstone (~210 Ma). (Photo by Jack Matthews; reproduced with permission.)



Figure 7.3.30: The Knox unconformity (a disconformity bounding the Sauk and Tippecanoe epeiric sequences) exposed on Isle La Motte, Lake Champlain, Vermont. Providence Island Formation of the Beekmantown Group (tidal dolomite of the Middle Ordovician) is overlain by lower Day Point Formation (also Middle Ordovician) from the Chazy Group, which is sandy limestone and carbonate sandstone. Note the karstic dissolution of the lower unit at the unconformity surface. (Photograph by Matthew Clapham; reproduced with permission.)

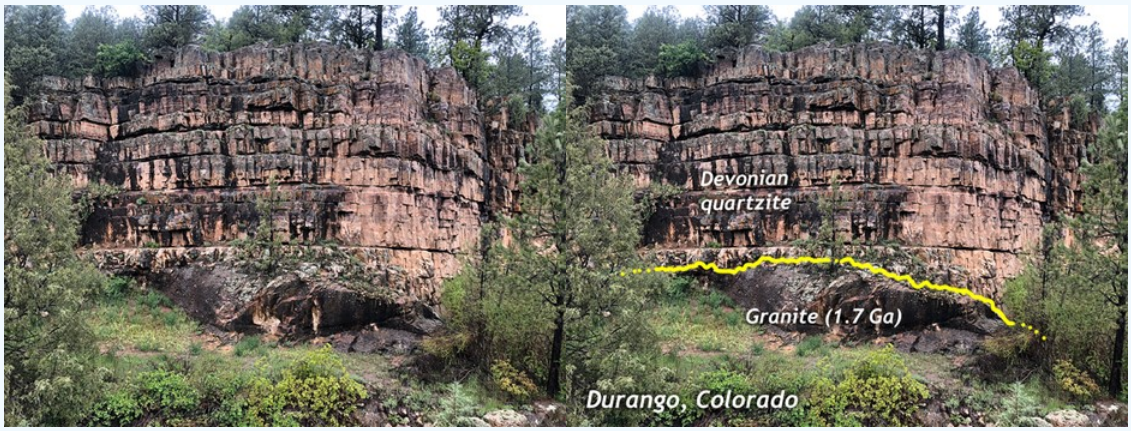


Figure 7.3.31: The “Great Unconformity” (really a nonconformity) north of Durango Colorado. 1.7 Ga Baker’s Bridge granite overlain by late Devonian Ignacio quartzite. (Photograph by Cody Mason; reproduced with permission.)

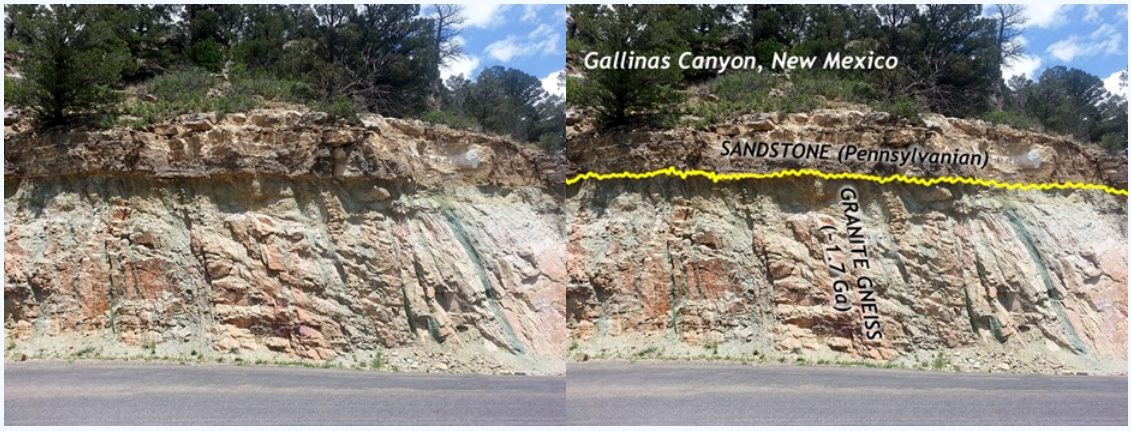


Figure 7.3.32: Nonconformity in Gallinas Canyon, New Mexico. (Photo by Howell Bosbyshell; reproduced with permission.)

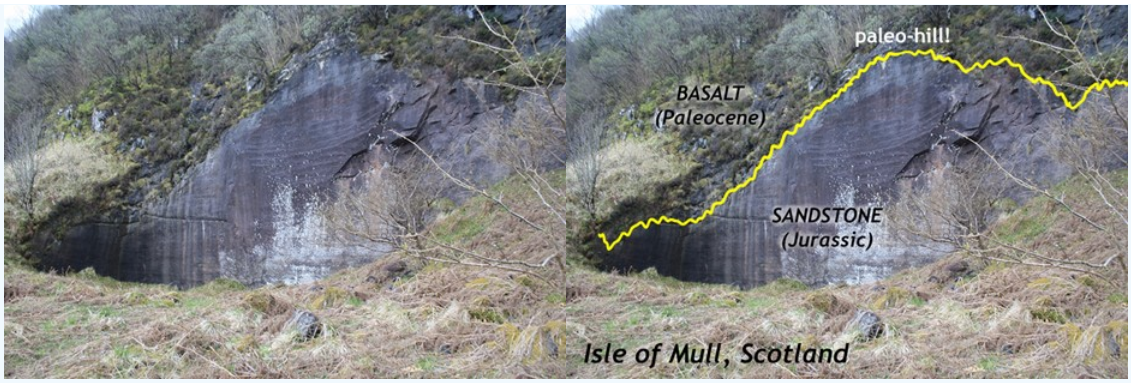


Figure 7.3.33: A paleo-hill of Jurassic sandstone, overlain by Paleocene volcanics mainly basalts. It was quarried for its rock used to make the approach to a local lighthouse on the Isle of Mull. (Photo by James Westland; reproduced with permission.)



Figure 7.3.34: Angular unconformity at El Chalten, Argentina. Jurassic volcanic strata overlie Paleozoic meta-sedimentary strata. (Photo by Donna Anderson; reproduced with permission.)

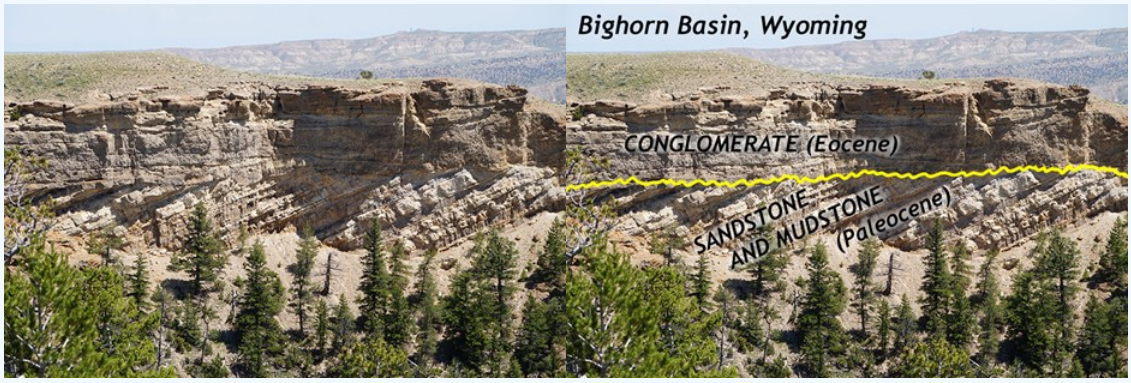


Figure 7.3.35: Dipping Paleocene interbedded sands and muds with overlying horizontal Eocene conglomerates. Bighorn Basin, Wyoming. (Photo by Amanda Owen; reproduced with permission.)

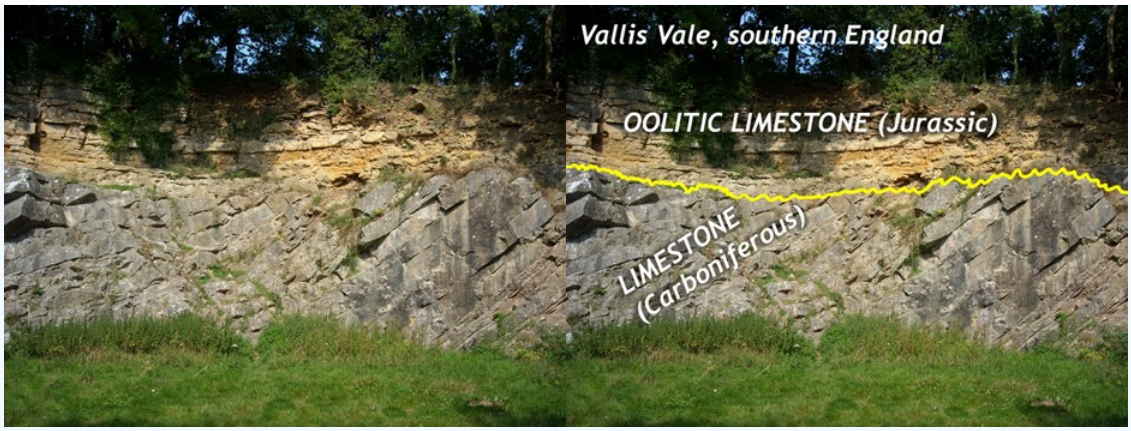


Figure 7.3.36: Probably one of the best examples of an unconformity in Southern England. An unconformity is a buried erosional surface separating strata of different ages; it indicates that sediment deposition was not continuous. In this unconformity at the very eastern end of the Mendips, younger strata of sedimentary Jurassic Inferior Oolite rest upon the eroded surface of tilted older Carboniferous Vallis Limestone. (Reproduced via a CC-by license; Photo by Vieve Forward, https://www.geograph.org.uk/photo/4165184)



Figure 7.3.37: At Chippewa Falls, Wisconsin, there is an unconformity shows a ≈1.8 Ga saprolitized trondhjemite (granitoid) nonconformably overlain by the Cambrian-age Mt. Simon Sandstone. (Photo by Scott Clark; reproduced with permission.)



Figure 7.3.38: Pleistocene terrace deposits over Miocene Astoria Fm. The latter was deposited between episodes of Columbia River flood basalts reaching the coast, which was the case with Otter Rock, the large headland in the mid-ground. (Photo by Lockwood DeWitt; reproduced with permission.)

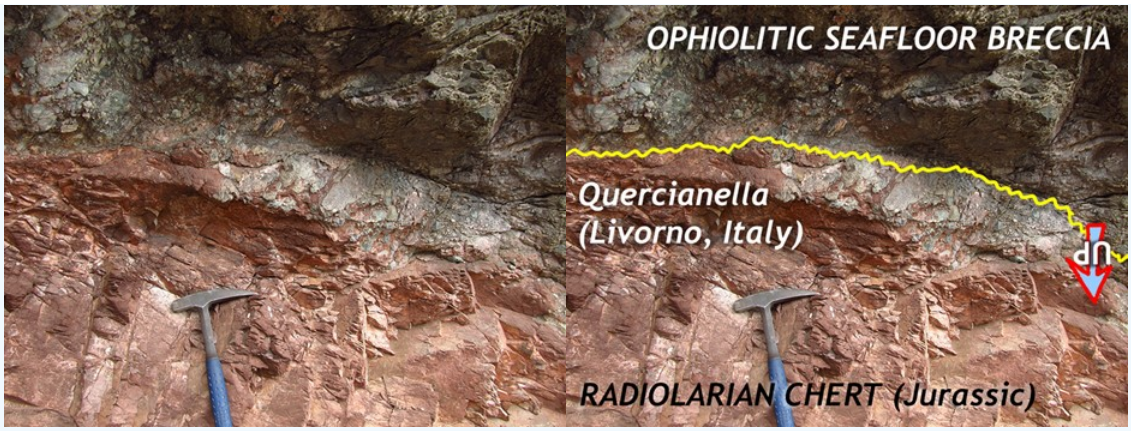


Figure 7.3.39: This is an upside-down unconformity! The rock above is ophiolitic seafloor breccia (160 Ma). The sediments at the bottom are Late Jurassic. Within the sediments there are also turbidites with inverted grading. The whole package has been flipped by tectonics (we are in the reverse limb of a major fold here). (Photo by Samuele Papeschi; reproduced with permission.)



Figure 7.3.40: A monocline overprinting an unconformity. Proterozoic basement rocks (purple, low in canyon) are overlain by the Chinle Formation, with the Wingate Sandstone as the cliff-forming unit. Colorado National Monument. (Photo by Robert Mahon; reproduced with permission.

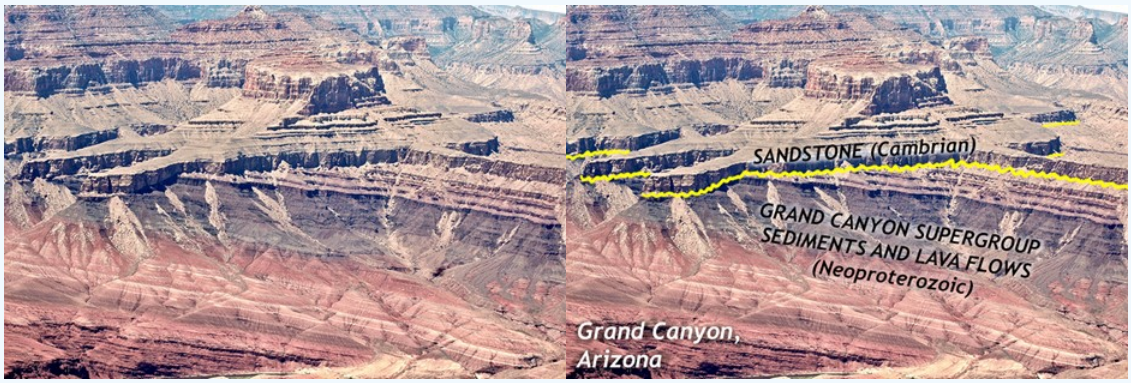


Figure 7.3.41: Angular unconformity in the Grand Canyon. Lower tilted layers are Neoproterozoic strata of the Grand Canyon supergroup (various sedimentary rock and some lava flows). The cliff-forming unit extending horizontally across the field of view is Cambrian-aged Tapeats Sandstone. (Photo by Zoltán Sylvester; reproduced with permission.)



Figure 7.3.42: Pleistocene terrace deposits over deltaic deposits of the Eocene Coaledo Fm, Shore Acres State Park, Oregon. (Photo by Lockwood DeWitt; reproduced with permission.)



Figure 7.3.43: Nonconformity in the Grand Canyon. Lower rock unit is Mesoproterozoic schist (gray) and granite (pink). The cliff-forming unit extending horizontally across the top of the field of view is Cambrian-aged Tapeats Sandstone. (Photo by Zoltán Sylvester; reproduced with permission.)



Figure 7.3.44: Nonconformity in the Grand Canyon. Lower rock unit is Mesoproterozoic schist (gray) and granite (pink). The cliff-forming unit extending horizontally across the top of the field of view is Cambrian-aged Tapeats Sandstone. (Photo by Zoltán Sylvester; reproduced with permission.)



Figure 7.3.45: At Gribun on the Isle of Mull is an angular unconformity. The underlying dipping rocks are Neoproterozoic psammites (meta-sandstones). There are horizontal Triassic conglomerates above. Most of the boulders on the shore are the conglomerate. Higher up, but not seen in this photo, there are also Paleocene basalts. (Photo by James Westland; reproduced with permission.)



Figure 7.3.46: The classic Scottish angular unconformity between Dalradian metasedimentary slates/phyllites/metalimestones and the overlying basal conglomerate of Late Devonian Old Red Sandstone. This exposure is on the Isle of Kerrera. Gylen Castle ruin for scale. (Photo by Roderick Brown; reproduced with permission.)



Figure 7.3.47: Basal Cambrian arenites (sandstones) on Ercall Granophyre (an igneous rock, like a granite). Ercall quarries, Shropshire. (Photo by Jon Radley, University of Birmingham; reproduced with permission.)

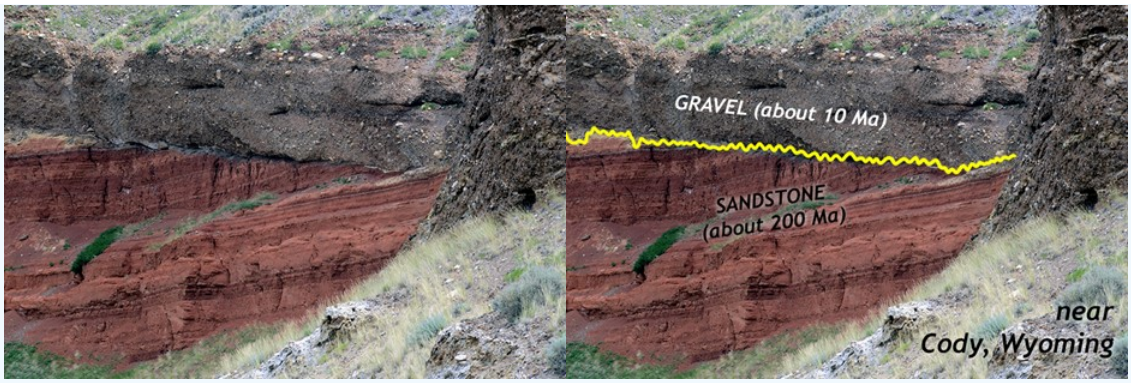


Figure 7.3.48: Angular unconformity near Cody, Wyoming. The lower unit is Triassic Chugwater Formation; the upper unit is Quaternary terrace gravels. (Photo by Ben Edwards; reproduced with permission.)

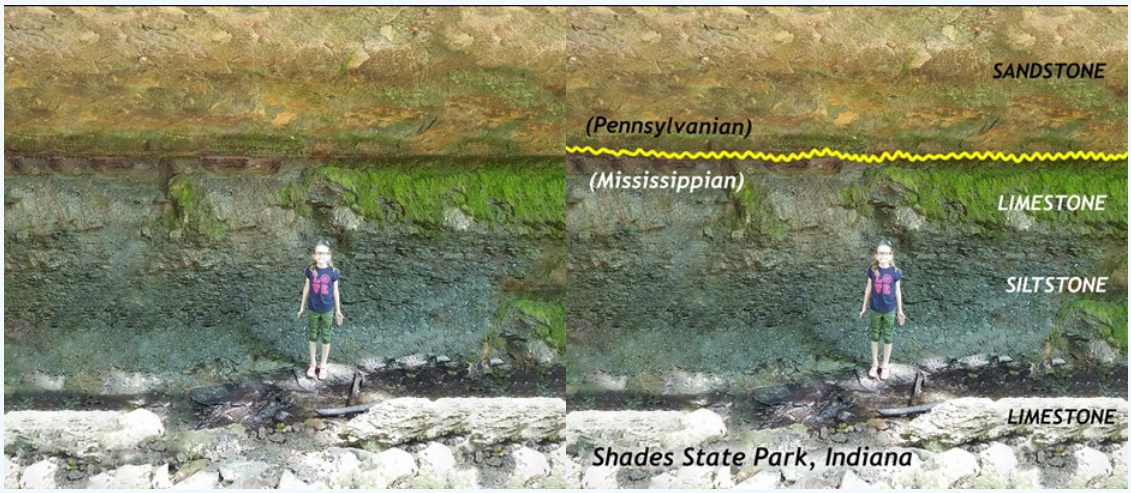


Figure 7.3.49: A gorge in Shades State Park, Indiana shows the Mansfield Sandstone (Pennsylvanian) lying unconformly over limestones and siltstones of the Mississippian aged Borden Group. Glacial waters cut into the sandstone, creating these gorges, exposing the strata. (Photo by Troy Simpson; reproduced here with permission)

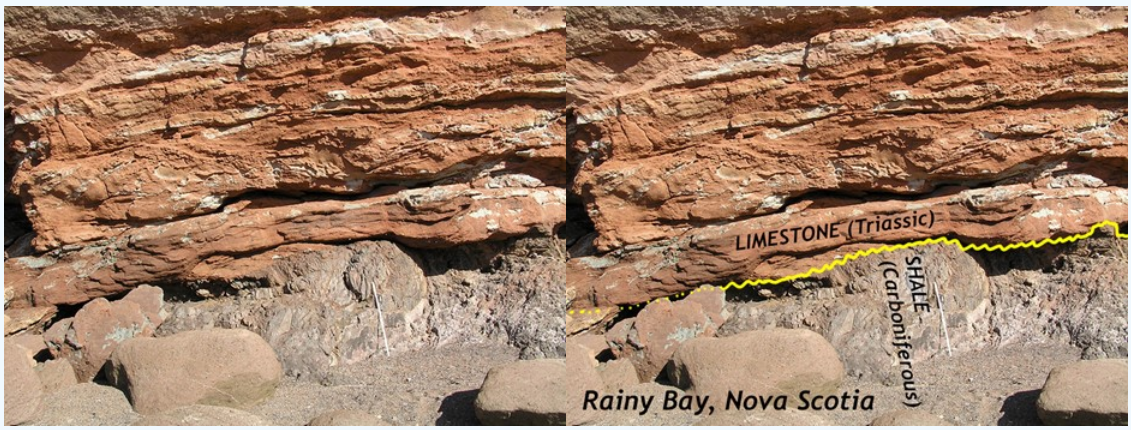


Figure 7.3.50: Gently dipping Triassic strata over vertical Carboniferous layers, Rainy Bay, Nova Scotia, Canada. (Photo by Jon Noad; reproduced with permission.)



Figure 7.3.51: 7,000-year-old Mazama Ash Bed overlying steeply dipping Carboniferous limestone. Exshaw, Alberta, Canada. (Photo by Jon Noad; reproduced with permission.)



Figure 7.3.52: A nonconformity exposed at Deer Creek in the Grand Canyon. The waterfall emerges from the contact between lower Mesoproterozoic basement rocks and overlying Cambrian Tapeats Sandstone. (Photo by Logan Wren Raming; reproduced with permission.)



Figure 7.3.53: The Kentland Disturbance Geologic Area is a theorized impact site where Ordovician and Silurian carbonate strata were forced to the surface by the impact. A quarry has existed here for a hundred years excavating rock aggregate and exposing folded and tilted strata with localized faults and telltale shattercones. The upended strata were then scraped over by Pleistocene glaciation, leaving a thin veneer of glacial deposits on top. (Photo by Troy Simpson; reproduced here with permission.)

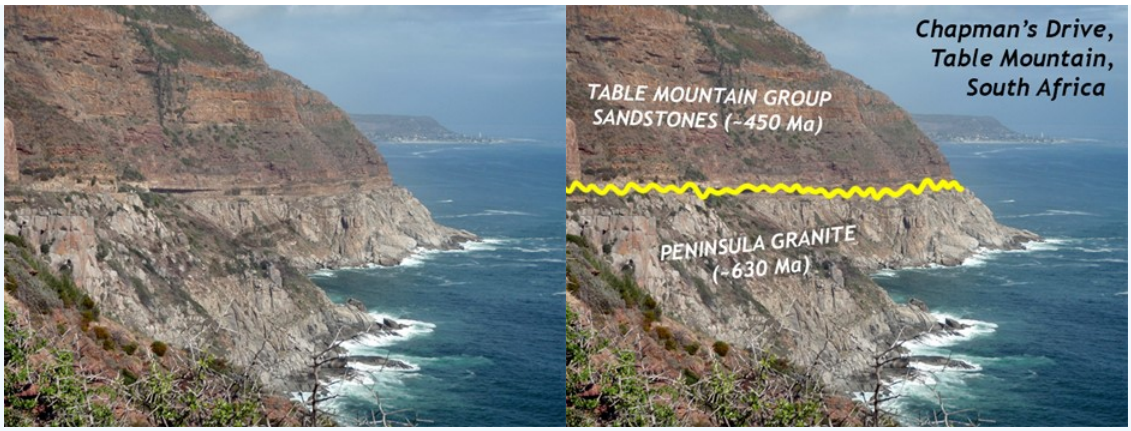


Figure 7.3.54: Chapman's Drive nonconformity, with Table Mountain Group sandstones (~450 Ma) overlying Peninsula Granite (~630 Ma). Cape Town, South Africa. (Photo by Evelyn Mervine; reproduced with permission.)



Figure 7.3.55: Annex Hill State Park in Minnesota shows two unconformities: an angular unconformity between Archean banded iron formation and overlaying Cretaceous shale, and a disconformity between the shale and overlying Pleistocene till.

Did I Get It? - Quiz

? Exercise 7.3.1

What kind of unconformity is shown here?



a. nonconformity

b. disconformity

c. angular unconformity

https://geo.libretexts.org/@go/page/22642

Answer

b. disconformity

Exercise 7.3.2

What kind of unconformity is shown here highlighted in yellow, separating black basalt over white chalk?



a. nonconformity

b. disconformity

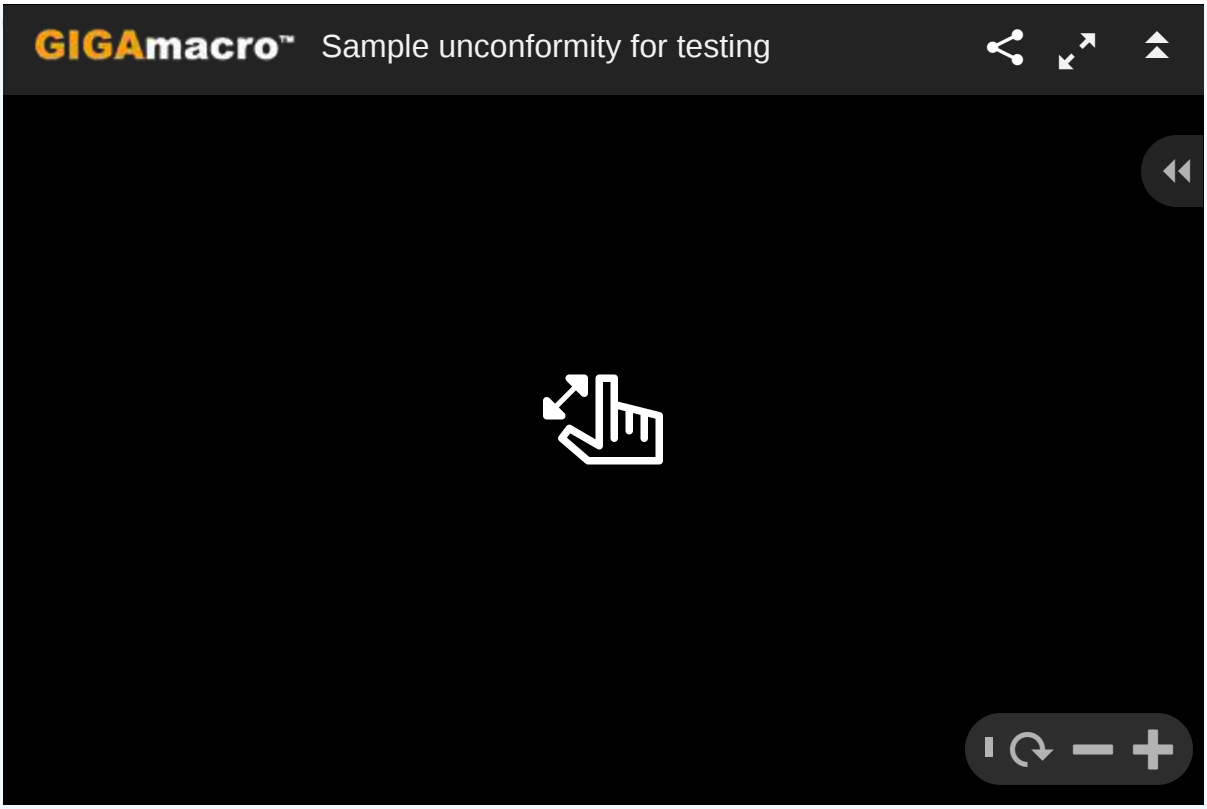
c. angular unconformity

Answer

b. disconformity

Exercise 7.3.3

What kind of unconformity is shown here?



a. nonconformity

b. disconformity

c. angular unconformity

Answer

c. angular unconformity

Exercise 7.3.4

What kind of unconformity is shown?



a. nonconformity

b. disconformity

c. angular unconformity

Answer

a. nonconformity

Exercise 7.3.5

What kind of unconformity is this?



a. nonconformity

b. disconformity

c. angular unconformity

Answer

a. nonconformity

Exercise 7.3.6

What kind of unconformity is shown here? How much time is missing?



a. This is a nonconformity. About one million years are missing.

b. This is a disconformity. About 75 billion years are missing.

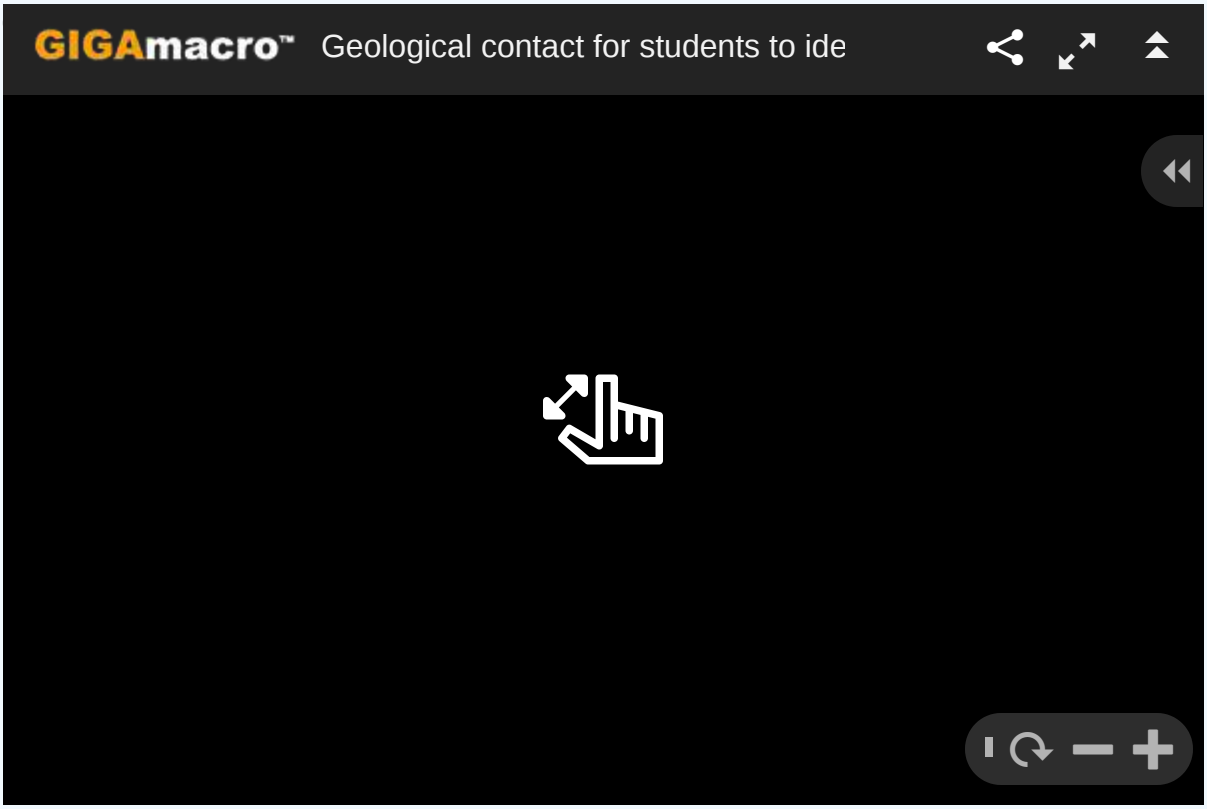
c. This is an angular unconformity. About half a billion years are missing.

Answer

c. This is an angular unconformity. About half a billion years are missing.

Exercise 7.3.7

Interpret this outcrop.



a. It's a nonconformity. It implies that (1) a pluton of magma was intruded, (2) it cooled down and crystallized, (3) it was tilted up and eroded at Earth's surface, (4) a deposit of sediment was laid down, (5) it was lithified, and (6) the whole package was once again raised up to Earth's surface where we can see it (and where it is eroding anew).

b. It's a disconformity. It implies that (1) an initial deposit of sediment was laid down, (2) it was lithified, (3) it was eroded at Earth's surface, (4) a second deposit of sediment was laid down, (5) it was lithified, and (6) the whole package was once again raised up to Earth's surface where we can see it (and where it is eroding anew).

c. It's an angular unconformity. It implies that (1) an initial deposit of sediment was laid down, (2) it was lithified, (3) it was tilted up by mountain-building and eroded at Earth's surface, (4) a second deposit of sediment was laid down, (5) it was lithified, and (6) the whole package was once again raised up to Earth's surface where we can see it (and where it is eroding anew).

Answer

c. It's an angular unconformity. It implies that (1) an initial deposit of sediment was laid down, (2) it was lithified, (3) it was tilted up by mountain-building and eroded at Earth's surface, (4) a second deposit of sediment was laid down, (5) it was lithified, and (6) the whole package was once again raised up to Earth's surface where we can see it (and where it is eroding anew).

Applications of Relative Dating

Example 7.3.2

Let’s take a look at an example from an exposure in Albersweiler, Germany. We usually can only capture several relative dating relationships in one picture. These relationships become much more complex in wider view or larger scale.





Figure 7.3.56: Geologists viewing an exposure in a quarry near Albersweiler, Germany. (Photo by: Paul R. Eizenhöfer, Eberhard Karls Universität Tübingen Germany. With permission from: https://eizenhoefer.wordpress.com/albersweilergermany)

The quarry exposure displays three different rock types. From field investigation we observe the following: 1) three rock types: a grey colored gneiss, a fine grained, black igneous rock, and dark red sandstone, 2) an erosional surface between the gneiss and the overlying sandstone, 3) baking of the gneiss adjacent the the contact with the igneous rock. How can we determine the order of events?

Solution

We can use several principles above help us solve this problem. First, the principle of superposition tells us that the older rocks will be on the bottom and the younger at the top. Therefore, the gneiss is most likely older than the sandstone. The cross-cutting and baked contacts tell us that the black, igneous rock is a dike (intrusion of magma). The erosional surface at the contact between the gneiss and sandstone tells us that an unconformity exists. What type of unconformity exists between metamorphic/igneous rock and sedimentary rock? This is specifically a **nonconformity**. From this evidence we can conclude the following:

Oldest: gneiss, igneous dike, nonconformity, sandstone **Youngest**

✔ Example 7.3.3

The cartoon below shows an imaginary sequence of rocks and geological events labeled A-J. Using the principles of superposition and cross-cutting relationships, can you reconstruct the geological history of this place based upon the information you have available?

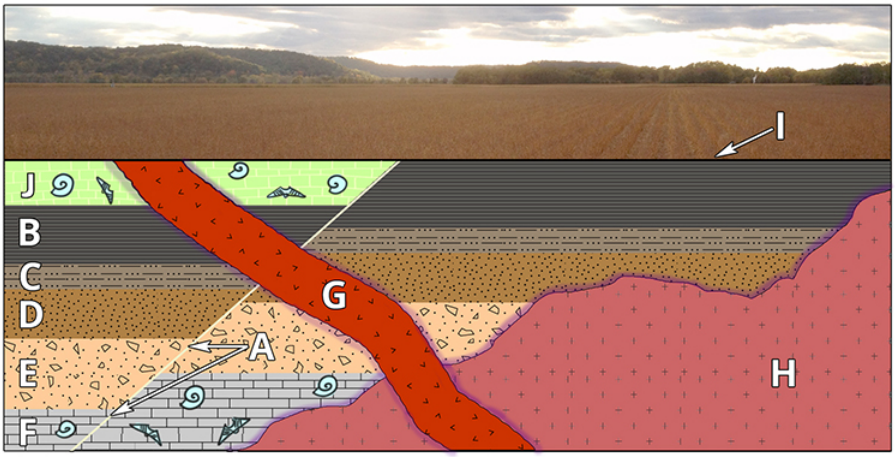


Figure $PageIndex{57}$: Relative dating relationships. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Original image by Jonathan R. Hendricks. Modified by Callan Bentley.)

Solution

Let's work through the imaginary example above.

First, we know from the principle of superposition that rock layer F is older than E, E is older than D, D is older than C, and C is older than B, and B is older than J.

Oldest F, E, D, C, B, J **Youngest**

Second, we observe that rock body H (which is an igneous intrusion) cuts into rock layers B-F, but not J. It is therefore younger than B-F and, with the data we have, we assume J as well (*), however H does not specifically cross-cut J. We can also observe the contact metamorphism "baked rim" around H so we know that B-F had to be there in order to get baked by the intrusion of H.

Oldest F, E, D, C, B, H, J* **Youngest**

Third, we observe that the fault A cuts across and displaces rock layers B-J. It is therefore younger than B-J. Because the fault *does not* cut across H, we do not know if it is older or younger than that rock unit.

Oldest F, E, D, C, B, J*, (H or A) **Youngest**

Fourth, we see that G, another igneous intrusion, cuts across A-J; it is therefore younger than all of these (note that G is not displaced by A, the fault).

Oldest F, E, D, C, B, J*, (H or A), G **Youngest**

Finally, we note an erosional surface, I, at the top of the sequence (and immediately below the corn field) that cuts both A and G. I is therefore younger than both A and G.

Putting this all together, we can determine the relative ages of these rock layers and geological events:

Oldest F, E, D, C, B, J*, (H or A), G, I **Youngest**

Given the information available, we cannot resolve whether H is older than A (or, vice versa). This problem could be resolved, however, if we were able to determine the age of the fossils and the also the age of the igneous intrusions. Read on!

Dating Rocks Using Fossils

Although the recognition of fossils goes back hundreds of years, the systematic cataloguing and assignment of relative ages to different organisms from the distant past—paleontology—only dates back to the earliest part of the 19th century. The oldest undisputed fossils are from rocks dated around 3.5 Ga, and although fossils this old are tiny, typically poorly preserved and are not useful for dating rocks, they can still provide important information about conditions at the time. The oldest well-understood fossils are from rocks dating back to around 600 Ma, and the sedimentary record from that time forward is rich in fossil remains that provide a detailed record of the history and evolution of life on Earth. However, as anyone who has gone hunting for fossils knows, that does not mean that all sedimentary rocks have visible fossils, or that they are easy to find. Fossils alone cannot provide us with numerical ages of rocks, but over the past century geologists have acquired enough isotopic dates from rocks associated with fossil-bearing rocks (such as igneous dikes cutting through sedimentary layers, or volcanic layers between sedimentary layers) to be able to put specific time limits on most fossils.

If we understand the sequence of evolution on Earth, we can apply knowledge to determining the relative ages of rocks. This is William Smith's principle of faunal succession, although of course it doesn't just apply to "fauna" (animals); it also applies to fossils of plants and other organisms.

Geologic Range

If we can identify a fossil to the species level, or at least to the genus level, and we know the time period when the organism lived, we can assign a range of time to the rock. That range might be several million years because some organisms survived for a very long time. If the rock we are studying has several types of fossils in it, and we can assign time ranges to several of them, we might be able to narrow the time range for the age of the rock considerably.

Here's an example. Some organisms survived for a very long time, and are not particularly useful for dating rocks. Sharks, for example, have been around for over 400 million years. This is not a useful range: if you find a shark fossil in a sedimentary layer, it only tells you, "This layer is 400 million years old, or younger." However, the great white shark has a range of 16 million years, so far, which is a much shorter amount of time. Organisms that lived for relatively short time periods are particularly useful for dating rocks, especially if they were distributed over a wide geographic area, and so can be used to compare rocks from different regions. These are known as **index fossils**. There is no specific limit on how short the time span has to be to qualify as an index fossil. Some lived for millions of years, and others for much less than a million years.

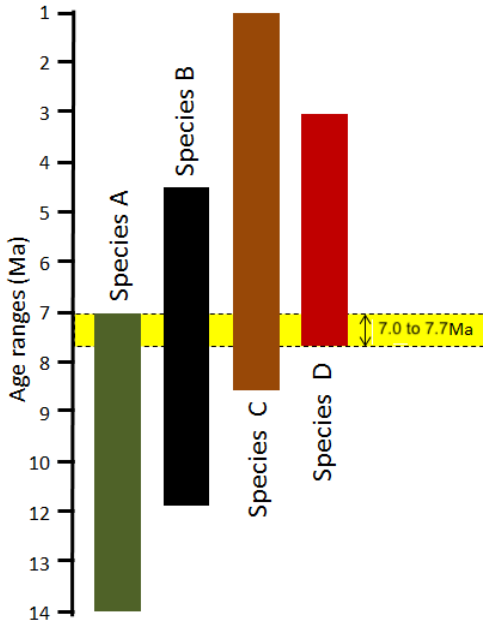


Figure 7.3.58: **Geologic Range Chart**. The application of bracketing to constrain the age of a rock based on several fossils. In this diagram, the colored bars represent the time range during which each of the four species (A, B, C, D) existed on Earth. Although each species lived for several million years, we can narrow down the likely age of the rock to just 700,000 years during which all four species coexisted. (CC BY 4.0; Steven Earle. From: https://opentextbc.ca/physicalgeolog...using-fossils/)

A geoscientist can also date rocks using the geologic ranges of an fossil assemblage from a single sedimentary layer. A fossil assemblage (or faunal assemblage) is a group of organisms that all found fossilized together in that layer. This application involves the bracketing of overlapping ranges to constrain the age of a rock based on several fossils. Consider the diagram left–four species of fossils (A, B, C, D) were collected from one stratum. The geologic ranges of each species are plotted side by side in a **geologic range chart**. Although each species lived for several million years, we can narrow down the likely age of the rock to just 700,000 years during which all four species coexisted. Comparison of these geologic ranges shows only a short time interval (from 7-7.7 Ma) when all four species overlap. Therefore, the age of the sedimentary layer from which these fossils were collected is constrained to the interval when all four species coexist.

Biozones

Some well-studied groups of organisms qualify as biozone fossils (a stratigraphic interval that can be defined on the basis of a specific fossil) because, although the genera and families lived over a long time, each species lived for a relatively short time and can be easily distinguished from others on the basis of specific features. For example, ammonoids have a distinctive feature known as the suture line—where the internal shell layers (septae) that separate the individual chambers meet the outer shell wall. These suture lines are sufficiently variable to identify species. These can be used to estimate the age of the rocks in which they are found. Generally, the more complex the suture pattern, the younger the fossil. Compare the sutures in the following fossil specimens. The fossil ammonoid on the left is Triassic in age, specifically living in the range between 251-237 Ma. The fossil ammonoid on the right is Cretaceous in age, specifically living in the range between 113-100 Ma. Select the play button to explore these fossils in 3D.

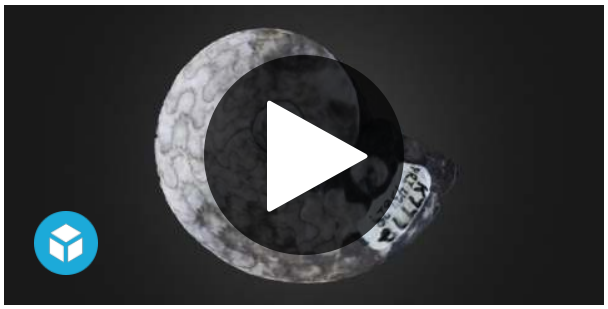


Figure 7.3.59: Fossil specimen of the ceratite ammonoid Gymnotoceras beachi from the Triassic Toad Formation of British Columbia (PRI 43629). Specimen is from the collections of the Paleontological Research Institution. Maximum diameter is approximately 4.5 cm. (CC0 Public Domain; Model by Emily Hauf.)

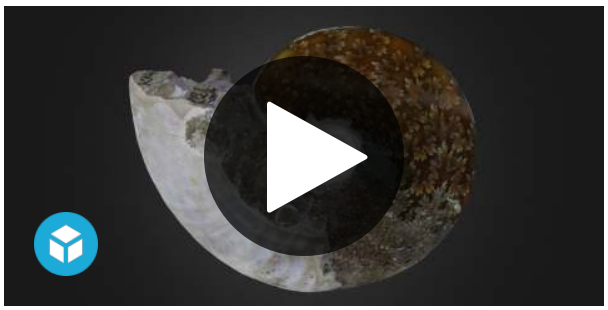


Figure 7.3.60: Fossil specimen of the ammonoid cephalopod Cleoniceras besairiei from the Cretaceous period. Specimen is from the collections of the Paleontological Research Institution, Ithaca, New York. Maximum diameter of specimen is approximately 12 cm. (CC0 Public Domain; Model by Emily Hauf.)

Foraminifera are small, carbonate-shelled marine organisms that originated during the Triassic and are still around today. They are extremely useful biozone fossils. As shown in the figure below, numerous different foraminifera lived during the Cretaceous. Some lasted for over 10 million years, but others for less than 1 million years. If the foraminifera in a rock can be identified to the species level, we can get a good idea of its age. The following is an abbreviated geologic time scale displaying select index foraminifera and age range from the Cretaceous and Paleogene time periods. The gigapixel photo that follows is of *Cribratina texana*, collected at "Fossil Hill," near the Chihuahua (Mexico) / Texas / New Mexico triple point border. Being that this is an index fossil, we can accurately date the age of the rock that hosts the fossil as being between 112-93.5 Ma.

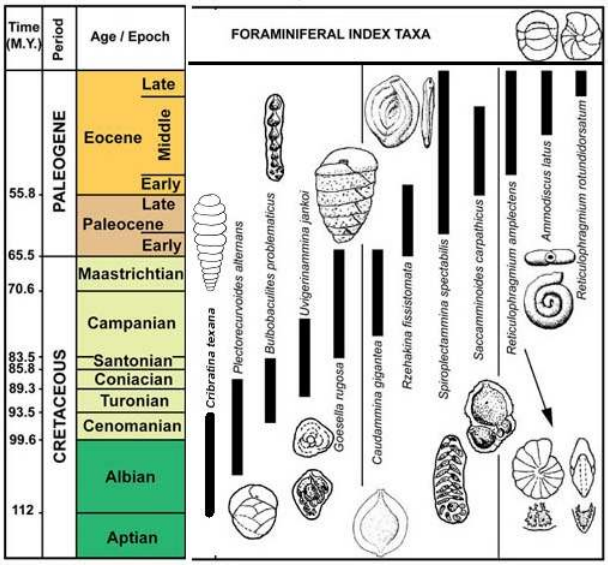
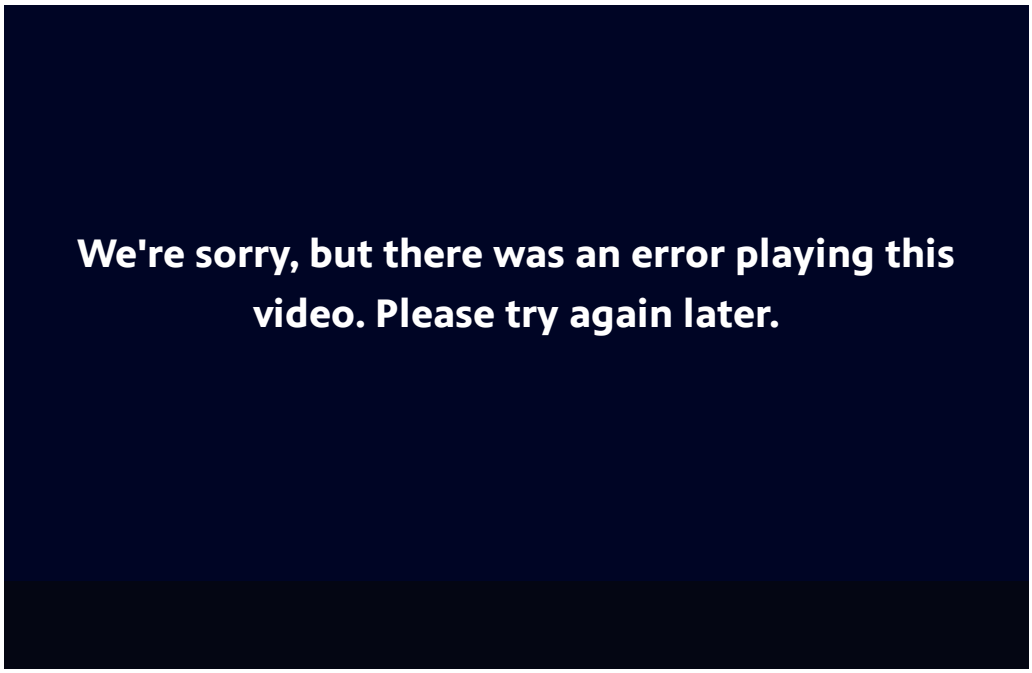


Figure 7.3.61: Select index foraminifera from the Cretaceous and Paleogene. (For educational purposes from the Geologic Timescale Foundation, https://timescalefoundation.org/. Modified by Shelley Jaye.)



Figure 7.3.62: Fossils of the benthic foraminiferid Cribratina, an index fossil for the Albian-Cenomanian age, collected at “Fossil Hill,” north of the Cristo Rey laccolith at the Chihuahua (Mexico) / Texas / New Mexico triple border. (Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.).)

Watch this video to learn about the discovery of conodonts and their use as index fossils.



Numerical (or Absolute) Dating

Numerical (also known as absolute) dating deals with assigning actual dates (in years before the present) to geological events. Contrast this with relative dating, which instead is concerned with determining the order of events in Earth’s past. The science of

absolute age dating is known as geochronology and the fundamental method of geochronology is radiometric dating (also known as isotopic dating).

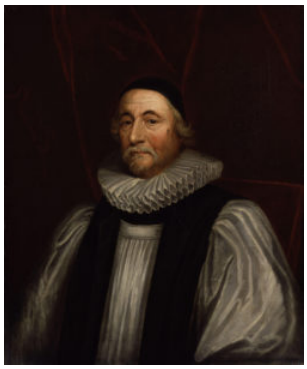


Figure 7.3.63: Archbishop James Ussher (1581-1656) (public domain; WikiMedia Commons).

Scholars and naturalists, understandably, have long been interested in knowing the absolute age of the Earth, as well as other important geological events. In 1650, Archbishop James Ussher (1581-1686) famously used the genealogy of the Old Testament of the Bible (e.g., Genesis, Chapter 5), and the human lifespans recorded in it, to estimate the age of the Earth. He concluded that the Earth was young in age, having formed in 4004 B.C., or about 6,000 years ago.

In the 1800's, practitioners of the young science of geology applied the uniformitarian views of Hutton and Lyell (discussed earlier in this chapter) to try to determine the age of the Earth. For example, some geologists observed how long it took for a given amount of sediment (say, a centimeter of sand) to accumulate in a modern habitat, then applied this rate to the total known thickness of sedimentary rocks. When they did this, they estimated that the Earth is many millions of years old.

We now know that this estimate is far, far too young. In part, this estimate is so low because these early geologists did not recognize that unconformities, which represent missing units of time, often caused by erosion, are rampant in the rock record, as well as the fact that some metamorphic rocks were once sedimentary, and thus left out of their calculations. Still, this estimate was on the order of millions of years, rather than Ussher's calculation of 6,000. Geologists were beginning to accept the views of Hutton that the Earth is unimaginably ancient.

What key discovery, then, allowed geologists to begin assigning absolute age dates to rocks and ultimately discover the age of the Earth? The answer is radioactivity.

Radiometric dating

Hypotheses of absolute ages of rocks (as well as the events that they represent) are determined from rates of **radioactive decay** of some isotopes of elements that occur naturally in rocks.

ELEMENTS AND ISOTOPES

In chemistry, an element is a particular kind of atom that is **defined by** the number of protons that it has in its nucleus. The number of protons equals the element's atomic number. Have a look at the periodic table of the elements below. Carbon's (C) atomic number is 6 because it has six protons in its nucleus; gold's (Au) atomic number is 79 because it has 79 protons in its nucleus.

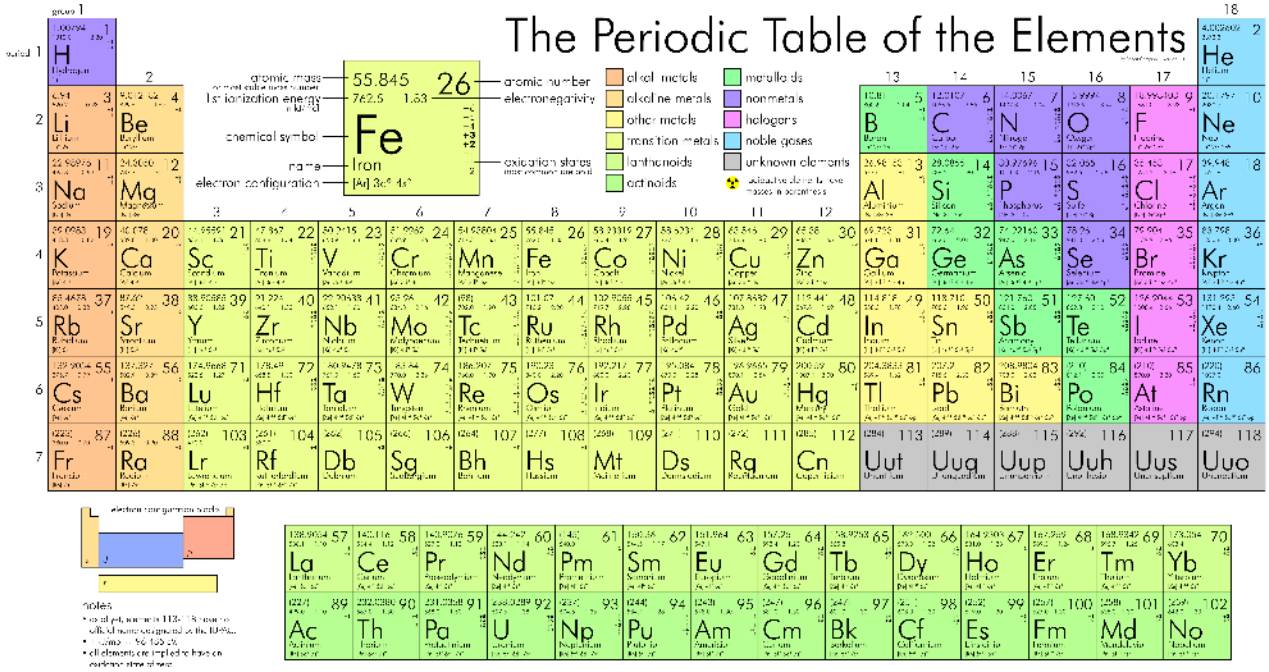


Figure 7.3.64: The periodic table of the elements of the world, showing element names, oxidation states, electron negativities, first ionization energies, and electron configurations. (CC0 1.0 Universal; Public domain from Wikimedia)

Even though individual elements always have the same number of protons, the number of neutrons in their nuclei sometimes varies. These variations are called **isotopes**. Isotopes of individual elements are defined by their **mass number**, which is simply the number of protons + the number of neutrons.

Consider, for example, the three different isotopes of Carbon:

- Carbon-12 (^{12}C): 6 protons, 6 neutrons
- Carbon-13 (^{13}C): 6 protons, 7 neutrons
- Carbon-14 (^{14}C): 6 protons, 8 neutrons

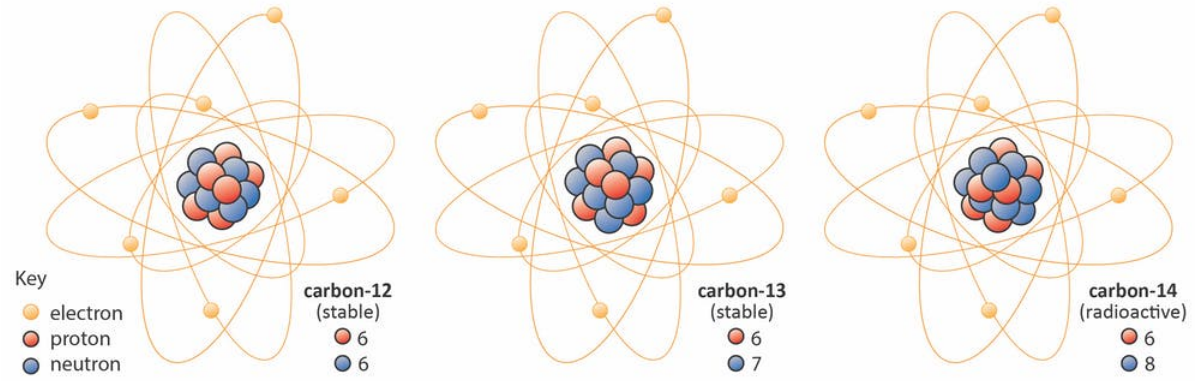


Figure 7.3.65: The three isotopes of carbon. (CC-ND 2.0; Tessa Koumoundouros from: https://theconversation.com/explainer-what-is-an-isotope-10688.)

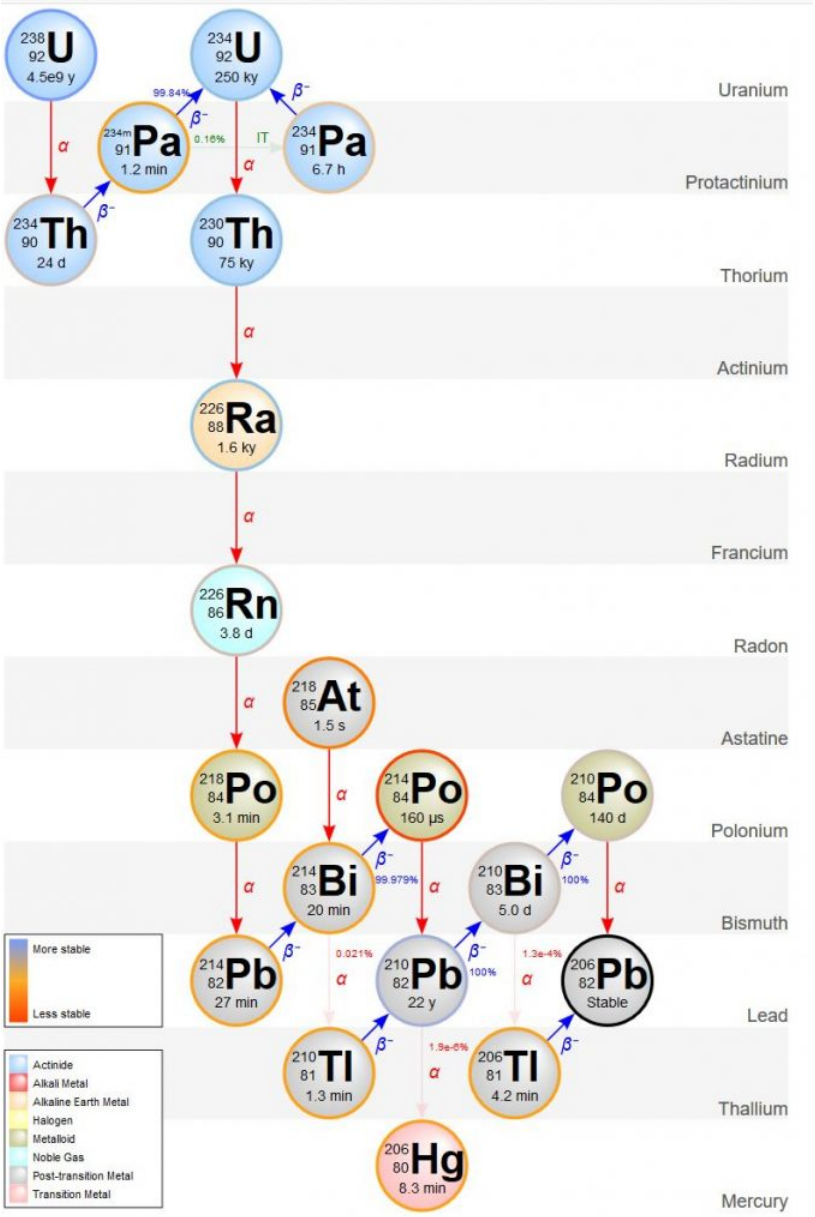


Figure 7.3.66: Decay chain of U-238 to stable Pb-206. (Creative Commons Attribution-NonCommercial-ShareAlike 4.0 International License; Salt Lake Community College, An Introduction to Geology, from: http://opengeology.org/textbook/7-geologic-time/#721_Radioactive_Decay/.)

Most isotopes are stable, meaning that they do not change, or decay, over time; they could remain the same stable isotope for eternity. Some isotopes, however, are unstable and undergo radical change through the process of radioactive decay. This involves the unstable isotopes shedding energy in the form of radiation and subatomic particles, causing their numbers of protons and neutrons to change. Once the number of protons change, the isotope is no longer the same element. The radioactive element will decay to form a different, stable element. The atomic nucleus that undergoes radioactive decay is known at the **parent** and the resulting stable element, the **daughter product** (or, decay product). Consider the example of the radioactive element $^{238}\mathrm{U}$ (Uranium-238) which will radioactively decay to $^{206}\mathrm{Pb}$ (Lead-206). This is not a one step process as indicated in the diagram below. We begin with $^{238}\mathrm{U}$ in the upper left. As the decay process releases particles and energy, $^{238}\mathrm{U}$ goes through a significant transformation, morphing into many other similarly radioactive elements prior to ending up as a stable isotope of lead, $^{206}\mathrm{Pb}$. Some steps take only fractions of a second, others may take thousands of years, however, relative to the pace of “geologic time,” this process happens very quickly.

How do we apply the radioactive decay of unstable isotopes to the dating of rock? We are not really dating the rock, we are dating a mineral in the rock. Radioactive elements become part of the mineral chemistry during the crystallization process. The key to understanding how this works is to realize that the physical chemistry of a growing mineral crystal can be 'picky' or it can be 'sloppy.' When 'picky' minerals are forming, their growing mineral crystal lattice can be very "exclusive" about what atoms it will let into its crystal structure. It accepts a select few kinds of atoms, but turns away most every other kind. The opposite would be sloppy – where all sorts of different atoms are permitted into the developing lattice. Sloppy minerals aren't going to be useful for geologic dating. We need picky minerals.

The ideal mineral for numerical dating accepts some atoms into its crystal that are radioactive parent isotopes. The parent will **decay at a known rate**, diminishing in number over time but causing their daughter (decay) products to accumulate over time in a one-for-one match. The ideal mineral loves bringing in the parent, but would never include the daughter in the developing crystal structure – like a "no children allowed" rule!

In some cases, the radioactive element in question is an isotope of an element essential to the mineral's growth; one of its "standard ingredients" – for instance, $^{40}\mathrm{K}$ getting incorporated into the 'potassium spot' in the potassium feldspar crystal ($\mathrm{KAlSi_3O_8}$), where most of the atoms in that position are stable (not radioactive) isotopes of the same element, $^{39}\mathrm{K}$ or $^{41}\mathrm{K}$ in this case. $^{40}\mathrm{K}$ is the radioactive parent. $^{40}\mathrm{K}$ will decay to form $^{40}\mathrm{Ar}$ as a stable daughter product. However, $^{40}\mathrm{Ar}$, due to its size, will never be found in the original mineral. As radioactive decay of $^{40}\mathrm{K} \rightarrow {}^{40}\mathrm{Ar}$ transpires over time, both elements remain trapped within the closed system of the formed mineral.

In other cases, we're focused on elemental impurities in the mineral crystal – atoms you won't find in the mineral's common chemical formula. Specifically, picky minerals will (a) let certain radioactive atoms into their crystals, but (b) at the same time, they actively exclude the stable daughter atoms that result when the radioactive parents break down. The parent isotope "fits" into the crystal, by virtue of its valence state and its atomic size. Conversely, the daughter isotope *doesn't* fit into the crystal, either because of its valence state or because of its atomic size, or both. An example of this occurs in the mineral zircon ($\mathrm{ZrSiO_4}$). Stable atoms of Zr dominate in the crystal lattice however radioactive $^{238}\mathrm{U}$ atoms can substitute for Zr. The daughter product of the decay of $^{238}\mathrm{U}$ is $^{206}\mathrm{Pb}$. Due to the atomic characteristics of Pb, there is no "happy place" for Pb in the crystal, so Pb atoms will never enter the zircon crystal lattice while it is forming. As radioactive decay of $^{238}\mathrm{U} \rightarrow {}^{206}\mathrm{Pb}$ occurs over time, both elements remain trapped within the closed system of the formed mineral.

The parent-daughter decay pair found within the mineral crystal starts off with 0% daughter isotopes, and 100% radioactive parent isotopes. Based on the previous discussion, we can be confident that any daughter isotopes we find in the mineral crystal got there due to radioactive decay of a parent isotope some time after the crystal originally formed, and they were NOT there from the very beginning. They form in place, due to the spontaneous breakdown of their radioactive parent isotopes, and they are "trapped" in a mineral crystal in which they do not belong. A comprehensive count of all the parent isotopes and all the daughter isotopes in that mineral system will sum to 100%.

Minerals that meet these criteria are very few and far between. The most applicable ones are listed below on the **Half-life Decay Constant Chart.**

HALF-LIFE DECAY CONSTANT CHART

Table 7.3.1: Half-Life Decay Constant Chart.

Mineral	Radioactive parent that is "allowed" in the crystal	Daughter isotope that results	Half-life decay constant of the parent isotope	Useful age range in geologic dating	Useful in Igneous rocks	Useful in Sedimentary rocks	Useful in Metamorphic rocks
Zircon (also Monazite)	^{238}U	^{206}Pb	4.5 billion years	More than 100 Ma	✓	✓*	✓**
	^{235}U	^{207}Pb	703.8 million years	More than 100 Ma	✓	✓*	✓**
	^{232}Th	^{208}Pb	14 billion years – longer than the age of the Universe!	More than 200 Ma	✓	✓*	✓**
Muscovite, biotite, hornblende, potassium feldspar	^{40}K	^{40}Ar	1.25 billion years	More than 100,000 years	✓		✓
Muscovite, biotite, potassium feldspar	^{87}Rb	^{87}Sr	48.8 billion years ~4 times older than the age of the Universe!	More than 100 Ma	✓		✓?

*as detrital grains within a sedimentary deposit: gives a maximum age for the deposit.
**as overgrowths on original grains – must be distinguished from the mineral's "core" during the dating process or as reset ages, if the rock has undergone very hot metamorphism (>800 °C).

Did you notice how ***few*** minerals there are? Of the thousands of minerals known, only a handful are common enough in rocks and 'picky' enough in their chemistry to be useful for isotopic dating. Our toolbox has only about a dozen tools in it. We can't numerically date quartz. We can't numerically date calcite. Same for olivine and pyroxene and most of the important rock forming minerals.

Two pieces of information are needed to calculate how old a given mineral is: the **ratio** of parent to daughter isotopes, and the **rate of decay**, or **half-life,** of the radioactive parent.

- The **ratio** of parent to daughter atoms is measured by using a mass spectrometer which is a very sensitive, high precision instrument that detects and separates atoms based on their mass.
- The **rate** at which a particular parent isotope decays into its daughter product is constant. This rate is determined in a laboratory setting. **A half-life is the amount of time needed for half of the parent atoms in a sample to decay into daughter products.** This is illustrated in the chart below.

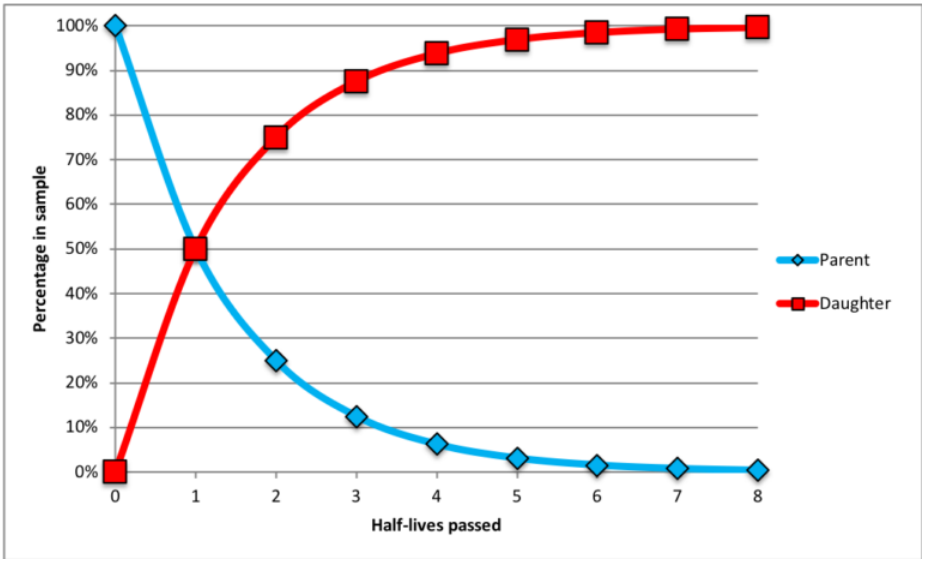


Figure 7.3.67: Radioactive Decay Curve diagram. Relationship between the amount of radioactive parent atoms in a sample relative to the number of daughter atoms over the passage of time, measured in half-lives. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

At the start time (zero half-lives passed), the sample consists of 100% parent atoms (blue diamonds); there are no daughter products (red squares) because no time has passed. After the passage of one half-life, 50% of the parent atoms have become daughter products. After two half-lives, 75% of the original parent atoms have been transformed into daughter products (thus, only 25% of the original parent atoms remain). After three half-lives, only 12.5% of the original parent atoms remain. As more half-lives pass, the number of parent atoms remaining approaches zero.

The ratio of parent atoms relative to daughter products in a sample will determine how many half-lives have passed since a mineral grain first formed. Consider the example shown below.

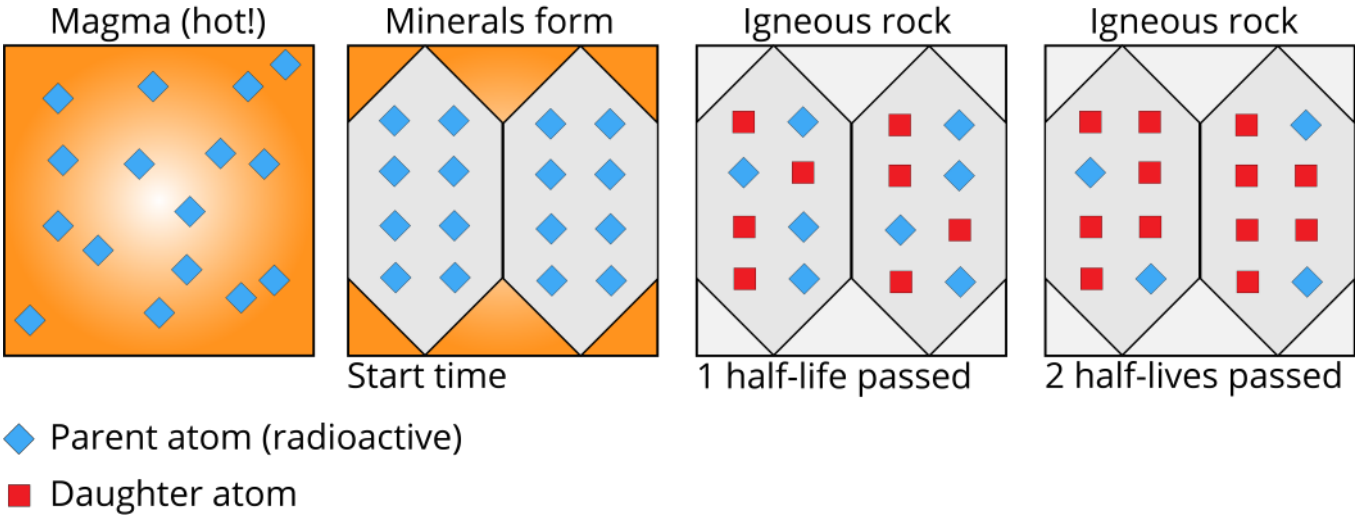


Figure 7.3.68: An example of how the initial number of radioactive parent atoms (blue diamonds) in two mineral grains (gray hexagons) changes over time (measured in half-lives) relative to the number of daughter products (red squares). (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

The left-most box in the figure above represents an initial state, with parent atoms distributed throughout molten rock (magma). As the magma cools, grains of different minerals begin to crystalize. Some of these minerals (represented above as gray hexagons) incorporate the radioactive parent atoms (blue diamonds) into their crystalline structures; this marks the initiation of the “half-life clock” (i.e., the start time, or time zero). After one half-life has passed, half (50%, or four) of the parent atoms in each mineral grain have been transformed into their daughter products (red squares). After two half-lives have passed, 75% (six) of the original parent

atoms in each grain have been transformed into daughter products. How many parent atoms would remain if three half-lives passed?

CALCULATING RADIOMETRIC DATES

The next step in radiometric dating involves converting the number of half-lives that have passed into a numerical (i.e., absolute) age. This is done by **multiplying** the **number of half-lives** that have passed by the **half-life decay constant** of the parent isotope (see the table above for decay constants).

Let's step through a simple problem to determine the age of the sample of granite below.

Granite contains the mineral potassium feldspar which can be dated using the radioactive decay pair of $^{40}\mathrm{K} \rightarrow {}^{40}\mathrm{Ar}$ [2] We send our sample of potassium feldspar off to the lab to be analyzed using the latest mass spectrometry methods to determine the ratio of parent isotope of $^{40}\mathrm{K}$ to daughter isotope of $^{40}\mathrm{Ar}$. Analysis reveals that 75% of our parent $^{40}\mathrm{K}$ remains and 25% of our daughter is present. We can use the half life chart (above) and the graph below to determine the age of the granite. The chart tells us that the half life of $^{40}\mathrm{K}$ is 1.25 billion years. We read the graph as follows:

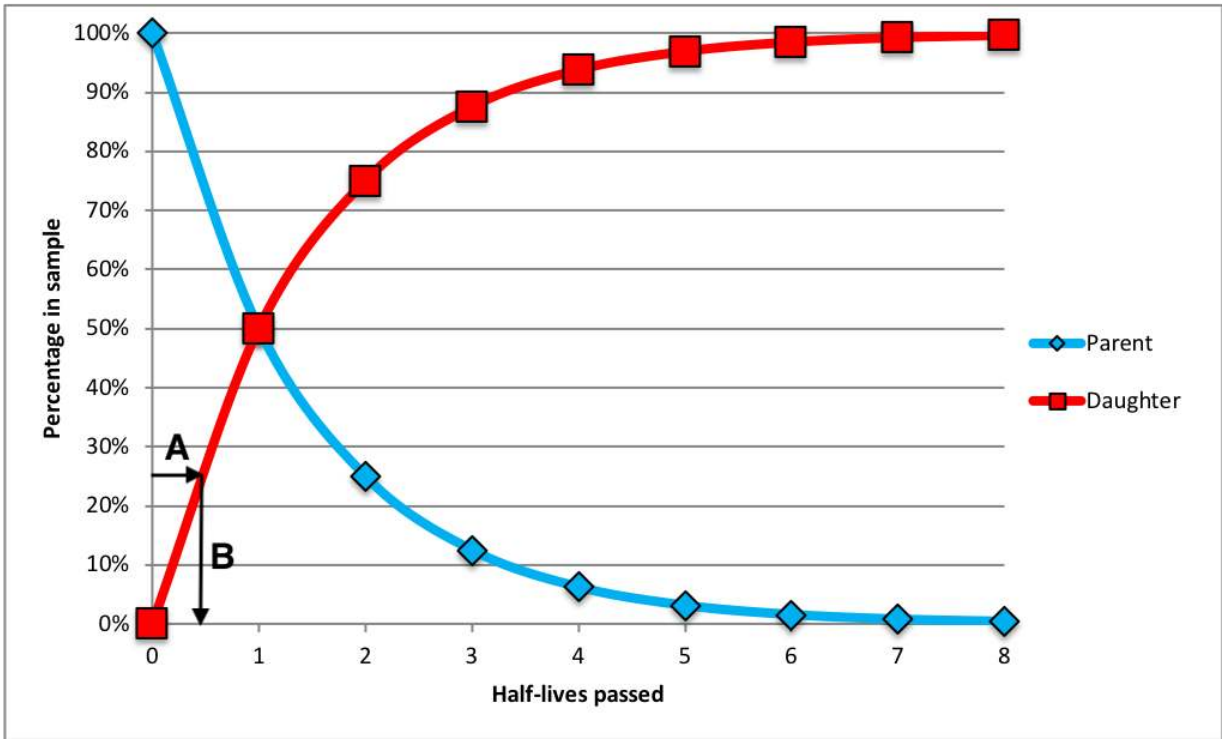


Figure 7.3.69: Radioactive Decay Curve Half Life Plot – example problem.

We can use either the percentage of the parent or the daughter; we will end up with the same answer. Choosing the daughter percentage of 25%, we find that along the left, vertical axis and then determine the intersection with the decay curve line for the daughter product (A). We drop down vertically from that point to the horizontal axis of half lives (B). The intersection of line B with the horizontal axis indicates that approximately 0.4 of one half life has passed since this sample originally formed. To determine the age of the sample in years, we simply multiply the half life decay constant for $^{40}\mathrm{K} \rightarrow {}^{40}\mathrm{Ar}$ (1.25 By) by the number of half-lives passed (0.4):

$$1.25 \text{ By X } 0.4 = 500 \text{ My}$$

To summarize, to determine the age of a sample you must:

1. Determine the percent of Parent/Daughter isotopes that exist within the sample.

2. Determine the number of half-lives that have passed by using the percent parent (or daughter) isotope and the **Radioactive Decay Curve diagram**.

3. Determine the age of the sample by multiplying the number of half-lives by the decay rate constant for the parent/daughter isotope pair found on the **Half-life Decay Constant chart**.

A WORD ABOUT CARBON-14 (^{14}C) DATING

Almost everyone has heard of Carbon-14 dating. ^{14}C does not appear on our Half-life Decay Constant chart above because it is not a useful isotope for dating rocks. ^{14}C has a very short half-life, geologically speaking. Its half-life is 5,730 years; compare that with the half-life data on the chart. After 5-6 half-lives (approximately 50,000 years), the amount of ^{14}C remaining is a sample becomes negligible and not useful for mineral/rock dating purposes. Carbon commonly exists in rock in the form of organic carbon and carbonate fossils. However the formation of lithified rock is in the order of millions of years therefore, any ^{14}C that originally existed in the sample has long since decayed to virtually nothing. As a result, ^{14}C's main use is determining archeological dates of wood, charcoal, seeds, pollen, pottery, paper, natural fabric and more recent bone and shell material.

MINERAL AGES VS. ROCK AGES

How does the mineral's numerical date relate to the rock's numerical date? What considerations must the historical geologist bear in mind as they make the transfer from the one to the other? It depends on the rock, and it depends on the mineral.

The simplest case is with igneous rocks, wherein all the minerals form at the time molten rock (magma or lava) crystallizes into solid rock. Bowen's crystallization sequence aside, this is more or less "instantaneous" from the perspective of most Historical Geology questions. So in igneous rocks, all the minerals are the same age as the rock. A zircon crystal (igneous mineral) date from a granite (igneous rock), therefore, may be taken as representative of the age of the granite. Exceptions would include xenoliths (pieces of older rock incorporated into igneous rocks) and xenocrysts (crystals from older rocks incorporated into igneous rocks).

In a clastic sedimentary rock, however, the situation is more complicated. Rocks like sandstone are made of sedimentary grains: small pieces of other rocks. Those other rocks must be older, in order to slough off pieces of themselves; pieces that can later be deposited with other small rock chunks and make a deposit of sand, sand that could then be lithified to make a sandstone. The minerals that make up a clastic sedimentary rock don't form in place: they originated elsewhere, some time before. So a zircon crystal in a sedimentary rock **does not** represent a maximum age for the sedimentary deposit: its age is the age of the igneous rock in which it formed, so the sedimentary deposit which contains a bit of rock weathered out of that original igneous rock must be younger than it. If the zircon is 2 billion years old, then the sedimentary deposit containing it must be younger than 2 billion years old. But it could be 1 billion years old, or 100 million, or 100 years old. It could be 20 minutes old, and still contain a 2-billion-year-old zircon.



Figure 7.3.70: Quartz-pebble metaconglomerate (Jack Hills Quartzite), the rock type that contains the Earth's oldest dated mineral grains (detrital zircon). (CC BY 2.0; James St. John – Quartz-pebble metaconglomerate (Jack Hills Quartzite, Archean, 2.65 to 3.05 Ga; Jack Hills, Western Australia))

These are called "detrital zircons" to distinguish our understanding of how they got into the sedimentary rock – they are survivors from older weathered-away igneous source rocks. They are part of the leftover scraps – the detritus, or waste – from the breakdown of older rocks. Detrital zircon have been very useful in tracking down the Earth's oldest rocks. At the beginning of this chapter, you learned that the Earth is 4.56 billion years old. As it turns out, the oldest dated mineral–a grain of zircon from the Jack Hills of Western Australia–is 4.404 billion years old. This zircon exists in metaconglomerate, a metamorphosed clastic sedimentary rock. Discovery of the age of this zircon tells us that it was derived from even older igneous rock which gives us a tremendous clue as to the composition of Earth's early, Hadean age, crust.

In a chemical sedimentary rock, where crystals are forming at the time of deposition, it would be plausible that a mineral date could be used to date the depositional age, but the problem is that none of the crystals in chemical sedimentary rocks are "picky" enough

about their isotopic inclusions for the technique to work. Calcite doesn't grab radioactive isotopes while excluding their daughter isotopes, and neither does halite or gypsum. Tough luck: we just can't date chemical sedimentary rocks using such a process. Next!

With metamorphic rocks, there's another layer of complication still: the mineral age can be representative of the rock's protolith, or the date of its metamorphism. It all depends on what mineral we're talking about, and what the protolith was. First, let's consider a metavolcanic rock – a metamorphic rock that started off as a volcanic (igneous) rock before it got metamorphosed. In this case, a zircon represents the age of the protolith (the age of the initial eruption), while the elevated temperature and pressure of metamorphism might encourage the growth of new metamorphic minerals. When a basalt is metamorphosed to make an amphibolite for instance, it grows new amphibole crystals. These amphiboles could be dated (using the K/Ar technique) in order to determine the age of metamorphism. Two minerals in one rock, each representative of a different process, happening at a distinct time. The careful geologist should pay attention to which mineral they are dating, and what its implications are for the history of that rock.

Next, consider a metasedimentary rock – a metamorphic rock with a sedimentary protolith. A metagraywacke is a metamorphosed graywacke: its protolith was clay-rich sandstone. In this case, a detrital zircon might well be a maximum age constraint of the original sedimentary deposit, as we just discussed a couple of paragraphs ago. Zircon has a very high closure temperature (link to resetting discussion below), so if the metamorphism was low- to medium-grade, the zircon's original igneous date could survive heating. But the same level of metamorphism in the same metagraywacke could easily be sufficient that the clay all recrystallizes to make muscovite mica, then we can use the mica to get the age of metamorphism. The zircon would have been there before metamorphism, but the mica grew as a consequence of metamorphism. Muscovite can be dated using K/Ar dating, or its derivative technique, Ar/Ar dating (link).

An example of this can be found in the Piedmont metamorphic rocks of Washington, DC. There, detrital zircons in metagraywackes of the Sykesville Formation give U/Pb ages of ~1 Ga, but the same rocks also host metamorphic muscovites that grew at 460 Ma, according to the K/Ar dating technique. The conclusion is that the muddy sand was deposited sometime after 1 Ga, but before 460 Ma, since it was metamorphosed then. The two dating techniques, on two different minerals, help "bracket" the age of the original deposit of sediment.

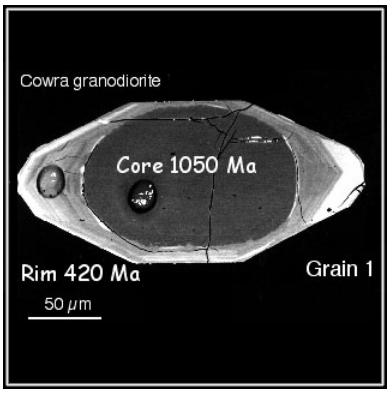


Figure 7.3.71: SHRIMP allowed in situ dating of discrete zones in zircon. This revealed for the first time the true complexities of the geochronology of this mineral as well as its ability to record individual geological events. (With permission from the Research School of Earth Sciences, Australian National University College of Science, http://shrimp.anu.edu.au/shrimp/history.php)

One additional wrinkle with metamorphic rocks is that sometimes you can get several dates out of a single mineral crystal. This happens because an original crystal can add new layers when it gets metamorphosed. Again, we turn to zircon for an example: if a granite pluton was to be metamorphosed and make a gneiss, then original (igneous) zircons in the rock might have the opportunity to acquire new zirconium, silicon, and oxygen atoms through metamorphic recrystallization. These new atoms get arranged in the zircon's distinctive crystal structure, accumulating on the outside of the old crystal. If you were to date this whole zircon crystal, you're going to get a hybridized age – a mix of the original plutonic age and the age of the metamorphic overgrowth. That would be a disaster!

Fortunately, there is a solution: SHRIMP to the rescue! SHRIMP is an acronym that stands for Sensitive High-Resolution Ion MicroProbe. It's a tool that allows the dating of different parts of a single zircon crystal. Although the crystal being examined may be the size of a grain of sand, the SHRIMP uses a thin laser beam to distinguish between the layers on the edge (youngest) versus

the layers in the middle (oldest). Incredibly, this allows us to extract multiple dates from a single mineral grain: one date for the igneous crystallization, and another date for a later metamorphic event. There are even examples where 10 or more dates have been extracted from a single grain of zircon!

Metamorphism can complicate accurately dating the age of a particular mineral. Increased pressure and temperature during metamorphism plus the addition/liberation of fluids and gases can result in ions becoming mobile in the metamorphic environment. We can return to a previous discussion on calculating radiometric dates to use as an example. Potassium feldspar is one of the most common minerals in the crust; common in both igneous and metamorphic rocks. Potassium feldspar typically begins its existence as a mineral formed during the crystallization of magma. It is a common constituent in the igneous rock, granite. The radioactive isotope, $^{40}\mathrm{K}$, can be included in the crystalline structure of potassium feldspar. Upon crystallization, the radioactive decay of $^{40}\mathrm{K}$ begins producing $^{40}\mathrm{Ar}$. Granite can be very accurately dated using this decay pair ($^{40}\mathrm{K}/^{40}\mathrm{Ar}$). However, most granite is exposed to metamorphic conditions because it typically forms in a cooling magma chamber, deep within the crust, in a tectonic environment that includes the compressive forces of plate collision. Granite, subjected to metamorphic conditions, does not change much in its mineralogy however the increased pressure and temperature conditions will realign minerals and allow trapped fluids and gasses to escape. The resulting metamorphic rock is a granitic gneiss. The realigned potassium feldspar minerals will remain in the granitic gneiss (see photo below) and still contain the decaying $^{40}\mathrm{K}$. However, the daughter product, $^{40}\mathrm{Ar}$, formed in the original granite, may have vacated the closed system during the metamorphic process thereby resetting the radiometric decay clock. Dating the potassium feldspar in the granitic gneiss would yield the date of metamorphism, not the original date potassium feldspar crystallization in the granite.

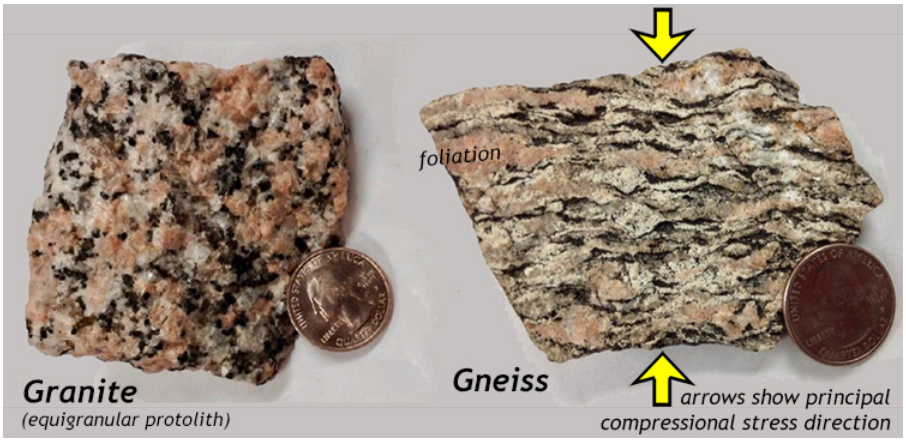


Figure 7.3.72: A comparison of granite and gneiss. The texture of the protolith granite is the same in every direction (equigranular) but the gneiss has a pronounced foliation, which we infer to have developed from squeezing (compressional stress) in a direction that was perpendicular to the resulting foliation.

Applications of all dating methods

Let's return to our Relative Dating cartoon to see how we can apply all relative and numerical dating methods to further our understanding of events in geological history. To summarize from our relative dating analysis above, here is what we know:

Oldest F, E, D, C, B, J*, (H or A), G, I **Youngest**. From the data given we could not tell the exact order in which fault A or igneous intrusion H occurred. We also assume* that F through J were in place prior to the intrusion of H however we do not know that since H does not cross-cut J.

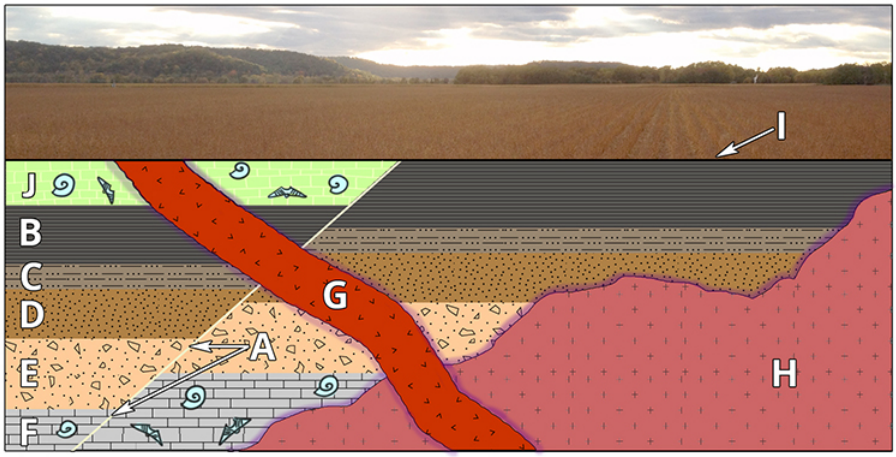


Figure 7.3.73: Relative dating relationships. Original image by Jonathan R. Hendricks. (This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License. Modified by Callan Bentley.)

We observe that there are fossils in both limestone units F and J. Using the principle of faunal succession and the paleontological tool of the fossil record we determine that a specific type of brachiopod fossil in layer F is from the genus *Mucrospirifer*. This brachiopod is an **index fossil** for the Devonian period during the Paleozoic Era of geologic time. Layer J reveals a cephalopod fossil from the genus *Domatoceras*. The geologic range for this fossil is the Pennsylvanian-Permian periods of the Paleozoic Era. We can add the following to our sequence of events:

Oldest F (Devonian), E, D, C, B, J* (Pennsylvanian-Permian), (H or A), G, I **Youngest**

The igneous intrusions G and H can provide us with numerical dates using isotopic analysis and our learned radiometric dating techniques. We will say that intrusion H is granite and intrusion G is basalt. The granite was analyzed using zircon crystals and the $^{238}U/\ ^{206}Pb$ isotopes which provided an age of 330 Ma +/- 5 Ma. The basalt was analyzed using $^{40}K/^{40}Ar$ of hornblende crystals yielding an age of 220 Ma +/- 8 Ma. Using the geological time scale below, we discover that the igneous intrusion H occurred during the Mississippian Period and G occurred during the Triassic Period. That provides us with some much needed timing from which we can now determine the sequence of all events.

Oldest F (Devonian), E, D, C, B, H (Mississippian), J (Pennsylvanian-Permian), G (Triassic), I **Youngest**

From this information can we now determine when Fault A occurred? Yes, we can!

Since Fault A cross-cuts J, we know that J is older than A. However, the intrusive dike G cross-cuts everything so we know that A is older than G. I is the erosional surface so it is the absolute youngest. So now, using all of our dating methods we can emphatically put all of the events in order:

Oldest F, E, D, C, B, H, J, A, G, I **Youngest**

Unraveling geologic events is a puzzle. Geologists find new pieces to the puzzle of Earth every day and continue to add pages to the book of our natural history. You can challenge yourself on a similar example in this case study: Extremadura.

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7.4: The Geological Time Scale

The geological time scale is one of the crowning achievements of science, and geology in particular. It is a reference and communication system for comparing rocks and fossils from throughout the world and is geology's equivalent of the periodic table of the elements.

The geological time scale is currently maintained by the International Commission on Stratigraphy (ICS), which is part of the International Union of Geological Sciences. The time scale is continuously being updated as geoscientists learn more about the timing and nature of past geological events. You can view the ICS time scale online.

The geological time scale provides a global summary of countless small-scale temporal correlations of rock layers made at local and regional scales. It is based almost entirely upon careful observations of the distributions of fossils in time and space. The dates were added much later following the advent of numerical dating techniques giving us the ability to bracket the ages of different organisms in relation to numerical dates provided by associated igneous activity.

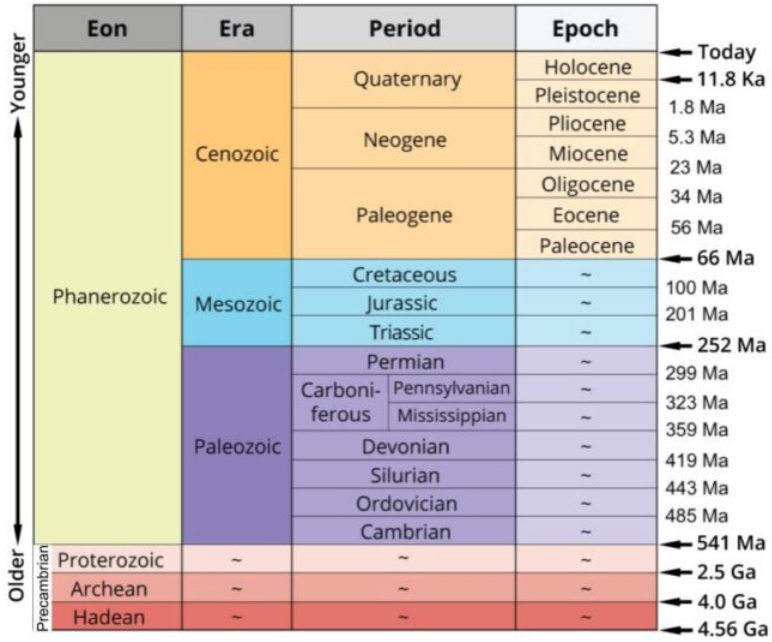


Figure 7.4.1: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

Time scale history

William Smith used the principle of faunal succession and correlation to great effect in his monumental solo project to create a geological map of England and Wales, published in 1815. Inset into Smith's great geological map is a small diagram showing a schematic geological cross-section extending from the Thames estuary of eastern England all the way to the west coast of Wales. Smith shows the sequence of rocks, from the oldest (Paleozoic age) rocks of Wales and western England, through the middle aged (Mesozoic) rocks of central England, to the youngest (Cenozoic age) rocks of the area around London. Smith did not put any numerical dates on these because he didn't know them. But using the principle of superposition and faunal succession he knew that, when tipped upright, this diagram represented a **stratigraphic column**, a sequence of rock in order from oldest, on the bottom, to youngest, at the top. And because almost every period of the Phanerozoic is represented along that section through Wales and England, it is a primitive geological time scale.

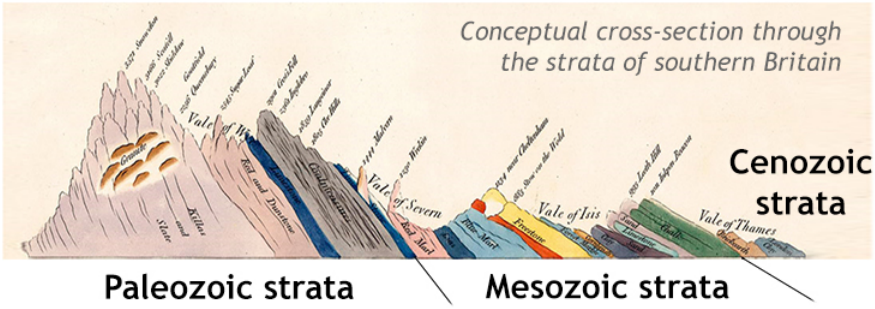


Figure 7.4.2: William Smith's "Sketch of the succession of strata and their relative altitudes," an inset on his geological map of England and Wales (with era names added). (Modified from a version by Steven Earle here: https://opentextbc.ca/physicalgeology2ed/chapter/8-1-the-geological-time-scale/)

The simple principles of superposition and faunal succession have provided geologists with the ability to reconstruct the geological history of Earth by correlating rocks and fossils from different places. This is critically important because no single place on Earth preserves a complete geological history, or even a small fraction of it. We only get a sense of the full story if we look everywhere, using multiple locations to piece together the puzzle of snapshots of the Earth through time.

The geological time scale provides a global summary of countless small-scale correlations of rock layers made at local and regional scales. It is based almost entirely upon careful observations of the distributions of fossils across time and space.

Another important principle of relative dating, the **principle of faunal succession**, which is the concept that fossils can occur in predictable sequences because of the time intervals in which they exist. Fossils can act as "timestamps" that reveal the age of the strata in which they are found.

Early on in this chapter we discussed the work of William Smith (1769 – 1839) as one of the founding fathers in our early understanding of geologic time in the 1700s and early 1800s. Smith's work in understanding early relative dating principles of superposition and faunal succession set the stage for the definition and naming of the periods of geologic time. British geologists initiated this task around 1820. Later, geologists from other European countries joined them. Many of the periods are named for places where rock of that age are found in Europe. A full list of meanings for all of the geologic time periods is found below.

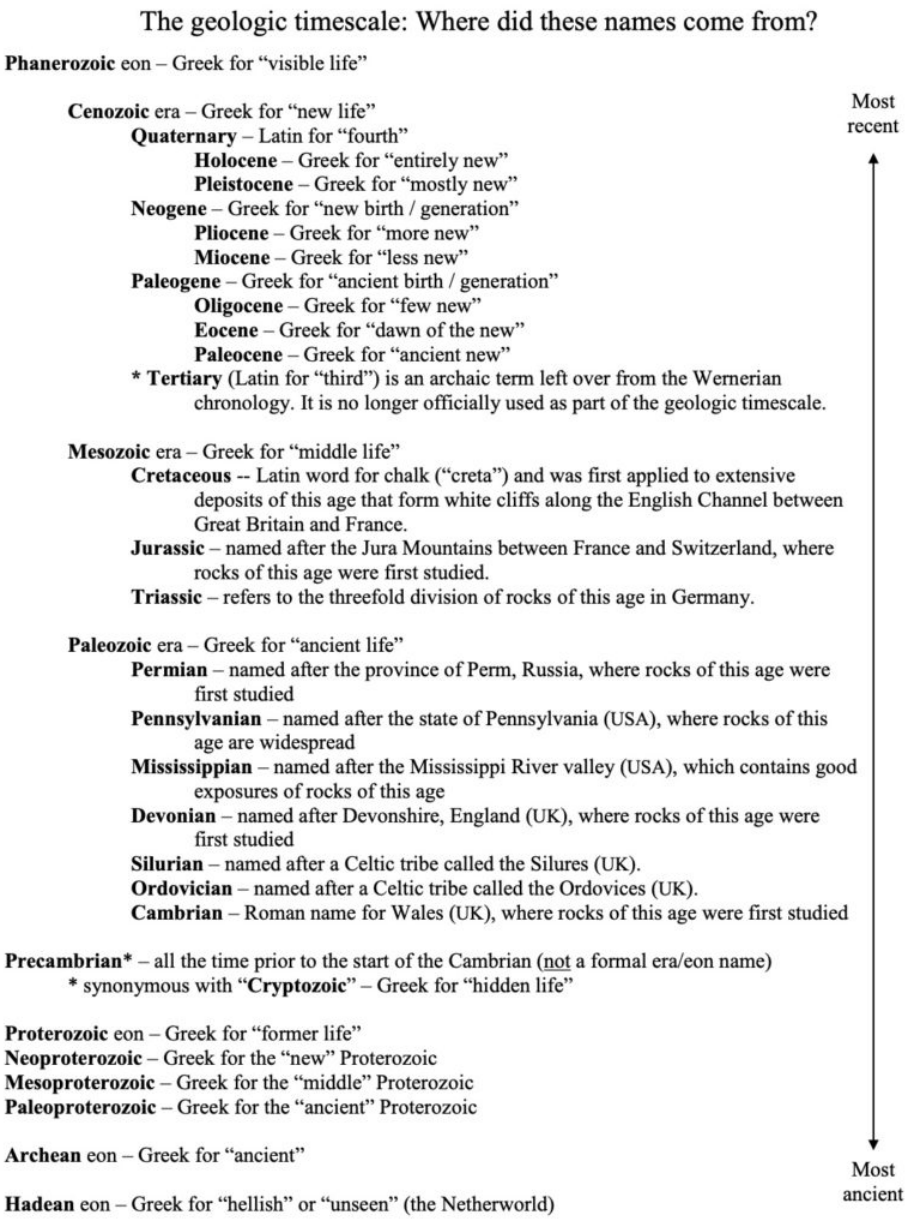


Figure 7.4.3: The history behind the naming of the various geological time periods. (Callan Bentley)

Learning the geological time scale

Because of its usefulness for communicating about events in Earth's history, it is important that all students of geology, paleontology, and evolutionary biology commit the geological time scale to memory. This is most easily done by first breaking the time scale into its component parts: eons, eras, periods, and epochs.

Eons

The eon is the broadest category of geological time. Earth's history is characterized by four eons; in order from oldest to youngest, these are the Hadean, Archean, Proterozoic, and Phanerozoic. Collectively, the Hadean, Archean, and Proterozoic are sometimes informally referred to as the "Precambrian." (The Cambrian period defines the beginning of the Phanerozoic eon; so, all rocks older than the Cambrian are Precambrian in age.)

We live during the Phanerozoic, which means "visible life." This is the interval of geological time characterized by abundant, complex fossilized remains. Being the youngest eon of time, it is also very well represented by rock at Earth's surface (because of

the Principle of Superposition). Because of these two factors, most paleontologists and geologists study fossils and rocks from the Phanerozoic eon.

Do not let the time scale at the top of this page give you a false impression, however, about the temporal duration of the Phanerozoic eon relative to the three older Precambrian eons. Note in the figure below the absolute ages of the boundaries separating each eon of time.

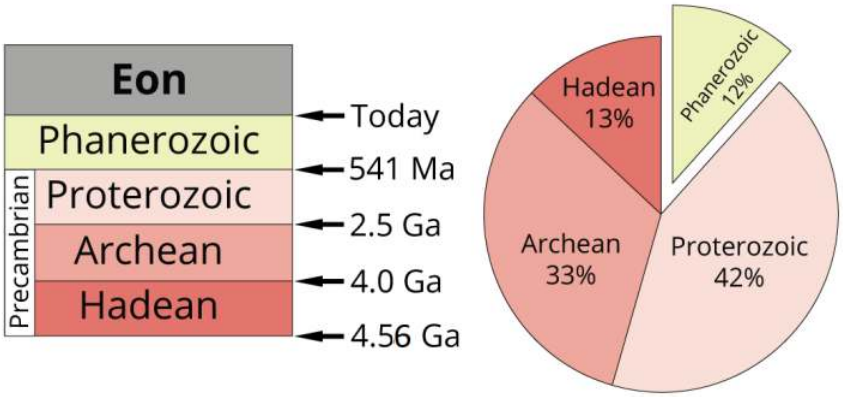


Figure 7.4.4: Left: the four eons of geological time. *Right:* the "Precambrian" eons (Hadean, Archean, and Proterozoic) represent 88% of geological time. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License.; Image by Jonathan R. Hendricks, https://www.digitalatlasofancientlife.org/learn/geological-time/geological-time-scale/.)

You have already learned that the Earth is 4.56 billion years old. The Phanerozoic eon began 541 million years ago (or, 0.541 billion years ago). Thus, the Phanerozoic eon represents a paltry 12% of Earth's history! Instead, most of Earth's history is represented by the three Precambrian eons. These older eons tell the story of Earth's beginning, life's origin, and the rise of complex life.

The Hadean and Archean are difficult eons to study, however, because they are exposed in very limited places on Earth's surface. (Since they are the oldest eons, rocks that are Hadean and Archean in age are often buried far below younger rocks at Earth's surface.) Proterozoic rocks–which span nearly 2 billion years (42% of Earth's history)–are much more accessible, but, until recently, have received significantly less attention from paleontologists than rocks from the younger, fossil-rich Phanerozoic eon. That is slowly beginning to change, however, as more clues about the origins of complex life begin to be revealed from Proterozoic-aged rocks.

Eras

Eons of geological time are subdivided into eras, which are the second-longest units of geological time. The Phanerozoic eon is divided into three eras: the Paleozoic, Mesozoic, and Cenozoic.

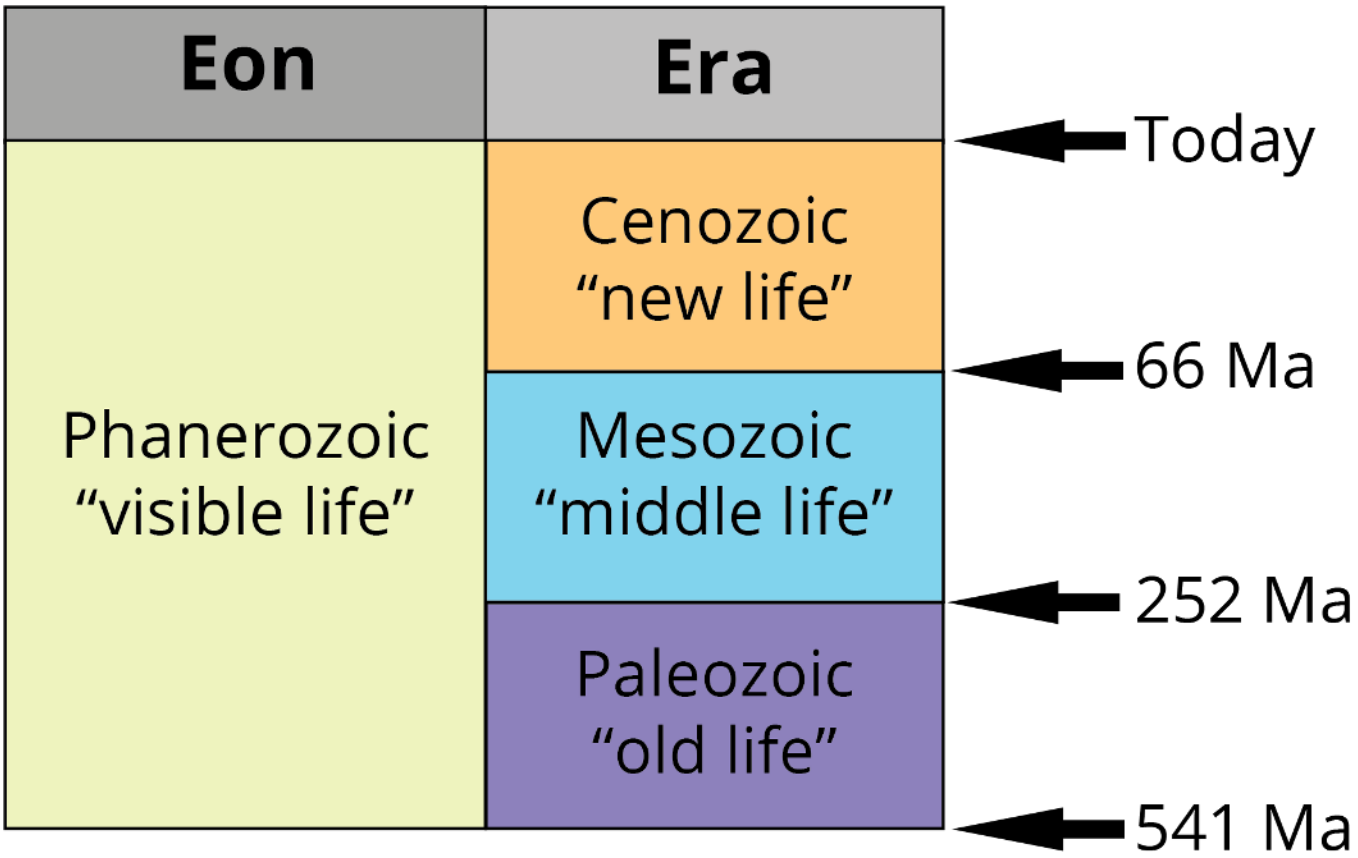


Figure 7.4.5: The three eras of the Phanerozoic eon. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

Most of our knowledge of the fossil record comes from the three eras of the Phanerozoic eon. The Paleozoic (“old life”) era is characterized by trilobites, the first four-limbed vertebrates, and the origin of land plants. The Mesozoic (“middle life”) era represents the “age of dinosaurs,” though also is noteworthy for the first appearances of mammals and flowering plants. Finally, the Cenozoic (“new life”) era is sometimes called the “age of mammals” and is the era during which we live today.

As temporal points of reference, it is worth memorizing the ages of the boundaries that separate the three eras of the Phanerozoic eon. Long before geologists knew these absolute age dates, they realized that the boundaries represent important events in the history of life: mass extinctions. For example, many fossils that are commonly found in the youngest Paleozoic rocks are not found in overlying Mesozoic rocks. Similarly, dinosaur fossils found in the youngest Mesozoic rocks are never again found in the overlying Cenozoic rocks. Paleontologists and geologists used these mass extinction events to define these (and other) boundaries within the Phanerozoic portion of the geological time scale. It is therefore no coincidence that some of the major boundaries coincide with mass extinction events!

The older Archean and Proterozoic eons are similarly divided into several eras. For example, the youngest era of the Proterozoic eon is called the Neoproterozoic. For the sake of simplicity, these older eras are not included on the time scale shown at the top of this page; they do, however, exist!

Periods

Just as eons are subdivided into eras, eras are subdivided into units of time called periods. The most well known of all geological periods is the Jurassic period of the Mesozoic era (the movie Jurassic Park, of course, has something to do with that).

The Paleozoic era is divided into six periods. From oldest to youngest, these are the Cambrian, Ordovician, Silurian, Devonian, Carboniferous, and Permian. Note that in the United States, the Carboniferous is divided into two separate periods: the Mississippian and the Pennsylvanian.

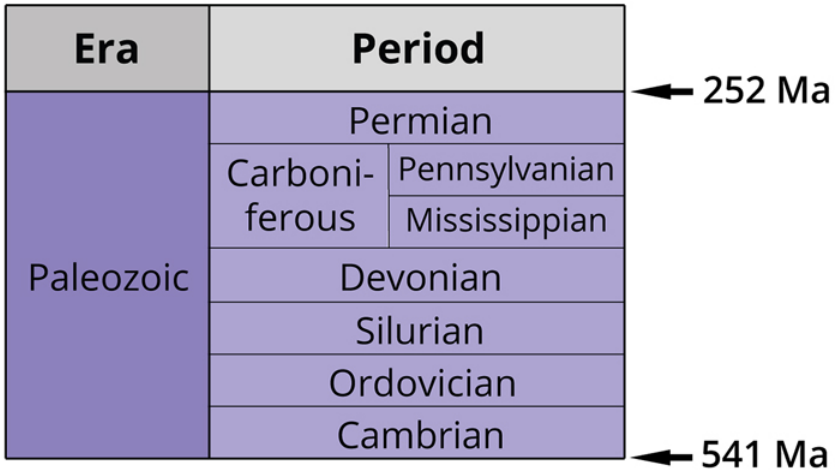


Figure 7.4.6: The seven periods of the Paleozoic era. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

The Mesozoic era is divided into the Triassic, Jurassic, and Cretaceous periods.

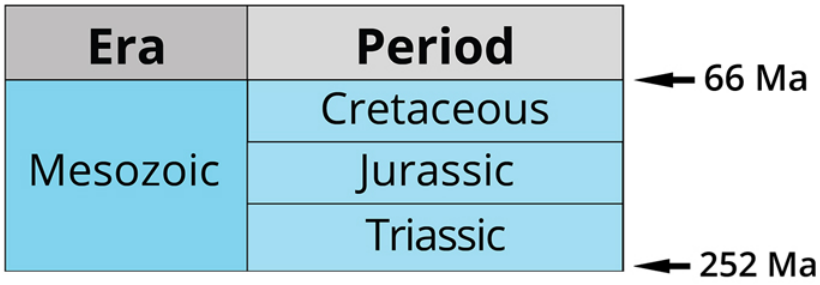


Figure 7.4.7: The three periods of the Mesozoic era. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

Finally, the Cenozoic era is divided into three periods: the Paleogene, Neogene, and Quaternary.

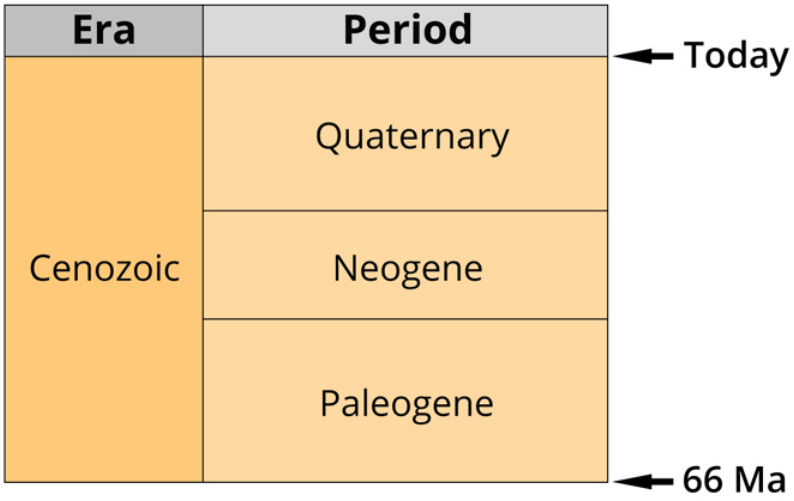


Figure 7.4.8: The three periods of the Cenozoic era. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

Epochs and Ages

Periods of geological time are subdivided into epochs. In turn, epochs are divided into even narrower units of time called ages. For the sake of simplicity, only the epochs of the Paleogene, Neogene, and Quaternary periods are shown on the time scale at the top of this page. It is important to note, however, that all of the periods of the Phanerozoic era are subdivided into the epochs and ages.

The Paleogene period is divided into–from oldest to youngest–the Paleocene, Eocene, and Oligocene epochs. The Neogene is divided into the Miocene and Pliocene epochs. Finally, the Quaternary is divided into the Pleistocene and Holocene epochs. Some geologists now think that–since humans are having such a notable impact on the Earth and its life–a new, youngest epoch should be added to the Quaternary: the Anthropocene. There is still considerable discussion in the geological community about whether this epoch should be added, as well as debate about what characteristics should define its beginning.

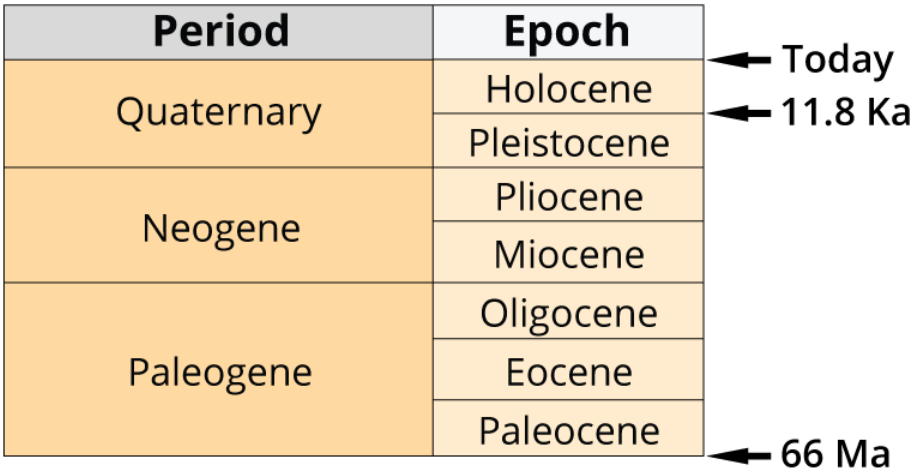


Figure 7.4.9: Epochs of the Paleogene, Neogene, and Quaternary periods. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Image by Jonathan R. Hendricks.)

Interacting with The Geological Time Scale

Spend some time getting to know the major events in the history of the Earth through this interactive Earth Viewer. Select “Info” and “Tutorial” to understand the full range of capabilities of this interactive to look at Earth throughout geologic time. The interactive opens on the Phanerozoic Eon; you can select “Ancient Earth” to view the globe and events of the Proterozoic Eon. The geologic time scale appears at the left with a silver slider that you can grab and peruse through time past. Select “View” to see overlays available for the globe and the time scale. The layer selected in the static image below is “Mass Extinctions.” Note how these correspond with the end of Eras and Periods during the Phanerozoic Eon (small yellow triangles). This view is focused on North America at the end of the Mesozoic Era which culminates with the mass extinction event of the dinosaurs.

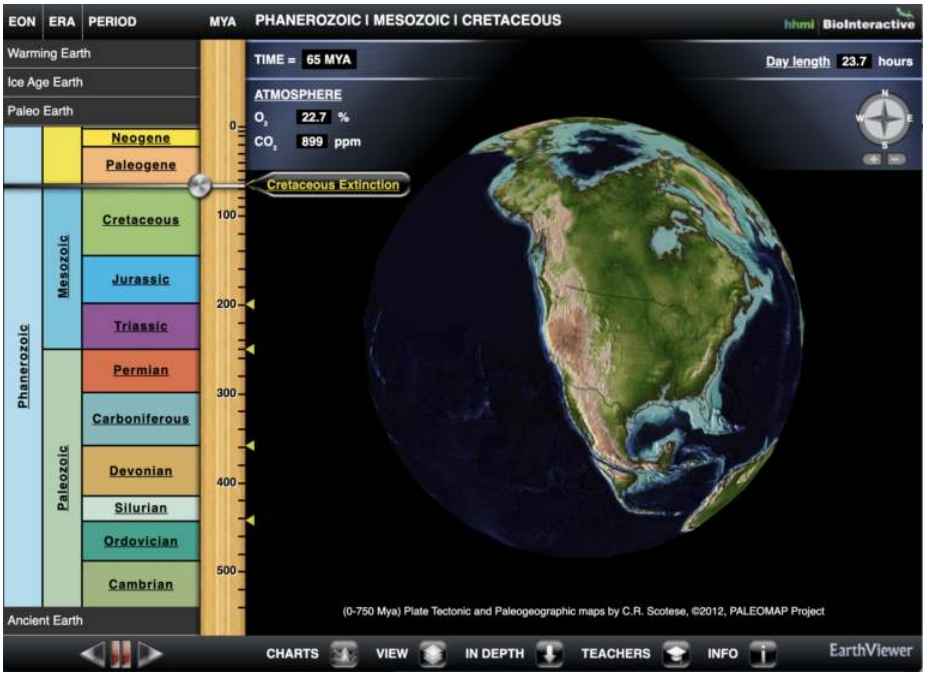


Figure 7.4.10: Static image of the Earth Viewer from Howard Hughes Medical Institute Biointeractive. (From: https://www.biointeractive.org/classroom-resources/earthviewer. Use for educational purposes.)

Now that you have learned about the hierarchical components of the geological time scale–eons, eras, periods, and epochs–consider again how all of these parts fit together. Note that some boundaries (those that follow horizontal lines on the time scale) are equivalent in age. For example, the Mesozoic-Cenozoic boundary is equivalent to the Cretaceous-Paleogene boundary (both have an age of 66 Ma). Similarly, the Paleogene-Neogene boundary is equivalent to the Oligocene-Miocene boundary.

It is much easier to memorize the time scale by first breaking it down into its component parts. Many geology students have developed clever mnemonic devices to help remember all of the names in the time scale and a quick Google search will provide many examples (which are often rather salty!) that you may find helpful as you memorize the time scale.

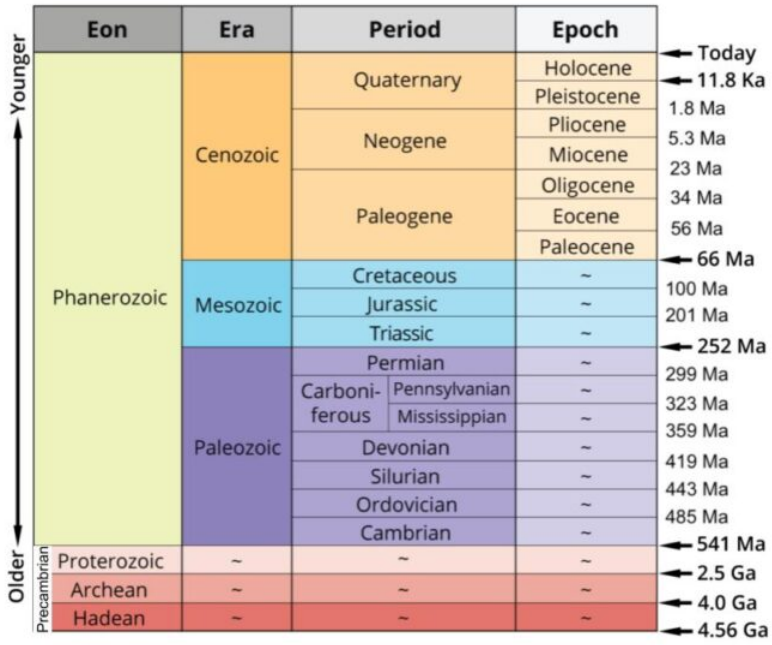


Figure 7.4.11: The geological time scale. (Creative Commons License. This work is licensed under a Creative Commons Attribution-ShareAlike 4.0 International License; Edited. Original image by Jonathan R. Hendricks.)

The geological time scale is a result of hundreds of years of investigation and remains very much a work in progress. Observe the geological time scale below, which is from a college-level geology textbook published in 1885 by geologist and conservationist Joseph LeConte (1823-1901) from the University of California, Berkeley. Many similarities exist with that presented above. Differences primarily exist with the oldest Eons. Our understanding of the occurrence, ordering and evolution of the species through time was remarkable even though the study of geology and paleontology was little more than a century old.

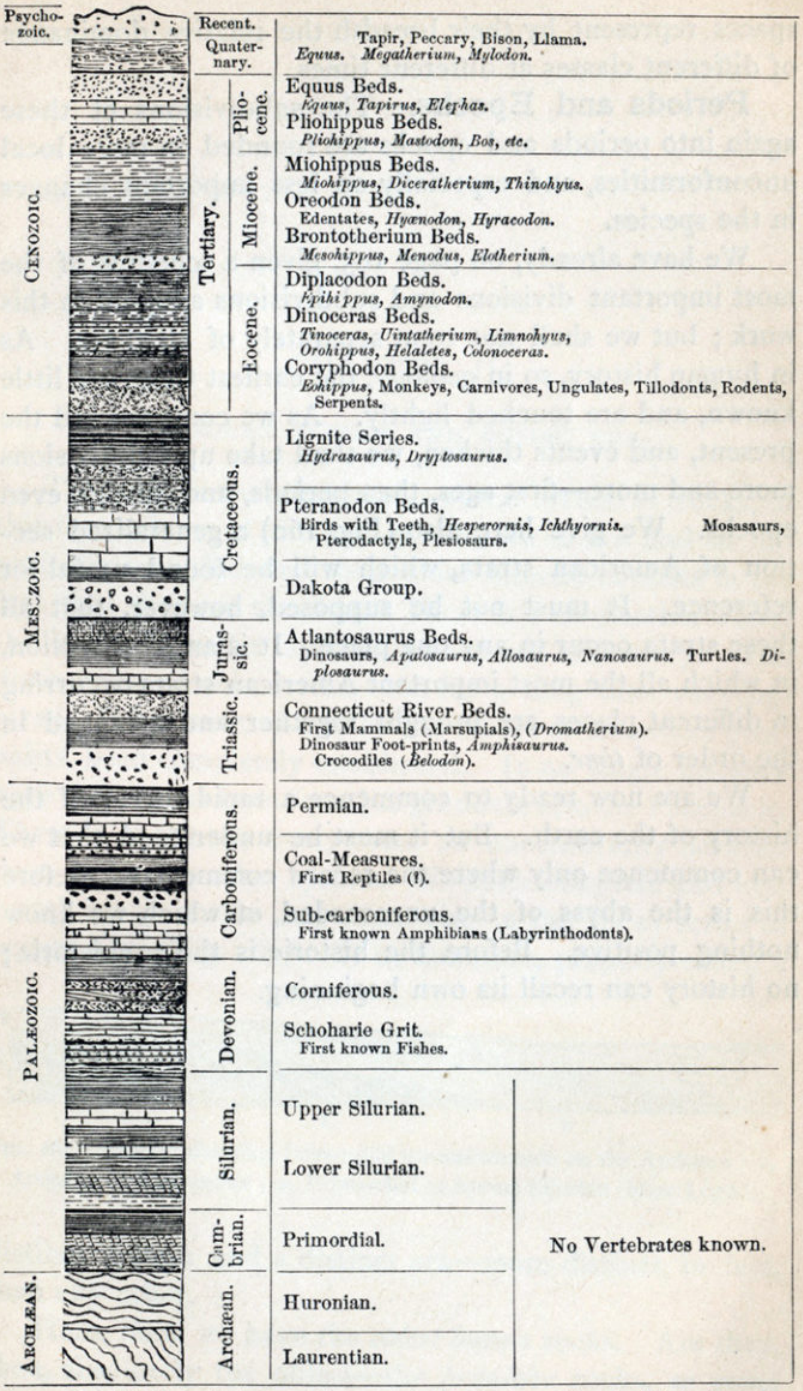


FIG. 160.—Section of the earth's crust, to illustrate vertebrate life in America. (Slightly modified from Marsh.)

Figure 7.4.12: Geological time scale from Le Conte (1885). (public domain)

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7.5: References

[1] Claude J. Allègre, Gérard Manhès, Christa Göpel, The age of the Earth, Geochimica et Cosmochimica Acta, Volume 59, Issue 8, 1995, Pages 1445-1456, ISSN 0016-7037, https://doi.org/10.1016/0016-7037(95)00054-4.

[2] Schreiner, G.D.L., Basson, H.H., Verbeek, A.A., "Potassium-argon dating of the Cape Granite and a granitized xenolith at Sea Point," *Talanta*, Volume 15, Issue 11, November 1968

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7.6: Further Reading

Bjornerud, Marcia (2018). *Timefulness: How thinking like a geologist can help save the world.* Princeton University Press. 219 pages.

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CHAPTER OVERVIEW

8: Evolution Part I - The Theory

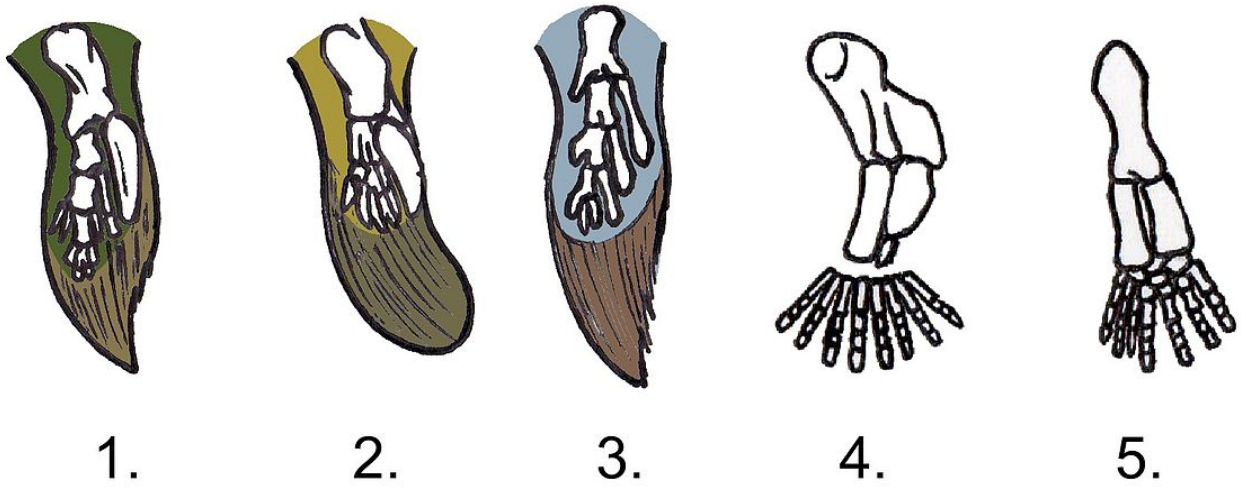


Figure 8.1: Comparison between the fins of lobe-finned fishes (Sarcopterygii) and legs of early tetrapods. 1. Tiktaalik. 2. Panderichthys. 3. Eusthenopteron. 4. Acanthostega.
5. Ichthyostega (hindleg). (Public domain image.)

Key Concepts

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

- Define biological evolution.
- Distinguish between macro and microevolution.
- Distinguish between different patterns that occur in nature as a result of evolutionary processes.
- Describe how life is organized using systematics.
- Identify some of the key ideas and players in the history of Evolutionary Theory.
- Identify the key paradigms in the development of Evolutionary Theory.
- Identify the ways that genes are passed on through various forms of inheritance to the next generation.

8.1: Evolution - Patterns in the Biosphere

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8.1: Evolution - Patterns in the Biosphere

The biosphere is replete with patterns. This is true not only in the ecosystems around us today, but also in our planet's fossil past. Some of these patterns include some of the basic things we learn about in introductory biology courses. These patterns force scientists to ask questions. Why are ecosystems arranged in a hierarchy with primary producers on the bottom and omnivorous organisms on top? Why is there so much shared DNA among life forms that seem so different, even across different environments? Why do so many life forms that live such different lives have such similar features? Why do humans and reptiles share four limbs? Why do both butterflies and birds have wings? Why do some features of organisms not have clear functions? Why do I need to put on a jacket, hat, and gloves to venture out into the winter cold?

One common pattern in the living world is governed by the mathematical Fibonacci Sequence. This simple sequence of numbers describes an amazing variety of phenomena, with names like the "golden ratio" and the "golden spiral." It can also be used to describe breeding patterns in organisms and subsequent geometric population growth. A Fibonacci Sequence describes a doubling with each calculation. Over time, it is possible to use this simple relationship to predict the number of organisms present after any number of generations, if resources are available and population growth is unchecked. Also predicted by this equation are other natural patterns, like the Fibonacci Spiral seen in the images below., The arrangement of seeds in a sunflower and the whorls of an ammonite are excellent examples of how this mathematical relationship governs some biological growth patterns. These spiral patterns are based upon a curve that, while growing, keeps its shape and central angle at all scales.

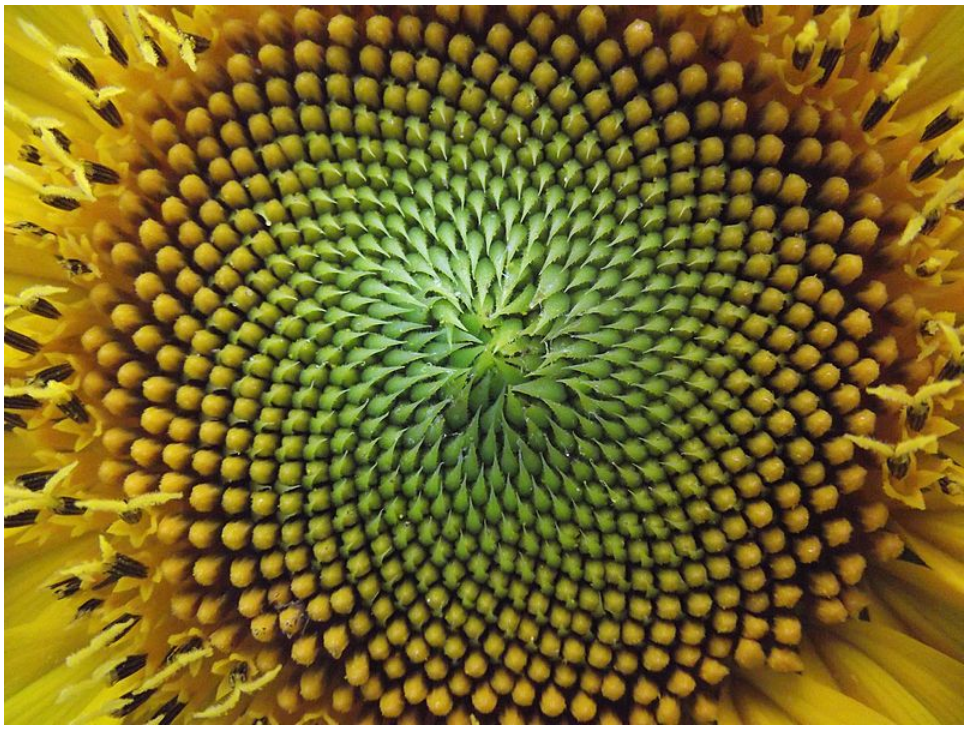


Figure 8.1.1: A Fibonacci Spiral in nature. In this case, within sunflower seeds (CCA ASA 4.0; Anna Benczur, Wikimedia Commons).

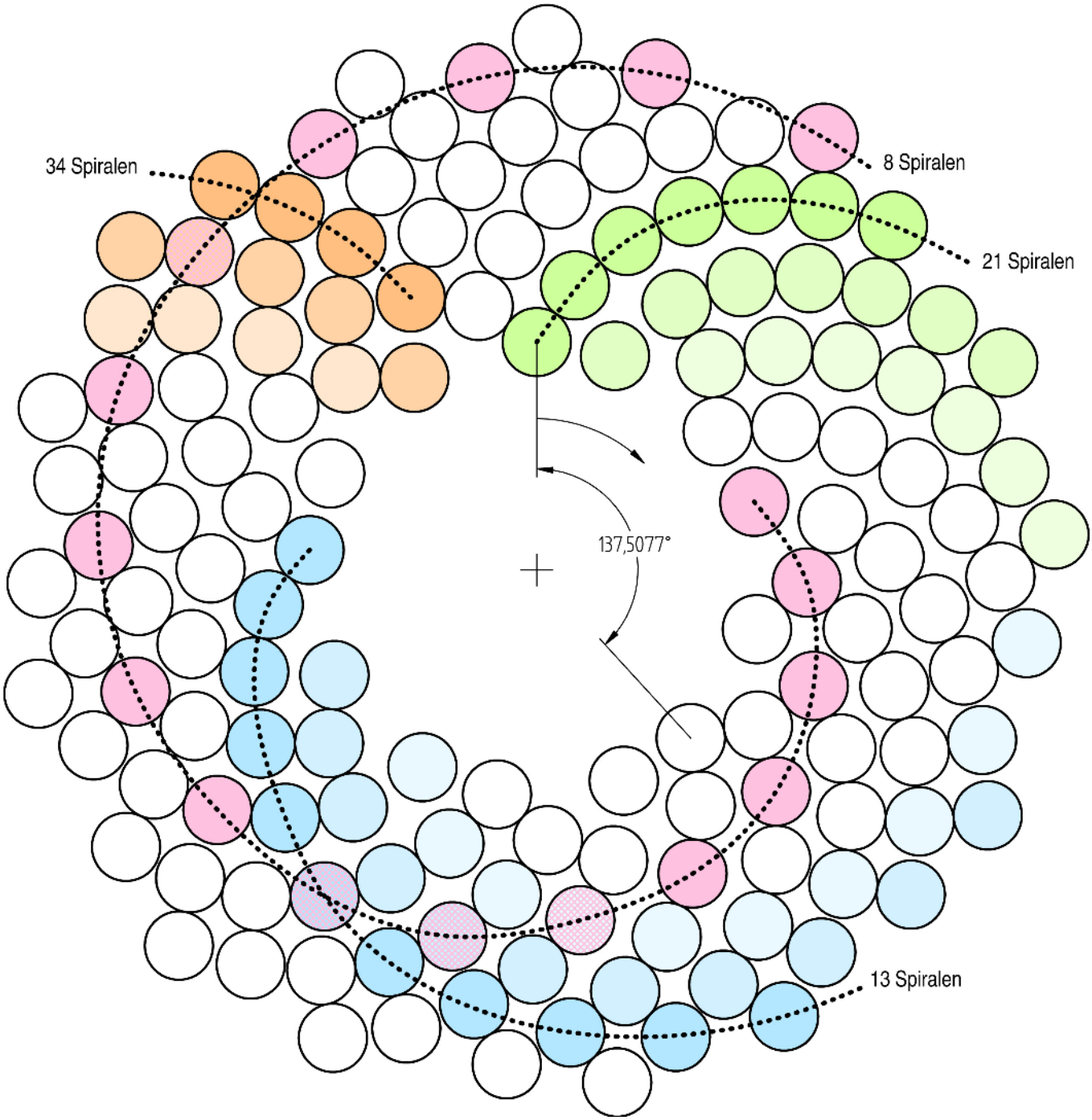


Figure 8.1.2: The Fibonacci Spiral, centered around the golden angle of 137.5 degrees, defines a sunflower's seed pattern (CC ACA 3.0; Andree Stephan, Wikimedia Commons)

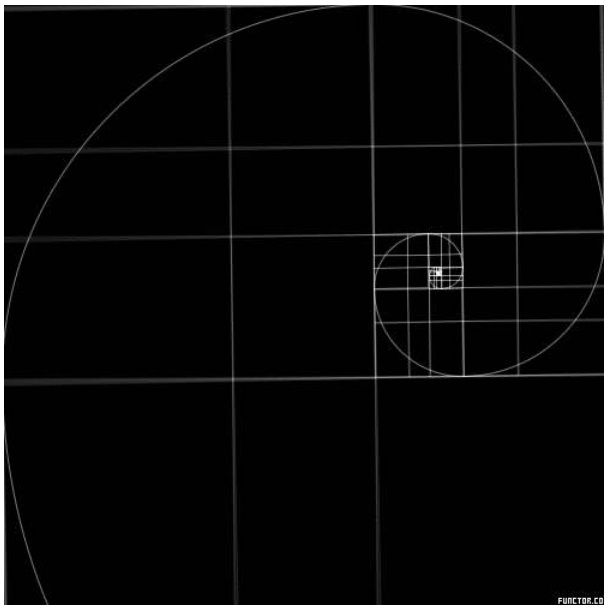


Figure 8.1.3: Animation of the Fibonacci Spiral. This pattern defines the whorls of an ammonite. It is a common pattern in nature and this is best explained through evolutionary connections among life forms.



Figure 8.1.4: Gigapan of fossils accumulated on the Cretaceous Interior Seaway seafloor at Fort Hays State University. Can you see fossils that display distinct patterns? Do you see a golden spiral emerge? (Image: Ron Schott)

Evolutionary theory presents us with answers to these questions. Using it as a framework, we can answer these questions and more. Evolution explains the observable patterns in the living world around us. It also explains the patterns we see in the fossil past.

Evolution means “change over time.” It is used in many ways and in many fields. Biological evolution is the mutation of genetic material through generations of offspring. Such evolution from one generation to another does not follow a trajectory. It is the result of the accumulation of changes in DNA. Such modifications occur at different rates in different groups of organisms and at different moments in time, but they are only observable when mutations in an organism’s DNA, its genotype, are physically

expressed through its phenotype, or appearance. Such changes in genes can lead to longer horns, more toes, improved leg bone structure for walking on dry land, and bipedalism in primates. These are just a few examples. Billions of years ago, when microbes were the only living things on the planet, such exotic future modifications were not inevitable or predictable. Our own species, Homo sapiens was never guaranteed existence. Still, we are here, and we are able to contemplate questions of origins. You are reading this chapter, and that is a result of evolutionary change! Ultimately, evolution tells us that we are all related. Yes, humans are related to one another, but humans are also related to sponges, flowering plants, and bacteria. All life on this planet is descended from a common ancestor, and was then teased out into millions of distinct evolutionary trajectories. Some of these probably look and act pretty similar to the most ancient ancestral forms of life (their features are said to be evolutionarily "conserved"), and some are extremely different – they are "highly derived." Descent with modification is the essential idea of evolution.

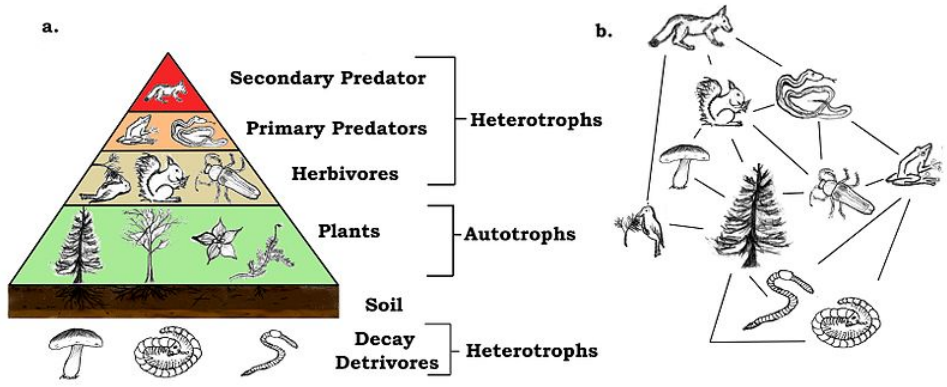


Figure 8.1.5: The trophic pyramid is based upon observations of food web, or energy transfer, as observed in living ecosystems. As a model, it explains the patterns of food production and predation we see in living systems. It can also be applied to fossil organisms because such energy transfer processes, from decomposing heterotrophs to omnivorous heterotrophs also apply to past ecosystems (CC BY 3.0; Thompsma, Wikimedia Commons) .

Biological evolution helps us understand the living world around us. It helps us understand ourselves as a species. It helps us explain the patterns we see in fossils contained in rocks. Evolutionary change is observable everywhere. The development of antibiotic resistance among bacteria. A stroll through a wooded area will reveal an ecosystem of food webs where an intricate daily dance is performed between organisms that have evolved together over time. A decomposing organism is broken down by other organisms specially adapted for that purpose. Parasites exist within your body, your skin, that you rely on for life but yet never meet as you would a friend or neighbor. Evolution can explain all of this and more. Because it is still occurring today, we can observe its processes in action. As a theory that has been itself evolving for more than a century, it is built upon a robust foundation of evidence from the fossil record, biochemistry, observations of naturalists, and observations still being made in labs, remote tropical forests, and even on other planets. Evolutionary theory binds all of these threads of evidence together into a coherent view of life.

> "There is grandeur in this view of life, with its several powers, having been originally breathed into a few forms or into one; and that, whilst this planet has gone cycling on according to the fixed law of gravity, from so simple a beginning endless forms most beautiful and most wonderful have been, and are being, evolved."
>
> – Charles Darwin, The Origin of Species (1859)

Systematics: Organizing Life

Macroevolution and Microevolution

The fossil record provides excellent evidence for the history of life. In particular, fossils reveal macroevolution, or how species have evolved over time. Microevolutionary change, or changes in gene frequency on very small scales, also occurred in fossils when they were alive. However, fossils do not preserve such change. Macroevolutionary change recorded by fossils shows visible variation in both physical features (body fossils) and behaviors (trace fossils), as well as the introduction of new organisms. We find that not only is all life related, but that life has evolved, through common biochemical means, over the course of more than four billion years. This evidence allows us to build models for how evolution occurred in the past and insights into how it is still

occurring today. One such model is the “Tree of Life.” Speciation, or origination, happens at different scales of time, depending on the time it takes to produce offspring. It happens very quickly in bacteria, but quite slowly in larger animals. Because macroevolutionary change can take a great deal of time, the best place to look for evidence of this change in the fossil record. Macroevolution can be thought of as the accumulation of microevolutionary changes, which include various kinds of genetic mutation processes and trait selection. These processes operate at much smaller scales and do not get recorded in fossilization. Despite not being visible in the fossil record, we can study how they work today, and apply uniformitarian principles to understand how they must have worked in the past.

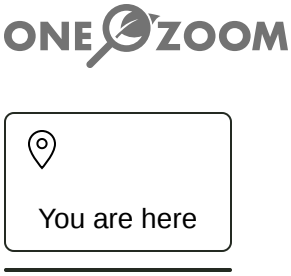


Figure 8.1.6: The OneZoom “Tree of Life Explorer”.

Linnaeus and His Method

People have long organized and classified organisms. The beginnings of recorded classification go back to Aristotle, with his observations of invertebrates while living on the islands of Greece. His method of classifying organisms began by separating animals from plants. He would go on to separate animals into two further groups that we still use loosely today, *Anhaima* (animals without blood, or invertebrates) and *Enhaima* (animals with blood, or vertebrates). This system would persist, like so much of Aristotelian thought, through the middle ages and right up to the time of Linnaeus. Classification of organisms in the years just prior to Linnaeaus’ work took place at a frenzied pace, with new specimens constantly showing up in Europe from the far corners of the world. The “Age of Exploration” (known in the non-western world as the “Age of Being Explored by Europeans”) was a time of great discovery, but each new discovery was as likely to bring confusion as confirmation. Classification was very disorganized, with resulting names of organisms becoming very long and inconsistent. By introducing the system of binomial nomenclature, which we still use today (Genus followed by species, e.g. *Homo sapiens*), Carl von Linné simplified and unified taxonomy. The system relies on shared characteristics, by analyzing what organisms have in common in order to create descriptive names and then organizing them into groups or “bins.”

Ultimately, understanding the complex cacophony of life is challenging. If evolutionary changes occur in predictable ways, then it should be possible to organize all of this into a “model” of life. Modern science owes its understanding of evolutionary processes to the earlier work of luminaries like Linnaeus, who tried to provide order to the living world around him. His approach to taxonomy, the scientific classification of organisms, is a nested hierarchy, with “Kingdoms” as at the highest taxonomic rank and the most inclusive category. Kingdoms get divided up into “Phyla,” then “Classes,” “Orders,” “Families,” “Genus,” and finally “Species.” Like many students, you may have once created a mnemonic device when you first learned this in school. In 1990, Carl Woese would add a higher taxonomic rank, the Domain, because of the discovery of three distinct forms of RNA that separate the archaebacteria, Eukarya, and Bacteria from one another (Figure below). Other than this change, The system has remained largely unchanged since Linnaeus created it. Though Linnaeus only used anatomy as the basis of his classification decisions, his scheme has largely been validated through insights later derived from DNA sequencing. In a few cases, small refinements have been provided through genetic insights, but overall, Linnaeus’ system provides clarity for the patterns we see!

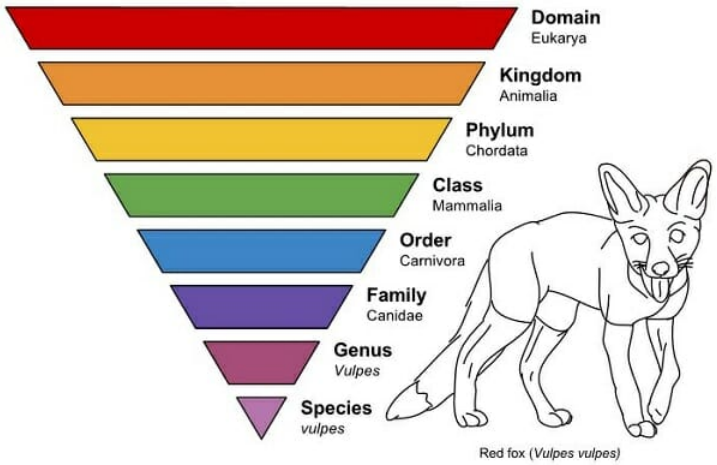


Figure 8.1.7: Red Fox classification according to the Linnaean System. Including the later addition of the "Domain", wolves belong to the Kingdom Animalia and as the system progresses, end with the genus and species names vulpes vulpes.

Lamarckism and Evolutionary Foundations

The theory of evolution provides a robust model for how life changes over time that allows scientists to not only describe what they observe occurring, but to also test ideas about processes related to speciation, genetic mutation, and similar important topics. However, descent with modification is just a part of the story. Determining a mechanism for understanding why some "modifications" helped create success and others did not would rely on uniting these ideas with the environment in which an organism lived. In late 18th century France, Jean-Baptiste Lamarck tried to explain such observations by attributing parents with the ability to pass on traits or behaviors through use or disuse of an organ, anatomical feature, or behavior picked up during their lifetime. This late 18th century idea would become quite a popular way to explain the length of a giraffe's neck, for instance, which would become progressively longer during its lifetime as it strained to reach high branches. This long neck in the parent could then be inherited by the giraffe's offspring. Continued use of this trait would enable further development of the feature. Overall, Lamarck viewed evolution as trending from more simple to more complex forms, because he believed in a naturalist origin of life through spontaneous generation. New life forms, such as microbes, evolved toward more complexity through the generations as they adapted to their environment. His ideas were a major step forward for two reasons. He correctly argued that life took its form through adaptation to the environment, an idea that was quite controversial at the time. Though aspects of his ideas would later be proven wrong, the concept of traits being inherited by subsequent generations through natural processes would be a major step forward.

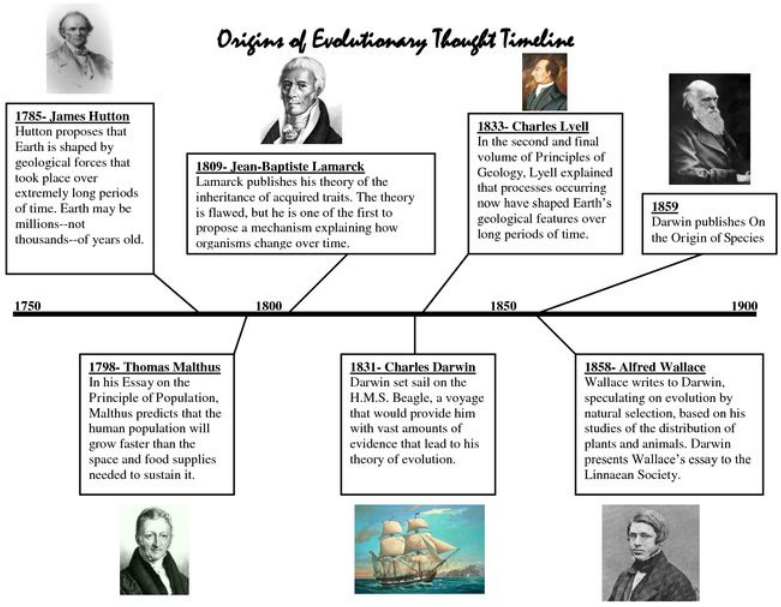


Figure 8.1.8: A brief timeline of evolutionary thought, from James Hutton to Charles Darwin and Alfred Russell Wallace.

Revolutionary Ideas: Darwin and Wallace

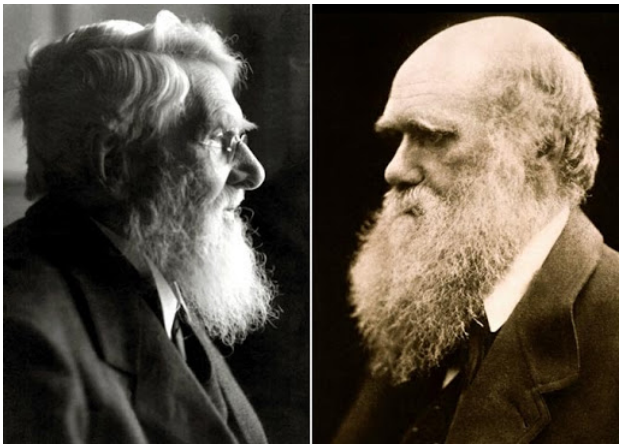


Figure 8.1.9: Photo of Alfred Russel Wallace (left) and Charles Darwin (right), co-discoverers of natural selection.

A New Paradigm Emerges: Darwin and Wallace

Scientific discovery often occurs over time, but always occurs through the accumulation of data. Thomas Kuhn (1962) outlined how scientific "revolutions" are structured as changes in a paradigm. A paradigm is a pattern of thinking. Old models and theories give way and evolve as new information arises. Sometimes, the entire pattern of thinking shifts in a major way, marking an entirely new way of approaching scientific problems. Charles Darwin and Alfred Russel Wallace both discovered, pretty much at the same time, the key process that would lead to a new paradigm in evolution. This key process was ***natural selection***, the variation in survival due to advantages and disadvantages in certain traits within an organism. Before Darwin and Wallace, Linnaeus' ideas showed the relatedness of life and Lamarck's ideas about inheritance supposed that life has changed over time. These ideas represented the paradigm of the time. Darwin's publication of "The Origin of Species" in 1859 initiated a major shift in the view of how traits are transmitted from parent to offspring. Rather than parents passing traits directly to offspring through use or disuse, Darwin and Wallace would argue that this occurred through natural selection. Their views on this were influenced by two major factors. First, Thomas Malthus' writings on economic ideas regarding population and resources emphasized that population growth eventually far outstrips available resources in an environment. Second, both Darwin and Wallace had opportunity to conduct extensive fieldwork in locations filled with biodiversity relatively untouched by intensive human development.

Charles Darwin famously spent five years (1831-1836) as a naturalist aboard *The Beagle*, a British ship with a mission to circumnavigate the globe. During this time, Darwin was able to spend extended periods of time on land, exploring regions around the South American coast, from Brazil to the Galapagos and beyond. He collected rocks, fossils, animal specimens, planet specimens, and more. Sending them on to England prior to his return, these collections made him famous and would become his life's work. Ultimately, the immense variation in plant and animal species across South America was puzzling. He found many completely different species living in environments that were very similar. Why was there such variation across distances if the environments were so similar? The classic example of this are his observations of finch variation across the Galapagos Islands (Darwin's Finches). The best way to characterize this variety was by invoking a common ancestor sometime in the past that, as its descendants moved from island to island, evolved into new species. Another example from his voyage came from his explorations of the greater rhea and lesser rhea, two flightless birds that local guides often ate as food in the area around Bahia Blanca, Patagonia. These two birds were very similar, but different species. Because one lived north of the Rio Negro (River) and the other south, their ancestors must have pursued separate evolutionary trajectories. Darwin's insights were not relegated only to living specimens. Darwin also collected many fossils. In subsequent study back home, he made connections between his living specimens and fossil ancestors. Such evidence would become very important for the development of his ideas around natural selection.

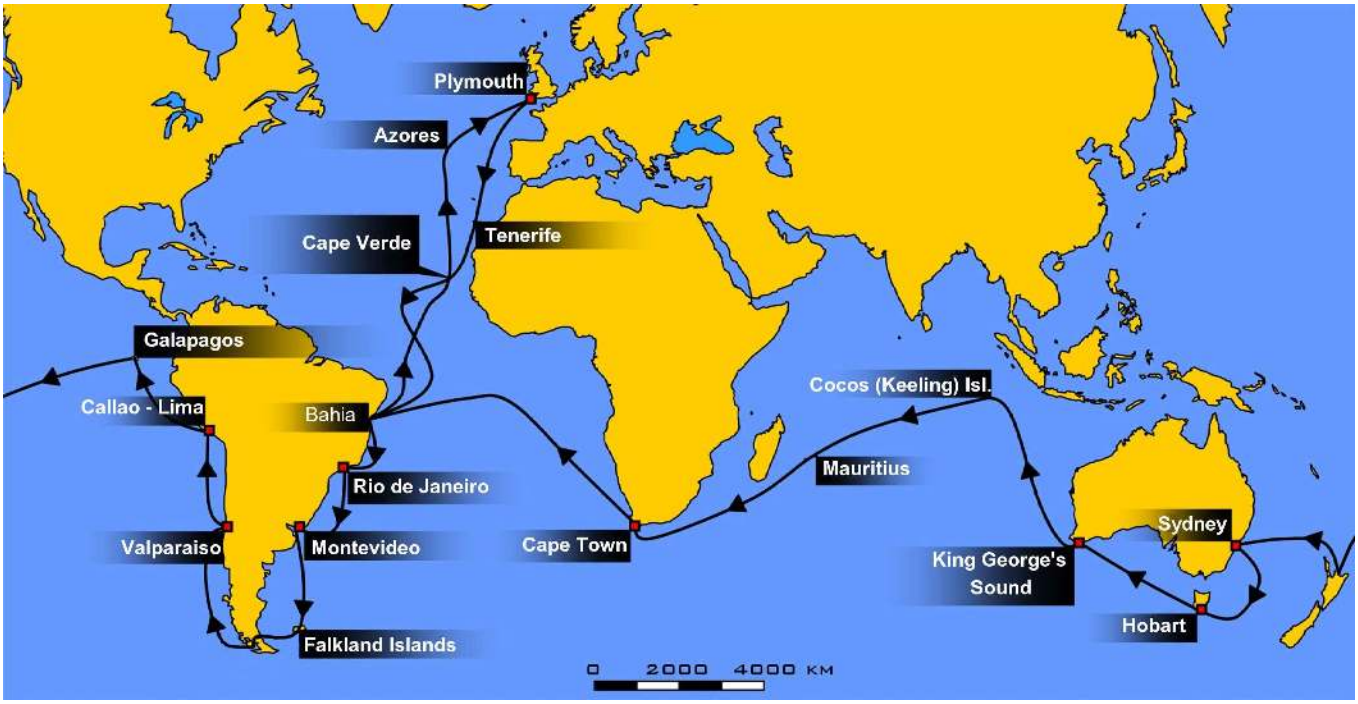


Figure 8.1.10: Stops along Charles Darwin's second voyage of "The Beagle" (1831-1836). These includes locations from Australia to Cape Town, South Africa, to South America and, eventually, back to England.

Alfred Russel Wallace conducted most of his work in two places. Like Darwin, he spent a great deal of time in South America, particularly Brazil. He also conducted field work in what is today Malaysia, Indonesia, and nearby places. Inspired by the field work of Darwin and others, he funded his expeditions, running between 1848 and 1862, through the sales of collected specimens. He methodically planned his field work to explore ideas he had surrounding biogeography, or the geographic distribution of species. If evolution occurs, he reasoned, closely related species should live close to one another. His field work played this out, noting that rivers and other bodies of water did indeed separate related but distinct species. His classic example, called the "Wallace Line," notes the separation of species of Asian and Australasian descent by the Sunda Straits.

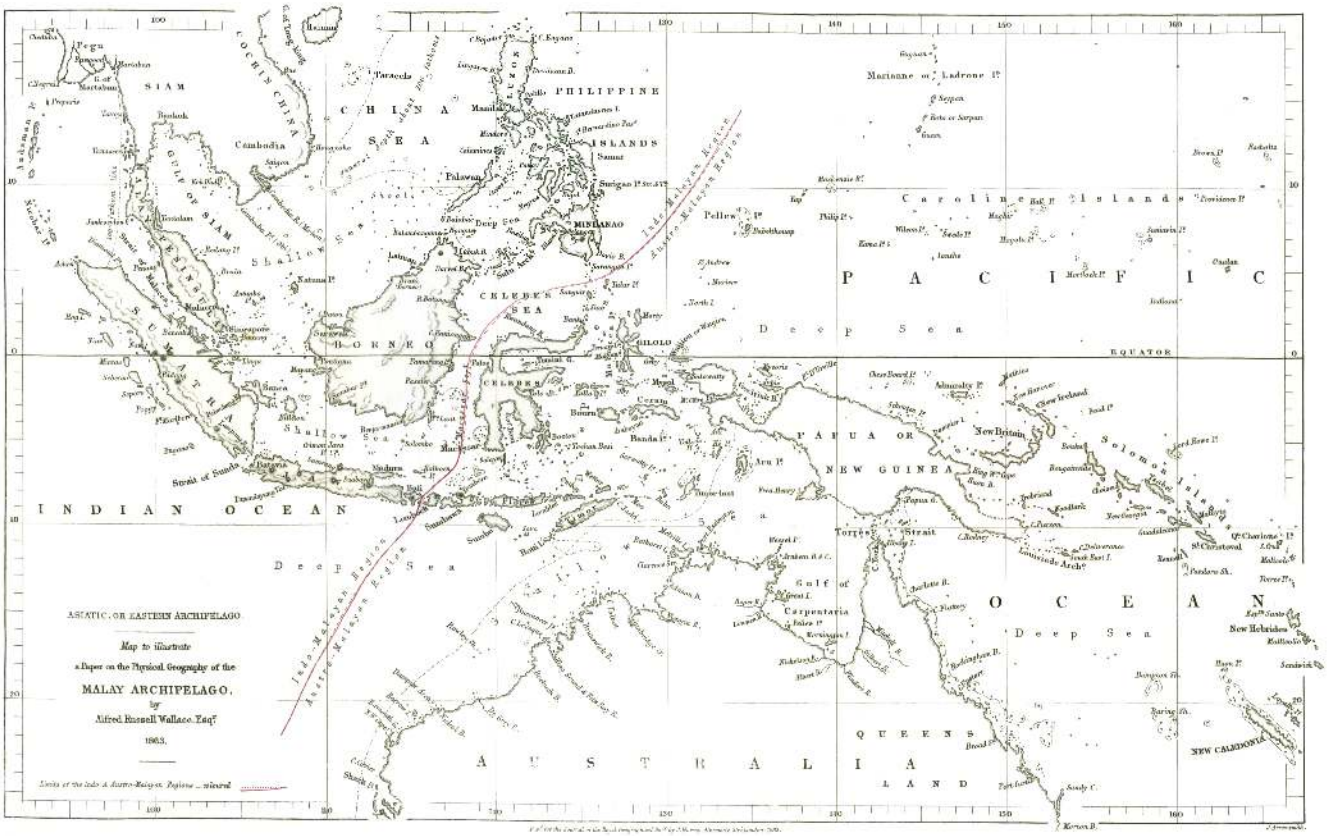


Figure 8.1.11: Map of Indonesia, Australia, and environs depicting "Wallace's Line". This line between landmasses separates Asian from Australasian faunas and was described by Wallace. The differences in species across this line provided Wallace with the patterns he used to describe his views on natural selection and evolution.

He noted that while some species likely descended from a common ancestor had evolved very differently on either side, certain species of flora were the same on both sides. These straits being about 35 km in width, the mobility of an organism was key. Plant pollen and seeds, blown by the wind or transported by water, allow growth on both sides of the strait. Large mammals and most birds, by contrast, could not traverse it. This separation, over time, would lead to new species where once an ancestral form existed. At the time, Wallace would not have known what we now know about the role of plate tectonics in this biogeographical separation. But, he was able to correctly observe here, and elsewhere in his work, that biological change occurs as new species arise from prior species. He recognized natural selection as the mechanism for these observations. Favorable traits, or traits that gave organisms an adaptive advantage, were retained while unfavorable traits would make organisms unfit for their environments.

The two men did correspond. While Darwin came to the idea before Wallace, it was in conversation with Wallace that natural selection, and the new paradigm that would follow, emerged. In 1858, they jointly read papers at a meeting of the Linnaean Society, which would mark the first public description of evolution through natural selection. This was followed one year later by Darwin's publication, *On the Origin of Species*. Darwin and Wallace described natural selection as the mechanism that makes survival to the next generation possible, though it does not mean that these offspring are necessarily the "fittest" to survive. Organisms just have to have enough favorable characteristics to be able to survive long enough to pass on their genes, in whatever condition, on to the next generation. Sometimes, genetic mutations passed on are benign. Other genetic changes passed on provide a favorable adaptation for offspring. At still other times, mutations passed on are detrimental to the survival of offspring or even the ability of a parent to reproduce at all.

Meanwhile, Back in Europe: Mendelian Inheritance and Genetics Take Shape

Even with this new understanding of natural selection, it would take an understanding of how genes work to explain how they are inherited by offspring. The foundations for understanding inheritance would come from the 1860s work of an isolated monk in a Czech monastery. While Darwin and Wallace toiled around the globe and in the heat of rainforests, an important part of evolutionary theory was taking shape through humble pea plants. Gregor Mendel, the son of a farmer, was very interested in plants.

Harnessing his curiosity and focusing on pea plants, he crossbred a wide array of varieties and recorded how traits were passed down in the next generation. Applying his mathematics training, Mendel was able to use statistics, applied to his pea plant populations, to predict which traits – smooth skin, wrinkled skin, and so on, would be inherited. As is often the case with new scientific insights, this work was not recognized for its importance right away. But, owing to the work, interests, and curiosity of this isolated monk, natural selection was bolstered with genetics. Natural selection would not only be attributable to environmental factors, but could now also be described by changes in how genes are inherited from one generation to the next.

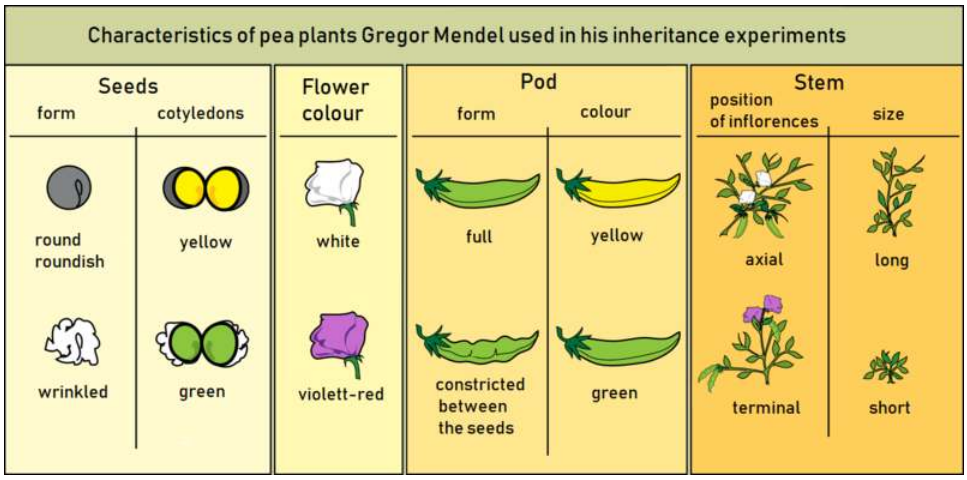


Figure 8.1.12: Characteristics of pea plants Mendel used in his inheritance experiments. These include seed form, cotyledon color, flower color, pod form and color, flower position on stem, and stem size.

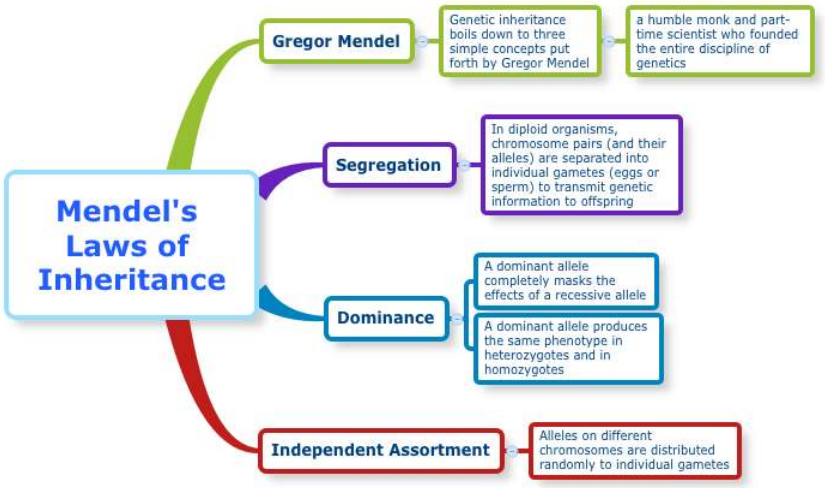


Figure 8.1.13: Mendel's laws of inheritance. (I will recreate this. I found it and it's okay, but I want to remove Mendel and the green arrow at least because it's redundant and doesn't really fit.)

There are three principles in Mendelian genetics. It is also worthy to note that, though the term "Laws" here are accepted within the scientific community, there is enough deviation from them that they may be more accurately referred to as "Principles." Genes come in two forms called alleles ("Law of Segregation"). Alleles can be dominant (expressed in an organism's appearance) and recessive (not-expressed). This is the "Law of Dominance". You may remember creating Punnett Squares in a previous biology class – as these are a tool that can be used to analyze the expression of one allele over another from one generation to the next. The selection of one allele of a gene over another, however, occurs independently of selection occurring among other genes ("Law of Independent Assortment"). These laws, however, do not describe how one allele is transmitted to the next generation, only that alleles exist, are treated independently of other genes during selection, and ultimately determine what variants of such genes are physically expressed. For instance, there are cases where one allele is not completely dominant over another, a situation called "incomplete dominance". Likewise, there are situations where the phenotype for both alleles is expressed. This is called "codominance". It is also known that some genes exist in nature with more than two alleles, or "multiple alleles". Finally, some traits are actually "cogenetic", such as the color of fruit fly eyes, where several genes contribute to a physically expressed variation.

[Image needed]

Figure 8.1.14: Fruit fly specimens displaying eye color variation.



Mutation, Migration, Drift, and Natural Selection

There are several key processes that drive evolutionary change at all scales. These are genetic mutation, gene migration, genetic drift, horizontal gene transfer, and natural selection.

Genetic mutations are simply changes in the sequence of nucleic bases in the coding portion of a DNA molecule. These can be caused by errors in replication or repair of a sequence as it is passed on to the next generation.

Gene migration, or gene flow works differently. Sometimes, a new, sexually reproducing, individual from a separate population is introduced into a new population of the same species. Introduced traits can include many things, such as new varieties of colors within an insect population. This new and unique genetic material mixes into the descendant populations and can then lead to unique and novel down-generation changes. Genetic material can also be exchanged with the donor population in the same manner as populations subject to gene migration are not necessarily isolated.

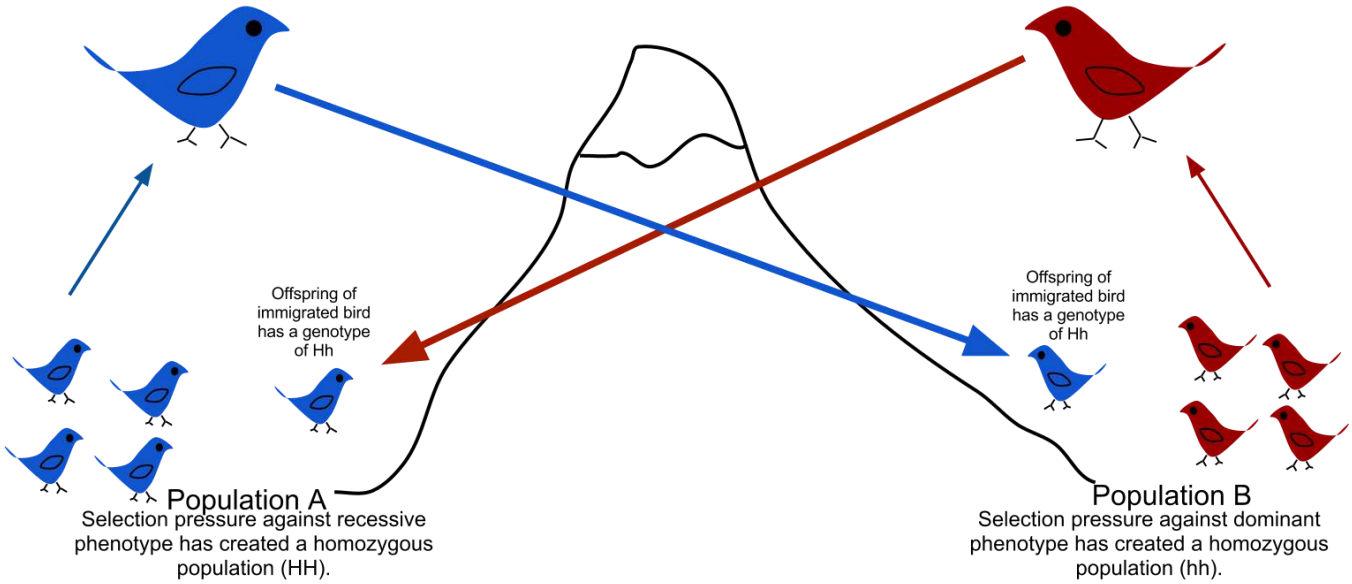


Figure 8.1.15: Gene flow, the transfer of alleles across populations, is depicted here showing birds from two populations living on either side of a mountain. Species A, the blue birds, contains the dominant alleles while species B, the red birds, contain the recessive alleles. A member of species A is able to incorporate it's allele for blue color into the red population through mating. This is one way that gene flow can work (CC BY-SA 3.0; Jessica Krueger, Wikimedia Commons)

Genetic Drift does not lead to adaptational changes in a population. Rather, it is the result of chance, most commonly because some individuals have left behind more offspring than others. It is a change in the frequency of an allele over time, and is completely independent of environmental changes. The genes of the following generations were passed on by those of the individuals fortunate enough to reproduce successfully, and not necessarily of the "fittest". One simple example comes in populations reduced to small numbers that then recover, such as the American Bison, which was nearly hunted to extinction. While its population is much larger than in the past, its genetic variation is much lower than it was 200 years ago (Ungerer et al., 2012). A simpler example might come among a human couple, each with different eye colors, say brown and blue, but where brown represents the dominant allele. Even if the chance of having brown eyes is 50%, all children may statistically end up with blue eyes, thereby eventually erasing that dominant allele. Such cases provide excellent examples of where phrases like "survival of the fittest" are at best an oversimplification of evolutionary processes.

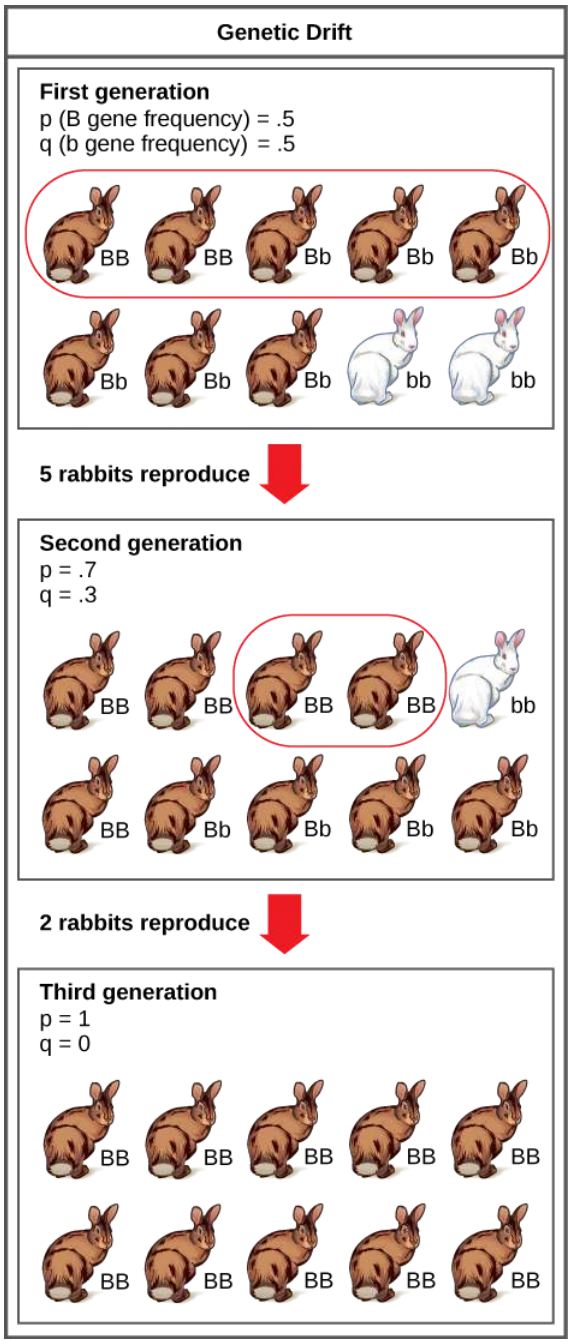


Figure 8.1.16: An example of genetic drift using rabbits. Genetic drift allows for the removal of an allele from a population due to chance. Even through the allele frequency for brown and white fur coats is equal in the first generation, the dominant allele (for a brown coat) quickly displays with more frequency in subsequent generations, eventually eliminating the recessive trait for white coats.

Two other important evolutionary mechanisms that also occur at the genetic level, but in different ways, are **horizontal gene migration** and **genetic symbiosis**. Gene transfer, as we usually think of it, is vertical. This means that genetic material is transferred from parent to offspring. Horizontal gene transfer occurs sideways (Figure left), as with viral transfer. It also occurs among bacteria and archaea that have no means of sexual reproduction. Most of us are familiar with viruses, as we suffer from them frequently and, if we are responsible and had responsible parents, have been vaccinated against them wherever possible. Bacteriophages are a kind of virus that replicates within bacteria and holds promise toward fighting antibiotic resistance (Bragg et al., 2014). Because they inject their genome into host cells where it can replicate, viruses are able to transfer DNA from related and unrelated organisms. This method of transfer is called transduction. Plasmids, which most often are transferred horizontally via a process called transformation, are genetic material that exists in a cell independent of chromosomal interaction. Transformation is

the transfer of genetic material between cells. The final method of horizontal transfer, which can occur via the transfer of plasmids or transposons (a chromosomal segment that can undergo a change of location), is conjugation, which occurs through direct physical contact between cells. Genes can also be transferred horizontally through gene symbiosis, which results from close ecosystem interactions between, at times, species that are evolutionarily separated a great deal. Endosymbiosis is thought to be the origin of key eukaryotic organelles such as mitochondria and chloroplasts (each of which brought its own DNA with it into the host cell). Fungi may transfer genetic material to an arthropod, such as an aphid, through exchanges of cellular material from close physical interactions. In any of these situations, horizontal gene transfer is very difficult to measure, but is also a key evolutionary mechanism that maintains diversity and novel mutation.

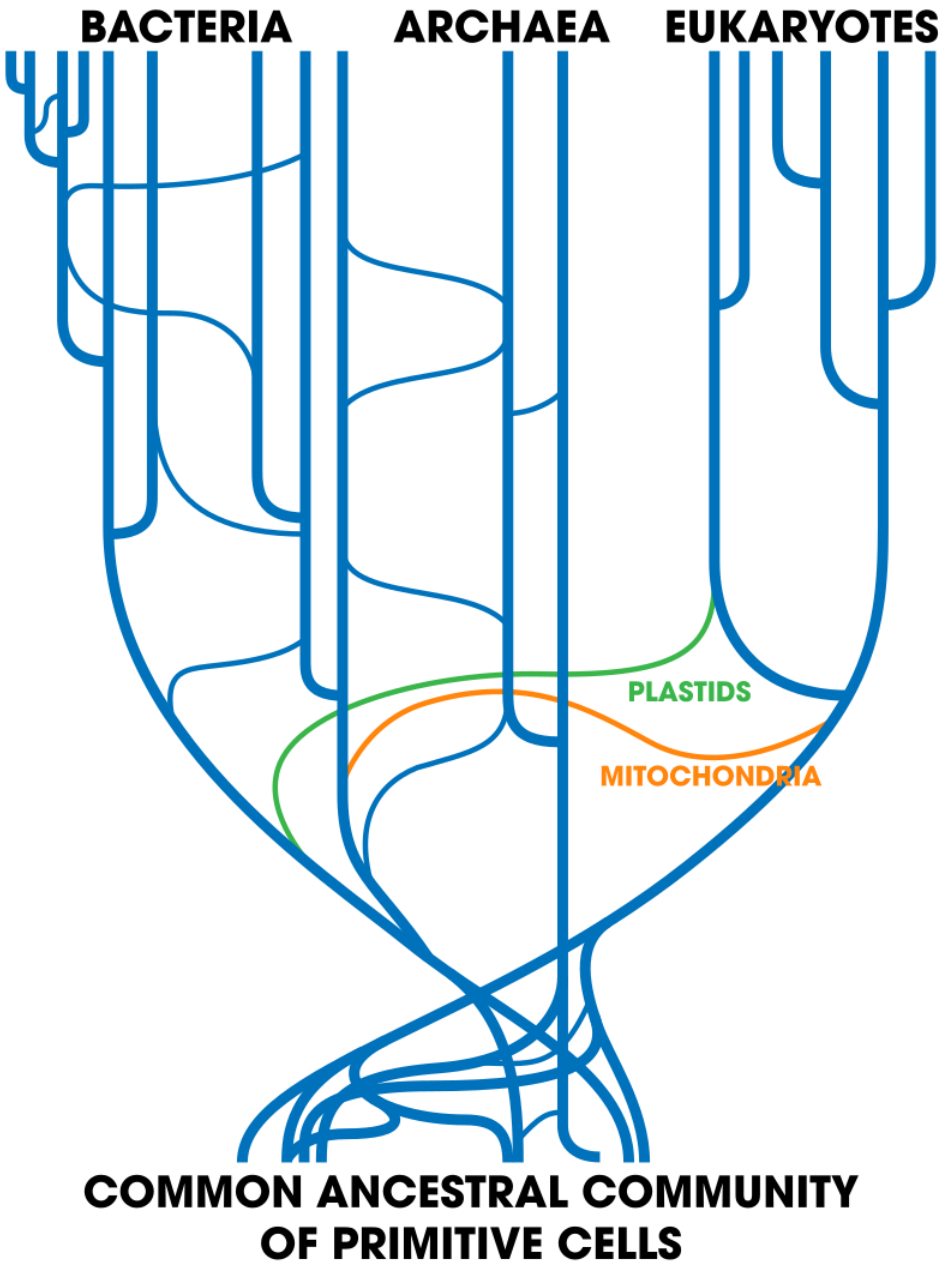


Figure 8.1.17: Horizontal gene transfer as a form of inheritance, in this case through the transfer of genetic material across Kingdoms through plastids and mitochondrial materials (Smets, 2005).

So far, all of the mechanisms of evolution discussed have been random in nature. Our final mechanism, natural selection, is very much not random. The genetic mutations produced by drift, horizontal migration, and others simply make an organism fit into its environment differently, which can lead to being passing such genes along or not. For natural selection to work, there are a number of important requirements that must be met. Within a population, there must be enough genetic variation to provide the flexibility

necessary for advantageous mutations. There must also be a system of heredity that allow offspring to inherit the genes of the parents. Ultimately, a population will necessarily experience differential reproduction, where not all individuals will survive to reproduce due to environmental pressures. These can be everyday pressures, such as predation, that exist in periods of environmental calm, or stasis. They can also occur as a result of rapid environmental change, such as a local volcanic eruption, an extinction-triggering asteroid impact, or our modern episode of anthropogenic climate change. Natural Selection assures that the individuals who survive are more likely to pass on their genes.

Forms of Natural Selection

Some of the most amazing, flamboyant, and beautiful things in the biological world are the result of sexual selection. **Sexual selection** is where a species uses physical features or physical prowess to gain the opportunity to mate with another and pass on their genetic material. Peacock tails, colorful feathers on many male birds, and even dancing among fruit flies can inspire attraction of the opposite gender. Such things grab attention! Sexual selection can also produce situations that are very detrimental to a long life. Organism can become more susceptible to predation. Such selection can require the organism to give up its life in order to pass on these genes. Examples of the latter include sexual cannibalism in male black widow spiders or in some species of mantids where sperm is not released until the male's head has been removed. In these mantids, this form of sexual selection may even be the main driver in the evolutionary change observed in their genitalia (Jensen et al., 2008).

Artificial Selection has been used for agriculture for thousands of years, as our species has continued to select for what are seen as favorable attributes. A classic example is the development of corn, or maize. Today, it is such a staple grain in our diet that you can find a corn product in nearly any food you purchase that has been processed. This includes meats, as many livestock are fed corn-based diets, even if they are not evolved to consume it well, such as with cattle. Native to Mesoamerica, the teosinte grass was used for its kernel and, over time, plants with plumper kernels were selected, eventually leading to the corn we see today. This form of plant domestication, an indigenous American export to the world, is now a staple everywhere.

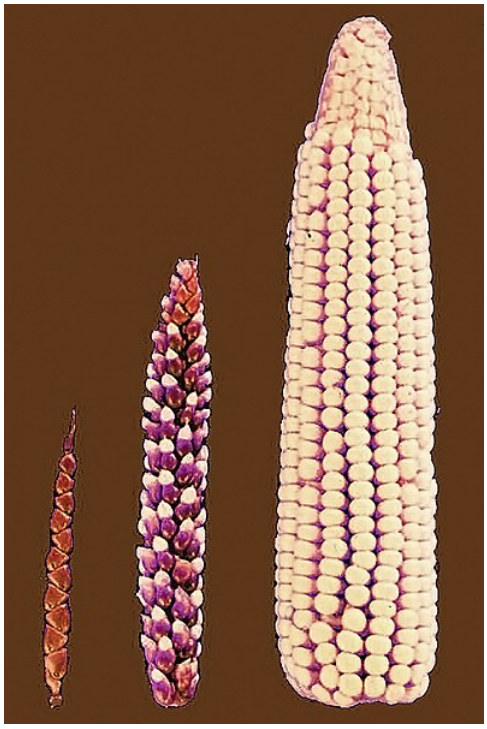


Figure 8.1.18: Teosinte development through artificial selection, or domestication, significantly increased kernal volume and number over time (By John Doebley, Wikimedia Commons)

Another critical example of artificial selection, or domestication, that has shaped the modern world in particular came from Normal Borlaug's work on wheat. Where ancient mesoamericans were selecting for corn based simply upon the appearance of a kernel, Borlaug, as a microbiologist, would select wheat strains based upon their DNA. The problems he was addressing with wheat were directly related to a need to increase yield, shorten the stem of the grain to improve its ability to hold up until harvest, and to make it more disease resistant. Through the use of genetics, this work addressed these problems in the short span of 20 years. The results

are still being experienced by people around the world today, through their ability to feed many more people than would have been possible otherwise (nobelprize.org, 2019). There are myriad examples of human-driven artificial selection that have resulted in the food we eat today. Artificial selection works by humans purposefully choosing individuals that will reproduce, based on which ones have characteristics valued by the farmer or researcher.

Coevolution

There are many examples of situations where two or more species will, as time passes, environments change, and natural selection occurs, affect one another's evolutionary trajectory. Most often, these occur through ecological relationships. These relationships can be mutually beneficial, competitive, predatory, or parasitic. Coevolution is common in nature and illustrates the web of complicated relationships that allow the biosphere to function.

A classic example of coevolution is one that we rely on daily, the interaction between flowering plants and pollinators, such as honeybees. Such coevolutionary relationships are mutualistic, meaning that both species benefit from the interaction. The bees gain food while the plants are able to spread their pollen. Another excellent example of coevolution exists between the acacia ant, *Pseudomyrmex ferruginea* and the various species of the genus *vachellia*, acacia, that it protects. The ant is an obligate organism, in that this form of mutualism is absolutely necessary for its ability to reproduce. The acacia plants provide food and thorns that serve as nests while the ants protect the acacia from herbivores. Such examples are not limited to the world of the living. The fossil record even bears out examples. A fabulous one, depicted in the image below, shows a platycerid snail fossilized near the anal vent of a crinoid. There are actually quite a few examples of this particular relationship, which suggests that it was quite normal for these snails to feast upon crinoid poop!



Figure 8.1.19: Coprophagus, or the eating of feces, by a Platycerid snail lodged near the anal vent of a crinoid. Ordovician, Kentucky (Rich Schrantz, 2000).

A Word About "Fitness"

Genetic change within populations acts randomly. Natural selection makes it more probable that individuals who are genetically prepared for current or new environmental realities will survive. They are the ones most likely to pass on the genes coded for the traits which permit survival. While genetic changes are random, natural selection is most definitely not random. Adaptations that make an organism fit for a cold climate, such as fur on a polar bear, will not likely be advantageous when that climate warms again at some point in the future. *Staphylococcus aureus*, a purveyor of infection throughout the human body, was famously stopped by the antibiotic Penicillin, developed by Alexander Fleming early in the last century. *S. aureus* roared back rapidly with resistance a few short years later and has now become so resistant to antibiotic treatment that some strains, such as "Methicillin-resistant *S. aureus*" (MRSA) are now feared in hospitals, where community infections can spread rapidly among healthy individuals (Chambers and DeLeo, 2010).



Evolution of a bacteria on a "Mega-Plate" petri dish, Koshony Lab, Harvard Medical School

Fitness is a situational circumstance, dependent upon being "good enough" for the environmental conditions of the moment, not the ability of an individual to "pull themselves up by their bootstraps", or run a marathon, or win in a fight, providing themselves some kind of position of advantage through their own will. Ultimately the fitness of an individual is expressed by adaptations, or features that provide some improved function in an environment that is produced through natural selection. There are many features, it should be noted, that are not adaptations. These include vestigial structures, non-functional adaptations left over from an ancestor. Another example of a non-adaptive trait includes the red color of your own blood, which is a result of its chemistry and not due to some kind of selection. ***Exaptations*** make up another category. These are features that may have formerly had an adaptive purpose, but was not produced by natural selection for its current use. These can also be both physical and behavioral. One behavioral exaptation example might be domesticated dogs licking the mouths of their mothers. Interpreted as a means to get a parent to regurgitate food for them, this same behavior in wild wolves toward a lead wolf is interpreted as showing submissiveness (Bauer and Smuts, 2007).

Some traits can also be maladaptive, or harmful to a species fitness. A 2019 study released by Pagano et al. suggests that the eustachian tube construction in Neanderthal humans, *Homo sapien neanderthalensis*, made it susceptible to ear infections. Their reconstruction of the neanderthal eustachian tube suggests that its horizontal construction, very similar to human infants, may very well have even contributed to their eventual extinction. Human infants are notoriously susceptible to ear infections because of this construction of the tube, which can lead to pneumonia and, if left untreated, even death. As infants grow toward adolescence and adulthood, the eustachian tube does not remain horizontal, allowing drainage to occur and less likelihood of infections. Interestingly, while many of us (particularly of European descent) carry neanderthal genetic variants within our genome, given that our our two sub-species are known to have interbred thousands of years ago, it is likely that this eustachian tube, arguably maladaptive in neanderthals, is a trait derived from a common ancestor. Ultimately, this trait may have reduced fitness for our nearest human cousins while remaining a relatively benign feature for our own species.

Modern Evolutionary Theory: "The Modern Synthesis"

The combination of Darwinian ideas with Mendel's genetics is referred to as the "Modern Synthesis". This term was coined by Julian Huxley in his 1942 book, "*Evolution: The Modern Synthesis*". Tying in malthusian ideas about population genetics with evolutionary theory, Huxley provided a framework that effectively brings together the macroevolutionary changes seen in the fossil record with the microevolutionary changes observed in nature and laboratories. However, this revolutionary model also provided the foundation for the rapid biological advances that would occur during the latter half of the 20th and into the 21st centuries. Some of these later developments and discoveries, such as horizontal gene transfer, developmental biology and embryology, etc. have yet to be fully integrated within this framework. Rather than totally replacing the modern synthesis with something new, work continues to discover ways to integrate such new fields and discoveries, ultimately towards finding a way to unite these ideas into a unified, single model. This can be referred to as neo-Darwinian evolutionary theory.

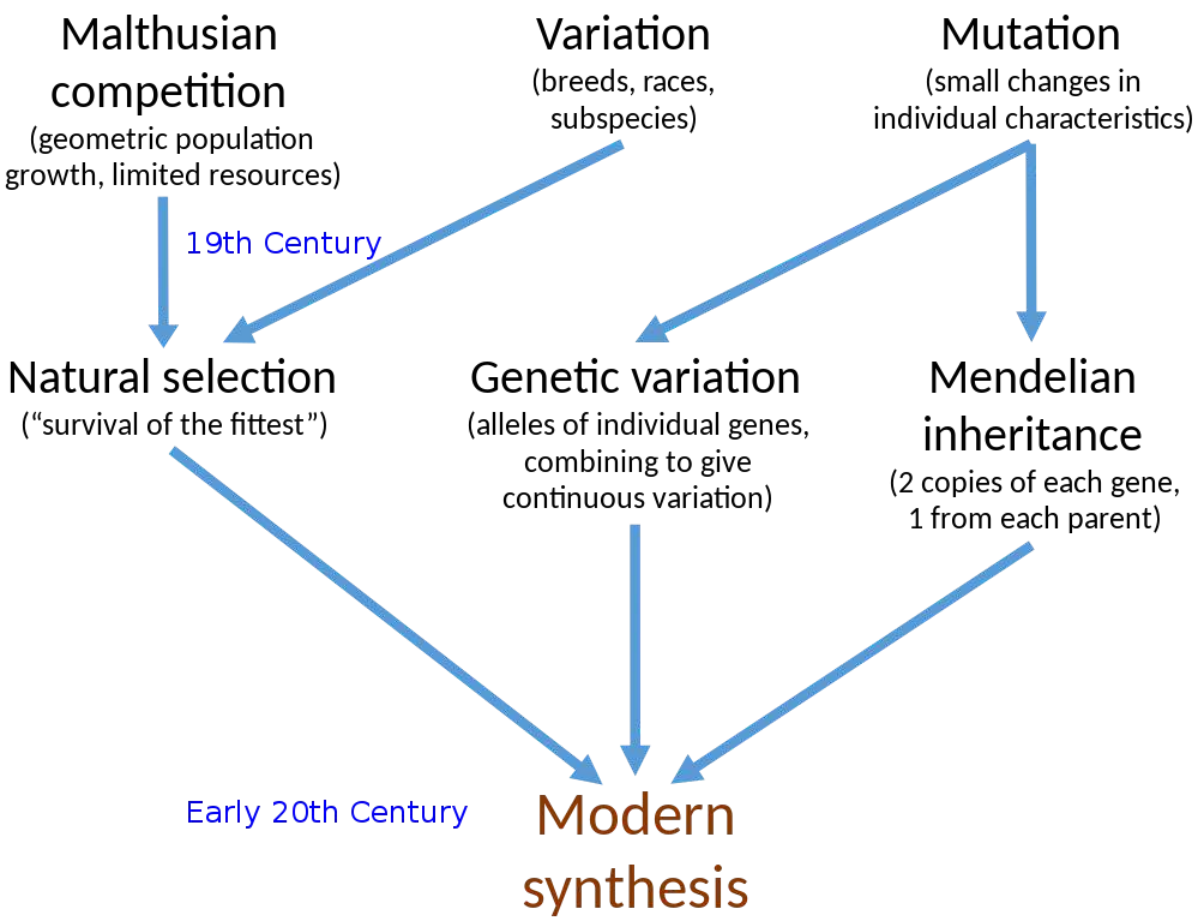


Figure 8.1.20: Diagram of the Modern Synthesis of evolutionary theory. The combination of principles undergirding population dynamics with understandings from genetics is synthesized into a larger and more coherent model of how organisms and populations evolve.

Evolution: Making Sense of Patterns

Like any scientific theory, evolutionary theory is robust, supported by multiple lines of evidence, explains observations make, and allows us to predict and test. As a theory, it does an excellent job of helping us make sense of the patterns we see around us in ecosystems, within populations of organisms, among natural systems, and within our own bodies. This cohesive framework is still being adjusted with new discoveries in different ways. It is also providing scientists with directions for research that shines light on the world around us but research that also leads to critical new medical treatments.

Additional Resources:

Tree of Life Navigators:

University of California Museum of Paleontology Taxon Lift Tree of Life

Wikispecies Taxon Navigator

Tree of Life Web Project

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CHAPTER OVERVIEW

9: Taphonomy - The Science of Death and Decay



Taphonomy is the subdiscipline of paleontology related to the processes of fossilization. This includes all things that happen to the remains of an organism after it dies until it is observed or collected by a geoscientist. This might include the decay of soft tissues, the separation of skeletal hard parts, and chemical changes to the organism after burial. By studying these processes, we can also uncover clues about the environment in which the organism lived, died, and was preserved.

9.1: Is This a Fossil?

9.2: Death to Fossil Transition

9.3: Modes of Fossil Preservation

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

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9.1: Is This a Fossil?



Figure 9.1.1: Archaeopteryx fossil

Consider the photo of *Archaeopteryx* on this page. This image likely represents what many people think of when they hear the word "fossil:" bones of an ancient organism preserved in rock. Indeed, it does include major components that define many fossils: easily preservable hard parts (bones), a medium of preservation (surrounding rock), and it's old. Now think about a turkey leg your dog buries in your backyard. Is that a fossil? No, the turkey leg isn't a fossil because it is too recent. Officially, fossils are remains of organisms over 11,700 years old, which marks the boundary between the Pleistocene and Holocene Epochs.

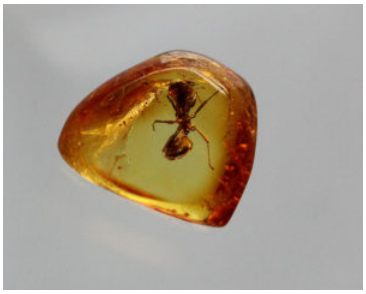


Figure 9.1.2: Fossil ant in Baltic amber, about 50 million years Eocene

Many fossils are body fossils, which includes both mineralized hard parts (shells, bone), and a diverse array of soft tissues (like skin, feathers, flowers) from a previously living organism. Trace fossils (such as burrows, nests, and coprolites) represent the activity of living organisms and are also found throughout the fossil record. Preservation of both body and trace fossils typically occurs in lithified sediment, but body fossils may also be trapped in amber (fossilized tree resin), glacial ice, tar pits, or even lava.

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9.2: Death to Fossil Transition

Decay of soft tissue



Figure 9.2.1: Otzi is a mummified man, who had been preserved in ice, and found high in the Alps between Austria and Italy.

When a living organism dies, soft tissues, as such muscles and internal organs, that make up the body of the organism will typically immediately begin to decay. Scavengers may take advantage of a feast. Bacteria and fungi will begin to break down the tissues to reuse available nutrients. Soft tissue decay happens fastest in the presence of warm, wet conditions with lots of hungry organisms around. Low temperatures, poor oxygen supply, and/or extremely high or low pH slow down rates of soft tissue decay, as these factors inhibit bacterial growth. Therefore, some of the best fossils that include unaltered soft tissues come from deposits that represent rapid burial which sealed the organism off from its environment. Other superb fossils including soft tissue remains are entombed in glacial ice. In some cases a thin film of carbon may be left behind as this rapid burial occurs, leaving a fossil preserved by carbonization.

Pre-burial decay of hard skeletal elements

For multicellular organisms with hard parts (shells, bones, teeth; collectively referred to as skeletal elements), it is the soft tissues that hold these hard pieces together. Once the skin, tendons, muscles, and other connective tissue decays away, there is little to keep the skeletal elements intact, and these shells and bones can disaggregate, becoming sedimentary particles.



Figure 9.2.2: Xenocrinus baeri fossil crinoids in fossiliferous limestone (Whitewater Formation, Upper Ordovician; northeastern Warren County, southwestern Ohio, USA)

Most of these skeletal elements are reworked by waves or other currents in the environments in which the organisms lived. If the environment in which the organism lived was prone to currents of running water, flooding, large storms, or turbidity currents, it is possible for the skeletal elements to be transported out of their original habitat.

In high energy environments, such as those with with active currents, it may become very easy for skeletal hard parts to undergo any (if not all) of the following **physical** processes prior to burial:

1. *Disarticulation* – Separation of valved shells or vertebrate skeletons into individual pieces.

2. *Fragmentation* – Physical breakage of shells or bone into smaller pieces.

3. *Abrasion* – Physical wearing away of shells or bone as they are jostled against each other or against particles of sediment in currents.

Generally, skeletal components that are thin, less dense, and elongate are more susceptible to physical breakage. Chunky, robust, dense skeletal elements are more likely to persist with minimal damage.

There are also some **biological** processes that can contribute to the physical weakening of skeletal components and increase the surface area for chemical processes to continue breaking down the hard parts, such as:

1. *Boring* – parasitic or predatory organisms create holes via chemical and physical means in shells or bone when the animal is alive. In marine settings, common borers include sponges and snails. After death, additional boring of shells and bones can occur by organisms that require hard substrates for growth.

2. *Encrustation* – organisms that need a hard surface on which to grow may populate the surface of the shells or bone (note for some shells, this may also occur while the shelled organism is still alive). Typical encrusting organisms in marine settings are sponges, bryozoans, tube worms, and barnacles.

Compare the three specimens below–click on the first two to rotate them and consider the differences between the first two well-preserved specimens and the bored and encrusted specimen.



Figure 9.2.3: Articulated Mercenaria sp. specimen



Figure 9.2.4: Disarticulated Mercenaria specimen



Figure 9.2.5: Sponge borings (Entobia made by the genus Cliona) and encrusters on a shell of the modern hard clam, Mercenaria mercenaria, from North Carolina. This encrustation and boring happened after the death of the clam when the shell was empty. (English Wikipedia [Public domain]; Photograph taken by Mark A. Wilson (Department of Geology, The College of Wooster), 2007.)

Post-burial decay of hard skeletal elements

After skeletal hard parts are buried, additional **chemical** changes will commonly occur to the shells or bone. Many of these processes are described in the next section on modes of preservation, and are related to the changing the mineralogy of the original hard parts via recrystallization, permineralization, replacement, or dissolution. These chemical changes occur due to groundwater seeping through the pores of the burial sediments, and increasing temperatures associated with deeper burial. As the fossil becomes more deeply buried, additional distortion and fragmentation of the material is possible due to the weight of the overlying sediment. Stretching and shearing of fossils due to applied tectonic stress may also occur.

Death to Fossil Transition - Quiz

? Exercise 9.2.1

Which of the following would easily be considered fossils?

a. Woolly mammoth bones

b. Civil War bullets

c. Beach shells

Answer

a. Woolly mammoth bones

? Exercise 9.2.2

Is this an example of biological, chemical, or physical weakening of a shell?



a. Chemical

b. Physical

c. Biological

Answer

c. Biological

? Exercise 9.2.3

Which of the following would NOT slow down soft tissue decay?

a. Low oxygen levels

b. Acidic water

c. High temperatures

Answer

c. High temperatures

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9.3: Modes of Fossil Preservation

Unaltered remains

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 shell valves 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.

Table 9.3.1

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



Figure 9.3.1: Wooly mammoth (Mammathus primigenius) hair.

In some exceptional cases, soft tissues, such as skin, feathers, or fur may be preserved. This type of preservation does not typically occur within sediments, but rather, 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 when that water evaporates, minerals grow in the available space. Typical infilling minerals include quartz, calcite, and iron oxides, which are all common cementing minerals for sedimentary rocks. 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 both fill the porous spaces in the plant tissues and replace the organic material of trees. Permineralized fossils tend to 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.



Figure 9.3.2: 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.

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.



Figure 9.3.3: Crane fly, Florissant Fossil Beds National Park

Replacement



Figure 9.3.4: Pyritized trilobite, Triarthrus eatoni, Upper Ordovician, New York

Increased pressure and temperature conditions in the deep burial of sediments liberate trapped fluids which contain 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 assist in reducing sulfur and precipitating the iron sulfide mineral, pyrite. Often, much of the detail of the original organism can be preserved via replacement.

Recrystallization

Recrystallization is an important process that preserves fossils made of aragonite, a common, but chemically unstable mineral over the 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 $CaCO_3$ and it sometimes appears as beautiful "mother of pearl" on the inside nacreous layer of shells. But outside of the organism's living tissue, aragonite is less chemically stable than calcite, and so it spontaneous 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 of the finer details as new crystals form.

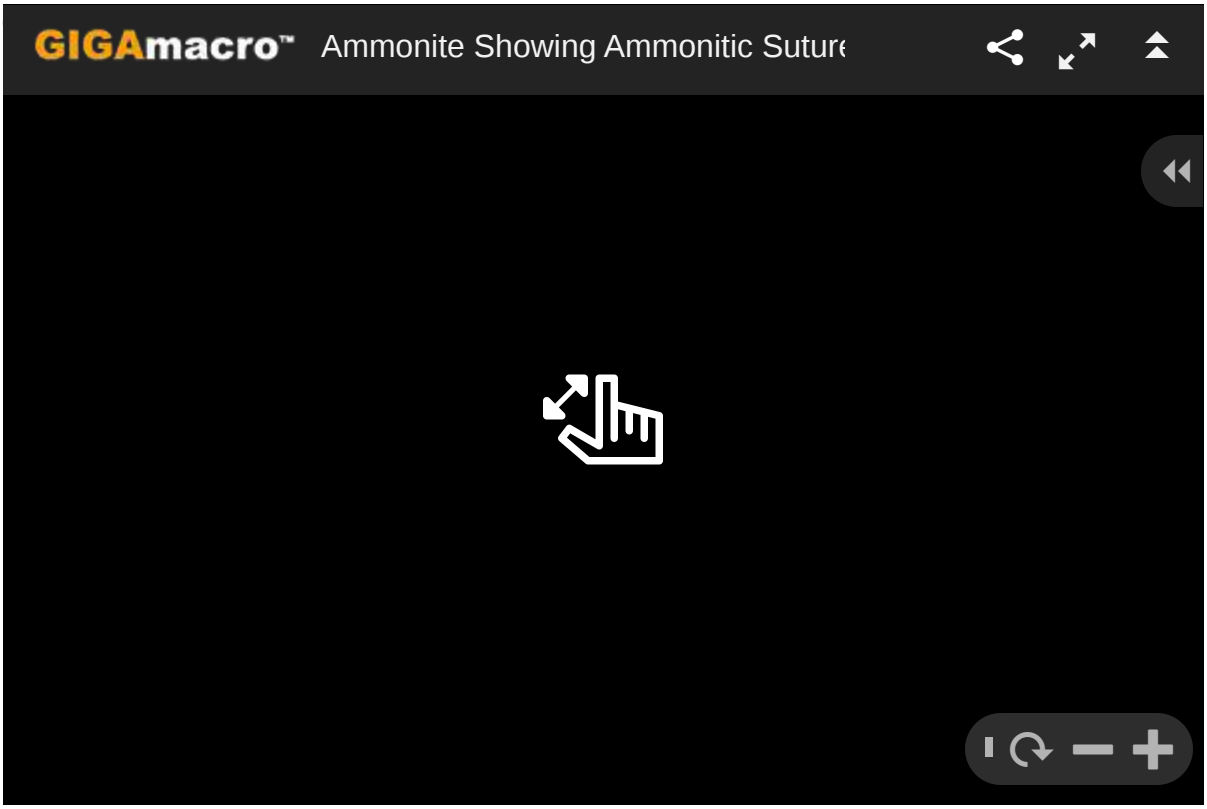


Figure 9.3.5: Iridescent mother of pearl on nautiloid shell

Internal molds



Figure 9.3.6: Internal mold of 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 though the bucket itself is gone–that's an internal mold. In geology, this mode of preservation is usually found in bivalves or brachiopods, or snails that have a single open chamber where sediment can get in and work its way into the entire space. This mode of preservation can reveal lovely impressions of internal features like muscle scars, but since most molluscs are more readily identified using external ornamentation, loss of the original shell can make things tricky for paleontologists.

In some cases, small cavities inside a shell may remain after sediment is washed into the shell. The sediment sinks to the bottom of the cavity and leaves a pocket of water above. Over time, these pockets can later close up with precipitated minerals, potentially serving as geopetal structures. Geopetal structures are "up indicators", and give geologists a good sense of which way was up when the bed was deposited, which can be helpful telling the stories of sedimentary layers that have been deformed by tectonic processes. In the image below, examine the difference in sizes of the infilled calcite crystals in the area outlined in pink versus the area outlined in orange. Which way was up?

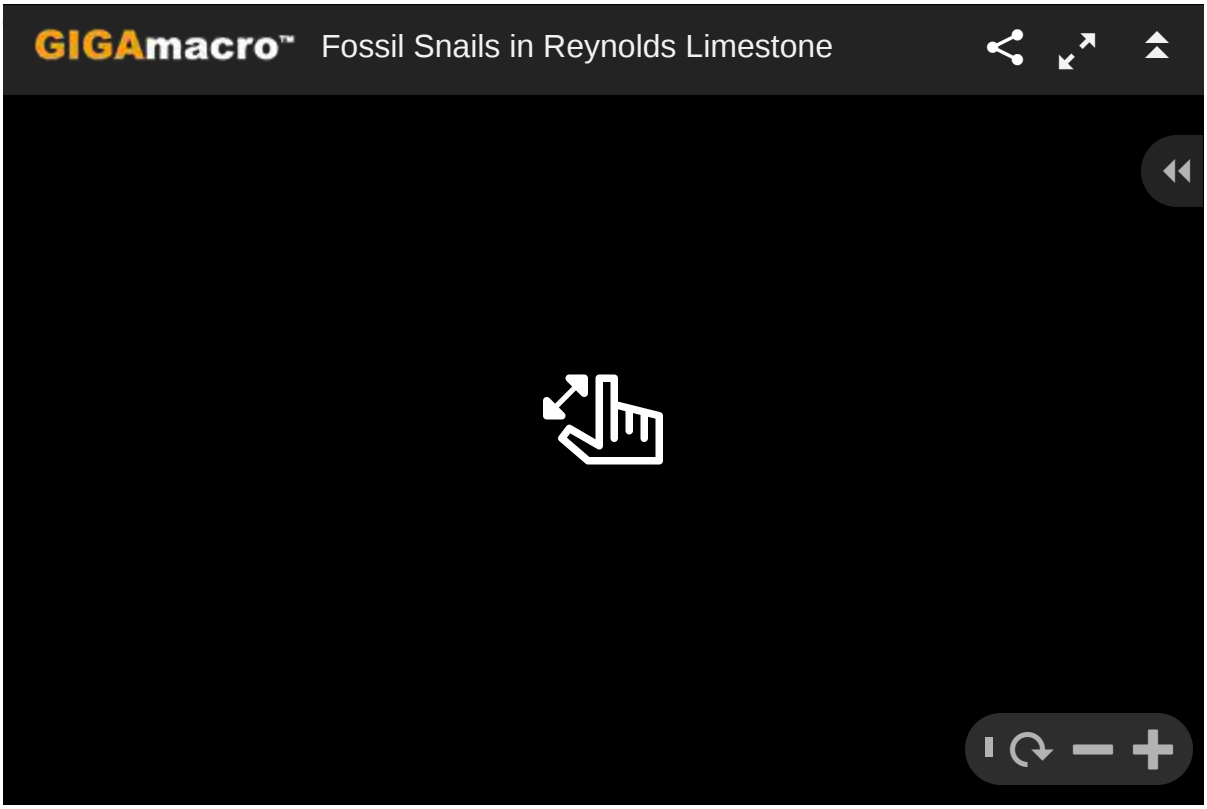


Figure 9.3.7: Geopetal structures in Reynolds Limestone

External molds

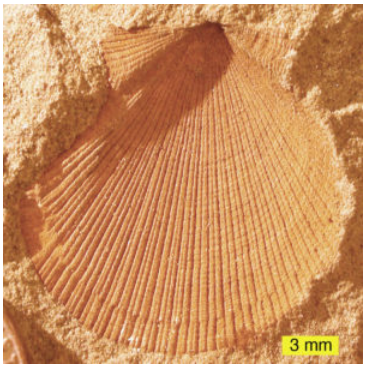


Figure 9.3.8: External mold of a bivalve from the Logan Formation, Lower Carboniferous, Ohio

An external mold is an impression of an organism's outside shape. This type of 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 away. When external molds are discovered, the observer sees a version of the exterior of that organism, but it is not an exact 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.



Figure 9.3.9: 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 bone. 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.

Another method of creating external molds is through bioimmuration. Bioimmuration comes in several forms, but consists of the creation of an external mold through the encrustation of other organisms on exterior surfaces.



Figure 9.3.10: Bioimmuration of a soft-bodied organism, Catellocaula (star-shapes) on bryozoan

For example, barnacles often encrust boats and if you were to remove the wood from the barnacle, an impression of the wood would remain on the bottom of the barnacle. Sometimes, whole successions of external encrusters will attach themselves. These can certainly include arthropods such as barnacles, but also encrusting corals, brachiopods, sponges, and many more organisms just looking for a place to attach. Bioimmuration as a form of external mold only happens when the original hard substrate dissolves, rots, or is otherwise removed, leaving behind the encrusting organisms and, very often, a history of successive encrustations that give important glimpses into unique paleoecological situations.



Figure 9.3.11: External mold (left) of a trilobite body fossil in a concretion

Casts



Figure 9.3.12: Sue, the most complete (and famous) Tyrranosaurus 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 off 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, it is a way to share specimens among museums, or the fossils may be currently be under study.

Modes of Fossil Preservation - Quiz

Exercise 9.3.1

What mode of preservation is this?



a. Internal mold

b. External mold

c. Cast

Answer

b. External mold

Exercise 9.3.2

The opalized belemnite (mollusc) in this photo is from the Coober Pedy mines in Australia. Opal is a form of silica (quartz mixed with water). What mode of preservation is this?



a. Replacement

b. Permineralization

c. Recrystallization

Answer

a. Replacement

Exercise 9.3.3

What mode of preservation is this?



a. Carbonization

b. Unaltered hard parts

c. Permineralization

Answer

a. Carbonization

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

As noted earlier, remains of organisms are not just found as body fossils. 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 were buried say anything about what you were doing at any point in time? Likewise, non-human organisms scurry, burrow, step, jump, run, walk, slither, and grow in a variety of ways. Traces of these actions are preserved in sediment and are studied under the branch of paleontology called ichnology.

Earliest trace fossils

Some of the earliest trace fossils are found in rocks that are around 3.5 billion years old, and preserved as structures referred to as stromatolites. These are macroscopic 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 convex through time, though the relief and size vary tremendously. Stromatolites require certain conditions to grow: (1) sunlight, and therefore shallow water, and (2) absence of grazing predators such as snails. They 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. The high salinities and powerful scouring tides of these environments generally 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 locomotion. 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 an ecological method that focuses on the behavior of the organism (see table).

Table 9.4.1: Trace fossil classifications *(after Seilacher, 1967)*.

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

Ethologic (behavior) class	Associated behavior	Example
Repichnia	Locomotion	 Cruziana trace fossil
Cubichnia	Resting	 Rusophycus trace fossil
Domichnia	Dwelling	 Thalassinoides trace fossil
Pascichnia	Grazing	 Helminthoposis trace fossil
Fodinichnia	Feeding burrows	 Zoophycos trace fossil



Figure 9.4.1: Skolithos trace fossil

An alternative trace fossil classification is similar to the Linnaean system, creating an ichnotaxa from an ichnospecies within an ichnogenus. An example of such an ichnotaxa is *Skolithos linearis*–a long, straight, vertical burrow. Over time, ichnologists, or paleontologists who study trace fossils, have been able to show relationships among traces and how those traces change shape over time, using cladistic methods similar to those applied to living organisms.Unfortunately, it is usually very difficult to associate a particular organism with a particular t race. However, in rare instances, organisms such as trilobites have been preserved within a trace they created (e.g., *Cruziana sp.*). The real utility of trace fossils is in their association with particular environments.

Some traces are found in beach facies. Some traces are characteristic of beaches, while other different traces are typical of deep-water muds, and yet others are found in terrestrial mudflats. Even though the shape and movements of particular organisms play a role in how a trace fossil is formed, the characteristics of the sediment are just as critical to creating traces. Using uniformitarianism, geoscientists can use the tools of “neoichnology”, or the study of modern traces, to help understand traces found in rocks and to interpret their likely environment of formation. Importantly, many traces are time-independent. Organisms burrow now just as they did 400 million years ago; though the organisms may be different, the action is the same (Seilacher, 1967). Thus, trace fossils are excellent facies indicators and provide fabulous insights into the organisms and environments that formed them, and have been grouped into categories known as ichnofacies. Each ichnofacies is created by a community of ichnotaxa.

Trace Fossils - Quiz

? Exercise 9.4.1

Which of the following are considered trace fossils?

a. Bones

b. Shells

c. Burrows

Answer

c. Burrows

? Exercise 9.4.2

True or false: It is always easy to match trace fossils with the organism that created them.

a. True

c. False

Answer

c. False

Further reading:

University of California Museum of Paleontology Fossils

Seilacher, A. (1967). “Bathymetry of trace fossils”. *Marine Geology*. **5**: 413–428. doi:10.1016/0025-3227(67)90051-5.

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CHAPTER OVERVIEW

10: Innovations of Life Through Time - Life Finds a Way

Based on current scientific knowledge, Earth is unique in its ability to host life in the diverse array seen throughout geologic time. With the processes of evolution as constraint and guide, life has developed numerous innovations that have allowed organisms of various kinds to make use of the available resources on this planet. This chapter explores the impact on some of these innovations to better understand the history of life on Earth.

10.1: In the Beginning

10.2: Phanerozoic Diversity of Life (According to the Hard Parts)

10.3: Further reading and resources

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10.1: In the Beginning

Conditions for first life



Figure 10.1.1: Artist rendition of Hadean Earth. (Artist: Raymond R. Kilburn, Jr. Used with permission.)

No one can say exactly when and where life began on Earth, but it is certain that the origin of life is one of the most critical, planet-changing events ever to occur. Based on scientific observation and reasoning, theories have been developed to explain the ideal conditions that allowed life to start. Conventional scientific thought suggests that extreme heat from formational, early meteorite bombardment of Earth during the Hadean would have made conditions at the surface too harsh for life to begin–literally, portions of the early crust melted from these collisions (e.g., Marchi et al., 2014). However, despite continued impacts during the Late Heavy Bombardment, some studies that suggest only a fraction of the planet's crust would have melted, leaving potential areas of refuge deep in the crust or within hydrothermal vents at the bottom of the oceans. Based on geochemical data, oceans are thought to have developed very early in Earth's history, by about 4.3 billion years ago (*i.e.*, during the early Hadean). Thus, while the possibility exists that life could have arisen during the Hadean, given the minimal availability of rocks from this interval, scientists in search of first life are typically working in rocks of Archean age.



Figure 10.1.2: Apex Chert of western Australia contains of of the oldest known fossils

Finding definitive evidence of the first life forms on Earth presents a challenge. While there are geochemical signatures to many early Archean rocks that suggest organisms may have been present, the organisms would be microscopic cells with no hard parts. Additionally, most of the Archean rocks that potentially preserve these fossils have been subjected to well over 3 billion years of tectonics and weathering. Schopf (1993) proposed they had discovered the oldest known fossils in the 3.5 billion year old (Archean) Apex Chert of western Australia. Debate quickly ensued in peer-reviewed literature, and many geoscientists now interpret the filamentous structures from the Apex Chert are the result of mineral growth around hydrothermal vents (e.g., Marshall et al., 2011). Currently, sulfur-metabolizing microfossils in the 3.4 billion year old (Archean) hydrothermal cherts of the Strelley Pool Formation of western Australia are considered the oldest definitive organic microfossils (Wacey et al., 2011; Alleon et al., 2018). However, Dodd et al. (2017) showed evidence of early life as tubular and filamentous structures near hydrothermal vent deposits of the 3.77-4.28 billion year old (Archean) rocks of the Nuvvuagittuq Belt in Quebec, Canada. While all of these findings

continue to be examined in further detail, there is a common thread–early life seems to have begun in association with the warm, chemically enriched environments of hydrothermal vents.

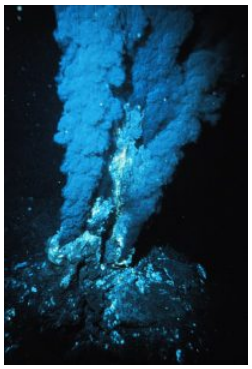


Figure 10.1.3: Black smoker at hydrothermal vent along mid-ocean ridge in the Atlantic Ocean. (P. Rona / OAR/National Undersea Research Program (NURP); NOAA / Public domain)

Today, hydrothermal vents are found in the deep ocean along mid-oceanic ridges at divergent plate boundaries and at volcanic hot spots. In these environments, cold water from the bottom of the ocean seeps along fissures found at the plate boundary. While underground, the water is heated by and mixes with fluids released from the rising magma. The super-heated water is able to carry many dissolved minerals and mobile ions. These ions are carried along with the water to the seafloor, and are then precipitated as mineral deposits around the vents. Sometimes these mineral deposits are infused into the surrounding deep sea sediments, and sometimes they build impressive “chimneys” of towering rock. Sulfur-based minerals are particularly common near the vents. The vents are surprisingly diverse in life, with sulfur-reducing bacteria extracting energy from compounds in the vented fluids. Using the principle of uniformitarianism, the first organisms at Hadean hydrothermal vents were therefore also likely sulfur-reducing bacteria. These bacteria were chemoautotrophs, meaning they extract the energy necessary for cellular functions from inorganic compounds. In the case of sulfur-reducing bacteria, they “breathe” sulfate (SO_4^{2-}), and reduce sulfur to hydrogen sulfide (H_2S). Sulfur-reducing bacteria are found today at the Mid-Atlantic Ridge, East Pacific Rise, and in pools associated with land-based hydrothermal activity like the hot springs and geysers of Yellowstone National Park in Wyoming. Some sulfur-reducing bacteria are also found in coastal environments–if you have ever noticed a rotten-egg odor in salt marshes, you’ve smelled the scent of chemoautotroph productivity! These prokaryotes belong to the Domain Archaea, which includes a diverse array of microbes that are able to live in chemically harsh conditions over a wide range of temperatures.

The prokaryotic pioneers

So the first life forms on Earth were likely sulfur-reducing bacteria, which are prokaryotes. Prokaryotes are structurally rather simple–they are single-celled, small in size (0.1 – 10.0 μm), and have no nucleus or other organelles. Make no mistake though, these simple organisms have changed the trajectory of Earth’s history in major ways. To quote Shakespeare, “Though she be but little, she is fierce!”

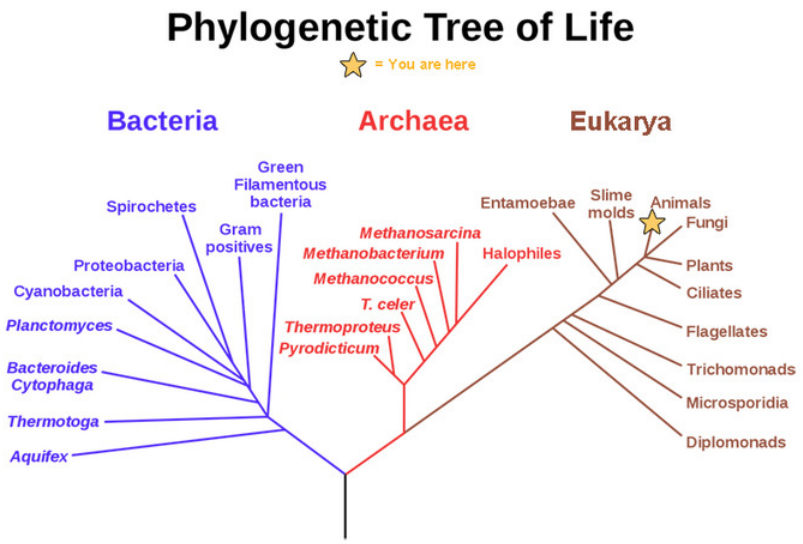


Figure 10.1.4: Phylogenetic tree of life showing the three major Domains. (Public Domain image.)

Two of the three Domains of life, Archaea and Bacteria, are prokaryotes. The third Domain, Eukarya, to which you belong, is discussed later in this chapter. Prokaryotes live on and in just about everything on Earth. They have been found in sulfuric volcanic lakes, hypersaline springs, glacial ice, growing on crystals deep in caves, and under rocks at the tops of mountains. In the human body, they outnumber human cells by 3 to 1. Rough estimates of modern oceanic prokaryotic diversity suggest more than 2 million species exist, which does not even include all the possible diversity on land. Consider how quickly bacteria evolve–check out this video that shows bacterial mutations against antibiotics in just 11 days! Then, remember that prokaryotes have been evolving for 3.5 billion years–there is no way the true diversity of prokaryotes will ever be known!



Figure 10.1.5: Stromatolite from the Archean Dresser Formation, Western Australia. (Creative Commons Attribution 2.0 Generic license.)

In many rocks of similar age as those preserving the Archean sulfur-reducing bacteria, macroscopic sedimentary structures known as stromatolites record the history of another important group–the cyanobacteria. Stromatolites are layered structures created when mats of cyanobacteria are smothered by fine calcite mud distributed by wave action in shallow marine environments. The bacteria grow through the mud layer to create a new mat, and the process repeats, producing a thinly laminated structure that is visible when lithified. Nutman et al. (2016) suggested they had found stromatolites in Greenland that were 3.7 billion years old (Archean), but this has been disputed by Allwood et al. (2018). The most widely accepted definitive stromatolites are those in the 3.5 billion year old (Archean) Dresser Formation in Australia.



Figure 10.1.6: Modern unicellular bacteria from a microbial mat in Guerrero Negro, Baja California, Mexico. (Public Domain via NASA.)

Cyanobacteria are photoautotrophs, meaning that they require sunlight to create their own food via photosynthesis. Therefore, these organisms are found in the shallow, photic zone of marine environments where sunlight can readily penetrate the water. Photosynthesis uses carbon dioxide (CO_2) as a source of carbon for cellular functions, and produces free oxygen (O_2) as a waste product. Cyanobacteria were the dominant life form on Earth for about 2 billion years, spanning the Archean and a major portion of the Proterozoic. As a result of all that photosynthesis, the amount of O_2 in Earth's atmosphere significantly increased over time in one of the most slow-building (but literally life-changing) events of all-time–the Great Oxidation Event. Banded iron formations (BIFs) are first preserved in the rock record around 2.4 billion years ago (Archean). BIFs consist of alternating layers of dark grey to black iron oxides (hematite and magnetite) with red to yellow chert.



Figure 10.1.7: Banded iron formation, Barberton Greenstone Belt, South Africa. (Creative Commons Attribution-Share Alike 1.0 Generic license.)

The interpretation of BIFs is that excess oxygen created by the photosynthesizing cyanobacteria in shallow oceans was available to combine with iron ions in seawater to produce the dark oxidized iron layers. The source of iron is likely a combination of both ions released from hydrothermal vents as well as input from chemical weathering of mafic-rich crustal rock from land. These distinct chemical sedimentary rocks were globally pervasive until about 1.8 billion years ago (Paleoproterozoic). BIFs decline in the rock record as the amount of oxygen in the atmosphere continued to climb. With more atmospheric oxygen available, iron ions were oxidized in terrestrial settings before they were transported to the oceans.

Eukaryotes: one small step for cells, one giant leap for life

A consistent pattern that we see in the rock record is that life modifies Earth systems, and Earth systems, in turn, modify life. The accumulation of oxygen in the atmosphere due to prokaryotic photosynthesis ultimately created a new resource that life could exploit as a source of energy. Around 1.8 billion years ago (Paleoproterozoic), definitive eukaryotes appear in the fossil record, armed with a new method of making energy for cellular function–aerobic respiration. Aerobic respiration requires oxygen to produce molecules of ATP which can then be stored and metabolized when needed to power a cell.

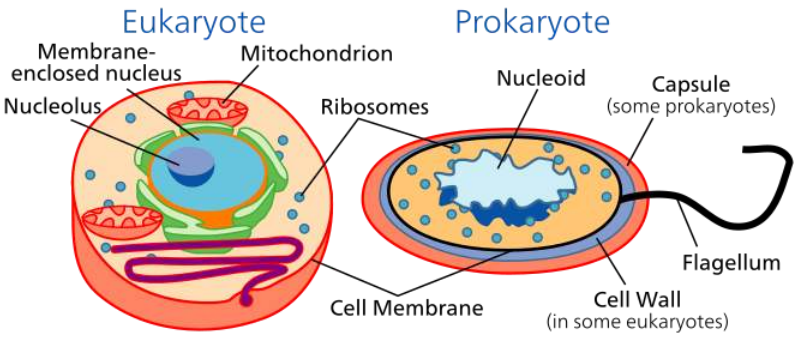


Figure 10.1.8: Prokaryotes versus eukaryotes. (Public Domain image; Science Primer (National Center for Biotechnology Information))

Eukaryotic cells contain a membrane-bound nucleus that houses the cell's DNA and a variety of other organelles that assist with cell activities. In particular, mitochondria are important organelles because they are associated with ATP production within all eukaryotic cells. In plants, chloroplasts are also necessary as the sites where photosynthesis occurs within the cell. Because they contain a variety of organelles, eukaryotic cells are generally one to two orders of magnitude larger than prokaryotic cells (eukaryotes are typically 10 – 100 μm).

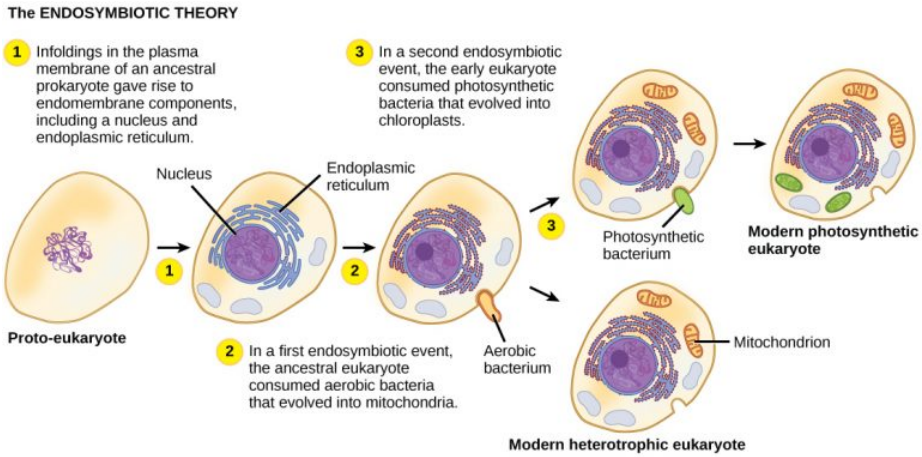


Figure 10.1.9: Schematic of endosymbiotic theory showing the evolution of eukaryotes from prokaryotes. (Image via Creative Commons Attribution-Share Alike 4.0 International license)

So, how did eukaryotic cells first evolve? Current science supports endosymbiosis theory (ET), which states that eukaryotic cells arose due to a mutually beneficial relationship where one prokaryote lives inside another. For example, one prokaryote engulfed another but did not digest it, allowing both cells to persist. The host cell offered protection, and the ingested cell provided energy to allow both cells to function. Over time, the cells co-evolved, and became reliant on each other as a single system: a cell with energy-producing organelles, mitochondria and chloroplasts. ET was first championed by Lynn Margulis in the 1960s, but took decades to be more widely accepted. One of the major lines of evidence in support of ET is that both mitochondria and chloroplasts possess their own DNA, with genetic signatures very similar to modern prokaryotes. Additionally, organisms without mitochondria, like the amoeba *Pelomyxa*, instead host symbiotic bacteria that serve the same function as mitochondria, but are not yet so wedded to their host that they cannot survive on their own.



Figure 10.1.10: Grypania spiralis fossils. (Creative Commons Attribution-Share Alike 3.0 Unported license.)

Today, eukaryotes include a wide variety of single-celled organisms, both with and without shells, like amoeba, foraminifera, coccolithophores, radiolaria, and diatoms. These are called protists. We classify them within the Kingdom Protista. All multicellular organisms are also eukaryotes, including all fungi, plants, and animals (Kingdoms Fungi, Plantae, and Animalia, respectively). The most widely accepted fossils that are interpreted as representing the first eukaryotes are acritarchs, a group of single-celled organisms that appear in the record about 1.8 billion years ago (Paleoproterozoic), well into the prokaryotic-driven Great Oxidation Event. Acritarchs are of unknown affinity, which means there is uncertainty as to where they fit on the tree of life. Acritarchs consist of a spherical sac-shaped cell ranging in size from 1.0 – 1000.0 μm, making them significantly larger than prokaryotes. The oldest forms are most commonly thought to be related to marine algae or dinoflagellates (a protist group). Also around 1.8 billion years ago (Paleoproterozoic), another fossil, *Grypania spiralis*, appears in the fossil record. This filamentous, spiral shaped fossil is commonly considered to be eukaryotic, primarily due to its size, leading some researchers adamantly debate that *Grypania* may also represent one of the first multicellular organisms on Earth.

Out onto the multicellular branches of life: the Ediacaran biota

Given that unicellular organisms had been quite successful for over a billion years, paleobiologists have wondered–what is the evolutionary benefit of a multicellular existence? Many scientists suggested that the typical pressures of natural selection, such as predation, could have played a role. In two interesting modern lab studies of unicellular green algae (Boraas et al., 1998; Herron et al., 2019), algal populations that grew into multicellular structures showed greater survival under predation. Multicellularity is likely an example of convergent evolution, occurring multiple times within different eukaryotic groups to create the diversity of multicellular eukaryotes that exist on Earth today.



Another potential evolutionary benefit shared by most multicellular organisms is the ability to reproduce sexually. Prokaryotes typically reproduce asexually by budding or fission, but the resulting organism is a genetic replicate unless a random mutation occurs. In the case of sexual reproduction, gametes (*e.g.*, sperm and egg cells) recombine DNA to make genetically distinct offspring. Sexual reproduction does have certain disadvantages. For example, it takes a significantly longer time to reproduce and grow population sizes when only half the population (*i.e.*, females) are capable of reproduction. However, the evolutionary trade-off from bringing together genes from two individuals can more rapidly produce advantageous traits and delete harmful mutations from a population. For example, the Red Queen hypothesis of sexual reproduction proposes that asexual organisms are more susceptible to parasitism due to their cloned DNA. Recombination of DNA through sex would allow an organism the genetic variability needed to defend against co-evolving parasites in the environment (Jokela et al., 2009). Given the incredible diversity of life on Earth, clearly both asexual and sexual reproduction are successful means to fully take advantages of Earth's resources.

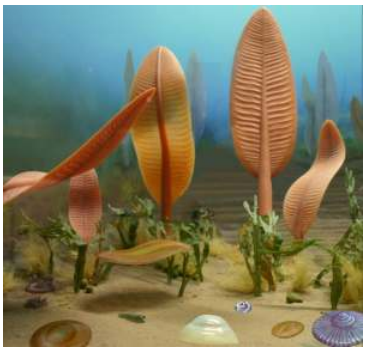


Figure 10.1.11: Artistic rendering of Ediacaran shallow seas. (Creative Commons Attribution-Share Alike 4.0 International license.)

As with other forms of early life, the origins of multicellular life are difficult to pinpoint given the lack of hard parts available for preservation. The fossil record of the Ediacaran Period (635-541 million years ago) at the end of Neoproterozoic best captures this next major innovation of life. The Ediacaran biota are a diverse assemblage of complex, macroscopic multicellular body and trace fossils. This significant increase in diversity follows an extensive period of glaciation (Snowball Earth) during the Cryogenian (Neoproterozoic). Prokaryotes likely contributed to eukaryotic success; this biota would have originated during a time when oxygen levels in the surface oceans were still increasing due to photosynthesizing cyanobacteria. Recent geochemical research (Bekker et al., 2017) also indicates bacteria levels were high in the Ediacaran seas, creating higher concentrations of dissolved organic matter as a food resource for these organisms. The warmer, oxygen- and food-rich seas of the Ediacaran may have set the conditions necessary for multicellular life to globally flourish.

Ediacaran fossils have been found on every continent except Antarctica. They are typically preserved within sandstones of shallow marine settings, like the famous deposits from the White Sea Coast of Russia, the Flinders Range of Australia, and Namibia. In a few places, however, fossils of the biota have been found in finer-grained sediments of deeper waters, such as the ash beds at Mistaken Point in southeastern Newfoundland. The biota primarily consists of impressions (both casts and external molds) of soft-bodied organisms that do not usually preserve well in coarser, sandy sediments, and many are even found underlying turbidity

current deposits, effectively smothered by the rapid burial. Some research suggests that the fine texture of pervasive microbial mats would aid in preservation of the fossil impressions.



Figure 10.1.12: Dickinsonia tenuis fossil. (Creative Commons Attribution-Share Alike 4.0 International license.)

The fossils reveal a few common shapes (also known as morphologies). Many are described as disc-shaped (*Cyclomedusa*, *Mawsonites*). Others, known as rangeomorphs, have a frond-like shape (*Rangea*, *Charnia*). Still others exhibit the first clear examples of bilateral symmetry (*Kimberella*). There are even some odd forms that have an apparent trilateral symmetry (*Tribrachidium*), which is unlike anything alive today. The coarse sediment in which these fossils are preserved has hidden much of their anatomic detail, thus where these organisms should be placed on the tree of life has been widely debated. Some researchers suggest the Ediacaran biota represents basal members of some certain invertebrate animal groups that take off in diversity during the subsequent Cambrian Explosion, such as cnidarians and mollusks. Others have argued for identification of various Ediacaran forms as giant protists, algae, worms, fungi, or in a phylum of their own, the Vendozoa. Recent chemical analyses of *Dickinsonia* fossils showed evidence of cholesteroids, molecular fossils of the compound cholesterol, which is only found in animal cells (Bobrovskiy et al., 2018). When compared to molecular analyses of the surrounding sediments, only stigmasteroids, molecular fossils indicative of the green algae of the microbial mats was found. This study supports the interpretation that at least some of the Ediacaran biota are some of the first true animals on Earth.



While many ichnotaxa from this time have been re-evaluated in recent years to be sedimentary structures or microbially related, most Ediacaran burrows are simple, horizontal trails and burrows, or rasping traces made in association with microbial mats. Rasping is a mode of feeding by scratching at a surface. Snails and sea urchins eat algae this way today. Toward the end of the Ediacaran, burrows begin to increase in complexity, with a shift to more vertical and branching shapes, suggesting animal motility was on the rise.

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10.2: Phanerozoic Diversity of Life (According to the Hard Parts)

Before and after: the exoskeleton appears

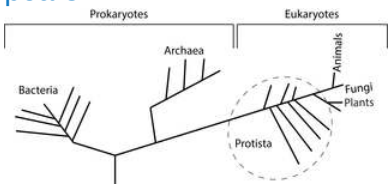


Figure 10.2.1: Eukarya is the Domain that includes most familiar organisms, including humans.

One of the most fascinating aspects of Earth history is the diversification of life through time. There are many ways to graphically represent this diversification. The “tree of life” diagram is one way to broadly represent phylogenetic relationships of the three major Domains of life, including the Archaea, Bacteria, and Eukarya. Within each of these main stems, further distinctions can be made as organisms are grouped with others that share similar physical traits, as discussed in the cladistics section of this textbook. However, one of the factors that is missing from the tree of life diagram is the aspect of time. As we have seen in the prior sections, it took a long time for life to separate into the three Domains, and even longer for Eukarya to split into the four Kingdoms: Protista, Fungi, Plantae, and Animalia. The next major innovation of life is so revolutionary that geoscientists literally draw a line in the sand(stone)–the Precambrian and everything after. During the Cambrian Period, which kicks off the Phanerozoic Eon, shelled organisms with an exoskeleton proliferate in the fossil record. This significant uptick in the diversity of marine life is known as the Cambrian Explosion. (Note there are small, shelled animals (metazoans) of uncertain affinities found in rocks of the Late Ediacaran known as the Tommotian fauna, including the genus *Cloudina*.)



Figure 10.2.2: Trilobites, like modern arthropods, molted their exoskeletons as they grew. (Creative Commons Attribution 2.0 Generic license.)

Why did animals develop hard, mineralized exoskeletons in abundance during the Cambrian? It was likely that shells were the evolutionary answer to a biological need. Increased animal diversity at this time would lead to greater competition for resources, and an external shell offers several evolutionary advantages: it offers protection from physical and chemical variations in the environment, such as changes in water temperature and salinity; it offers protection from biological predators; and, it allows muscles additional mechanical leverage during locomotion, making it easier to burrow, for example. Being able to withstand seasonal shifts in currents, ward off predators, or exploit food sources below the seafloor would increase chances of survival. Interestingly, the boundary between the Proterozoic and Phanerozoic is marked by a shift in seawater chemistry that impacted carbonate mineral deposition in the oceans (Wood et al., 2017). During the Ediacaran (Neoproterozoic), dolomite ($CaMg(CO_3)_2$) was the dominant carbonate deposited in the oceans, but increased erosion of continental crustal rocks and chemical weathering led to an influx of calcium ions to the seas during the early stages of the Cambrian. Higher calcium levels allowed for easier precipitation of aragonite and calcite, the two polymorphs of $CaCO_3$ that most marine organisms use to create their shells. As the Cambrian and Ordovician progressed, a wide array of body plans developed among representatives of most of today’s major invertebrate phyla, including sponges, corals, brachiopods, bryozoans, arthropods, annelids, mollusks, and echinoderms.

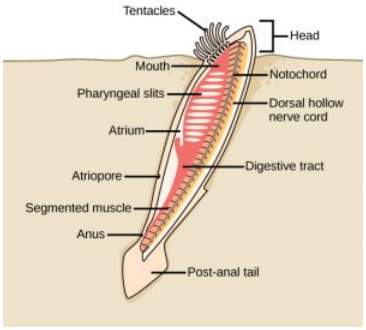


Figure 10.2.3: Modern lancelet showing position of notochord along dorsal side of body. (Creative Commons Attribution 4.0 International license.)

Internal skeletons, or endoskeletons, also have their origins in the Cambrian Period. The first endoskeletons were not mineralized bone, but rather, strong, flexible tissue known as cartilage. This is the same material that makes up shark skeletons and your own ears and nose today. Some of the first vertebrates are fossils from the early Cambrian Chengjiang locale in China. These vertebrates were marine organisms, eel-like in shape, with elongate, tapered bodies that possess a notochord and simple vertebrae. A notochord is a skeletal structure shared by all members of the phylum Chordata at some stage in their life cycle. The notochord helps to support muscles and define characteristic, mirroring, bilateral symmetry on either side of the chordate body's midline. For many chordates, additional strength, flexibility and support are provided with the bones of the vertebral column. Note the spinal cord found in vertebrates is part of the nervous system, which is a completely different set of tissues from the notochord and vertebrae, which are part of the skeletal system.

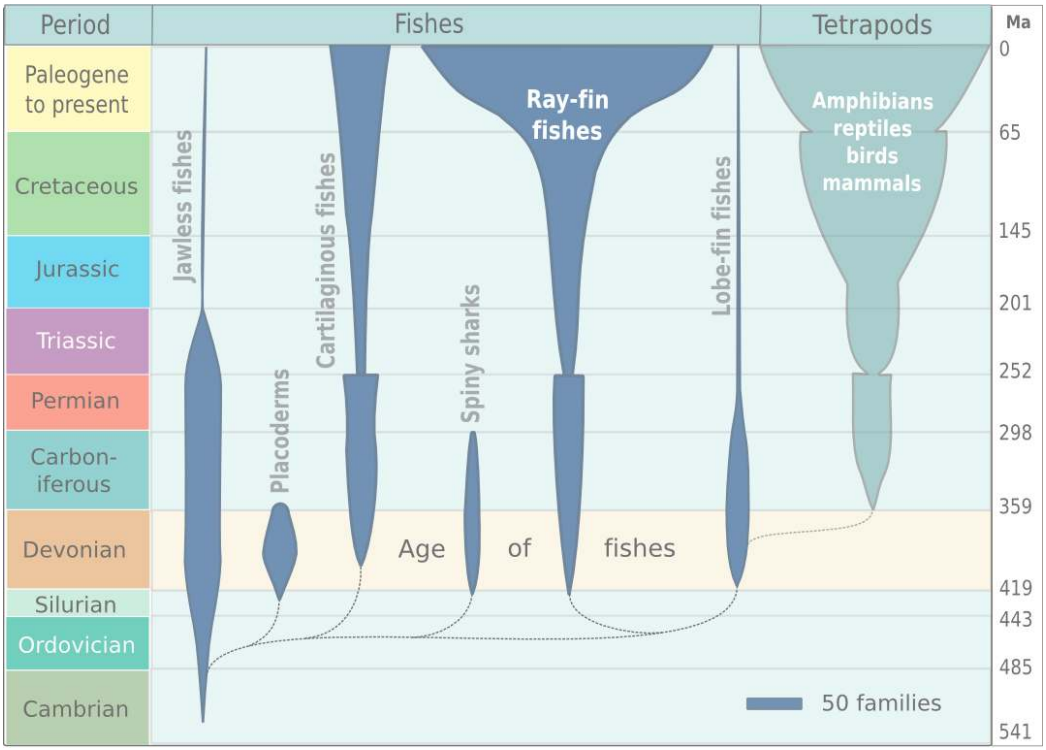


Figure 10.2.4: Spindle diagram of fish diversity through time. Vertical length of shapes indicate geologic range of species while the width represents diversity. (Creative Commons Attribution-Share Alike 3.0 Unported license.)

Throughout the Cambrian and into the Silurian, jawless fish (agnathans) and armored fish (ostracoderms) diversified. Interestingly, while the development of body external plates of the ostracoderms has traditionally been thought to be a classic response of protection to predation, there is also the added benefit of ion storage. In particular, bone stores phosphorous which is used by muscles for quick bursts of energy, like when swimming quickly away from predators as well. Jawed fish, including cartilaginous fish like sharks and rays (chondrichthyes), and bony fish (osteichthyes) appeared during the Silurian and radiated during the Devonian.

Here come the plants: landward, ho!

As human animals, we tend to focus on the parts of Earth history that tell our story; however, animals are not the only life form that moved to land during the Paleozoic, and in fact, plants beat our ancestors to the punch.





Figure 10.2.5: Moss (a bryophyte) with spore releasing structures. (Creative Commons Attribution-Share Alike 4.0 International license.)

The first multicellular plants are marine green and red algaes that were likely derived from photosynthetic cyanobacteria during the Neoproterozoic. The first land plants evolved around the early Ordovician, when fossil evidence of the first bryophytes are found. Bryophytes include plants like mosses, hornworts, and liverworts, that are commonly found in moist areas. They do not rely on a root system for uptake of water and nutrients, rather taking these in via their leaves, and reproduce via spores. Later in the Ordovician, the first true vascular plants (trachaeophytes) evolved, and by the Devonian, this group looked like what you think of when you think of a plant–leafy and green, possessing a lignin-reinforced stem and roots that conduct water and nutrients through their tissues, and using seeds for reproduction.

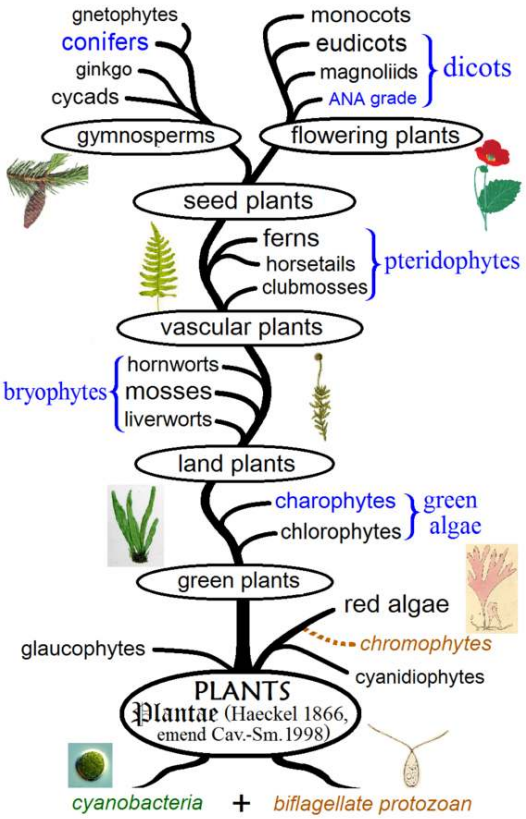


Figure 10.2.6: Diagram showing the phylogeny of major plant groups. (Creative Commons Attribution-Share Alike 4.0 International license.)

Imagine land on Earth before plants. It would have been lots of bare rock, very much like deserts or areas after volcanic eruptions today. The first liverworts and mosses to colonize land would have begun to significantly impact physical and chemical weathering of terrestrial landscapes. Plants can assist in the development of soils from bedrock as they extract nutrients, retain water, and produce organic matter. Later, root systems of vascular plants could wedge into cracks in rock in search of water, physically creating more surface area for physical and chemical weathering to occur. However, given enough time and spread of populations, plants and their roots can also add stability to slopes, lead to the development of new ecological environments (forests, swamps), and provide new resources (leaves, seeds) for other life forms to put to evolutionary use.

Next up: invertebrate animals invade terrestrial environments

Invertebrate groups diversified significantly throughout the Paleozoic oceans (see diversity section below), increasing the competition for resources. As time went by, there was an increase in ecological tiering above and below the seafloor, as animals evolved forms that allow for deeper burrowing and higher extension away from the sediment-water interface. It was only a matter of time until life found a way to exploit the vast expanse of geographic space on land. This transition to land is thought to have occurred during the Devonian period, when global changes in oxygen levels in the ocean were in decline. This dysoxia was in part driven by excess runoff of organic matter from land due to dying plants, which triggered large scale algal and other plankton blooms in the oceans. An abundance of plankton uses up dissolved oxygen in the oceans, leaving less behind for the larger ocean dwellers. Today, “dead zones” develop at the mouths of modern rivers where concentrations of nutrients from over-fertilization of farms and other industrial processes create plankton blooms. This phenomenon ultimately may have been connected to the mass extinctions at the end of the Devonian.



Figure 10.2.7: Modern horseshoe crab showing specialized appendages and book gills. (Creative Commons CC0 1.0 Universal Public Domain Dedication)

It is important to recognize the transition to land by complex organisms was not instantaneous. Aquatic invertebrates would have had an advantage due to their exoskeletons, which allow for protection and moisture retention as these organisms adapted to freshwater ecosystems, like rivers and lakes, and eventually land. It is not surprising then that arthropods are the first aquatic invertebrate group to venture into the terrestrial realm. Out of all of the invertebrate phyla existing at the time, arthropods had the body plan with the greatest evolutionary flexibility–in particular, lots of segmented appendages that were highly specialized for different types of movement such as walking, burrowing, swimming, and feeding. Aquatic arthropods take in oxygen via book gills (a structure that looks similar to folded or overlapping sheets of paper) which are easily visible on horseshoe crabs you might find along the East Coast of the United States today. The first land arthropods would have developed internal book lungs from these structures, which we see in modern arachnids, like spiders and scorpions. Around the Late Silurian to Early Devonian, terrestrial fossils of tiny spider-like arthropods often found in association with plant fossils, scorpions, and myriapods (clade including millipedes, centipedes, eventually insects) have been found. In time, some groups of arthropods, including insects developed tracheae, a system of tube-like structures that directly deliver oxygen from external pores in the exoskeleton to tissues of the body. And so, the rest of us animals have come to live in a bug's world.



Figure 10.2.8: Large fossilized Meganeura dragonfly. (Creative Commons Attribution-Share Alike 3.0 Unported, 2.5 Generic, 2.0 Generic and 1.0 Generic license.)

Coevolution among insects and with other animals, coevolution with plants, adaptation to freshwater, and variable predation in larval versus adult forms are just some of the means by which insects have spread across the terrestrial realm. Perhaps most importantly though was the adaptation to the sky–arthropods are the only phyla of invertebrates to develop wings and flight. The first flying insects show up in the fossil record of the Carboniferous. While there is still debate in scientific circles, it is generally assumed wings represent another modification of appendages that were given a molecular and mechanical boost from high oxygen levels in muscles achieved via tracheae. The development of flight also corresponds to increases in atmospheric oxygen on a global level during the Carboniferous as vascular plants diversified, grew into large populations, and pumped out oxygen as a byproduct of photosynthesis. This rapid diversification of insects into all available ecological niches is known as an adaptive radiation, and this pattern is repeated by other terrestrial animal groups later in geologic time. Today, there are about one million described insect species, representing almost 75% of all known animals.

Vertebrates take their first steps

While arthropods are crawling and scurrying their way onto land, vertebrates were keeping an eye on their progress–literally! An interesting avenue of recent research suggests that vertebrates begin to show critical increases in eye size and a shift in eye position

from the side to the top of the head during the interval leading up to their transition to land (MacIver et al., 2017). These researchers hypothesize that these changes in eyesight would have yielded better vision over longer distances, allowing these animals a better sightline to prey on land.



Figure 10.2.9: Comparison between the fins of lobe-finned fishes (Sarcopterygii) and legs of early tetrapods. 1. Tiktaalik. 2. Panderichthys. 3. Eusthenopteron. 4. Acanthostega.5. Ichthyostega (hindleg). (Public domain image.)

Other structural changes were also required to the vertebrate body plan to produce success on land. The first land-based animals are known collectively as tetrapods, referring to those animals moving about on four feet. Without the buoyancy provided by water, the bones of tetrapods would need to be more robust to provide more structural support to the body and allow for larger muscles to develop to aid in locomotion. Bony, lobe-finned fish (those in the sarcopterygian class) possess the basic vertebrate body plan in their appendages, and this same

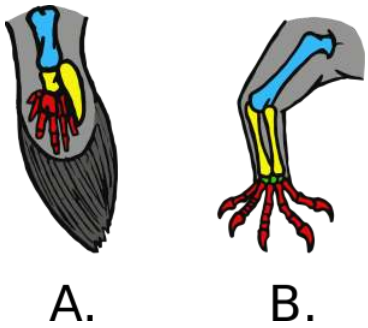


Figure 10.2.10: Comparison between the fins of A., Crossopterygiian fishes and the legs of B., tetrapods. Bones which correspond to each other have the same color. (Public domain image.)

musculoskeletal template in their ancestors would have allowed for the evolution of muscular arms and legs. Specifically, their fins exhibit the same sequence of bones as all other tetrapod limbs: one bone at the point of body attachment, followed by a joint, two bones, another joint, then lots of little bones. Other changes are not particularly well-fossilized, but would also have been key to success on land: development of a three-chambered heart and more complex lungs to extract oxygen from air, continual modification to auditory systems to allow for better hearing in air, and development of more waterproof skin to retain tissue moisture.

Paleontologists who study the colonization of land are commonly looking for what are known as transitional fossils that show these gradual changes over time. One of the most famous transitional fossils of the Late Devonian is *Tiktaalik*, a fish discovered by a team of researchers headed by Neil Shubin and Ted Daeschler.



This fossil shows clear traits of being fish-like while also showing some of the necessary adaptations to being able to move about without the buoyancy of water. *Tiktaalik* possessed stronger limbs including hip-like structures on the hind limbs, a neck that allowed for swiveling motions, and large eye sockets on the top of the head. Interestingly, an increase in size and shift in the position of spiracles (skull openings that allow air to pass across a fish's gills) toward the back and side of the skull in several transitional fossils (*Eusthenopteron*, *Icthyostega*, and *Tiktaalik*) suggest a positioning more in line with the location of lungs.

Tetrapods and plants: conquering land, sea, and air



Figure 10.2.11: Skeleton of Ichthyostega, one of the earliest amphibians. (Creative Commons Attribution-Share Alike 3.0 Unported license)

The first tetrapods on land during the Devonian quickly evolve characteristics that are reminiscent of today's amphibians, although modern amphibian groups generally evolved during the Triassic. While these animals began to spread into various environments on the continents as adults, there would be some limitations due to key aspects of their life cycle–amphibians, like fish, require water for reproduction. The eggs of these animals must be laid in water because they do not have any sort of protective outer coating to protect the egg from drying out. Additionally, early life stages after hatching (for example, a tadpole stage in frogs) require water until the organism develops enough to survive on land.

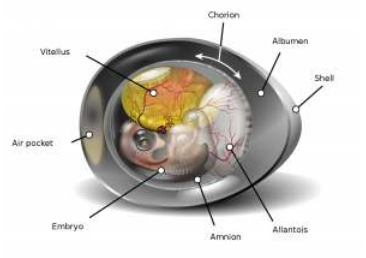


Figure 10.2.12: Modern chicken egg showing structures of amniotic egg. (Creative Commons Attribution-Share Alike 4.0 International license.)

During the Carboniferous, a major innovation arose in tetrapod groups–the amniotic egg. An amniotic egg protects a developing embryo and its required nourishment (yolk sac) within several protective layers. These layers provide protection from environmental elements, and also allow for the transmission of gases needed for respiration and a place to store metabolic wastes. Whether the outermost shell layer is leathery or more hardened, this change to eggs allowed tetrapods to spread out into different inland environments, further away from permanent water sources. Amniotes, as a clade, are represented by reptiles (including dinosaurs and birds) and mammals.

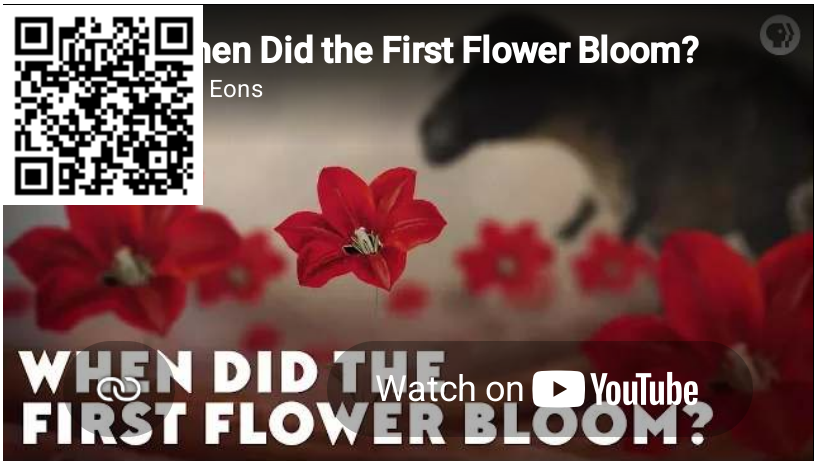
This ability to exploit more terrestrial environments allowed amniote groups to diversify rapidly and persist for millions of years, with natural selection aided by the constant shift of tectonic plates and resulting climate patterns. During the Late Carboniferous, two distinct lineages of tetrapods arose–the sauropsids and synapsids. Within the sauropsid clade, an enormous diversity of body forms and life habits arose, particularly after the end-Permian extinction cleared out a variety of ecological niches and allowed reptiles to undergo an adaptive radiation. This lineage leads to reptilian groups like crocodilians, turtles, snakes, dinosaurs, and dinosaurs' cousins, birds. The end-Cretaceous mass extinction wiped out a significant number of the reptilian groups, effectively wiping the slate clean for emerging mammalian groups to repopulate ecological niches on land, air, and sea.



Figure 10.2.13: Cynognathus, a cynodont from the Triassic of South Africa, pencil drawing by N. Tamura. (Creative Commons Attribution-Share Alike 3.0 Unported license.)

The Late Carboniferous synapsids began with a reptilian-like, low-slung body plan, and a distinctive single opening in the skull behind the eye socket. Many of the larger early members of the synapsid group such as *Dimetrodon* and *Cynognathus* were greatly affected by the end-Permian and end-Triassic mass extinctions. Eventually, changes within this clade led to the various groups of mammals we see today, including monotremes (egg-laying mammals, like platypus), marsupials (pouched mammals, like koalas and kangaroos) and placental mammals (most mammals, including such diverse groups as cows, whales, cats, and humans).

While vertebrate groups on land were evolving throughout the Mesozoic, so were vascular plants. The ability to reproduce via seeds had developed during the Devonian among gymnosperms (seed ferns, and eventually conifers, cycads, and ginkgos), but it was millions of years later, during the Cretaceous, when flowering plants (angiosperms) evolved. Angiosperms are the most common plants on Earth today and includes grasses, common trees like oaks and maples, herbs, and all of our domesticated crops, such as corn, rice, and soybeans.



Gymnosperms primarily reproduce via wind-blown pollen released from cones. Flowers on angiosperms are reproductive structures that allow for pollen distribution by organisms (e.g., insects like bees) and are often colorful, showy features that attract organisms for gamete dispersal. Gymnosperm seeds are literally “naked seeds”–the seed is not encased in an ovary. However, on angiosperms, (“vessel seeds”) the plant’s seeds are encased in ovaries we know as fruit. So the more fruit that is produced and eaten, the more seeds are dispersed via animal dung, continuing the plant population, and potentially altering the plant’s geographic distribution. These tightly woven relationships between plants, pollinators, and dispersers represent some of the most elegant examples of coevolution throughout geologic time.

The interactions of living organisms clearly play a key role in how Earth has changed during its 4.6 billion year history. Likewise, interactions with the atmosphere, hydrosphere, and geosphere have shaped and pruned the tree of life. As humans, we are part of the unique, weird, beautiful, fascinating,opportunistic, innovative, and simply wonderful story of life on Earth.

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CHAPTER OVERVIEW

11: Mass Extinctions

11.1: Adaptation's dark twin

11.2: "The Big Five"

11.3: Other 'Mass' Extinctions

11.4: Further reading

11.S: Summary

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11.1: Adaptation's dark twin

Natural selection and related processes control the evolution of organic life. Elsewhere in this text, we have explored how random mutations in the genetic material inside organisms' cells produces variations in their size, shape, and abilities, some of which will provide a selective advantage over other individuals. In those circumstances, the mutated individual is likely to enjoy enhanced survival, and thus enhanced reproduction, which passes the gene for the new trait on to the next generation. The new trait is an **adaptation**. Through repeated cycles of copying and culling, the adaptive gene spreads through the population and comes to dominate. If the changes are sufficiently pronounced, we call the organism by a different name; we infer that **speciation** (or "origination") has occurred.

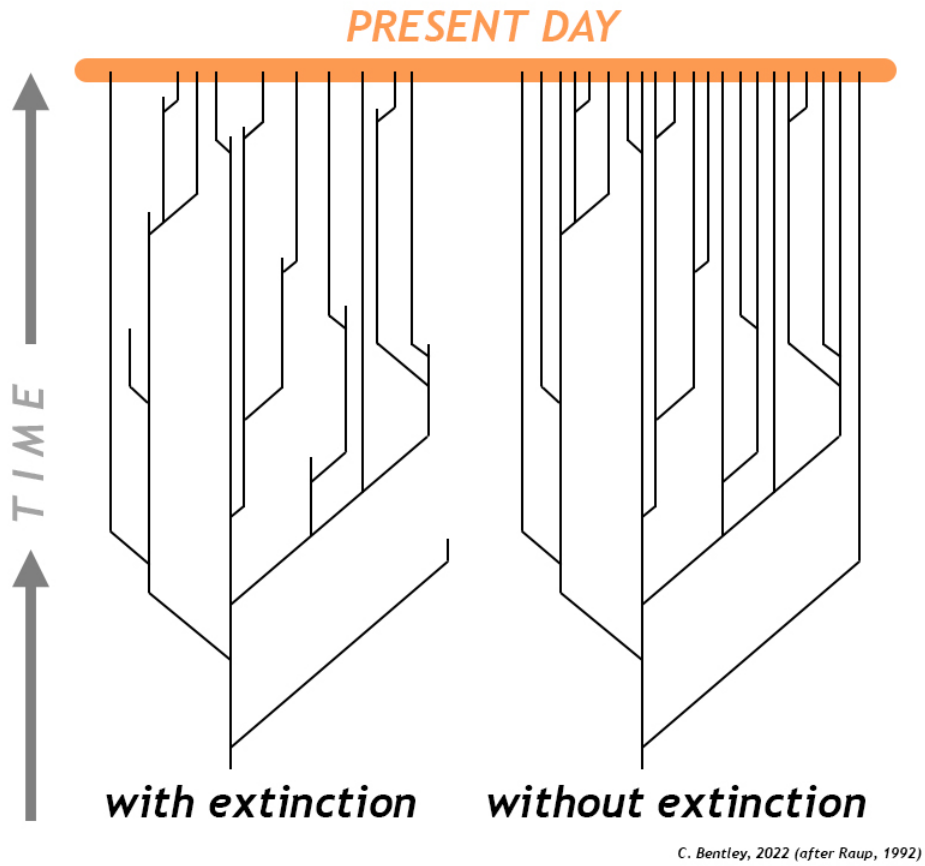


Figure 11.1.1: Theoretical comparison between evolution "tree of life" diagrams with and without the "pruning" effect of extinction. (Callan Bentley redrawing of an original diagram by David Raup.)

However, this narrative emphasizes the *spread* of a new trait (and the gene that codes for it). Equally critical in the history of life is the *cancellation* of some old traits (and the genes that code for them). Without extinction, there would likely be insufficient ecological "space" available for new species. The pattern we see in the fossil record is not one of continuous diversification with new species being added, but none ever removed. Instead, the average species lasts a few million years, and then vanishes forever from the face of the planet. It goes **extinct**. A handful of species have persisted with relatively minimal change (sometimes we call them "living fossils"), but most lineages show changes through time, gaining some species but losing others. As you look around the world today and (rightly) marvel at the diversity of organisms who share your planet with you, it is sobering to realize that Earth's deep time hosted many, many more — species that got snuffed out somewhere along the way. Estimates are that 99.9% of all species that ever existed on Earth no longer exist — the vast majority have gone extinct. Most of these occur as part of the normal "background extinction" rate, a regular coming-and-going of species due to habitat loss, changing physical environments, and biologic pressures over time.

Periodically, entire lineages will go extinct. The trilobites, for instance, were exceptionally diverse and widespread, but they never made it out of Paleozoic Era. Similarly, sauropod dinosaurs are a Mesozoic-only group. When multiple lines of descent all

terminate at the same moment in geologic time, we call it a **mass extinction**. Mass extinctions potentially imply a common cause for the synchronous extinctions of many unrelated groups. We study mass extinctions because they are fascinating and dramatic, but also because we worry about our own future. The study of mass extinctions demonstrate that Earth's biosphere can periodically be wrecked, and then healed, but healed with a new suite of organisms. Put another way, every time there is a mass extinction episode, it is followed by a time when the survivors diversify and adapt, taking advantage of the lack of competition. This is an **adaptive radiation**. For instance, after the reptilian dinosaurs, mosasaurs, and pterosaurs went extinct at the end of the Cretaceous, mammals underwent an adaptive radiation, diversifying to fill all sorts of ecological niches on land, sea, and even in the air.

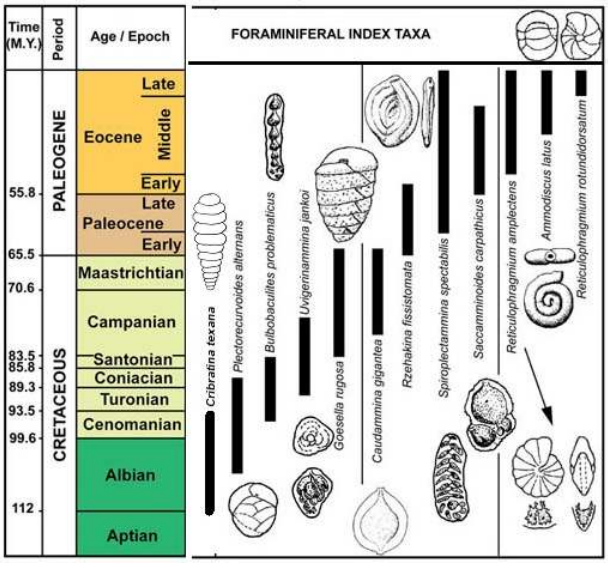


Figure 11.1.2: Geologic ranges for select index foraminifera from the Cretaceous and Paleogene. (For educational purposes from the Geologic Timescale Foundation. Modified by Shelley Jaye.)

We know about mass extinctions because of careful study of the appearance of fossil forms in the stratigraphic record. Each species has a **geologic range** defined by its first appearance (oldest stratum in which it is known) and its last appearance (youngest stratum in which it is known). Like any scientific conclusion, these ranges are always subject to change with new data (such as in the famous case of the coelacanth), but for the most part the ranges are well established, having stood repeated testing over the past two centuries of paleontological documentation.

Mass extinctions lead to declines in both diversity and abundance of organisms, are typically global scale events, and cut across taxonomic groups affecting organisms in a wide array of environmental habitats.

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11.2: "The Big Five"

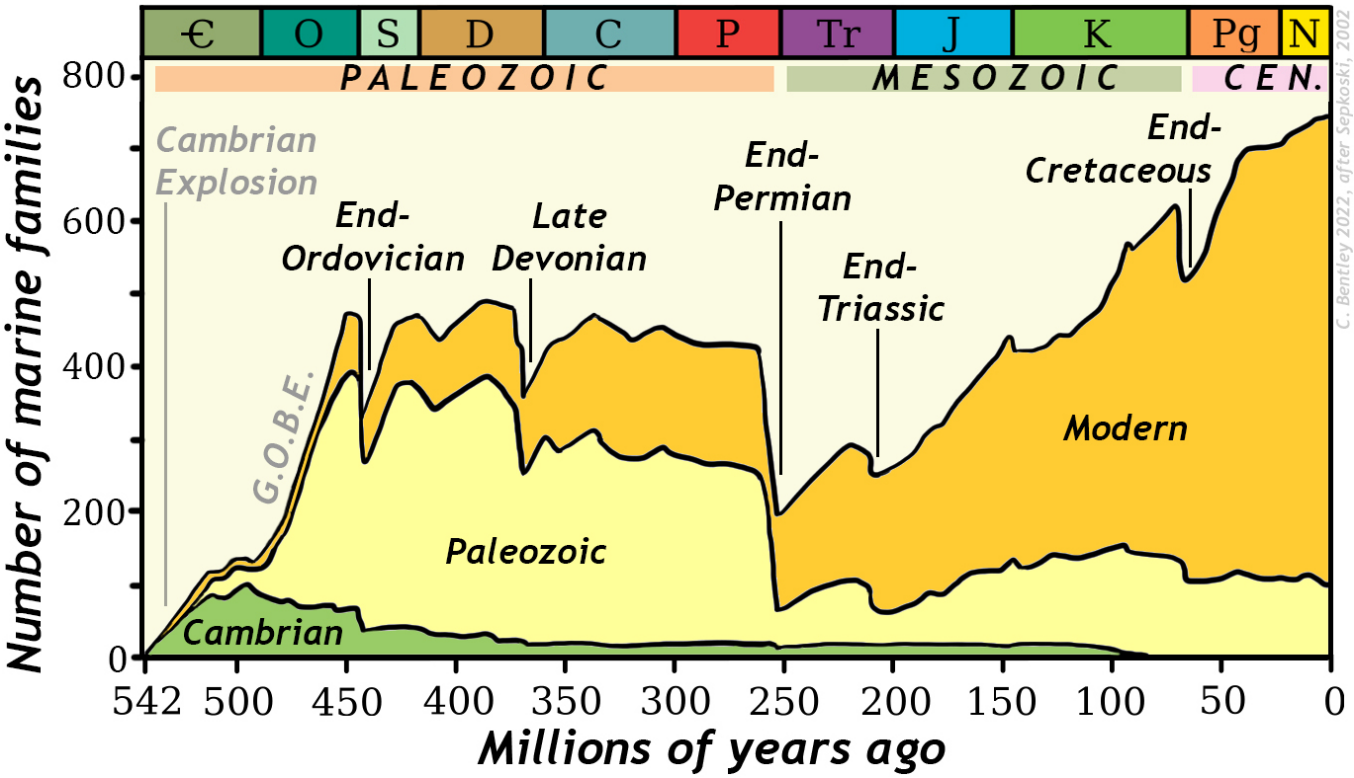


Figure \(\PageIndex{1}\): Marine animal family diversity over the Phanerozoic Eon of geologic time. The “Big Five” mass extinctions are indicated, as well as the Cambrian Explosion and Great Ordovician Biodiversification Event. (Graph by Callan Bentley (2022), after Sepkoski (1984, 2002).)

Traditionally, historical geologists have recognized five major episodes of mass extinction from the fossil record. They show as big spikes on the graph above. Plotted a different way, they show as sharp declines on the diversity through time graph. Chronologically, they are the end-Ordovician extinction, the late Devonian, the end-Permian (which closes the Paleozoic Era), the end-Triassic, and the end-Cretaceous (which closes the Mesozoic Era). Each has its own idiosyncratic circumstances, cast of characters, and potentially unique causes for so much dying. Let us now recap each of them.

Extinction is going on all the time. In a way, it’s arbitrary to select a handful for consideration, both because of the incomplete nature of the fossil record (such as microbial turnover during the Great Oxygenation Event) and background extinctions (such as the spikes you see in the Cambrian above, which seem big because the total known number of fossil organisms was itself low).

End-Ordovician extinction

At the end of the Ordovician period, the only animals on Earth were invertebrates living in the ocean. Certain varieties of mollusks, trilobites, graptolites, eurypterids, brachiopods, conodonts, corals, echinoderms, and other groups all went extinct about 443 Ma. These were all key members of the Paleozoic Fauna. Of the species paleontologists have documented, about 52% of them disappeared at the Ordovician/Silurian boundary, never to return. It is worth emphasizing that — uniquely among the Big Five — the end-Ordovician extinction was entirely limited to marine organisms, since land had yet to be colonized at that time.

The Ordovician/Silurian boundary was defined by Charles Lapworth with the stratotype at Dob’s Linn, Scotland, which is still used as a GSSP today. Below are some of the graptolites Lapworth used as index fossils to distinguish between the two periods in the black shale there:



(Callan Bentley GIGAmacro)

After the mass extinction was over, it took 50 million years for Earth's oceans to recover their former levels of diversity.

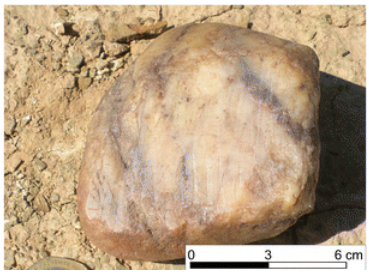


Figure \(\PageIndex{2}\): Faceted and striated clast extracted from Ordovician strata in Arabia. (Modified from Figure 3 of Masri (2017).)

The cause of the late Ordovician extinction is inferred to likely be **global cooling**. There is evidence of glaciation during the late Ordovician in the southern supercontinent Gondwana: this is known from sedimentological documentation of tillites bearing faceted, striated clasts, and dropstones in what is today South America, Arabia and Africa). There are also vast areas where the glaciers appear to have bulldozed pre-existing unlithified sediments, deforming them as the ice flowed along. Using isotopic evidence, Finnegan *et al.* (2011) have estimated that global seawater temperatures dropped 5 \(^{\circ}\)C during the latest Ordovician. Five degrees may not sound like much, but because of water's exceptional heat capacity and the huge volume of Earth's seawater, it represents a truly enormous amount of heat energy lost from the world ocean. Glass sponges, a deep-water siliceous sponge-like animal, are well-adapted to cold conditions, and they do well in the aftermath of the end-Ordovician extinction.

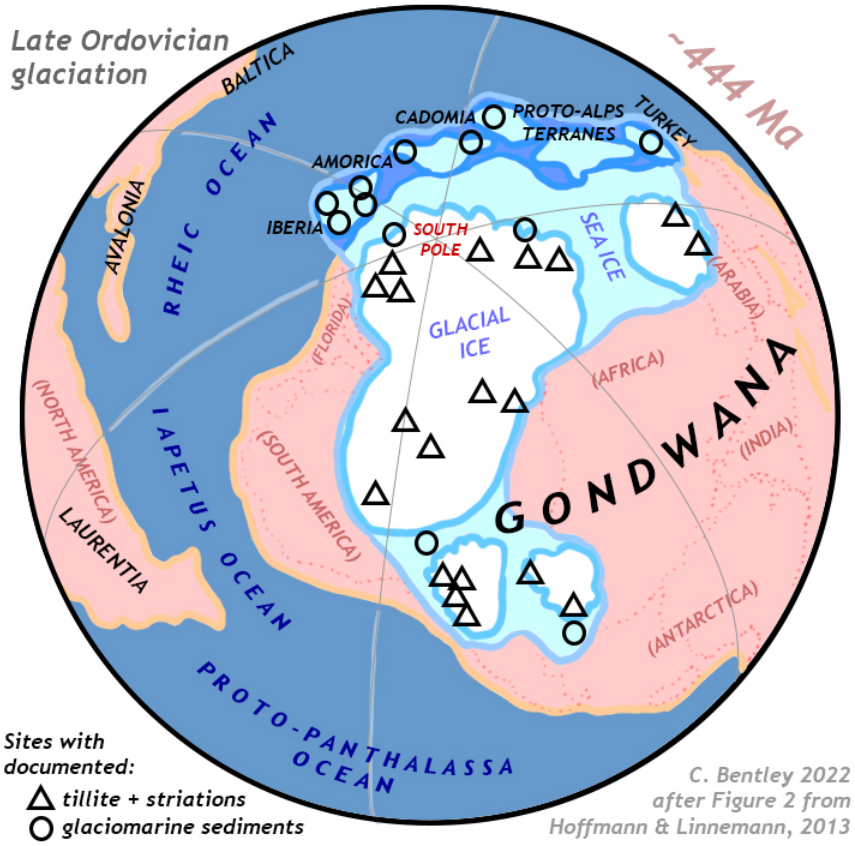


Figure \(\PageIndex{3}\): Map showing the inferred extent of end-Ordovician glaciation in Gondwana. (Map by Callan Bentley, redrawn after an original by Hoffmann and Linnemann, 2013.)

How could a cooling climate trigger a mass extinction? One key variable is a change in the ocean temperature, with negative impacts for those organisms who are specialized for warmer waters. This is particularly acute in the tropical/equatorial regions of the planet's ocean. But another key impact is in the change in sea level that accompanies the build up of glacial ice mass. Glaciers are made of water, and that water has to come from somewhere. As we are currently seeing in modern times, melting glaciers add water to the ocean, which raises sea level. The reverse is also true: at the height of the most recent ice age, sea level was ~120 meters lower, and the shoreline was at the edge of the continental shelf. This is bad news for marine invertebrates, most of which live in the shallow marine realm. Dropping sea level by more than 330 feet reduces the amount of habitat they have in which to dwell, and increases competition between individuals.

What caused the cooling climate? It appears to be a coincidence of several events: (1) the position of Gondwana over the south polar region, drifting there due to plate tectonics, and (2) a lowering of greenhouse gas levels through some perturbation to the climate. We can break condition "2" down further into a couple of hypotheses: (2a) enhanced silicate weathering by rainwater carbonic acid due to an episode of mountain-building in ancestral North America, the Taconian Orogeny, and (2b) reduced sunlight influx due to stratospheric dust kicked up by a series of meteorite impacts. These aren't necessarily mutually exclusive, of course. Let's briefly examine each.

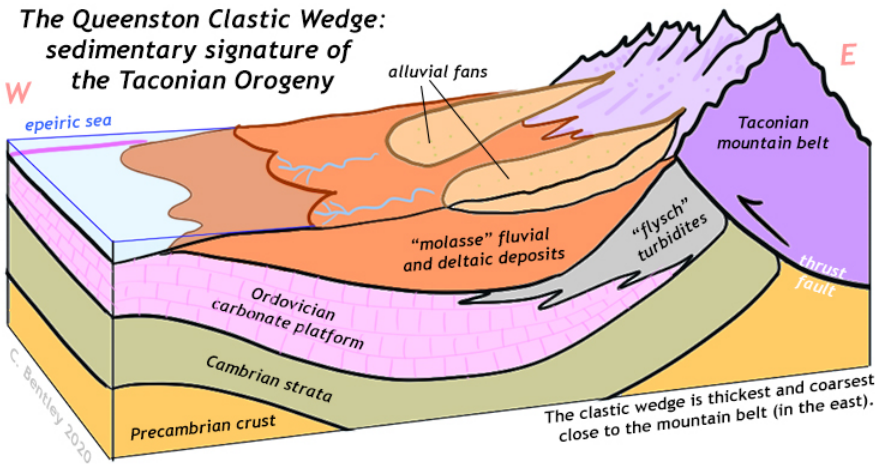


Figure \(\PageIndex{4}\): The Taconian Orogeny shed off a large amount of clastic sediment, reflecting the weathering and erosion of the late Ordovician-aged Taconian Mountain Range. Much of this sediment was clay, which requires carbonic acid to produce, removing \(\ce{CO2}\) from the atmosphere. (Callan Bentley cartoon.)

The Taconian Orogeny, first phase of the three-part suite of Appalachian Orogenies, was caused by the collision of a volcanic island arc in the Iapetus Ocean with the Iapetan margin of ancestral North America. This raised a big chain of mountains that stretched from Maine to Georgia, and shed a tremendous quantity of clastic sediment, including a lot of mud. Mud is made mainly of clay minerals, which are produced through the hydrolysis of feldspars. Mountain belts put a lot of feldspar into the path of chemical weathering, which gobbles up \(\ce{CO2}\) from the atmosphere.

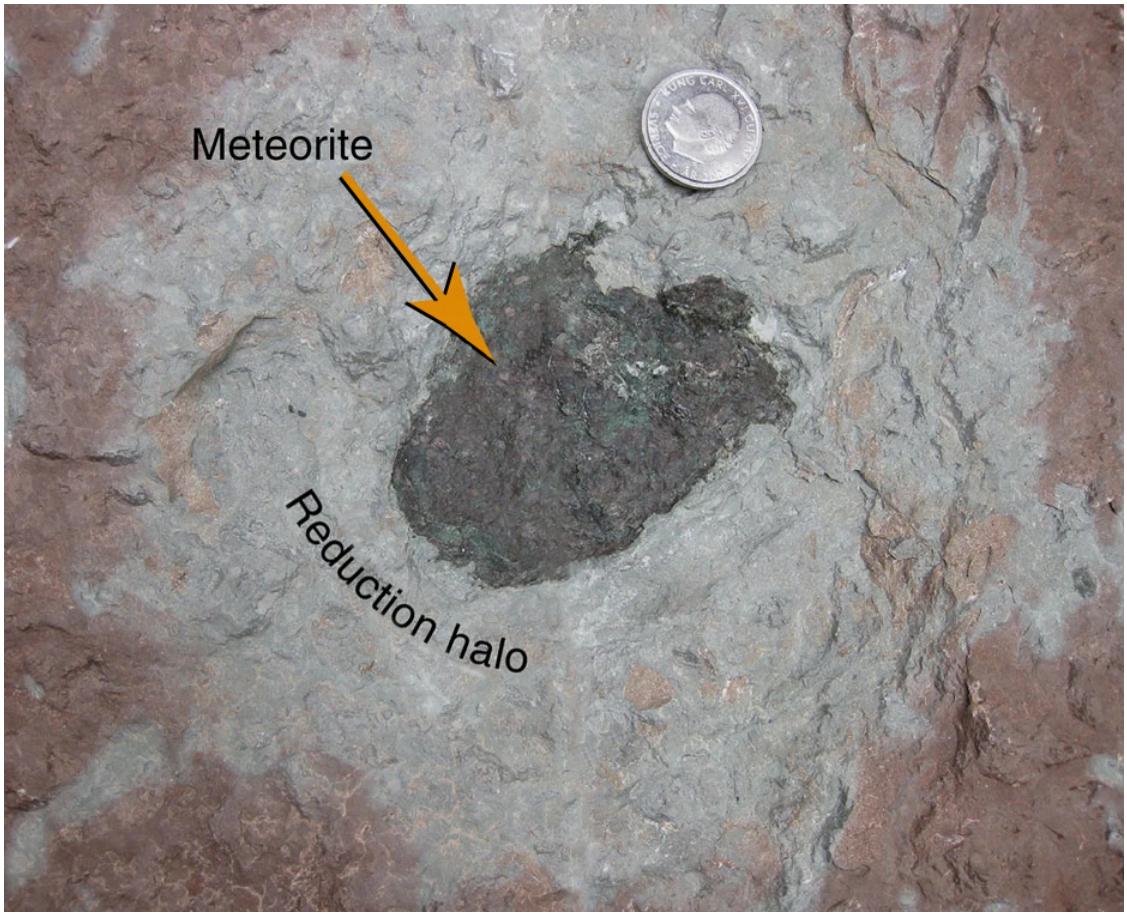


Figure \(\PageIndex{5}\): The Österplana 065 meteorite is preserved as an inclusion in Ordovician limestone, Sweden. (Modified (enhanced!) by CB from an original by Shmitz, et al. 2016.)

Another possibility has an extraterrestrial component. For a long time, it was observed in limestone quarries in Sweden that there were a lot of **meteorites** included in those carbonate strata. More widely and generally, sedimentary rocks of that time contained a lot of chromite grains that were attributed to meteors. These meteorites generally share a common composition; they are classified as "L-chondrites," and are all inferred to come from the breakup of a single chondritic parent body, floating somewhere out in space, at about 468 Ma. There are a lot of them; the profusion of meteorites and chromite grains has been dubbed the Ordovician Meteor Event. The idea is that these meteorites would have kicked up a lot of dust. If that dust reached the stratosphere, where there is no rain or weather, then it could have blocked incoming sunlight. A reduction in the flux of sunlight would have acted to cool the climate, regardless of what greenhouse gases were doing at the time of impact.

So, to summarize: it looks like the first of the Big Five mass extinctions was driven by global cooling.

Late Devonian extinction

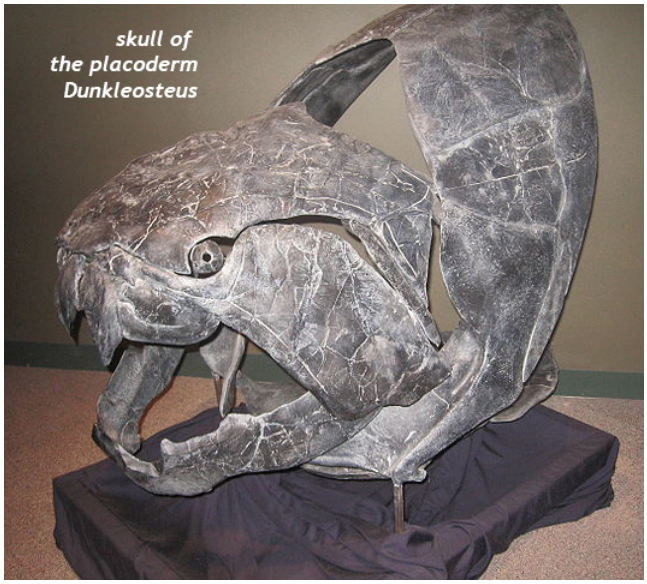


Figure \(\PageIndex{6}\): Dunkleosteus, a predatory placoderm from the Devonian. (Public domain.)

After the relatively brief Silurian period (total duration only 20 million years), the planet entered the Devonian period, sometimes dubbed "the age of fishes." This was a time of great radiation among the fishes, producing the first jawless fishes, sharks, placoderms, the first ray-finned fish, and the lobe-finned fishes, some of whom were almost certainly the ancestors of the humans reading this case study.

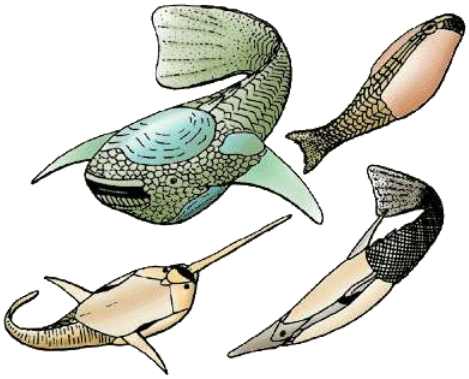


Figure \(\PageIndex{7}\): Jawless fish that lived during the Silurian and Devonian periods: tolypelepids (top right), pteraspidids (bottom right), protopteraspidids (bottom left) and the huge psammosteids (top left), which are the only members of the group to survive the late Devonian mass extinction. (CC-BY; Drawing by Phillipe Janvier (1997).)

Toward the end of the Devonian, things started getting rough for organisms on Earth, but unlike the end-Ordovician, end-Permian, end-Triassic, and end-Cretaceous extinctions, this one took a while. It was spread out over more than 20 million years, maybe as

many as 25 million. It's not the called the "end"-Devonian, for that reason. ***It took a while.*** It got worse, and then a little better, and then worse again. The two biggest die-offs occurred at 374 Ma and 359 Ma.



Figure \(\PageIndex{8}\): Atrypa reticularis, from Devonian strata in Germany. (CC-BY; Dwergenpaartje photo, via Wikimedia.)

By the end of it, though, 99% of reefs had been destroyed, the diverse fish world had been severely cut back, and many marine groups had been severely pruned back or eliminated altogether. Among reef-builders, the tabulate and rugose corals got slammed. Conservative estimates put the carnage at a 40% loss of taxa. The atrypid brachiopods, for instance, were a diverse tropical group of articulate brachiopods that were extinguished. Moving into their shallow-sea habitat after the extinction were silica-skeleton sponges, which otherwise prefer cooler, deeper water.

One of the first clues that things were going awry in the late Devonian is a profusion of black shales. Black shale is rich in organic carbon.



Figure \(\PageIndex{9}\): Watkins Glen, near Ithaca, New York, is a gorge cut into Devonian black shale of the Genesee Group. (CC-BY; Photograph by Barbara Friedman, via Flickr.)

Carbon loves to react with oxygen, if there's any oxygen around to react with. If there is, it becomes \(\ce{CO2}\) gas, and diffuses into the atmosphere (and is thus *__not__* preserved in sedimentary rock). But if muddy sediment is deposited under low-oxygen or anoxic (no oxygen) conditions, any organic carbon will be interred in the sedimentary record, giving its host rock a dark gray or black color.

So why was so much of the ocean anoxic in the late Devonian? The answer may surprise you: ***trees***.



Figure \(\PageIndex{10}\): The roots of this tree are demonstrating the destructive power of root wedging. Though this picture is an anthropogenic “rock” (asphalt), it works on natural rock as well. (Public domain image.)

Yes, you read that right: trees. During the Silurian, the first plants [LINK TO PLANTS CASE STUDY HERE] had colonized wet areas of the shoreline of freshwater areas. Mosses and liverworts can’t do much in terms of churning into their sedimentary substrate, but trees can. During the Devonian, Earth’s land surfaces saw the growth of the first forests. The trees in these forests had roots, and those roots probed down into the ground, helping break up rocks below, and thereby releasing their nutrients at an accelerated rate. Organic acids produced by the trees probably helped accelerated chemical weathering, too.

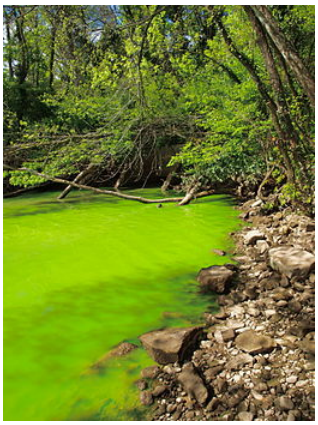


Figure \(\PageIndex{11}\): Eutrophication of the Potomac River adjacent to a wastewater input. (CC-BY; Photo by Alexandr Trubetskoy.)

While the trees doubtless consumed some of these liberated nutrients themselves, others were washed away by streams and rivers, ending up in the oceans. Today, agricultural use of phosphorus and nitrogen fertilizers produces a similar kind of nutrient-rich runoff. When it reaches the ocean, this “vitamin soup” jacks up the rate of reproduction of algae; a process called **eutrophication**. At this point, it all sounds well and good. The issue that makes it deadly enough to cause a mass extinction, though, is that when those supercharged populations of algae die, they die in great numbers. The rotting those billions of dead algae cells consumes all the available oxygen in the water, creating “dead zones” which lack dissolved oxygen. Any fish swimming into such a dead zone must turn around immediately, or asphyxiate. Sadly, this happens regularly today in places like the Gulf of Mexico, where the Mississippi River efficiently delivers nitrogen and phosphorus washed off farm fields in the central United States. Human sewage is another rich source of nutrients; when it enters freshwater lakes or rivers, it can drive this same harmful “boom and bust” cycle that depletes the water of oxygen.

Once the supply of oxygen was exhausted, remaining unrotted dead algae rained down on the seafloor in profusion. Hence: a lot of black (organic-rich) shale. You may have heard of fracking, the process by which humans use high-pressure fluids to shatter organic-rich rocks deep underground, allowing natural gas to flow out. The targets for those fracking operations are mostly Devonian black shales. The Bakken Shale in North Dakota, and the Marcellus Shale in western Pennsylvania are today enticing targets for natural gas exploration because of events that occurred during the late Devonian mass extinction.

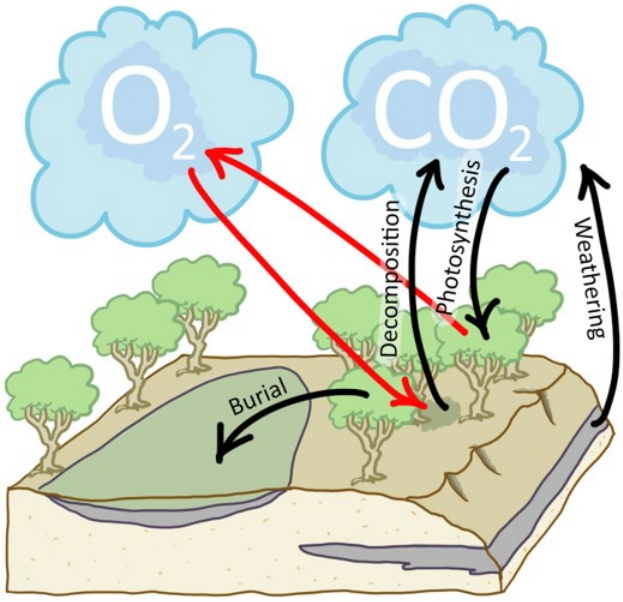


Figure \(\PageIndex{12}\): Carbon and oxygen fluxes between plants, the atmosphere, and the geosphere. Carbon burial during the late Devonian drew down atmospheric \(\ce{CO2}\) levels. (Callan Bentley cartoon.)

So that is part of the story, but producing thousands of cubic miles of black shale has implications for the planet’s climate, too. Removal of all that organic carbon from the biosphere means that the atmosphere will be relatively depleted in \(\ce{CO2}\), which is of course a greenhouse gas, helping warm the climate. When \(\ce{CO2}\) levels are drawn down, there’s less heat retained by the planet, and global cooling can result.

As with the end-Ordovician extinction, there is evidence for a new build-up of glacial ice during the late Devonian. In the Appalachian basin of Maryland, Pennsylvania, and West Virginia, distinctive glaciogenic sedimentary layers can be observed from the late Devonian. Given that paleomagnetic reconstructions put this part of ancestral North America about 30\(^{\circ}\) south of the equator at the time, this implies fairly cold conditions indeed!

As with Snowball Earth, the end-Ordovician glaciation, or the Pleistocene “ice age,” the evidence of glaciation can be seen in faceted, striated clasts, tillites, and dropstones.



Figure \(\PageIndex{13}\): Outsized granite cobble in black shale of the latest Devonian Rockwell Formation, Grant County, West Virginia. (Callan Bentley photo.)



Figure \(\PageIndex{14}\): The “Robinson boulder,” (Logan Branch, Morehead, Kentucky) is a faceted, granite dropstone estimated to weigh 3 tons. It is embedded in the anoxic Cleveland Member of the Ohio Shale (late Devonian). Note upturned layers or “mud drape” along edges of the boulder. Hammer for scale. (From Ettensohn, et al. (2020); reproduced with permission.)

Just to bring the discussion of the late Devonian mass extinction full circle, consider anew the black shales laid down on the seafloor, robbing the Devonian atmosphere of its warming \(\ce{CO2}\). As we extract and burn natural gas from these shales today so that we may produce energy, we ***finally*** oxidize all that fossil carbon, which had remained sequestered underground for more than 300 million years and would have continued to do so if industrialized humans hadn't shown up. The \(\ce{CO2}\)-induced warmth that would have saved the Devonian biota from an ice age is now being unleashed on the Holocene (and maybe inducing the Anthropocene...).

End-Permian extinction



Figure \(\PageIndex{15}\): Fusulinid foram fossil from Permian limestones in west Texas. (Callan Bentley photo.)

The end-Permian mass extinction was the most extreme of any in Earth history. It's sometimes dubbed "The Great Dying," with 62% of marine genera going extinct, as well as severe impacts among terrestrial biota. Perhaps only 17% of species on Earth survived it. It's the sharp cliff at the end of the "Paleozoic Plateau." It's the closest our biosphere has ever come to being deleted. Writer Peter Brannen astutely called it "**the worst thing that ever happened**."

By the end of the Paleozoic, life had rebounded from the late Devonian extinction. In the sea, crinoids and their cousins the blastoids were having a heyday. Tabulate and rugose corals had made great progress in building up reefs in the late Paleozoic ocean, reefs that almost certainly served as shelter and nurseries for many other species, as modern scleractinian coral reefs nurture fish, shrimp, crabs, *etc*. Giant single-celled fusilinid foraminiferids were rolling around the seafloor like fat grains of rice.



Figure \(\PageIndex{16}\): Dimetrodon reconstruction. The posture is controversial, but here shown upright, based on possible Dimetrodon trackways. (Max Bellomio imagery, via Wikimedia.)

On land, the *Glossopteris* seed fern reigned supreme, at least over the southern supercontinent Gondwana. In its shade, the waterproof amphibian descendants called reptiles had diversified and spread into a great number of ecological roles. Dominant among them were the **synapsids**, those that used to be dubbed "mammal-like reptiles." In some cases, these animals could grow to large sizes. You would not have wanted to meet *Dimetrodon* in a dark alley. The sail-like "fin" rising from its back is its most eye-catching feature, but its mouth was full of sharp teeth.

Also worth noting are the cynodonts, which are probably the direct ancestors of the mammal lineage. Cynodonts were not as large as the **therapsids**, though, and many of them were hungry. With massive heads bearing powerful jaws and saber-like teeth, *Gorgonops* and its kin were fearsome beasts. The therapsids had several advantages over their less derived pelycosaur relatives. They had a fully-upright stance, with their limbs always beneath the body. There is also potential evidence that they were endothermic (also called 'warm-blooded') and had at least some hair. Many species show 'whisker pits' in the skull, which is good bony evidence of hairs that were once anchored in these pits.

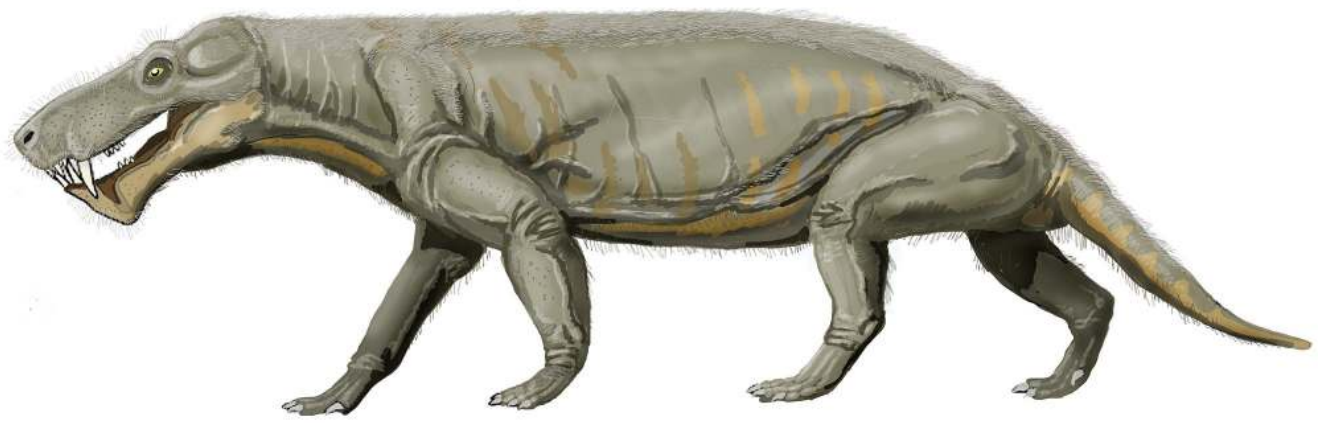


Figure \(\PageIndex{17}\): Gorgonops, a large Permian predatory therapsid. (Reconstruction by Dmitry Bogdanov via Wikimedia.)

Another group of reptiles had a relatively subdued presence during the Carboniferous and Permian: the **sauropsids**. These animals included the ancestors of modern lizards, snakes, crocodiles, and birds, as well as the dinosaurs who would dominate the Mesozoic.

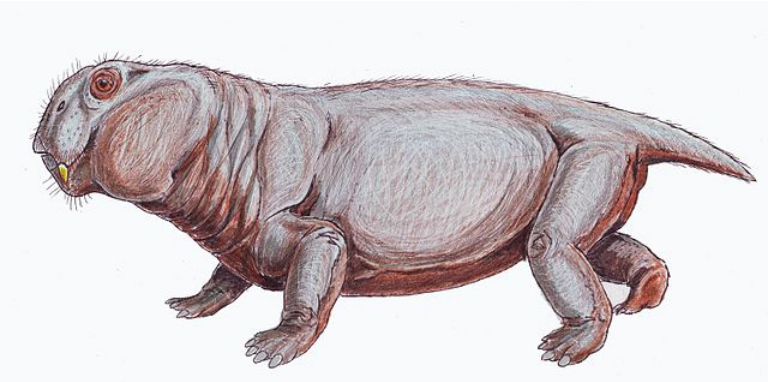


Figure \(\PageIndex{18}\): Lystrosaurus murrayi, about the size of a small dog. (CC-BY; Sketch by Dmitry Bogdanov, via Wikimedia.)

Gorgonops and *Dimetrodon* went extinct at the end of the Permian, along with 2/3 of therapsids. Some cynodonts survived, and these diversified during the Mesozoic that followed, some of them ultimately leading to you and me. One that practically overran the world in the early Triassic was *Lystrosaurus*, whose populations rose unchecked after the the demise of all its predators. But in the Mesozoic, it was the sauropsids that diversified most successfully, and led to dinosaurs, plesiosaurs, ichthyosaurs, pterosaurs, and mosasaurs. They were the big winners from the end-Permian extinction. But there were *so many* losers...



Figure \(\PageIndex{19}\): El Capitan and the Guadalupe Mountains of west Texas, a Permian reef complex and the namesake of the Guadalupian epoch of the Permian period. (Public domain; NPS photo.)

There were two pulses to the end-Permian extinction. Because they were separated by ~8 million years, it has become commonplace in recent years to hear them discussed independently: the Capitanian (or 'end-Guadalupian') at 260 Ma and the true end-Permian event at 252 Ma. Both appear to have the same general cause: global warming.

The careful study of variations in oxygen isotopes in strata spanning the Permian-Triassic boundary shows an lightening of the oxygen content of the oceans. This correlates with a big boost in heat energy, making \(\ce{^{18}O}\)-based water molecules as likely to evaporate as \(\ce{^{16}O}\)-based water molecules. Using the ratio of \(\ce{^{16}O}\) to \(\ce{^{18}O}\), it has been calculated that the global ocean warmed by about 6 \(^{\circ}\)C. As with the discussion of 'cooling by 5 \(^{\circ}\)C' in the end-Ordovician extinction above, please realize that it takes a truly enormous amount of energy to shift ocean temperatures by six whole degrees Celsius. That extra energy would have been radiated to space, had it not been trapped by extra \(\ce{CO2}\) in the late Permian atmosphere.

Now let's consider where that extra \(\ce{CO2}\) came from, and then examine the havoc it caused in the ocean.

There were two major large igneous provinces that erupted during the late Permian. The first was in China: the **Emeishan Traps**, where "Traps" derives from the Swedish word *trappa*, meaning "stairsteps." This refers to the step-like landscape where the many layers of a vast flood basalt province get etched out. (The Deccan Traps in India have a similar morphology, as do the lava layers of the Columbia Plateau in eastern Washington state.) The Emeishan basalts erupted from 265 to 259 Ma.



Figure \(\PageIndex{20}\): Putorana Plateau, Russia, part of the Siberian Traps flood basalt province. (CC-BY; Photo by Александр Лещёнок, via Wikimedia.)

The second, much larger, was in Siberia. Today we call it the **Siberian Traps**. The Siberian Traps represents one of the biggest eruptive episodes in Earth history. Presumably the result of a mantle plume slamming into the bottom of the north Asian continent, the volcanic rocks of the Siberian Traps have been shown to span the Permian-Triassic boundary, including the mass extinction at 252 Ma.

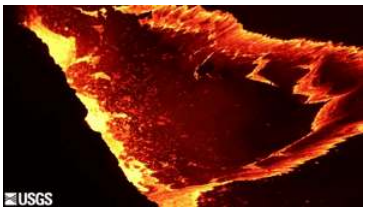


Figure \(\PageIndex{21}\): Roughly what Siberia looked like 252 million years ago. (From a video by the USGS.)

Size estimates of the Siberian Traps boggle the mind: A volume of more than 1.5 million cubic kilometers of lava was erupted, smothering roughly 2 million square kilometers of northern Siberia. That is enough basaltic lava to bury the entire contiguous United States to a depth of half a mile. It's enough basalt to make three towers, each with a footprint of 1 square kilometer, that reach to the Moon.

So much lava being erupted carries with it a commensurate amount of volcanic gas. Most of that gas is water vapor, which is rapidly removed from the atmosphere via precipitation. But the second most common volcanic gas is \(\ce{CO2}\), and you know what that does.

Not only that, but one of the layers that the Siberian Traps basalt passed through on its way from the mantle to the surface was a vast Carboniferous-aged coal field (the world's largest) in the Tunguska Basin, which caught fire and burned, adding even more \(\ce{CO2}\) to the total that was released! Estimates are that the level of \(\ce{CO2}\) in the atmosphere reached to at least 3000 ppm, and maybe as high as 30,000 ppm. Average estimates for the end-Permian are **~8000 ppm**. (For reference, in 2022, the world was at about 420 ppm, up from ~280 ppm in pre-Industrial times.) So the end of the Permian had **twenty times as much** \(\bf{\ce{CO2}}\) in its atmosphere as we do today, an insane level that would have warmed the planet to an astonishing degree.

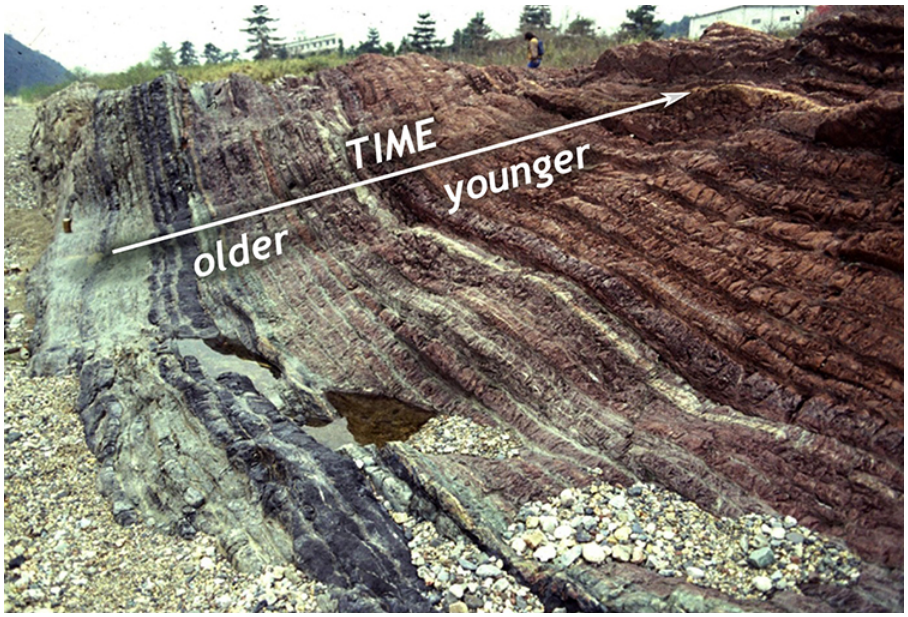


Figure \(\PageIndex{22}\): Outcrops in Inuyama, Japan, show a Triassic sequence of sedimentary strata that record changes in deep ocean oxygen levels after the end-Permian mass extinction event. The strata have been tectonically rotated due to accretion with the Japanese mainland. (Photo by Yukio Isozaki; reproduced with permission.)

A warm climate doesn't necessarily spell the end of the world, but in this case, the **oceans** played a key role. In the modern world, with cold polar regions and warm tropics, our ocean vigorously circulates in a vast, floppy loop. The energy that drives this flow of water comes from the difference in temperature (and thus density) of the water. When warm water grows cold, it gets more dense, and it sinks. Sinking water pulls other water after it, and a flow is induced. But in the Permian, as the polar oceans warmed up to roughly the same temperatures as the tropics, this flow was reduced, and — apparently — stymied completely. In short, the oceans became stagnant.

As with the late Devonian, this produced **anoxia** in the oceans, which is recorded as usual by jet black sedimentary deposits. The anoxia was at first most pronounced in the deep ocean, but it appears to have eventually invaded the coastal shallows. This is where most marine life makes its home, and that was of course quite disastrous.

Worse, though, is what happened when the Sun's rays hit the anoxic shallow water. This is the rare combination of factors that promotes the growth of sulfur-reducing bacteria, which release **hydrogen sulfide gas** (\(\ce{H2S}\)) as a waste product. You know \(\ce{H2S}\): it's the stink of rotten eggs — or the nauseating stench of really gross farts. In low levels, it's revolting and annoying, but in high levels, it's deadly. The appearance in seafloor sediments of little raspberry-shaped clusters of pyrite (\(\ce{FeS2}\)) suggest that the ocean at the end of the Permian was quite rich in sulfur. The inference is that these distinctive purple bacteria must have bloomed in profusion without any of their usual microbial competitors.



Figure \(\PageIndex{23}\): Purple sulfur bacteria from a modern euxinic lake in British Columbia. (J. Glass photo; permission requested.)

Thus the oceans were not only anoxic, but also probably also **euxinic** — the word that describes being thoroughly infused with \(\ce{H2S}\). In 2005, Lee Kump and colleagues suggested something bold – that the oceans were so euxinic during the end-Permian that \(\ce{H2S}\) began to exsolve out of the sea and into the air, where it acted like an agent of chemical warfare, gassing terrestrial animals to death.

Another potential "kill mechanism" that applies to terrestrial species has its roots in a layer of rock salt that the Siberian Traps eruption also pierced, exploded, and scattered as it erupted. The chemicals that were released are thought to be the sort that destroy ozone (\(\ce{O3}\)), which humans have done using the famous chlorofluorocarbons (CFCs) that were banned by the Montreal Protocol. If the stratosphere's **ozone layer** was severely damaged at the end of the Permian, that would have increased the flux of ultraviolet light (particularly UV-B) to the surface of the planet. This deadly radiation would have potentially driven up rates of mutations and cancers for any organism that was exposed to it (*i.e.*, only those who lived on land).



Figure \(\PageIndex{24}\): A layer of limestone made of nothing but shells of the clam *Claraia*. (Modified from a photo by Toedi3614, via Wikimedia.)

In the aftermath, the world lay obliterated, an apocalyptical denuding of the biosphere that probably reeked of rotten eggs. On land, the formerly terrifying forms of *Dimetrodon* and *Gorgonops* lay still; corpses rotting away in the foul air. The forests of sphenopsids and lycopsids were done forever. From the rotting foliage, a few therapsids nosed out and sniffed the air. In the sea, there were no more blastoids, no more fusulinid forams, no more trilobites or rugose corals. No more eurypterids; no more goniatitic ammonoids. Brachiopods were greatly reduced in number and diversity (a 96% decline). They would be replaced by the bivalves in the years to come. Still depleted in oxygen and utterly emptied out, the early Triassic seas were overtaken by one genus of clam that was adapted to low-oxygen conditions. *Claraia* spread far and wide, unimpeded by any competition. It's so omnipresent and numerous that as a monoculture, it makes an excellent index fossil for the first part of the Triassic period: a bizarro "Clamworld."

This was the one big reset of Earth's biosphere –phyla that had dominated the oceans for 200 million years were gone, and oceans would never look the same. The survivors went on to diversify into the Modern Fauna; mollusks would now reign supreme, scleractinian corals would build reefs, and echinoids, sea stars and sea cucumbers would now represent the echinoderms (replacing crinoids and blastoids).



Figure \(\PageIndex{25}\): The Palisades cliff above the Hudson River in New Jersey is the surface expression of a vast near-surface mafic sill. This massive slug of basalt was intruded into the Newark Basin during the breakup of Pangaea. (CC-BY; Photo by Anita Gould, via Flickr.)

Therapsids flourished in the Triassic, now that their synapsid competitors had been suffocated. These crocodilian relatives were attended by the first dinosaurs, but dinosaurs were the sideshow rather than the main act. Only after the end-Triassic extinction could the dinosaurs surge forward and dominate the Mesozoic land surface. But first, their crocodilian overlords must be cleared away.

In the sea, about half of the Triassic genera didn't make it into the Jurassic. Almost every kind of ammonoid and bivalve was decimated. Ammonoids had evolved a ceratitic morphology, which is an index fossil for the Triassic. Also in the seas were a menagerie of marine reptiles, many of whom were snuffed out by the end-Triassic extinction event. Walrus-like reptiles called placodonts and the seal-like nothosaurs totally perished, while some of their ichthyosaur and plesiosaur cousins eked a way through, and would re-diversify in the Jurassic and Cretaceous.

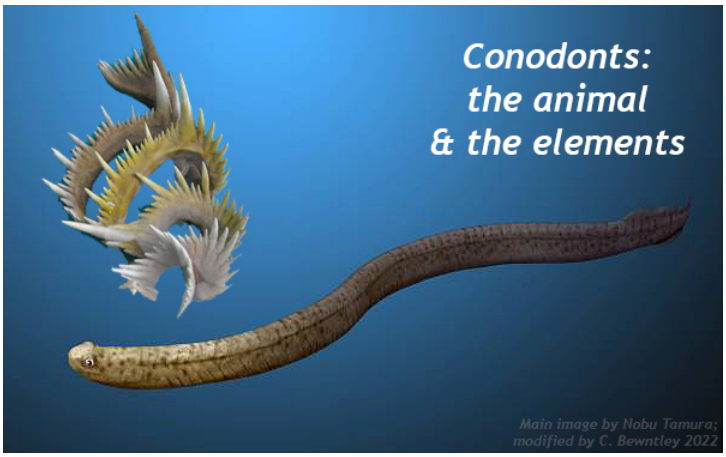


Figure \(\PageIndex{26}\): Conodonts were eel-like jawless fish that left behind copious prickly “elements” as index fossils. (Both CC-BY; Photo illustration by C. Bentley (2022), using the background illustration by Nobu Tamura and element photo by Wikipek, both via Wikimedia.)

One small animal of note, the conodonts (a kind of thin, eel-like jawless fish) had been a fixture of Earth’s oceans from the Cambrian onward. They were practically as common in the oceans as water molecules, and left their jaws and gill support structures behind as an uniquely valuable index fossil. But the end-Triassic got them, too. Their long run on Earth came to an abrupt end at 202 Ma.

Now that you know who died and who survived, let’s examine the “why?”

Like the end-Permian crisis that preceded the Triassic, the mass extinction that brought it to an end appears to be a case of (1) “large volcanic province erupts,” (2) “\(\ce{CO2}\) levels spike,” and (3) “global warming ensues.” In this case, a hot spot is not to blame.

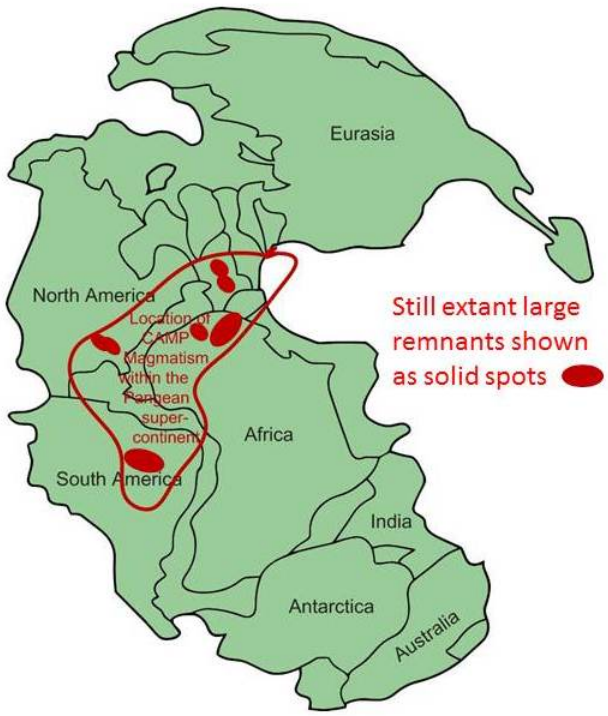


Figure \(\PageIndex{27}\): Location of mafic igneous rocks of CAMP in four continents, positioned at the time of Pangaea’s breakup. (Williamborg image, after Blackburn (2013), via Wikimedia.)

Instead, it was the titanic breakup of the most gargantuan supercontinent ever. As **rifting split Pangaea** and opened the spindly Atlantic Ocean down its center, the underlying crust was thinned dramatically. In what is today eastern North America, a series of

rift basins — the Newark Supergroup basins — opened up along normal-faulted hinges, dropping down like trap doors. These basins did an excellent job catching immature sediment from the surrounding highlands, but they also released the pressure on the hot mantle beneath, which partially melted and made its way upward along the fault system. What issued forth was an enormous amount of basaltic magma. And it wasn't only in North America; huge quantities of basalt and gabbro were produced in adjacent regions of northeastern South America, northwestern Africa, and southwestern Europe. Though its various major volcanic centers have now been separated by seafloor spreading of the Atlantic Ocean, they were once adjacent. Today we call it all the Central Atlantic Magmatic Province (**CAMP**).

Ultimately, CAMP may have erupted even more basalt than the Siberian Traps, but a good amount of it has since weathered away. Its total volume is estimated at something between 2 and 3 million cubic kilometers of mafic igneous rock. CAMP rocks cover 11 million square kilometers. The eruptions came in four major pulses over the span of about 600,000 years. That's a lot of magma making it to the surface (and degassing \(\ce{CO2}\)) in a relatively short amount of time.

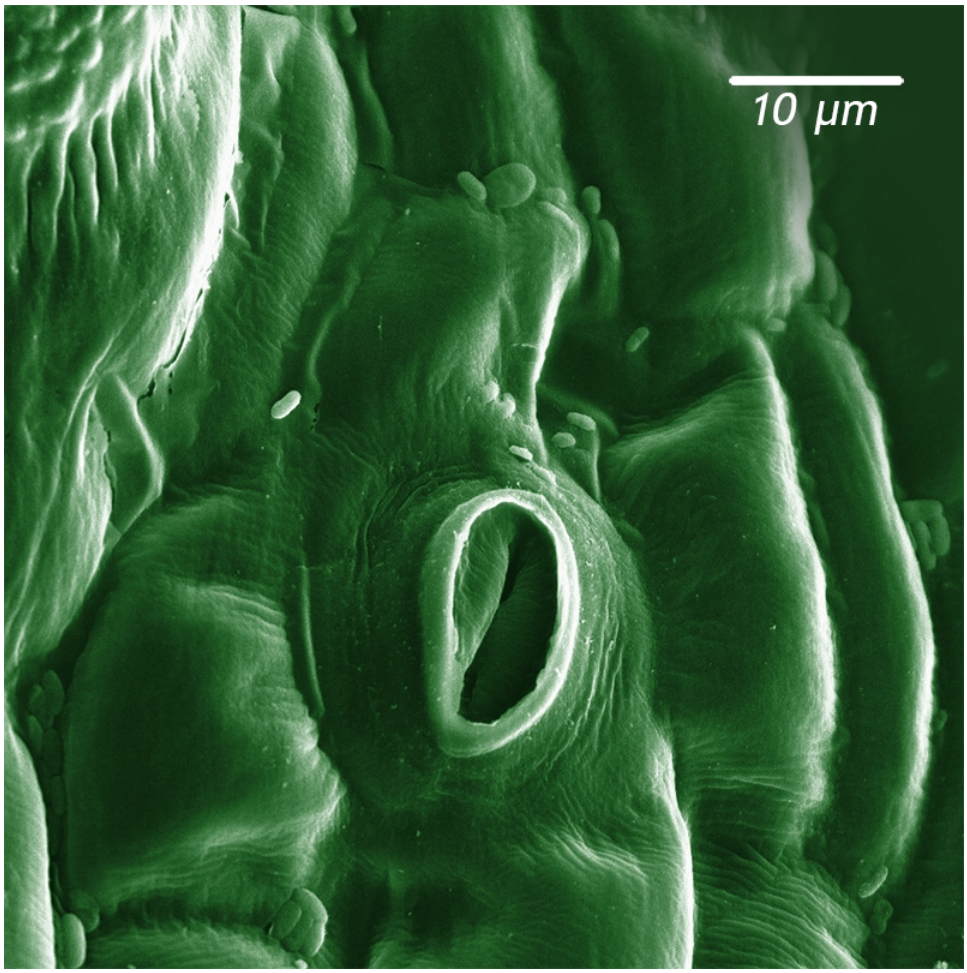


Figure \(\PageIndex{28}\): Colorized scanning electron micrograph of a single stoma on the underside of a tomato leaf. Here, the guard cells are relaxed, and the stoma is closed. (Modified from a public domain image via Wikimedia.)

An interesting study about the Triassic extinction uses plant fossils [LINK TO FOSSIL PLANT CASE STUDY] as a proxy for examining \(\ce{CO2}\) levels at the time. On the underside of plant leaves are little holes that allow gas exchange with the atmosphere. These holes, called **stomata** (singular: *stoma*), are like little doors that let \(\ce{CO2}\) in to enable photosynthesis and waste \(\ce{O2}\) out after glucose has been made. In modern plants, the number of stomata inversely correlates with atmospheric \(\ce{CO2}\) levels. The higher the \(\ce{CO2}\) level, the fewer stomata are needed to get what the plant needs. Under conditions of lower \(\ce{CO2}\), plants require more stomata to let in sufficient air to extract the same absolute amount of \(\ce{CO2}\). So simply by collecting fossil leaves and counting their stomata with a microscope, we can arrive at a sense of how \(\ce{CO2}\) levels varied across the Triassic/Jurassic transition. Jennifer McElwain and colleagues (1999) analyzed fossil leaves from end-Triassic strata in Greenland and Sweden. They found that the stomata density decreased dramatically at the end-Triassic,

commensurate with an atmospheric \(\bf{\ce{CO2}}\) **level that was seven or eight times as high** as we have today. That means a 3-4 \(^{\circ}\)C jump in temperature, above and beyond temperatures that were already ~3 \(^{\circ}\)C warmer than today's. *[If you're interested in this technique, the Smithsonian has a citizen science project for counting stomata that you can help out with.]*

With the planet's biota reset once again, it was finally time for the dinosaurs to have their day.

End-Cretaceous extinction



Figure \(\PageIndex{29}\): "Sue," the most complete *Tyrannosaurus rex* fossil, on display at the Field Museum, Chicago.

Dinosaurs came in two major groups, the saurischians and the ornithischians. With the crocodile-like Triassic therapsids now mostly dead, the dinosaurs diversified and persisted for tens of millions of years. They dominated terrestrial ecosystems in a pervasive and enduring fashion. Other reptiles colonized the air (pterosaurs, and eventually birds too) and the sea (plesiosaurs, ichthyosaurs, huge turtles like *Archelon*, and eventually mosasaurs too). Giant reptiles were everywhere during the Jurassic and Cretaceous. The world was their oyster.



Figure \(\PageIndex{30}\): Display in the Natural History Museum of London, showing ammonoids as the ideal index fossils. (Callan Bentley photo.)

Speaking of oysters, the sea had plenty of those during the middle and late Mesozoic. Some of them encrusted the shells of clams called inoceramids that had shells a meter in diameter. Another bizarre bivalve group was the rudists, which resembled Sesame Street's Oscar the Grouch in their trashcan-like morphology. Swimming among them were spiral-shelled cephalopods: the ammonitic ammonoids, which bore distinctly fractal-looking suture patterns between segments of their shells. They make for superb index fossils for the Jurassic and Cretaceous. But along with the dinosaurs and other giant reptiles, their time on Earth came to an end on one very bad day.

Before we get to that day, there are several items that need mentioning. First off, the Cretaceous was **hot** in general, all around the world. There is no evidence of glacial ice anywhere on the planet during this "hothouse" climate. This, combined with a vigorous rate of seafloor spreading, pushed oceans up onto the continents. In North America, this allowed the Western Interior Seaway to develop as the Sevier Orogeny produced an adjacent foreland basin.



Figure \(\PageIndex{31}\): India's Deccan Traps are made of many layers of flood basalts. (CC-BY; Photograph by Nichap, via Wikimedia.)

Then things started going wrong. The first, you'll be shocked to hear, was the eruption of another large igneous province. This one was in western India: the **Deccan Traps**. As with the Emeishan Traps and the Siberian Traps discussed above, this is a landscape of step-like mesas and plateaus, as a thick stack of basalt flows has been dissected by more recent streams. Like them, it is astoundingly large: 1 million cubic kilometers of basalt, slathering half a million square kilometers of western India (it may have originally been much higher in volume and area, but as with CAMP, it looks like much has been eaten away by subsequent erosion). The eruptions were relatively rapid – from beginning to end, it may have lasted only 30,000 years. At the time of its eruption, India was its own small continent, surrounded by the Indian Ocean — *i.e.*, this was prior to its collision with southern Eurasia. It is inferred that the eruptions were triggered by the island continent's drift over the Réunion hotspot, a long-lived welt in the Earth's mantle that is still producing lava today, at namesake Réunion Island itself.



Figure \(\PageIndex{32}\): The Cretaceous-Paleogene boundary at Bottaccione Gorge, near Gubbio, central Apennines of Italy. Pencil for scale. (Callan Bentley photo.)

Up until 1980, that alone would have been viewed as entirely sufficient to explain the end-Cretaceous (a.k.a. “K/T” or “K/Pg”) extinction. But in 1980, a son/father team published a paradigm-shifting paper in the journal *Science* that dredged up the specter of catastrophism anew. Walter Alvarez and his father Luis Alvarez, with others, published a bombshell of a paper that laid out the case for an extraterrestrial impact as the cause of the end-Cretaceous extinction. The younger Alvarez was working as a geologist in the central range of mountains that run down the spine of Italy, the Apennines. He studied the stratigraphy of the limestones and shales there, and one day got to wondering about the nature of the boundary between the Cretaceous and the Tertiary*. He noticed that it really came down to a single layer of clay between two ~4 cm thick limestone strata. He had a really great outcrop in Bottaccione Gorge, a canyon near the town of Gubbio. Below this horizon, Alvarez saw distinctive Cretaceous radiolarian fossils. Above it, *none*.

* When Alvarez did this work, “Tertiary” was still used as a period by the International Commission on Stratigraphy. Only more recently (2003) did “Paleogene” replace it.

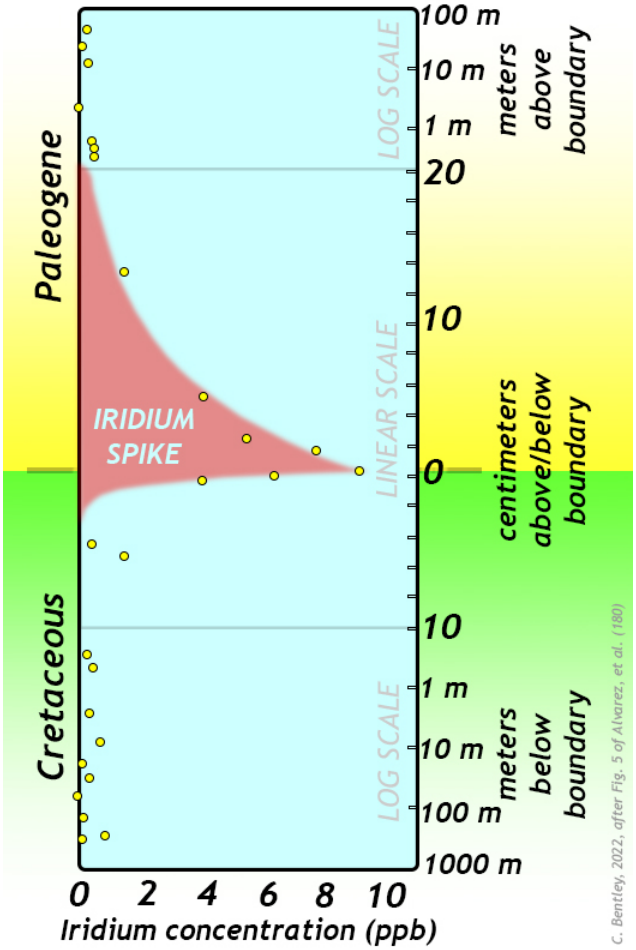


Figure \(\PageIndex{33}\): The iridium anomaly in Italian limestone/shale strata, presented by the Alvarez, et al. (1980) paper. (Graph redrawn by C. Bentley (2022))

It occurred to him to wonder how much time was packed into that clay layer. Was it 10,000 years, or 100,000, or something else? His father, a Nobel-prize-winning physicist, suggested examining the clay layer from the perspective of the small but (presumed-) constant flux of micrometeorites to Earth. Luis Alvarez advised his son to look for something distinct to outer space sources, that would have built up gradually, and thus could be used to calibrate how much time had elapsed while the clay was deposited. The elder Alvarez suggested the element **iridium**, which (he hypothesized) should have a more or less constant flux to Earth's surface.

But the amount of iridium they actually measured were out of this world. Literally.

There was so much iridium packed into that clay layer that the Alvarez team found another hypothesis more likely: this had to be fallout from a single **giant meteorite impact** somewhere else in the world.

Though many geologists initially were skeptical of this "catastrophe" of a hypothesis, they soon got busy testing it, by checking the Cretaceous/Paleogene (K/Pg) boundary elsewhere in the world, to see if other locations also shared this iridium anomaly. Sure enough: they showed the iridium anomaly, too! In addition, they started documenting additional evidence of an impact: (1) spherules and (2) shocked quartz.

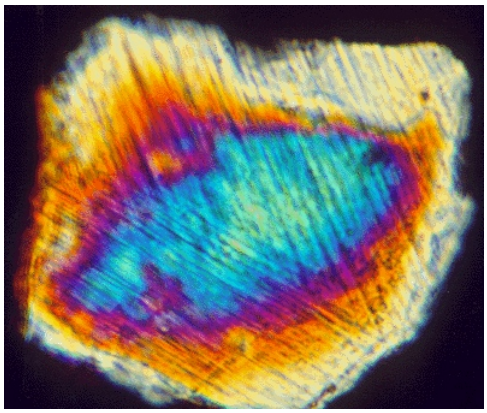


Figure \(\PageIndex{34}\): Grain of shocked quartz viewed under cross-polarized light with a petrographic microscope. Note the planar deformation features. (Public domain; Photograph by Glen A. Izett, USGS.)

Spherules are little glass droplets that form when rock, melted instantaneously by the tremendous heat of the impact, is flung into the air as billions of tiny droplets. These cool and solidify before they rain out onto the landscape: little glass beads that speak of unimaginable violence. **Shocked quartz** is a little bit different: it starts off as a solid crystal of the common mineral quartz, but then it deforms as the powerful shock wave of the impact passes through it, wrecking its previously precise crystal lattice but not destroying the integrity of the grain by melting or disintegrating it. These distinctive features were first identified in the aftermath of nuclear bomb testing, and were decisively associated with extraterrestrial impacts by the legendary Gene Shoemaker, the "grandfather of impact geology."

The K/Pg boundary layer thickened toward the Americas, and toward the Gulf of Mexico even more. In their search for oil and gas deposits, the Mexican state petroleum company, Pemex, had mapped a vast circular structure that was mostly under the Gulf of Mexico, but also had about a third of its area lapping over onto the thumb-like Yucatán Peninsula. At first, it was inferred to be a big, extinct volcano, but in 1981, geophysicists Glen Penfield and Antonio Camargo suggested it was an impact crater instead, and indeed that it was the one responsible for the end-Cretaceous extinction. One reporter, Carlos Byars, wrote up the announcement for the *Houston Chronicle*. Somehow, almost no one noticed.

Luckily, a decade later, Byars was covering a geology conference where the crater's location was still an enigma. In an astonishing moment, the journalist Byars let the scientists in on what to him must have seemed like "old news" — and now the geological community had found its ground zero, its Big Kahuna.

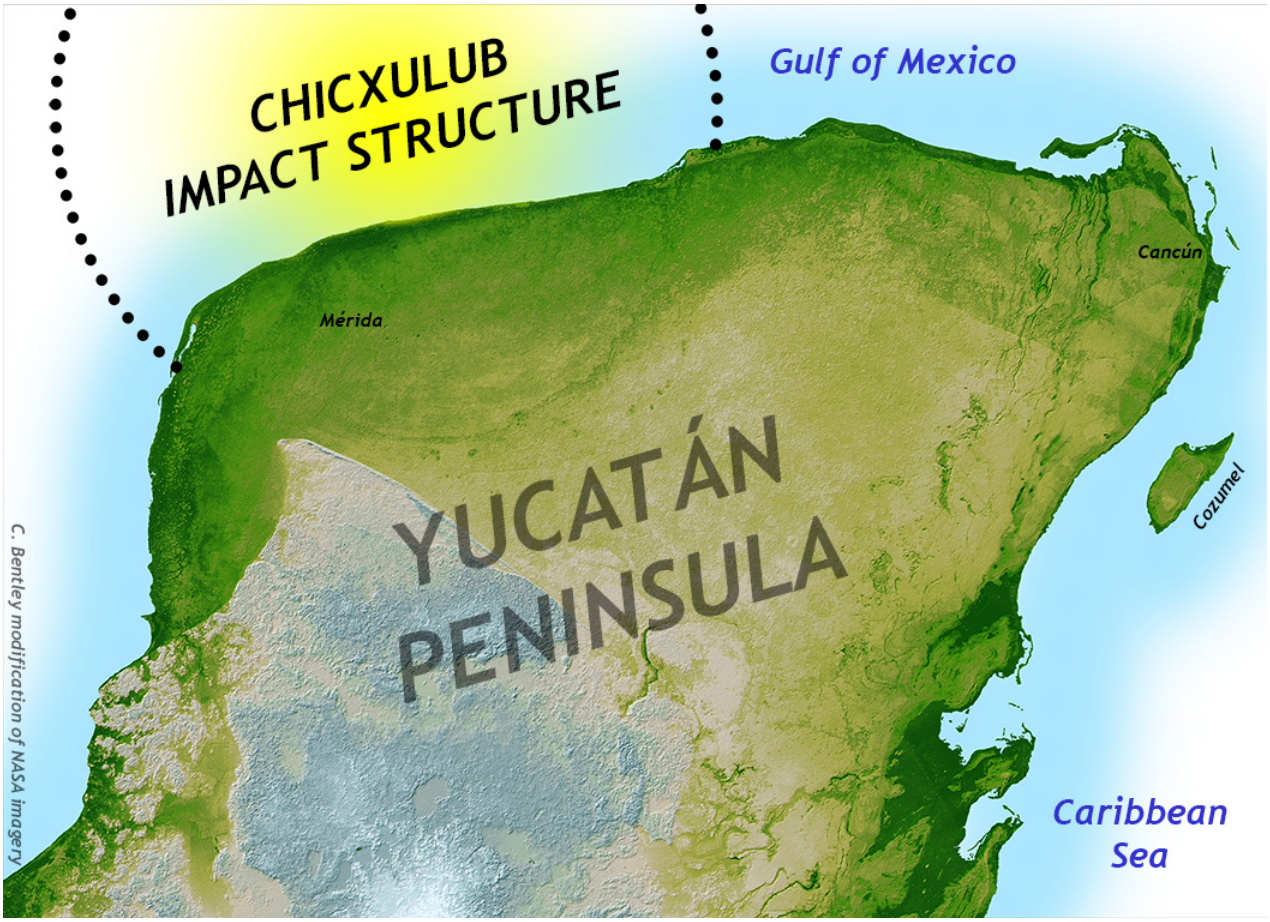


Figure \(\PageIndex{35}\): The position of the Chicxulub impact structure in the southern Gulf of Mexico and northern Yucatan Peninsula. (C. Bentley annotation of a NASA image.)

The crater was centered on a small port town called **Chicxulub**. Thus, the structure has come to be known by that name. The structure is about 180 kilometers (110 miles) in diameter and 20 kilometers (12 miles) deep, but it is now mostly hidden from view underneath the subsequent 65 million years worth of sedimentation – limestone, mostly. These overlying limestones have slumped downward a bit over the course of the Cenozoic, allowing a distinctive ring-shaped scarp to develop in the overlying limestone of the Yucatán. All along this ring are water-filled sinkholes, called cenotes. The entire arc of cenotes is a candidate World Heritage Site, not only for its relevance to the end-Cretaceous extinction, but also because the great pre-Columbian civilization of the Mayans utilized these cenotes for access to water.

Impact experts can model the size of the impacting object (a meteorite or a comet, or more agnostically named, a bolide) by the size of the crater it leaves behind. In this case, the impactor would have been about ten kilometers across, roughly the diameter of San Francisco, or Mount Everest. The energy released would have been something on the order of **100 million megatons** of TNT. If you've seen footage of the Chelyabinsk meteorite's detonation in 2013 over Russia, you've seen what *half* a megaton looks like. Just multiply that by 200,000,000.

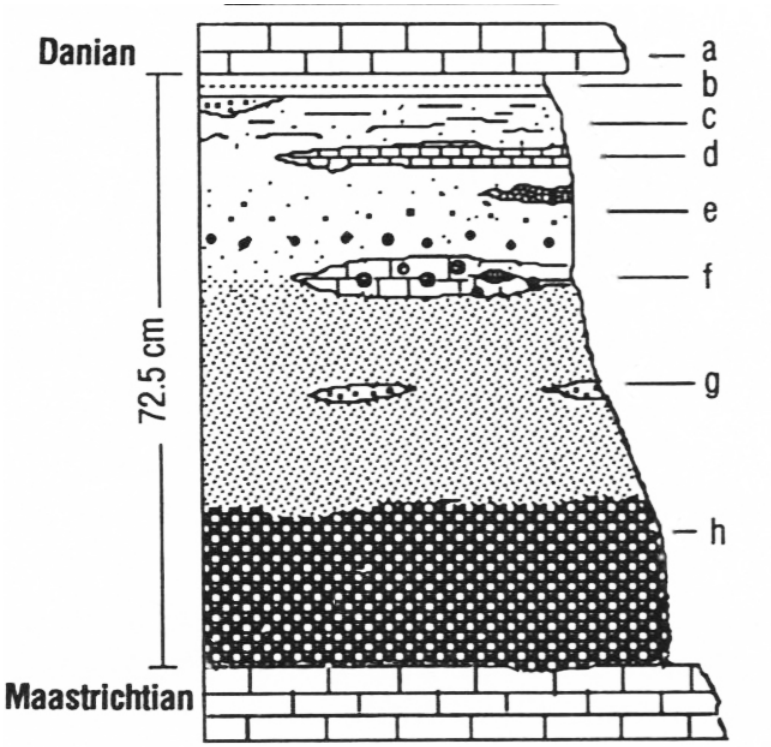


Figure \(\PageIndex{36}\): Part of Figure 1 of Maurrasse & Sen (1991), documenting key stratigraphic features of the K/Pg boundary at Beloc, Haiti. From bottom to top, the strata above the last Cretaceous limestone [Maastrichtian] include (h) coarse, brown, spherule-rich ejecta layer, now converted mostly to smectite clay, (g) fine spherules in sandy marl, (f) coarse-spherule-bearing marl lens, (e) graded sandy marl with small lens of micrite, (d) white chalk lens, (c) laminated sandy marl with small lenses of coarse spherules, (b) fine clay with greatest iridium concentration, (a) earliest Paleogene limestone [Danian].

The “discovery” of the Chicxulub stimulated geological interest immensely – the key elements of the story were now in place. Jan Smit added another: **tsunami deposits** at the K/Pg boundary at sites in Mexico and Texas. Floretin Maurrasse also documented a substantial tsunami deposit on the island of Haiti.

Recently, Robert DePalma and colleagues (*including Walter Alvarez and Jan Smit*) published a new site they call “Tanis” in North Dakota. It exposes strata of the Hell Creek Formation. Though the Hell Creek is a widespread regional geological unit, Tanis holds an incredible concentration of fossils that appear to have been deposited within hours of the impact. There are many superlative features of Tanis, but one of the most evocative is the presence of meteorite impact evidence inside the organisms fossilized. Specifically, fish breathe with gills, and those gills strain oxygen from the water. At Tanis, about half of the **fish fossils have impact spherules caught in their gills**, suggesting they were alive and breathing as impact debris was raining down upon their waterway. This astounding details suggests an amazing level of precision for the date of the Tanis deposits – they literally show **the very last day** of the Cretaceous period.

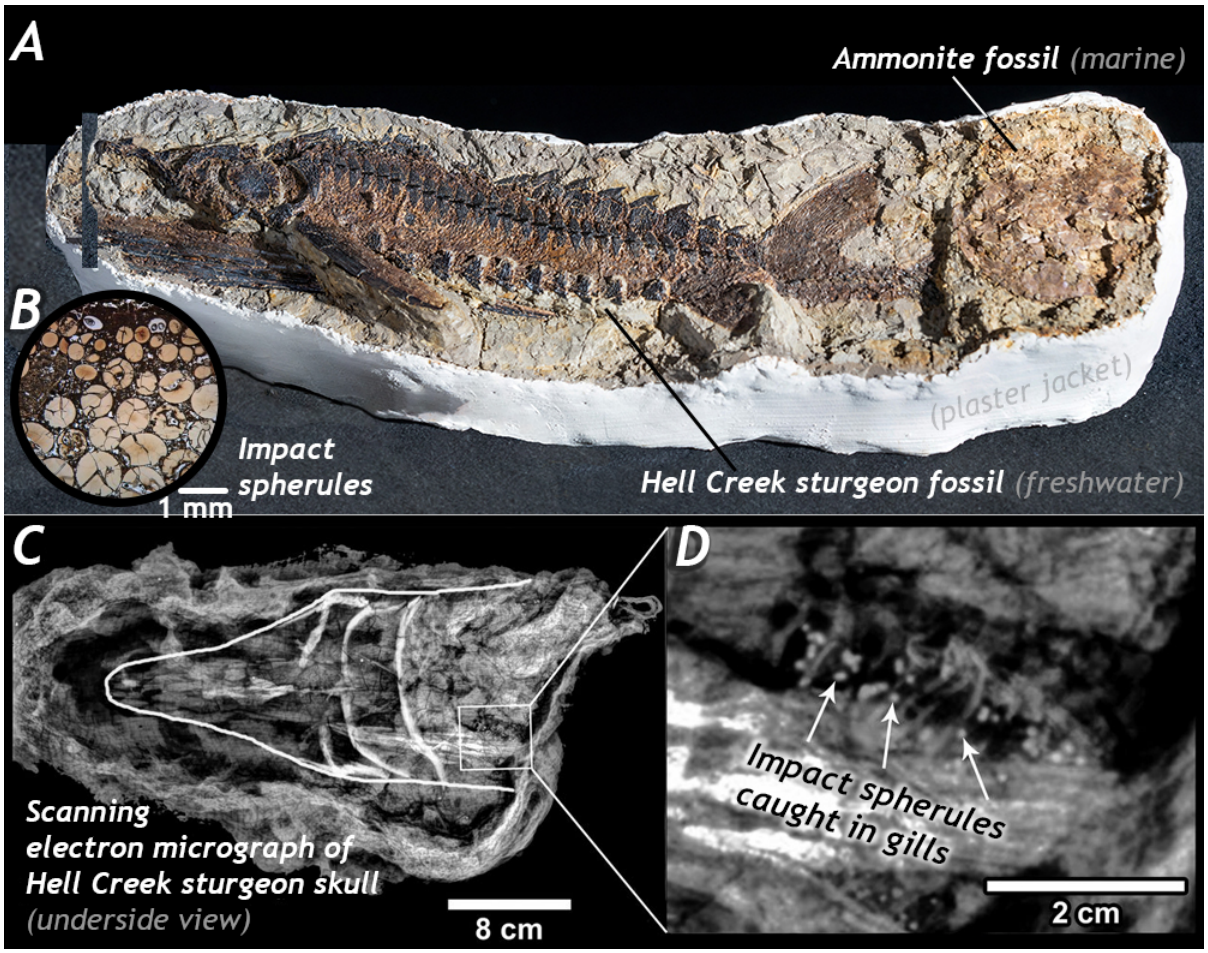


Figure \(\PageIndex{37}\): Photo illustration showing key details from the Tanis site. (A) shows a fossil sturgeon from the site adjacent to an ammonite; a freshwater species and a marine species deposited at the same moment, in the same stratum (modified from a NASA photo), (B) impact spherules from Tanis, (C) a scanning electron micrograph of the underside of a sturgeon skull, with (D) showing a zoomed-in look at the gills of this poor fish, clogged with impact spherules. (Images B, C, and D are from Figures 4 and 6 of DePalma, et al. (2019).)

So there is now substantial evidence for a major extraterrestrial impact at the very, very end of the Cretaceous. But how did that kill half of life on Earth? Various kill mechanisms have been proposed, including from the radiative heat of the impact, blunt force trauma from its shockwave, subsequent forest fires, violent tsunamis, ocean acidification, and finally a prolonged “nuclear winter” from small particles suspended in the stratosphere. Upending the food chain, some or all of these traumas destabilized the biosphere and ended the Mesozoic Era. What followed was the Age of Mammals, the Cenozoic — the time that would ultimately lead to us.

It’s worth noting that dinosaurs aren’t all gone... **birds** are with us today, and like mammals and plants (particularly angiosperms), the birds have undergone adaptive radiations to lead to the familiar ecosystems we see across the modern Earth.

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11.3: Other 'Mass' Extinctions

Peter Brannen's excellent book *The Ends of the World* recaps the Big Five, but also looks in detail at numerous other mass extinction events, including some in the Precambrian and some that have yet to occur (*i.e.*, in Earth's future). A few notable examples outside the traditional Big Five are briefly described below:

The Great Oxidation "Event"

Mass extinctions are relatively easy to detect when the organisms dying (and replacing them) leave behind body fossils with hard parts. The picture is much more difficult to discern when all organisms were microscopic. In the early Proterozoic Eon, Earth transitioned from an oxygen-free atmosphere to one with a decent amount of free oxygen (O_2). Though this is called the Great Oxidation "Event," [GOE] it is not a very sudden event at all. Traditionally, it is viewed as the span from 2.4 Ga to 2.0 Ga: that's 400 million years, a very drawn-out "event." Other names include the Oxygen Crisis, the Oxygen Catastrophe, and the Oxygen Revolution, each of which puts its own spin on interpreting the rise of this important gas.



Figure 11.3.1: Deformed banded iron formation, Soudan, Minnesota. Chemical sedimentary rocks like these imply an oxygen-free atmosphere. (Callan Bentley photo)

Prior to the GOE, the planet's atmosphere lacked significant free oxygen. We know this because of unusual chemical sedimentary rocks that could only form in an ocean that is largely oxygen free (**banded iron formations** [BIFs]), as well as grains of minerals like pyrite and uraninite in clastic sedimentary rocks. These signatures of the absence of oxidizing conditions imply a very different early Earth.

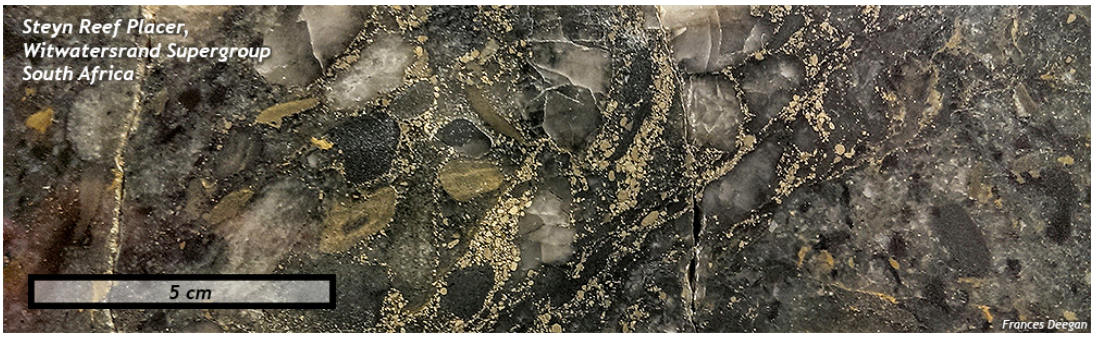


Figure 11.3.2: Detrital pyrite in conglomerate of the Steyn Reef Placer, Witwatersrand Supergroup of South Africa. (Photo by Frances Deegan; reproduced with permission.)

Those anoxic conditions are preserved in Archean strata and those of the early Proterozoic. But by the middle of the Proterozoic, the BIFs and **detrital pyrites** peter out, and we see the first appearance of oxidized terrestrial sediments (red beds). These are river

and floodplain sediments deposited in contact with copious oxygen – enough O_2 to rust their iron content through and through.



Figure 11.3.3: Modern stromatolites, Shark Bay, Australia. (Photo by Alicejmichel via Wikimedia)

So where did all this free oxygen come from, and what was the impact on the biosphere? Oxygen is a waste product of photosynthesis, so any photosynthetic organism produces it in proportion to the amount of glucose it generates. Of particular note are cyanobacteria, which are preserved from Archean and younger times as **stromatolites**. Long before true plants evolved, photosynthetic cyanobacteria was busy for billions of years pumping oxygen out into the oceans (and thence by diffusion into the atmosphere).

This oxygen, being the highly reactive element it is, immediately reacted with whatever was suitable and available: carbon, for instance, making CO_2. Or silicon, making SiO_2. Or maybe iron dissolved in the seawater; that would oxidize and settle out as magnetite or hematite. But the rates of production matter: if oxygen was being churned out at a rate greater than the availability of these other reactive elements to buffer its rise, then it eventually began to accumulate in the atmosphere, available for reactions, but finding insufficient reactive elements to bond with. It built up and built up.

But while life was chemically fine for cyanobacteria, surrounded by their waste. Sulfur-reducing bacteria, for instance (the ones implicated above in end-Permian euxinia) cannot survive in the presence of free oxygen; it's a deadly poison to them. They are not alone: many groups of microbes cannot tolerate free oxygen. Some persist today in isolated low-oxygen settings such as swamp muck, but many others probably went extinct as the GOE took hold. How many exactly, and what were their characteristics? We'll never know: microbes just don't fossilize with the level of detail we would need to answer those questions. It seems reasonable to infer it was a great many, but we must become comfortable with the uncertainty about the specifics.

The Cambrian explosion

The Cambrian Explosion is justifiably celebrated as a great blossoming of diverse animal life forms, well recorded in the geologic record by their profusion of shells and other hard parts. But it was preceded by a time when soft bodied, sessile animals were widespread across the globe. These were the Ediacaran biota, a diverse group of organisms presumed to be animals. Some resembled jellyfish, and others looked similar to modern sea pens. Some might have been the ancestors of mollusks. But the suddenness of the Cambrian Explosion implies that it was an adaptive radiation, and therefore took place in an ecological space that had recently been cleared out of its previous inhabitants. The Ediacara had no chance against their Cambrian competitors, not only because they lacked armor, but also because the Cambrian critters destroyed their ecosystem.

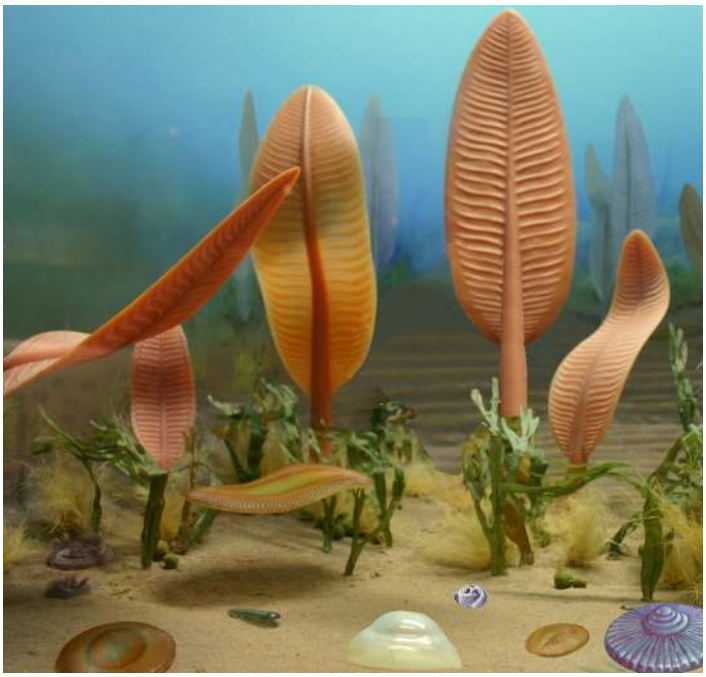


Figure 11.3.4: Artistic rendering of Ediacaran shallow seas. (Creative Commons Attribution-Share Alike 4.0 International license.)

The Ediacaran period was a time of microbial mat-dominated ecospace, apparently with very few organisms exploiting higher levels in the water column, nor were there any burrowing very deeply into the sedimentary substrate below. The lack of bioturbation in Ediacaran sediments suggests it was a “sealing” ecosystem, largely anchored to the sediment/water interface, and rarely deviating from it.

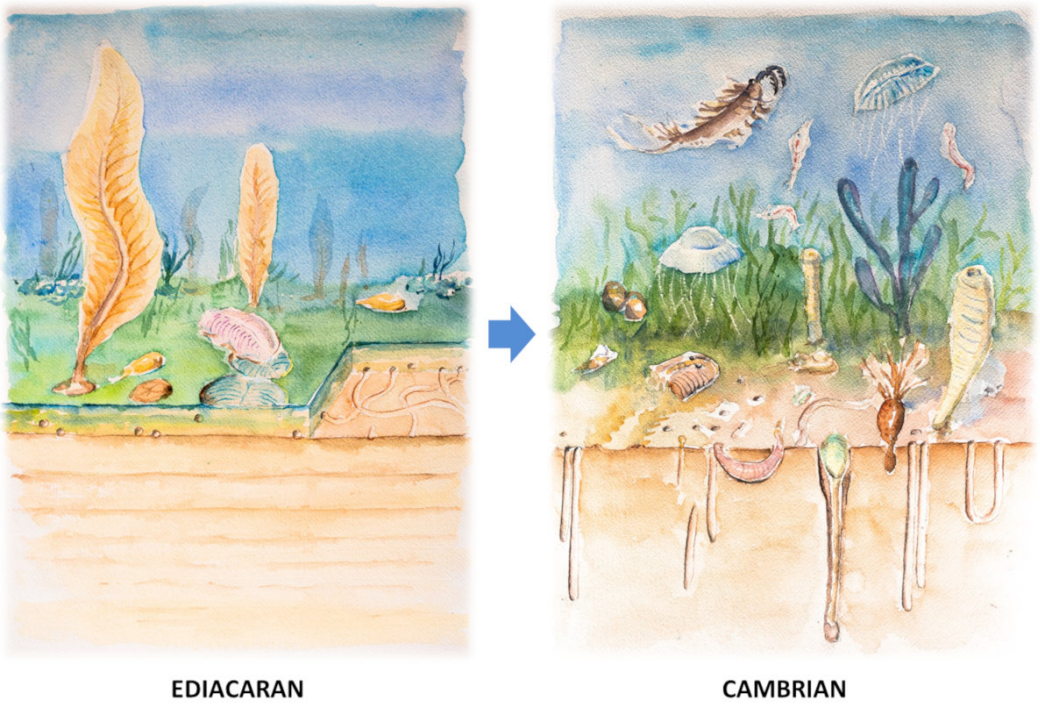


Figure 11.3.5: The fundamentally different ecosystem structure of the Ediacaran biota as compared to the Cambrian Explosion. (Watercolor painting by Richard Bligny for an article by Jean Vannier; inspired by an original in Fedonkin et al. 2007.)

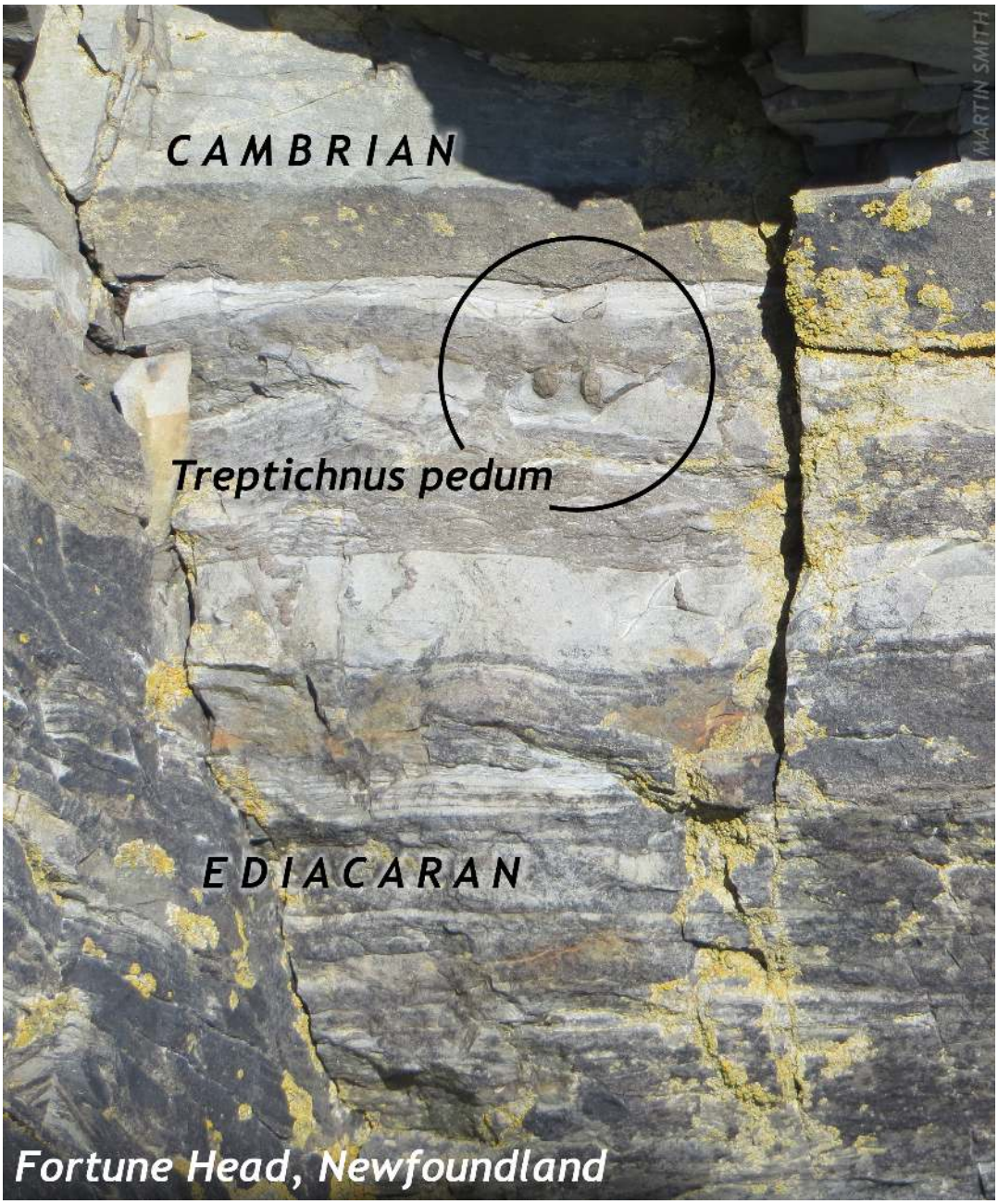


Figure 11.3.6: Treptichnus pedum defines the Ediacaran / Cambrian transition at the GSSP at Fortune Head, Newfoundland. (C. Bentley annotation of Martin Smith photo via Wikimedia.)

Once vascular body plans and hard parts had evolved, however, the diverse animals of the Cambrian were freed from being bound to the bottom of the sea. Some swam high above, such as the fearsome predator *Anomalocaris*, while many others burrowed into the sediment below. The latter group includes inarticulate brachiopods and various worms. These organisms dug into the sediment, using its bulk as protections against predation. In so doing, they churned up the sediment, disturbing the continuity of the bedding. This **bioturbation** is a "trace fossil" signal of the Cambrian Explosion. Indeed, the very oldest beds of the Cambrian are defined by the first appearance of the trace fossil *Treptichnus pedum*. (These burrows are presumed to be the probing burrows of priapulid worms, which continue to make similar traces today.)

By all means, celebrate the glories of the Cambrian Explosion, but spare a thought for the poor, doomed Ediacara that came before!

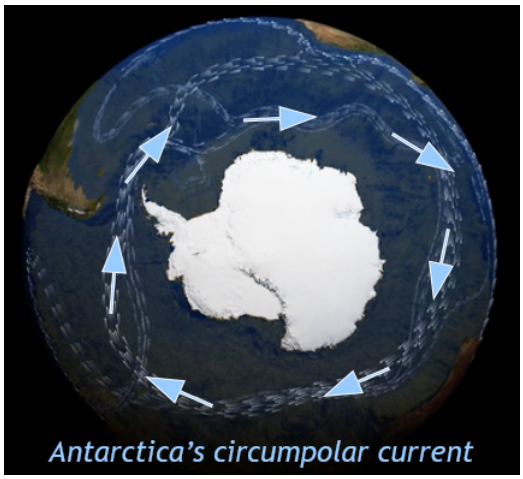


Figure 11.3.7: Antarctica's circumpolar current keeps it cold. (Modified by CB from a NASA public domain image.)

Halfway through the Cenozoic Era, Earth's climate shifted into a cooler state. The most obvious cause for this cooling was the separation of Antarctica from southern Australia and southern South America. This divergence opened up a swath of open ocean that encircled Antarctica on all sides. A new ocean circulation pattern developed, the circum-polar current, which wrapped around Antarctica like a doughnut. These endlessly orbiting waters cooled and cooled, isolating Antarctica from the rest of the world. In as little time as 100,000 years, deep sea ocean temperatures dropped globally by 4°-5°C. By 30 Ma, there were new glaciers growing on the southern continent.

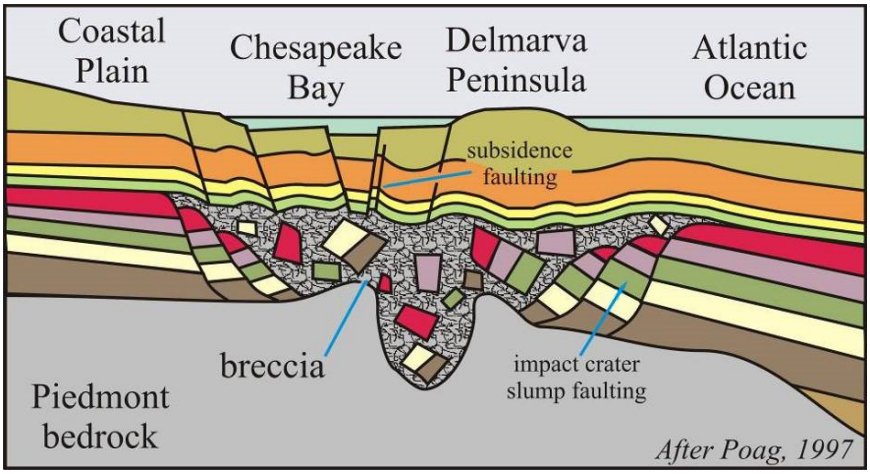


Figure 11.3.8: Schematic cross-section through the Chesapeake Bay Impact Structure, Virginia, USA. (Callan Bentley cartoon (2003) after Wylie Poag original (1997).)

Another series of events that may be relevant are at least three substantial bolide impacts in the northern hemisphere: Popigai in Siberia (35.7 ± 0.2 Ma), Toms Canyon offshore from New Jersey (~35 Ma), and Chesapeake Bay, Virginia (35.5 ± 0.3 Ma). As mentioned in the discussion of the end-Cretaceous impact, dust and debris may be thrown up to the stratosphere by an impact. Suspended in this rain-free portion of the atmosphere, the particles can screen out a portion of incoming solar radiation, reducing global temperatures.

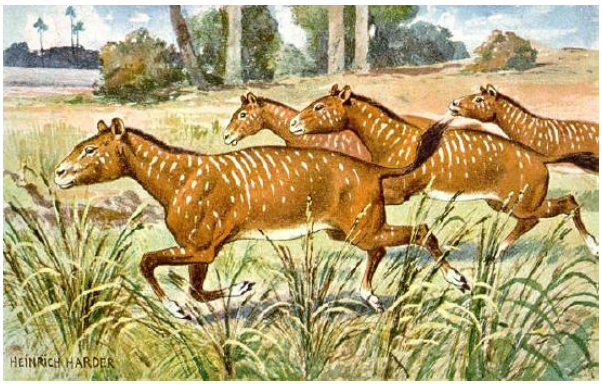


Figure 11.3.9: Mesohippus, an ancestral horse, at the time of expanding grasslands at the Eocene/Oligocene boundary. (Public domain; Art by Heinrich Harder (1858-1935), via Wikimedia.)

This shift in the climate resulting in a turnover of organisms, both in the sea and on the land. In the oceans, spiky foraminiferida were replaced by forms with less surface area. Bivalves adapted to warm conditions died out, replaced by more cold-adapted species. On land, grasslands increased, taking over land previously covered by forest. Mammals living on land had to adapt to these cooler, drier conditions – and we see a transition from browsers (that eat leaves off bushes and trees) to grazers (that "mow" the grass) at this time. Horses, rhinos, camels, and antelopes all spread at this time.

Pleistocene megafauna

Unlike the preceding examples, the most recent mass extinction skipped sea life, skipped plant life, skipped small mammals, and skipped nocturnal animals. It just focused on large mammals and birds. The dead include giant ground sloths, mammoths and mastodons, glyptodonts, huge bears, saber-toothed cats of several varieties, and (in Australia) marsupial equivalents of many of these. There were even vultures of unusual size, who presumably fed on the corpses of these great beasts. These last presumably enjoyed a great glut for a short while, and then saw their smorgasbord go bare. Other huge birds included the phorusrhacids, the elephant bird, and the moas.

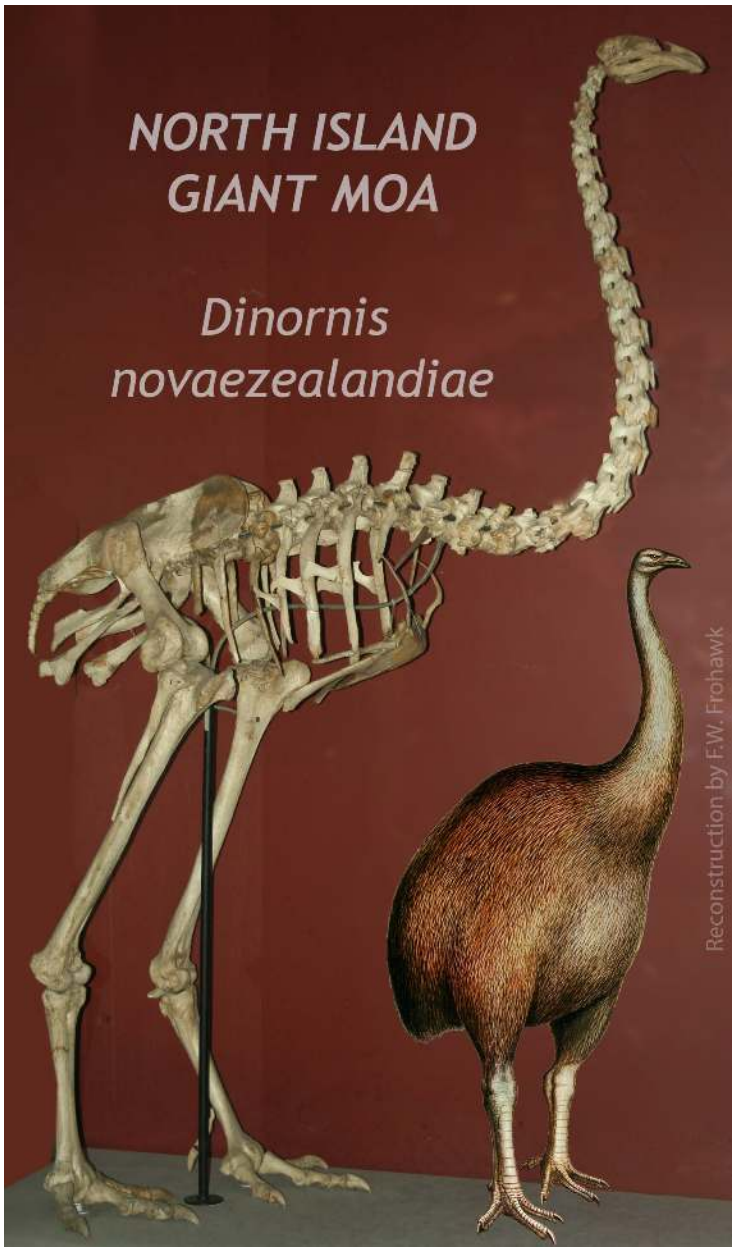


Figure 11.3.10: This moa was one of several dozen species that inhabited New Zealand prior to being killed off by humans. (Photo mashup of public domain imagery from Wikipedia.)

Various causes have been proposed for the extinction of these “megafauna,” from a new ice age to a meteorite impact. The ice age hypothesis is particularly weak, however, since the most recent Pleistocene glacial advance was just one in a series of dozens that stretch back in time to ~2.6 Ma. Why these animals did fine for the first several dozen but succumbed only at the most recent one is not explained. Evidence for a meteorite impact has been documented in recent years, but while it corresponds in timing with the North American extinction of the megafauna, it post-dates the Australian extinction event, and pre-dates the Madagascar extinction event.



Figure 11.3.11: A Woolly mammoth (*Mammuthus*, left) and an American mastodon (*Mammut*, right) facing each other, showing the physical differences between the two animals. (CC-BY; Image by Dantheman9758, via Wikimedia.)

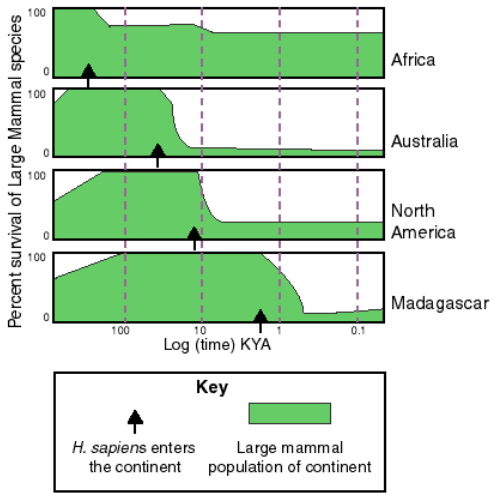


Figure 11.3.12: Timing of Pleistocene megafauna extinctions relative to humanity's arrival, organized by landmass. (Public domain; Diagram by ElinWhitneySmith (2006), after Martin (1989), via Wikimedia.)

Key to understanding the Pleistocene megafauna extinctions is the fact that the timing for the extinctions varied from continent to continent. Further, it seems to be strongly tied to the arrival of the first humans to arrive there. It happened earliest in Australia (~45,000 years ago), and most recently in the Americas (14,000 to 13,000 years ago), and more recently still in isolated islands, such as Madagascar or New Zealand. The Pleistocene "overkill" hypothesis (first articulated by Paul Martin) links the predatory habits of our own species with the loss of these giant beasts. This explanation is controversial, as some people see it as unfairly maligning indigenous people around the world.

The one place where the Pleistocene megafauna extinctions happened ***least*** was the one place where the megafauna had hundreds of thousands of years to get used to *Homo sapiens*: Africa. Africa is famous today for its surviving megafauna: elephants, hippos, rhinos, lions, and giraffes are the sizes and sorts of animals that used to be found on the other continents, too. This is a classic example of the adage "the exception that proves the rule" — where "the rule" is that when humans arrived in a place that was home to animals with no natural fear of humans, the humans rapidly killed the animals. Africa is "the exception," as this is humanity's birthplace, and the animals there have evolved in tandem with humans, developing a healthy aversion to them.

The modern biodiversity crisis

Right now, you and I may be living through another mass extinction, one brought on by humans' severe impact on the natural world. Dubbed "the Sixth Extinction" to denote its severity as approaching the Big Five, the modern biodiversity crisis is harder to get a handle on, since we are in the middle of it. The tools we use to assess mass extinctions recorded in the deep time record of rocks don't work for the present. So this is a trickier comparison to make. But there is a strong case that humanity's influence has been both geological in scale and negative in effect. Some geoscientists have teamed up with environmental activists to suggest that we should consider the current moment as a new, distinct geologic epoch, the Anthropocene.

Elizabeth Kolbert's Pulitzer-prize-winning book *The Sixth Extinction* is a well-written account of the various relevant factors. Uniquely among books about the current environmental crisis, it is exceptionally well grounded in the geological context. She notes numerous factors that contribute to the modern crisis:

- Habitat destruction and fragmentation
- Invasive species
- Disease
- Pollution
- Climate change

Rates of extinction documented in the modern day (using extant species and a historical timeframe) match or even exceed rates calculated from fossil turnover during the Big Five mass extinctions. However, these accelerated rates have only been operating for a geologically brief amount of time (a few centuries). The numbers are alarming, but it's not an ideal scientific contrast, since all the confounding variables haven't been controlled for. It's comparing "apples to oranges." That said, what other choice do we have? Humans cannot help the fact that they live in the present.

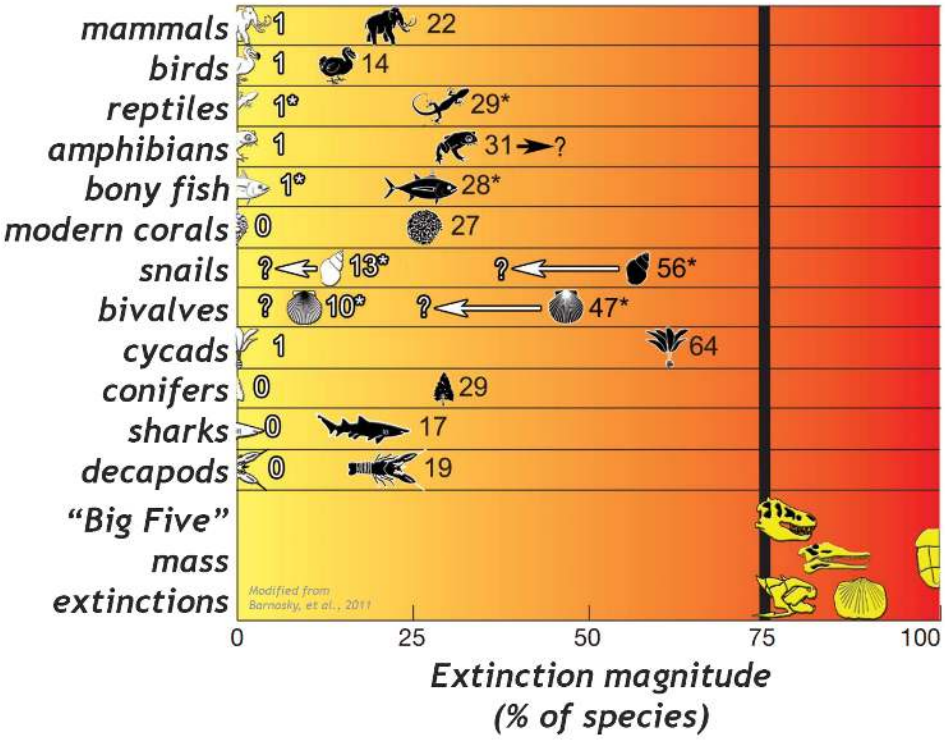


Figure 11.3.13: A comparison by Barnosky, et al. (2011) shows rates of species loss in modern groups, as compared to the "Big Five" in the fossil record. Numbers next to each icon indicate % of species. White icons indicate species 'extinct' + 'extinct in the wild' over the past 500 years. Black icons add currently 'threatened' species to those in white; the amphibian percentage may be as high as 43%.

In an attempt to assess the actual levels of extinction that have so far occurred and can reasonably be "blamed" on humanity, Barnosky *et al*. (2011) don't actually find very many. They examined species documented by the main international body the focuses on conservation biology, the IUCN. (*This is the group that maintains the "red list" of endangered species.*) How many of those species have gone extinct in the past 500 years? How does that compare with extinction rates determined from fossils in stratified rocks at each of the "Big Five?" Thankfully, they show that there is a pretty big discrepancy between modern numbers and ancient numbers. But the question of time hangs over this comparison – 500 years is almost nothing in the geological sense of time. It is hard to compare a modern rate of 1 species per 500 years with an ancient rate of 75% of all species per 100,000 years. The timescales are too radically discrepant.

If the modern rates play out for another 500 years (or 1000), will they cross the threshold of being a mass extinction? If the answer is yes, and we wait until then to decide it's actually a big problem, then it would be too late. Asking the question now allows us to take action to prevent a sixth mass extinction from becoming a reality.

The last life on Earth

Far in the future, perhaps 2.5 billion years from now (maybe sooner), life on Earth will cease. This will be the final mass extinction our planet is capable of pulling off. Two sets of processes need to be considered, those of Earth's interior and those of our star, the Sun. Plate tectonics is responsible for continuously revitalizing Earth's surface, but it is driven by heat from the planet's interior. Heat production is maintained through time by radioactive decay of unstable parent isotopes, such as $^{40}\mathrm{K}$, $^{232}\mathrm{Th}$, $^{87}\mathrm{Rb}$, and $^{238}\mathrm{U}$. But there are fewer and fewer of these isotopes every day. Our planet is cooling off, and eventually it may get too cool to maintain plate motions. Estimates of this sad date put it about 1.45 billion years in our future. A cooler core means less liquid outer core through time, as the solid inner core grows larger. Eventually, when the core becomes completely solid, the magnetic geodynamo of the outer core will cease, and our planet will lose its atmosphere-protecting magnetic field. This exposes the atmosphere (and the surface hydrosphere) to erosion by the solar wind. Life won't last much longer in those conditions – Earth would increasingly come to resemble Mars.

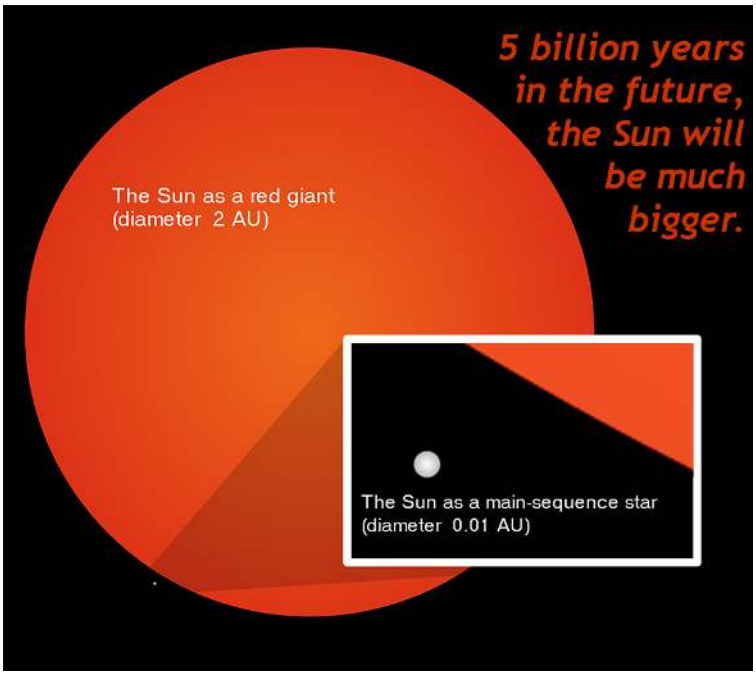


Figure 11.3.14: The Sun is forecast to grow in volume in the far distant future. This does not bode well for life on Earth. (C. Bentley modification of an original by Oona Räisänen, via Wikimedia.)

Regardless of these "local" details, the fate of our planet ultimately hinges on the future the Sun. The Sun is a G-type star, and those grow in volume through time as they burn through their hydrogen fuel. Ironically, as they lose mass, they swell up to a larger size. Less mass means less gravity holding them in a tight, dense ball; older stars have relaxed and stretched out into space.

In our solar system, that will mean that Mercury and Venus are likely fated to be consumed by the Sun as it expands outward. Earth's location is likely about where the surface of the Sun will reach in ~5 billion years when it becomes a elderly **red giant**, but a reduction in the Sun's mass also means a reduction in the Sun's gravitational hold on Earth, and so the planet will likely shift into a more distal orbit, further out from where we currently orbit. Nonetheless, the larger, closer Sun will fry the Earth, evaporating away its oceans and sterilizing its surface.

Franck, et al. (2006) made estimates of how Earth's temperature will respond to this swollen star, and compared these estimates of its future temperature to the tolerances seen in modern examples of various domains and phyla. Multicellular life is estimated to have 800 million years left, according to that study, or possibly up to 1.2 billion years, while single-celled eukaryotes will last longer, perhaps 1.3 to 1.5 billion more years. Prokaryotes are hardiest of all at high temperatures; it is thought they will persist until about 1.6 billion years from now.

And that will just about do it for life on Earth.

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11.S: Summary

To recap these mass extinctions, here is a summary table, in chronological order:

Table 11.*S*.1: Summary table.

Name	Date	% extinct* (taxonomic loss)	Putative cause
Great Oxygenation Event	2.4-2.0 Ga	¯_(ツ)_/¯	Cyanobacterial pollution of the atmosphere
Cambrian Explosion	~539 Ma	*all the Ediacara*	Hard parts arms race
Late Ordovician	~445 Ma	52%	Global cooling
Late Devonian	~372 Ma	40%	Excessive carbon burial due to the advent of trees + global cooling
End-Permian	260-252 Ma	83%	Ocean anoxia induced by eruption of the Siberian Traps
End-Triassic	201 Ma	73%	Global warming induced by the eruption of CAMP
End-Cretaceous	66 Ma	40%	Meteorite impact ± Deccan Traps
Eocene-Oligocene	34 Ma	15%	Global cooling
Pleistocene	18-10 ka	*(just megafauna)*	Human overhunting
The end of all life on Earth	1.6 billion years in the future (?)	100%	Sun expanding & overheating the planet

*Most numerical estimates from McGhee, et al. (2013).

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CHAPTER OVERVIEW

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12.1: Stratigraphy

> “Time, which measures everything in our idea, and is often deficient to our schemes, is to nature endless and as nothing; it cannot limit that by which alone it had existence; and as the natural course of time, which to us seems infinite, cannot be bounded by any operation that may have an end, the progress of things upon this globe, that is, the course of nature, cannot be limited by time, which must proceed in a continual succession.”
>
> James Hutton, “Theory of the Earth”, 1788

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12.2: The Story of Basins Past, Present, and Future

Plates move and collide, mountains rise, sea level ebbs and flows. All the while, chemical and physical weathering processes, buoyed by important interactions between the Earth's spheres, break rock down in place. After it breaks down, erosional forces powered by gravity and aided by wind, water, and ice provide the energy and work necessary to move weathered materials downhill, downstream, into a new place of rest. The story of these weathered sediments, large and small, has great variety. Numerous variables determine the outcome for these sediments. Latitude, the parent rock that weathered, the climate and degree of physical and chemical weathering, life forms that interacted with the grains, and so many more ultimately determine the story eventually told by the layers of sediment. As these layers accumulate, so does the tableau of stories they tell.

Strata

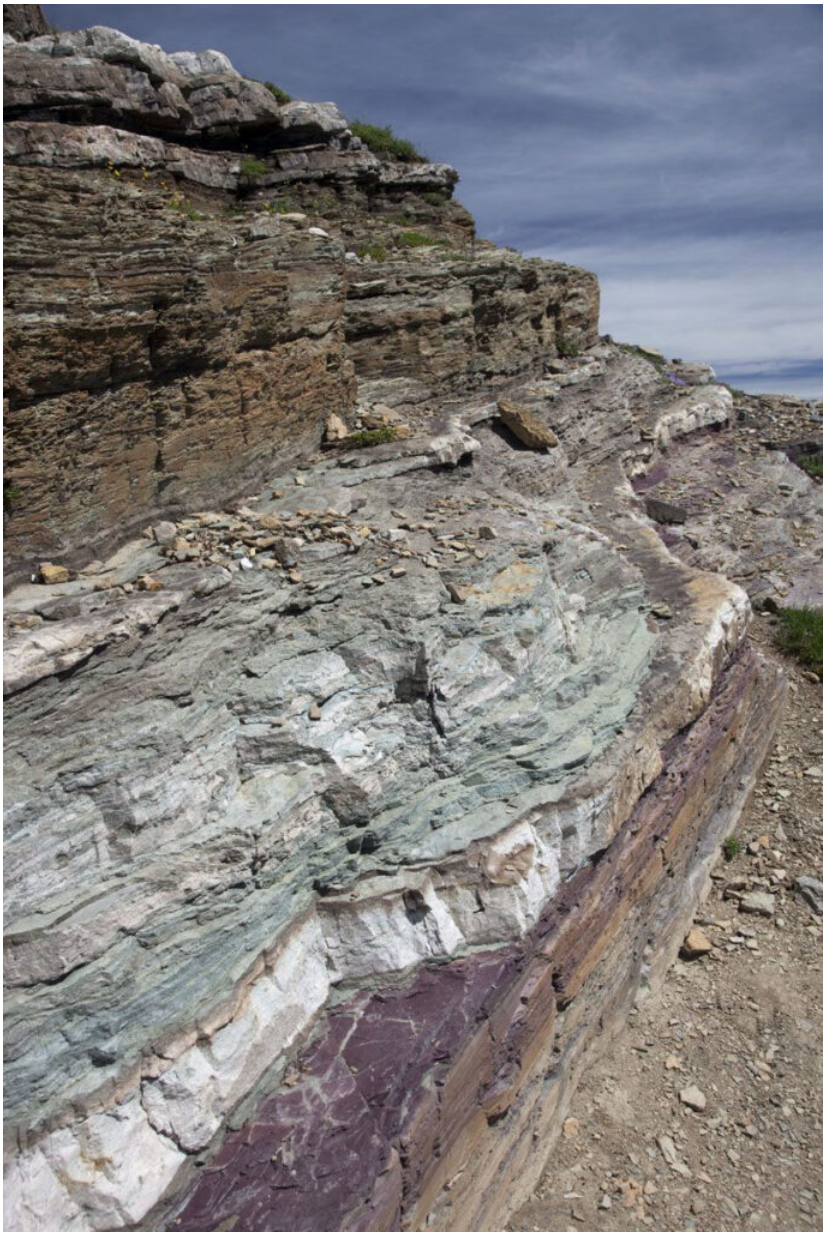


Figure 12.2.1: Brown, Green, white, and red sedimentary strata, indicating different levels of oxidation of disseminated iron in the rock. Proterozoic Spokane Formation, Glacier National Park, Montana. (With permission for educational purposes from Marli Miller Photography.)

Sedimentary rocks come in strata, or layers, much like a nice birthday cake or pages in a storybook, and stratigraphy is the study of the record of these layers. These rocks are formed from sediments originally deposited in a basin. A basin is an area of depressed

elevation generally surrounded by land at higher elevations. Basins can be on dry land or be filled with freshwater or saltwater. Over time, the land surrounding a basin erodes and the resulting sediment begins to fill the basin. From the point of origin in the highlands to the point of deposition in the basin, a great deal can happen to this sediment. Weathering, in various physical and chemical forms, takes its toll. By the time of their final moment of deposition, these sediments record a memory of their former existence as rocks, of their transport journey, and of processes that occur in the basin. Much of this sediment is solid and arrives as particles. Some of it arrives in aqueous solution, becoming the salts we experience in the seas around us. Some of it forms in the bodies of organisms, such as carbon-rich plant tissues or the skeletal remains of animals.

Depending upon the distance particles travel, some will remain the size of massive boulders and some will be reduced to the size of tiny flakes of clay. The range of grain sizes, or sorting, is the result of weathering and subsequent changes in energy as the particles move downslope. Some are well-rounded, others still retain their rough edges. Some sediment, found far from its source, will be composed primarily of quartz, with less stable minerals having rotted away. Other sedimentary deposits will still contain unweathered minerals like micas and feldspars. Once deposited, sediment and rock will begin to undergo physical and chemical processes associated with the local environment. Soil formation will begin on land. In marine settings, chemical processes associated with water chemistry and biogenic materials will occur, adding onto the earlier story of transport. These layers of sediment, accumulating over time, are the subject of stratigraphy.

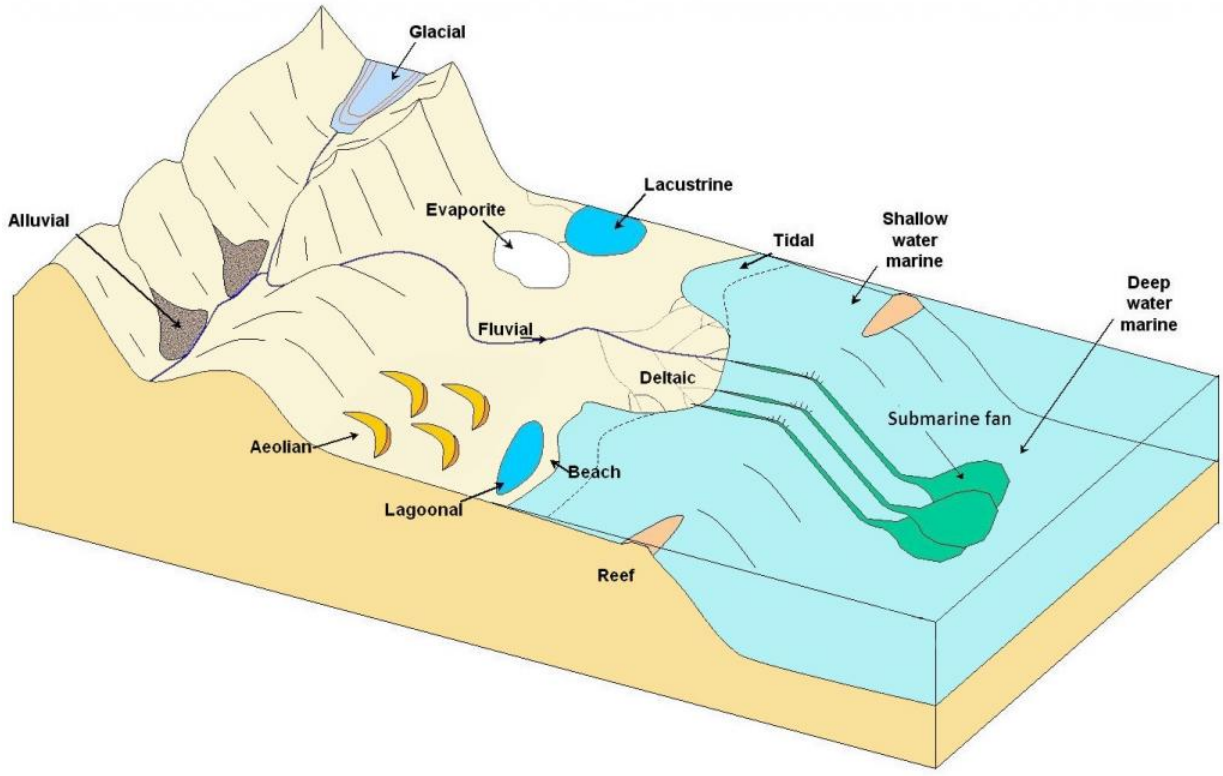


Figure 12.2.2: Sedimentary facies commonly found in a variety of basins.

The journey a sediment grain takes ultimately ends with deposition in a basin. Throughout that journey, sediments will take up residence for a time, waiting for a sufficiently energetic event to move them further along toward this end. The collective experience of sedimentation along these myriad pathways creates situations, as illustrated above, where different depositional environments are created. Geologists refer to the unique sedimentary characteristics of these varying environments with the term "facies". A sedimentary facies is defined as the body of characteristics of a sediment, or rock, that are indicative of the environment of deposition at the time. Real-time facies migrate, or change, at a location as conditions change. All of the environments in the figure above are dynamic in nature. We can also assume, using uniformitarianism, that similar environments have existed in the past, producing rocks in the geologic record that record similar characteristics we see in their modern counterparts. As sediments deposit and facies migrate, strata are generated. Geologists call such strata, beds. As beds representing similar environments accumulate and lithify, they will come down to us as mappable units called formations.

All of this action ultimately occurs in sedimentary basins.

Basins

Basins can be thought of as a library filled with books, which are the layers of rock that accumulate. These layers tell the story of global geologic changes, regional changes, and changes within the basin itself. There is great variety among sediments and other materials deposited in basins and in the layers, or strata, that they create. This variety and layering provides the equivalent of text on the pages of a novel or even in this textbook. Individual pages are the strata, or beds, and their characteristic portion of the story on that page are the facies. These pages are laid down horizontally, building up vertically over time, reflecting the order in which they occurred (superposition). Eventually, these pages (beds) accumulate into books (formations). Processes that occur after deposition, such as lithification, deformation, and further weathering and erosion can then make changes to this record that are akin to the deterioration seen in ancient Egyptian papyrus, or in a book whose pages and cover have been subject to the ravages of the elements. Strata can be torn (faulting), wrinkled (folding), or younger material may be injected (igneous intrusions). Depending on the context, these modifications may provide some confusion or clarification to the overall story. Over time, some of the material that makes up the strata and rock of this "book" can be lost or altered (erosion, metamorphism, secondary chemical alteration, etc.). Sometimes, even entire pages, or beds, are removed. Entire books, or formations, can be truncated, leaving gaps in the storyline (unconformities). Stratigraphers grapple with a record that was incomplete to begin with, and is further reduced over geologic time. Still, the remaining story these layers of sediment have to tell is powerful and important. It is of critical value in providing a more complete understanding of the history of our planet, including your own home landscape. Geologists who study this record are called stratigraphers. They are the historians within the geosciences.

Consider the office bookcase below as a stratigraphic metaphor:

Facies – The words, notations, etc. on an individual page of a book;

Beds = Pages within periodicals or books;

Formations = Books full of pages, come complete, some perhaps incomplete;

Groups of Formations = Shelves of books with similar topics (assuming they are organized as in a library!);

Bookcase = Stratigraphic record of the entire basin.

The limitation of this metaphor is that most of the "stratigraphy" you see here is vertical, implying a great deal of geologic alteration. Putting that aside and remembering the "Principle of Original Horizontality", we know that if these were rocks, they would be stacked differently.



The gigapan below depicts the Hampshire Formation. It was taken along Corridor H (US 48) in West Virginia. The image depicts a single formation. If you observe this formation, akin to a book, you may first see the varying color. These are beds. Some of these beds, or pages, could be broken down further into smaller beds. Ultimately, these beds are all related because they were deposited in similar depositional environments, as illustrated by their facies.



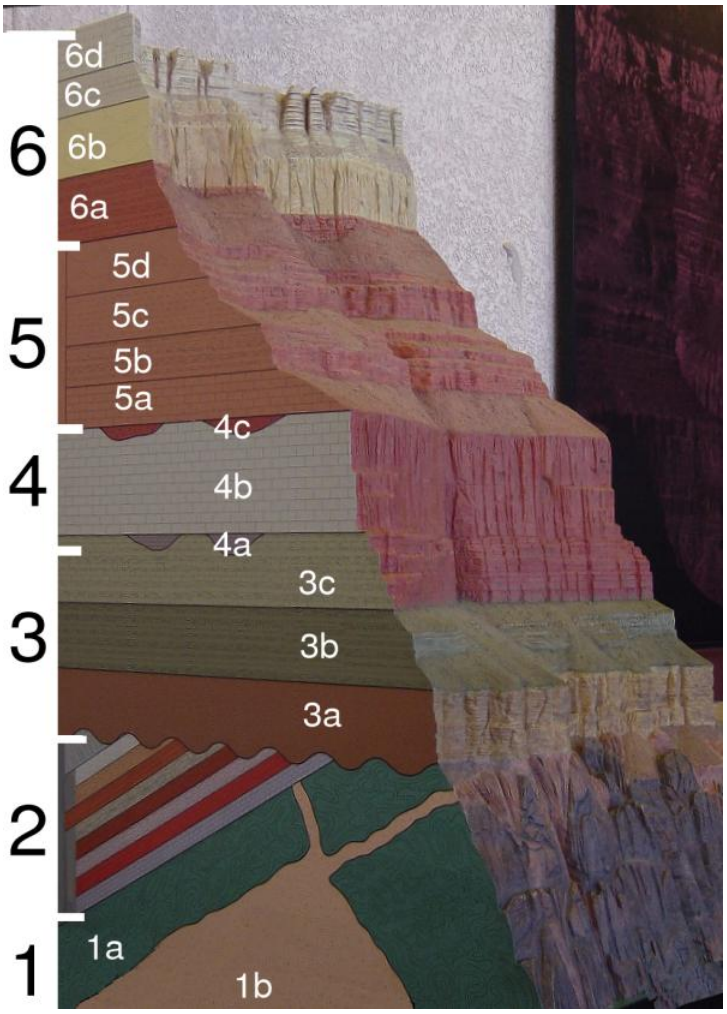


Figure 12.2.3: A geologic model of the Grand Canyon. Formations are indicated with numbers and letters (ie 6a through 6d are all formations, or books of rock). Formations are then assembled into Groups, where appropriate (ie 1 through 6). Groups are like shelves of similar books. (CC BY-SA 3.0; The original uploader was Maveric149 at English Wikipedia. – Transferred from en.Wikipedia to Commons, https://commons.wikimedia.org/w/index.php?curid=2257589.)

Layers of sediment are not static recipients of the story only. They are also participants. As we will learn in the next section, they are recording their story and being acted upon by outside forces such as tectonic change, sea level change, and lithification. But, they are also having their own impact on tectonics and sea level. Layers of sediment also play an important role in the record of geologic history that is important to understand also. As these layers accumulate in basins, they record the history of an area. Stratigraphy is the study of that history.

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12.3: Basins, Sea Level, and Accommodation Space

What is a Basin?

Sedimentary basins are geographic depressions where sediment accumulates. Some basins are epicontinental; they reside on top of the continental platform. Other basins are tectonic, located around the margins of the continents, after the continental shelf break. The power of stratigraphy lies in its ability to help scientists understand the history of basins in general. Basins sometimes contain water and sometimes do not. If the basin is dry, as with Death Valley, California, then the primary control on the creation of stratigraphic sequences would be changes in tectonics, such as the uplift or subsidence of the land itself. If the basin is connected to a body of water, then the cyclical rise and fall of this water will be recorded in the material deposited in the basin. Climatic and astronomical factors can cause changes regionally or locally that can affect the level of the water. The melting of ice or the thermal expansion of water are among the variables that can cause sea level to rise in a basin. Such changes typically do not occur within the lifespan of a single human. The record left behind is the pattern of sedimentary deposits. This pattern is preserved when the sediments become sedimentary rocks, and reading it is how we understand the history contained in this deep time.

The long-stable igneous and metamorphic cores of the original continents are referred to as cratons. Some portions of these cratons are exposed at the surface and some other portions are covered with blankets of sedimentary rock. The exposed portions are called shields. The portion of the craton that is buried is referred to as the platform. It is on these platforms where basins exist and where sedimentary strata accumulate.

First, sediments accumulate as individual beds. Eventually, groups of beds that are widespread enough to be geologically long-lasting will be referred to as formations. Platforms make up the continental areas in which formations accumulate. The stacking of formations is like adding pages to a book. Some platform environments are dominated by carbonate sedimentation while others are dominated by siliciclastic sedimentation.

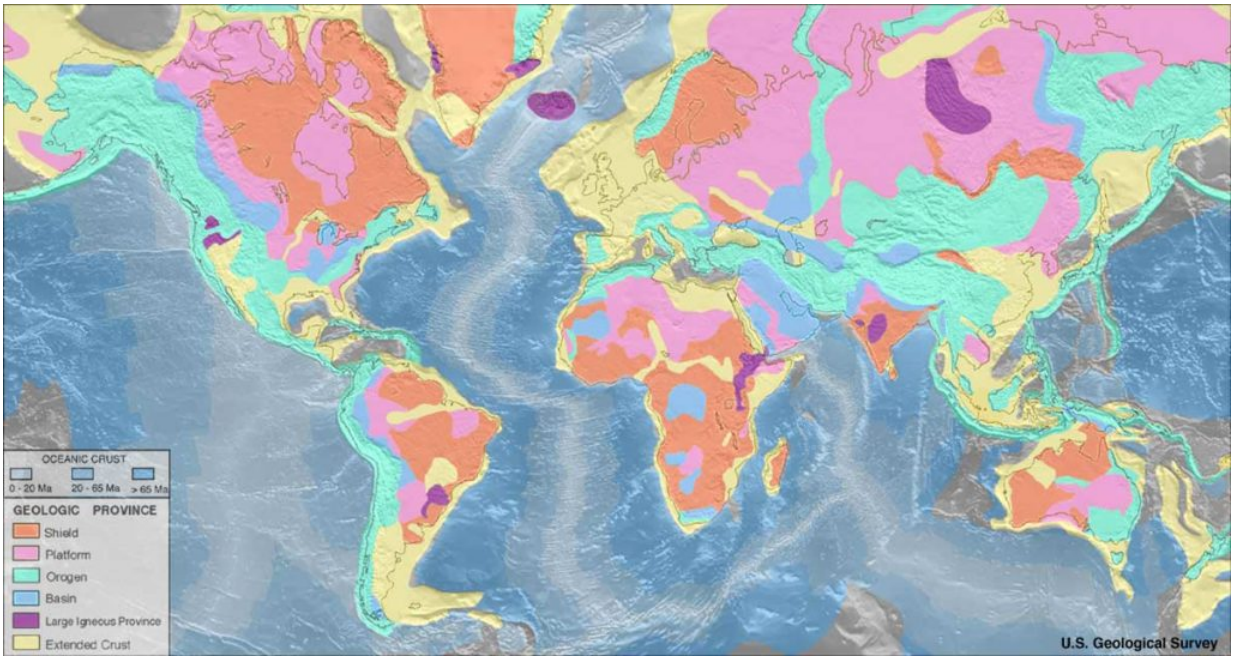


Figure 12.3.1: USGS image depicting shields, platforms, basins, recent orogens (mountain belts), and other features. The cratons are made up of the shield and platform areas. The basins, in blue, are just some of the many sedimentary basins that exist globally. In modern as in ancient times, the oceans themselves are basins.

All basins have limited space. The interplay between tectonic changes, global sea level, local water level, and sediment accumulation all affect changes in space within any given basin. This "accommodation space" is the volume of available within a basin to accommodate influxes of sediment. This concept will be dealt with in a bit more detail in chapter on "Sequence Stratigraphy." Accommodation space is not static. It adjusts as changes occur in tectonics, sedimentation, water level, and global (eustatic) sea level. As Earth's systems interact over time, the ability of a basin to accumulate sediment or even the amount of sediment being produced fluctuates. This is true of all basins.

Types of Basins

Several types of basins develop in different tectonic situations, each bearing the distinctive signatures of the conditions under which it originated. Let's now examine the sedimentary basins that form in divergent tectonic settings, convergent boundaries, and along strike-slip fault systems.

Rift Basins

As divergent boundaries rip apart a plate, a rift basin is produced. Similarly, at convergent subduction zones, rifting in a backarc basin can open up accomodation space for sediment. The character of such basins evolves significantly as the rift widens. Basins created by rifting occur along the edges of continents and well within the interior of the continent. At first, the basin will be intracratonic. This kind of basin is landlocked. Subsidence, a drop in land elevation, will occur as the rift develops due to thinning of the lithosphere. Once the topographically low rift develops a body of water, the nature of the depositional setting will also change. The result is an epicontinental basin.

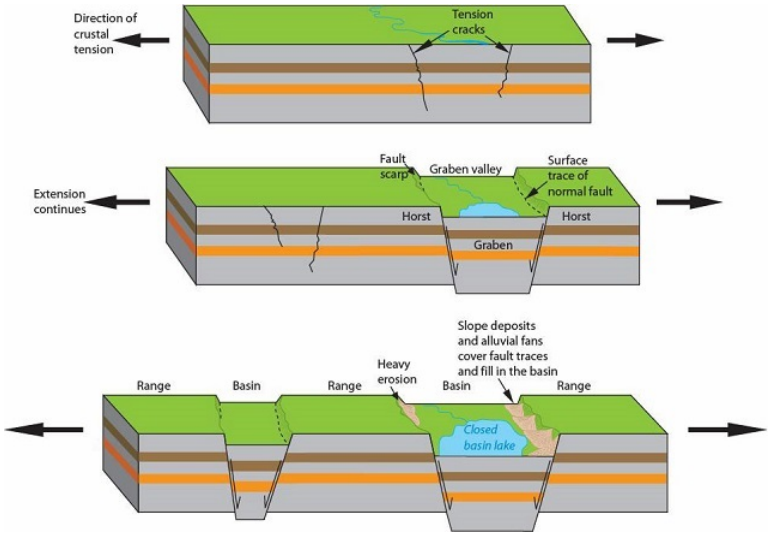


Figure 12.3.2: The development of a classic extensional basin structure. (Source: National Park Service, Trista L. Thornberry-Ehrlich, Colorado State University.)

In these locations, basins form as the graben subsides during the ripping apart of the crust. At this point, the basin may be entirely devoid of water and in fact will stay dry unless it intersects with a body of water or is not located in a wet climate.

In the case of the East African Rift, the only bodies of water within the rift valley itself are lakes fed by rivers. The thinning of the lithosphere in such basins can lead to volcanism. The East African Rift, as depicted below in two images, is an excellent example of an intracontinental rift basin that contains active volcanism. Famous mountains like Mt. Kilimanjaro, Mt. Kenya, Oldoin'yo Lengai, and others have important places in the cultural traditions and mythology of the people from these areas. They also present an interesting array of igneous materials.

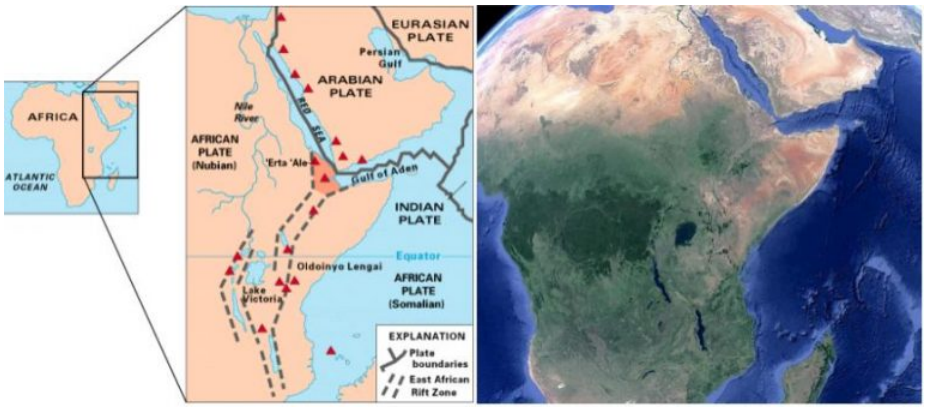
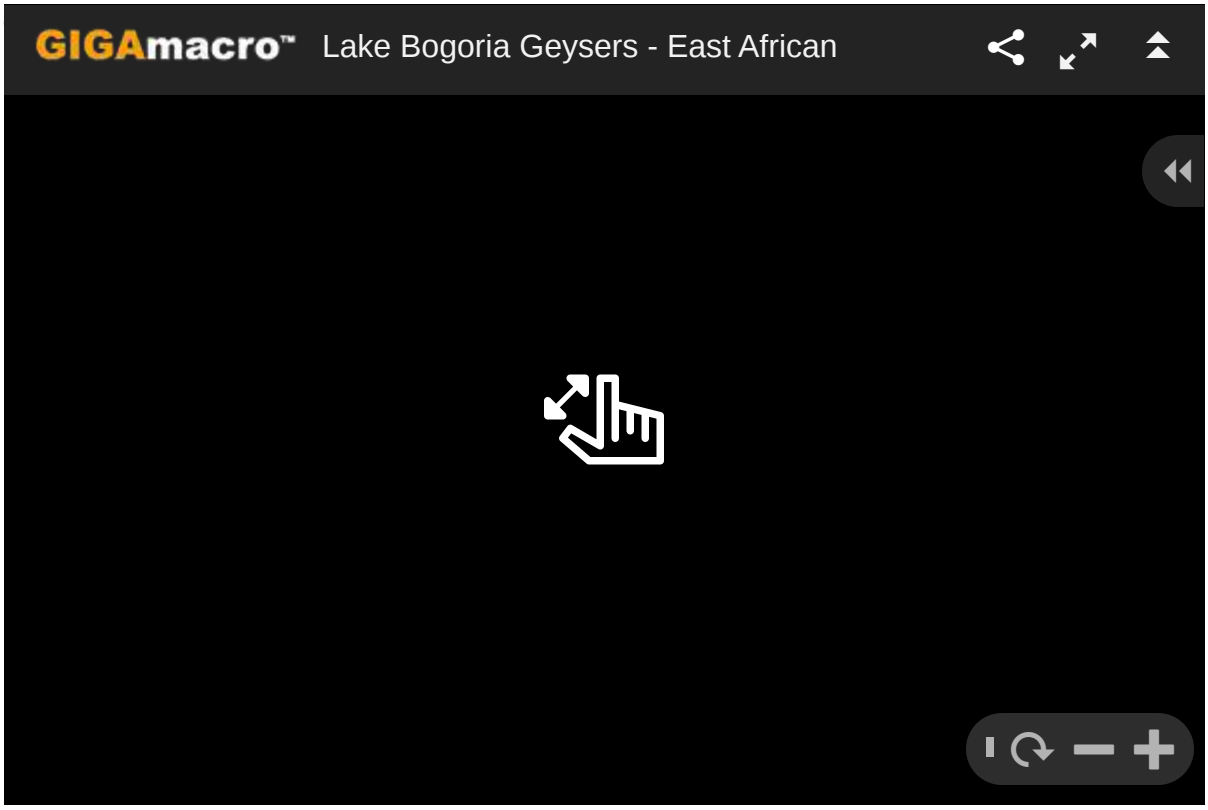


Figure 12.3.3: The East African rift zone includes the terrestrial rift basin as well as two other basins filled with water, the Red Sea and the Gulf of Aden. The unique geography of eastern Africa is due to the geology of the East African Rift. Extending from the southern tip of the Red Sea to the country of Mozambique, where Lake Malawi, the southern most linear lake feature in this image, can be found.



GigaPan of Lake Borgoria with Geysers

Passive Margins

Our modern ocean basins are the ultimate in sedimentary basins. Many of them, the Atlantic Basin in particular, are the result of ancient rifting. These ancient rift margins are also referred to as "passive margins". Major tectonic activity is generally absent at these locations. These are basins where active sedimentation is occurring. The characteristic of the sedimentation is, as always, dependent upon the environment, its energy regime, and other factors. The character of local sediment is indicative of its facies. As such, what you find at a passive margin beach, like Myrtle Beach, SC, is very different than what you find well off of the continental shelf in the abyssal plain, where you might find a siliceous or calcareous ooze.

The GlobSed map below, compiled by the Center for Earth Evolution and Dynamics, illustrates the current positions of epicontinental marine basins around the globe today. It also highlights their thickness. Note that many of the very thickly sedimented basins are where major rivers flow into the ocean. The Gulf of Mexico is one, but also the Mediterranean Sea, and the

Bay of Bengal. Nearly every continental margin outside of the Pacific Ocean (and even some of those are!) are examples of very ancient rift basins. It is worth noting that land-based rifting, such as in East Africa, is not highlighted here due to the nature of this data.

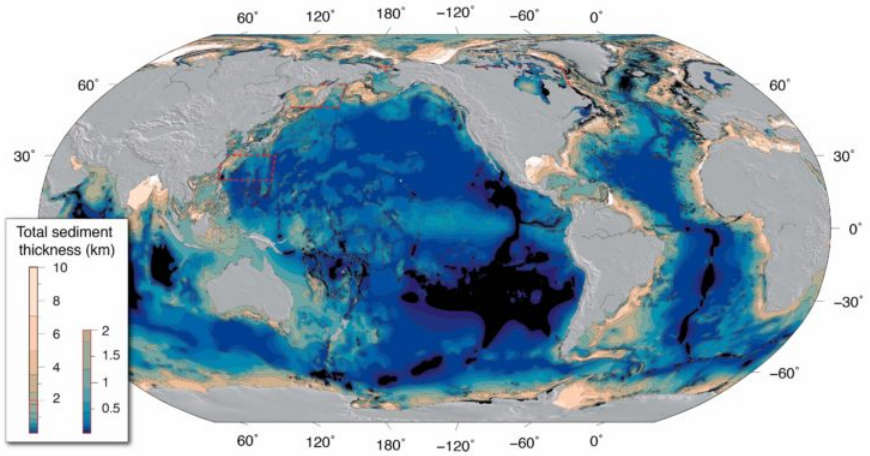


Figure 12.3.4: Total sediment thickness in the world's ocean basins, NOAA (Straume, E.O., Gaina, C., Medvedev, S., Hochmuth, K., Gohl, K., Whittaker, J. M., et al. (2019). GlobSed: Updated total sediment thickness in the world's oceans. Geochemistry, Geophysics, Geosystems, 20. DOI: 10.1029/2018GC008115)

Sedimentation in Rift Basins

All sediments that have accumulated in the basin have resulted from the weathering of the horsts that form the surrounding mountains, from the materials erupted by volcanoes, or from the marine or freshwater itself that has filled the basin. Rift valley basin sediments are a mix sediments from all of these sources. Sedimentary beds formed from surrounding mountains are dominated by quartz and feldspar, often producing arkosic sandstones. The East African rift meets up with two other rift zones in its northeastern-most area, the Gulf of Aden and the Red Sea, both of which are flooded rift valleys. These areas record not only the sediments weathered during the early opening of the rift, but also now the substantial accumulations of material deposited in these seas. The shorelines of the Gulf of Aden and the Red Sea, like the shorelines of all landmasses on the margins of rift zones, represent passive-margin deposition.

The east coast of the United States is an example of a very ancient passive margin, a remnant of the rifting of Pangaea. Key evidence for this not only comes in the form of the Mid-Atlantic Ridge itself, but in the array of Triassic/Jurassic rift basins that run parallel along the east coast from Nova Scotia south into North Carolina. Over the intervening millions of years, there has been ample time for coastal wave action along the Coastal Plain of these states to winnow out most everything but quartz sand. Eventually, such passive margin sands will produce excellent sandstone rich in quartz and poor in other minerals. These are referred to as quartz arenites.

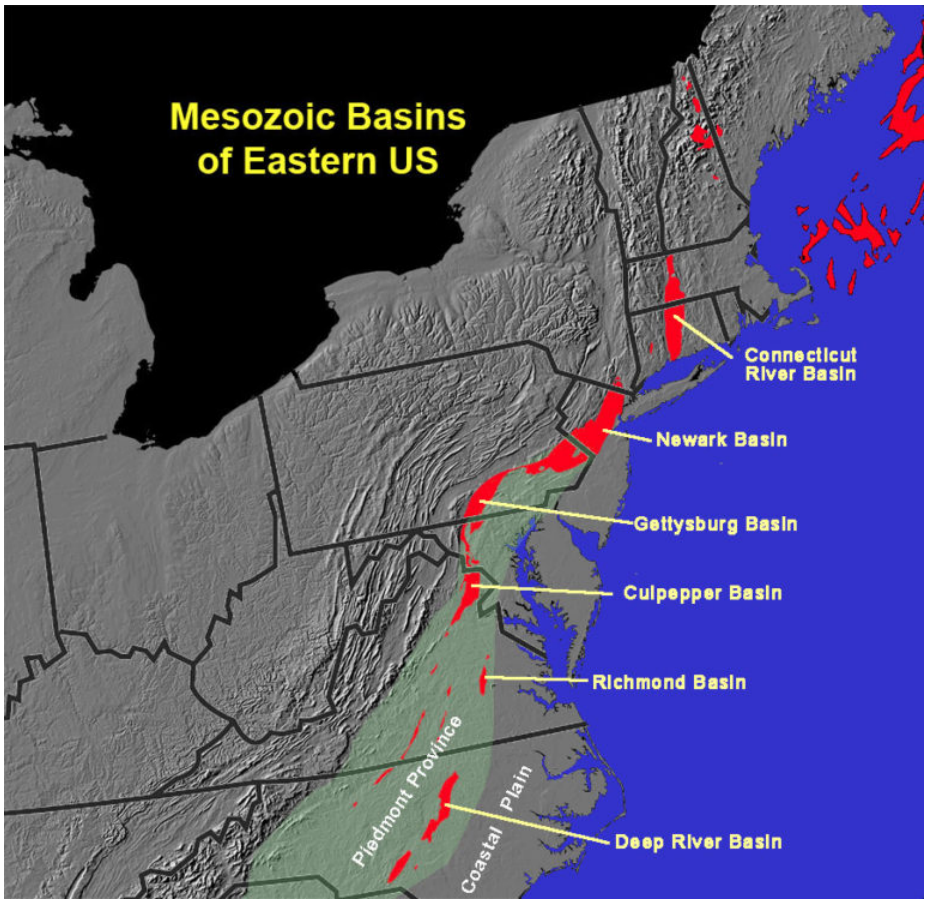


Figure 12.3.5: Mesozoic rift basins of eastern North America. These basins are filled with Triassic and Jurassic age sediment that sometimes contains evidence of dinosaur bodies (body fossils) or activities (trace fossils).

Not all rift zone basins flood with water, and none rift forever. Some rift basins fail and stop rifting. Others continue rifting for hundreds of millions of years, forming ocean basins like the Atlantic Basin. At some point, a rift basin will cease its stretching due to changes in tectonic stress. When this happens, space within the basin will be reduced as eroding sediments continue to fill the basin until there is no “accommodation space” left, barring further tectonic or isostatic changes. While the basic storyline is very similar for any kind of rift basin, the types of sediments deposited and resulting secondary changes will define the story in much greater detail.

Subduction-Related Basins

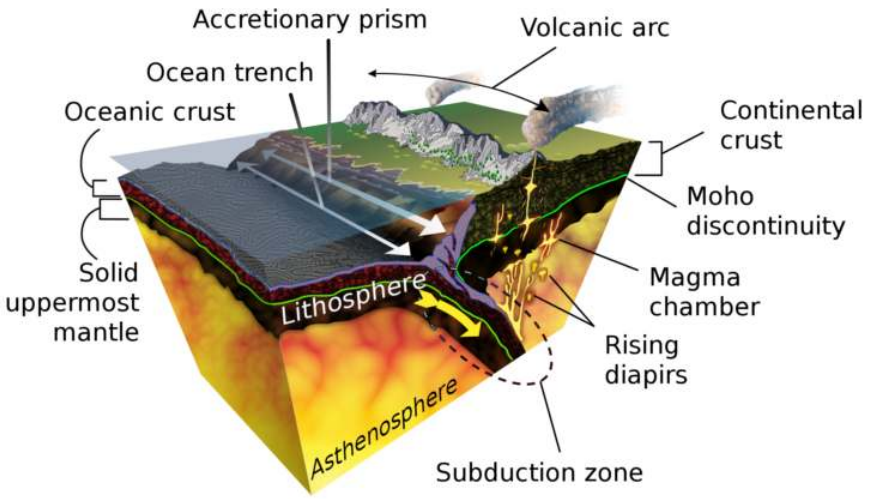


Figure 12.3.6: Subduction-related basin. Ocean floor sediments and eroded terrestrial sediments mix with planktonic sediments to create a unique mix of materials in these basins.

At subduction zones, there are three basins where sediments can accumulate. These include forearc basins, backarc basins, and the trench. Trench basins are deepwater settings formed where the subducting and overriding plates meet. Forearc basins form between the volcanic arc and the trench. Backarc basins form opposite the subduction zone, on the other side of the volcanic arc. Because of the tectonics involved, backarc basins tend to be extensional in nature. Such basins were discussed in the section on rift basins above. In the cross section below, the setting is an oceanic island arc and the backarc basin has thinned enough to develop into a rift basin. Backarc basins need not be submerged by water, as they can also occur on land.

Sediments in Subduction Zone Basins

Trench basins source their sediment from three sources: those sediments produced in the water column, those weathered from the overriding plate, and those scraped off of the subducting plate. Some sediments are generated in the water column by organisms that then rain down into the trench. These are generally siliceous, but can also be carbonate, if the depth is not too great. The balance of the sediment comes from land, sometimes via submarine mass movements like turbidity currents. Finally, sediment is conveyed to the trench atop the subducting plate, sometimes riding passively over great distances.

FOREARC BASINS

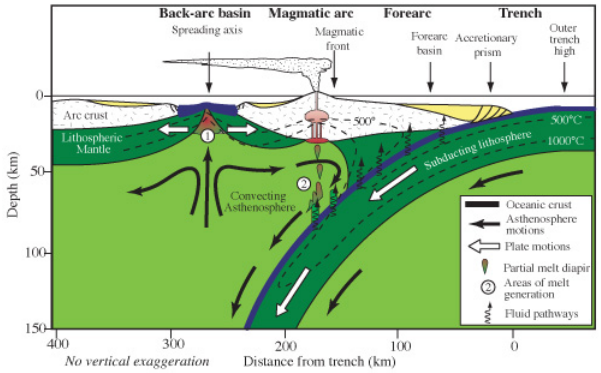


Figure 12.3.7: A cross section of a subduction zone, including both a forearc and backarc basin. Backarc basins can eventually develop into divergent margins over time and under the right conditions.

Forearc basins form between the accretionary wedge and the volcanic arc areas of a subduction zone and can occur both on land and underwater. The location of the foreland basin is dependent upon whether the subduction occurs between oceanic and continental lithosphere, or oceanic and oceanic lithosphere. In all cases, the basin is the result of downwarping, or flexure, of lithosphere created by the subducting plate. Such basins will contain a mix of siliciclastic material weathered off the volcanic mountains on its margin, volcanic deposits formed during eruptions of the volcanoes on the overriding plate, along with

biochemical sediments from life forms living within any water in the basin. Lithic sandstones are common in such basins because sediments do not travel long distances before deposition. An excellent modern example of a forearc basin can be seen in the Central Andean Forearc in South America. At this location, there are actually two forearc basins, the Atacama Bench and the Iquique Terrace.

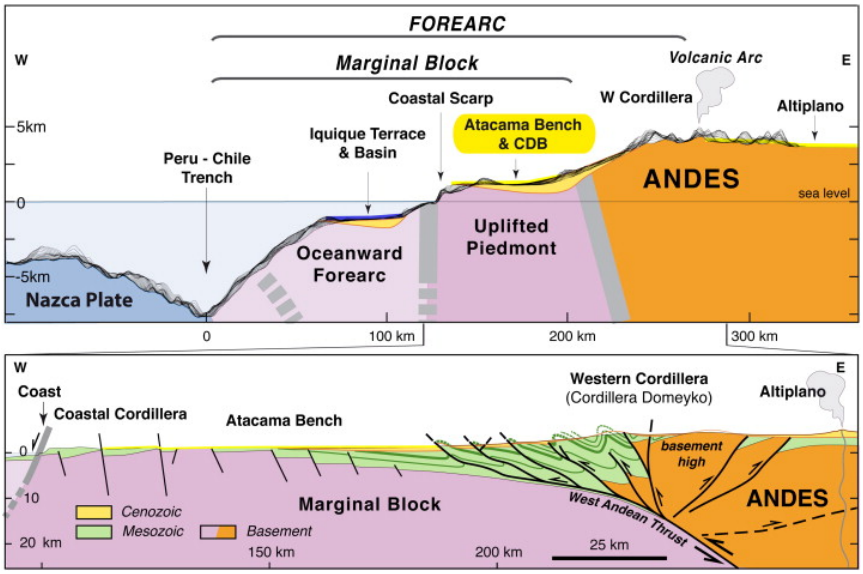


Figure 12.3.8: Simplified geology of the Central Andean Margin. Two foreland basins can be seen here, the Iquique Terrace and Atacama Bench. Both are the result of downward flexure of the lithosphere resulting from subduction of the Nazca Plate beneath the South American Plate (Used with permission: Armijo et al., 2015).

Because of the compression involved with the tectonic processes in these areas, the sediments have often undergone intense metamorphism, leading to large amounts of secondary mineralization. These metamorphic signatures, such as those contained in the Franciscan Blueschist below, are important for historical context. Before metamorphism, these were ocean floor sediments, deposited in a basin. They were then scraped up as subduction occurred and altered to their present form. The metamorphism itself is a part of the history of these sediments, even if the rock is no longer sedimentary. It is a great example of how non-sedimentary rocks are also important players in teasing out the stories rocks have to tell.



Franciscan Blueschist GigaPan

FORELAND BASINS

Foreland basins are the inland portion of the backarc region at subduction zones. They form parallel to mountain belts. They result from the massive crustal thickening and the downwarping of this increased mass on the lithosphere. This process is called lithospheric flexure. These basins vary in width according to the ability of the underlying lithosphere to flex. In all situations, the primary sediment flux into foreland basins is material eroded from the nearby mountain belt. Sedimentary strata are much thicker when proximal, or close to the mountains, and thinner in more distal, or far away, areas from the mountains. The thickness of the strata of sediment contained in such environments can provide information about the depth and lateral extent of the ancient basin. Studying these details helps sedimentologists apply these same principles to ancient foreland basin environments, resulting in more accurate models of basin development.

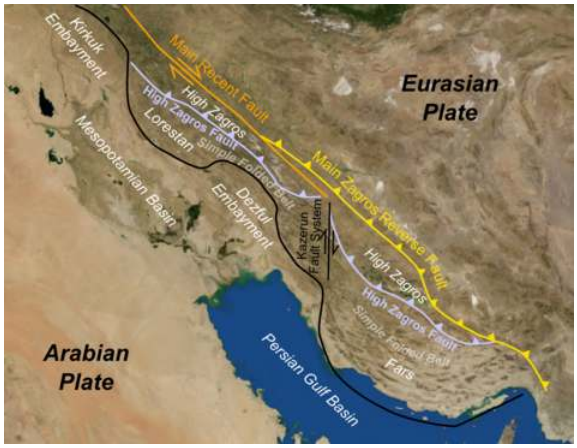


Figure 12.3.9: The Zagros Fold and Thrust Belt of Iran, Iraq, and environs. The Persian Gulf and Mesopotamian Basins are the foreland basins for this complex.

There are many excellent modern examples of active foreland basin deposition. For example, the Persian Gulf is the foreland basin for the Zagros Mountains of Iran.

Many ancient examples of foreland basins exist across North America. During the development of the Appalachians, the successive Taconian, Acadian, and Alleghenian Orogenies left behind a record of successive foreland basins deposits. These “sequences” of sediments form the Valley & Ridge province of the Appalachian mountain belt, and will be discussed later in this chapter. Their sedimentary record is prevalent along the eastern coast of North America, from Maritime Canada to northern Alabama and Georgia. Beginning with the limestones of the “Great American Carbonate Bank” of the Cambrian and Ordovician Periods, these sedimentary strata represent thousands of feet of deposition over several hundred million years. The most recent layers, deposited as the ancient supercontinent of Pangaea developed during the Permian Period, make up the hills west of the Appalachian Mountain chain. These Appalachian foreland basin strata contain economically important deposits of coal and natural gas. These deposits are evidence of the ancient seas and swamps that once existed in these regions.



Figure 12.3.10: Appalachian Foreland Basin during the Middle Devonian Acadian Orogeny. This portion of Laurentia (ancestral North America) was replete with foreland basins resulting from the Acadian Orogeny tectonic activity.

In the Cretaceous North American west, another important foreland basin developed that is known as the Western Interior Seaway. This seaway not only preserves some of the most marvelous marine invertebrate and reptilian fossils from that period, but also some of the iconic and important dinosaur fauna that lived in coastal environments of the time.



Figure 12.3.11: The Western Interior Seaway, visible on this globe over the North American states of Colorado, Texas, Oklahoma, Kansas, etc., is a foreland basin developed during the Sevier Orogeny.

Strike-Slip Basins

Strike-slip basins, or pull-apart basins, form due to the pulling apart and resulting subsidence of land along strike-slip faults. They can also form along transform boundaries. As shearing forces rip the lithosphere, blocks of lithosphere can detach and drop down relative to the surrounding land. This lateral and diagonal motion and associated subsidence should not be confused with the subsidence of horst and graben structures discussed earlier in rift basins. Good modern examples of such basins occur along the San Andreas Fault Zone in California and the North Anatolian Fault Zone in northern Turkey. In the San Andreas Zone, a classic result is the generation of the Gulf of California, running between the Baja Peninsula and the northwestern Mexican state of Sonora. This basin is the result of a series of pull-apart strike-slip basins that have effectively led to enough subsidence that the region has been infilled with significant amounts of water, forming the narrow Gulf seaway. Smaller examples can be seen along the Hayward Fault in Fremont, California.

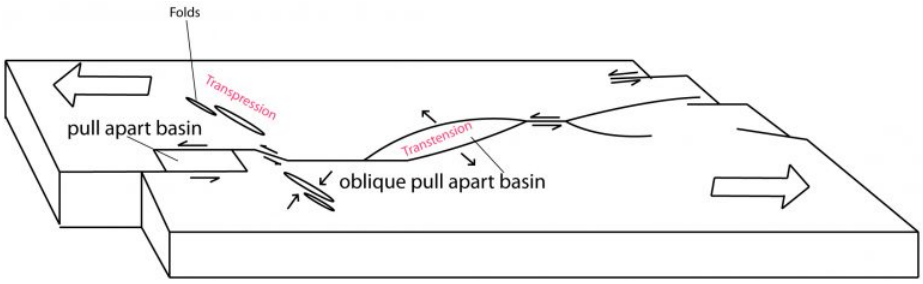


Figure 12.3.12: Strike-slip, or “pull-apart” basin situation. Where lateral plate or fault movement occurs, basins can develop due to the lateral or oblique (diagonal) motion created (CC BY-SA 3.0; By Wtbarron – Own work, https://commons.wikimedia.org/w/index.php?curid=32163143).

Sediments in Strike-Slip Basins

Sediments in strike-slip, or pull-apart, basins will typically only consist of the materials weathered from the uplifted sides of the basin, bounded by faults. In the case of the Gulf of California, sediments will also include marine biogenically-derived sediments and sediments derived from shoreline processes.

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12.4: Basic Principles of Stratigraphy

The history of a basin is recorded in its strata

Throughout the Earth's 4.5 billion year history, the interactions between the geosphere, hydrosphere, atmosphere, exosphere, and biosphere have evolved. These systems have evolved together throughout Earth's past and, over time, the results of this interplay has also changed. The introduction of the biosphere during the Archean Eon, for example, had a significant effect on how these systems have interacted with one another. The interplay between these spheres and the progressive nature of their development together has left a long rock record. While the record is incomplete, much of this long history is recorded in basins as sedimentary rocks. Sediments, deposited in basins and containing the evidence of this past, accumulate under the force of gravity in layers. It is these layers that make up the strata of stratigraphic record. These records vary across time, geography, and tectonic situation, but there are some basic principles, a kind of "rock grammar," that guide our interpretation of their story. These tools of stratigraphy are the geoscience equivalent of the Rosetta Stone, allowing us to peer into past worlds and translate what we find.

Strata and Stratigraphy: The Pages of Earth's History

Sedimentary layers are called strata. Strata form the stratigraphic record of the basin. The word stratigraphy comes from Latin roots that can literally be translated as "layer writing." Weathered sediments are eroded into basins and deposited in laterally extensive, horizontal layers. These strata form bedding and beds. Beds of strata accumulate into formations, the functional unit of stratigraphy. Formations, as collections of strata deposited in related environments, represent a collective facies. A facies is a body of rock with specific characteristics, such as the quartz sands and cross-bedding structures that might make up a sandstone that was once a beach. The "principle of original horizontality" describes this tendency for strata to form in a flat orientation. All sediments that eventually lithify into rock formations were originally laid down in a horizontal fashion. As these horizontal strata accumulate in the basin, younger strata will lie on top of older strata. This is called the "principle of superposition." Both of these principles were first formulated to Nicholas Steno in 17th Century Italy, and were the result of years of observations he made of rock strata in the Italian Alps. Horizontal and successively younger strata will retain much information about not only the environments in which they were deposited, but also the original rocks from which they weathered before that. Each layer of strata contains a detailed story.

The gigapan image below illustrates these layers, or strata, very nicely. This formation is no longer horizontal, but was originally. It was tilted by tectonic activity. Prior to that activity, these layers of alternating sandstones, siltstones, and shales were deposited deep under water as multiple series of submarine landslides. These are called "Bouma Sequences", and begin with the massive light tan sandstones which become gradually finer in grain size upward through the package. Each package is a new deposit, a new landslide at the time. The oldest rocks in this image are in the lower right and the youngest in the upper left.



Another very critical aspect of stratigraphy is understanding that depositional environments, preserved in vertically stacked strata occurring in geologic contact with one another were also at one point, laterally adjacent depositional environments. This is called "Walther's principle." Basins are not static and sea levels change for a variety of reasons. As they do this, environments that exist in time next to one another will migrate laterally. Over time, they will stack on top of one another. Eventually, this creates the sequences we see of rocks at outcrops, similar to the layers of cake in the image below.

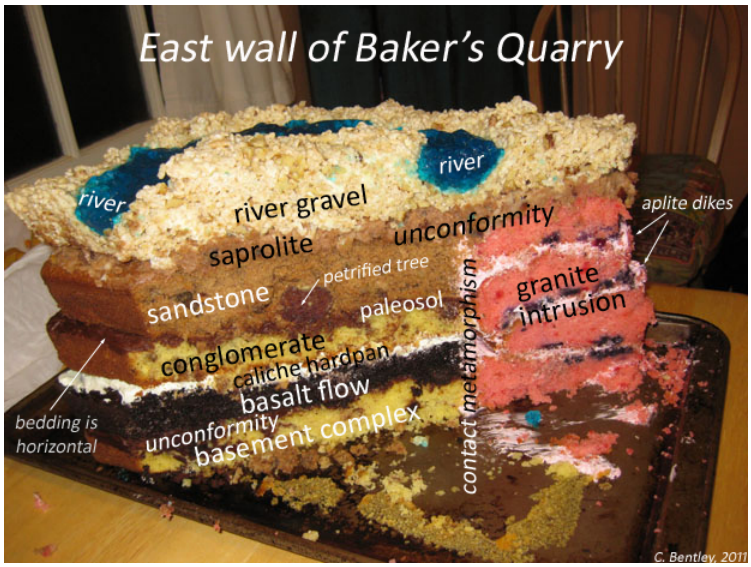


Figure 12.4.1: "East Wall of Baker's Quarry". This layer cake is another ripe metaphor for stratigraphy (Source: Callan Bentley, AGU Blogs. For a detailed analysis of this stratigraphy, click here).

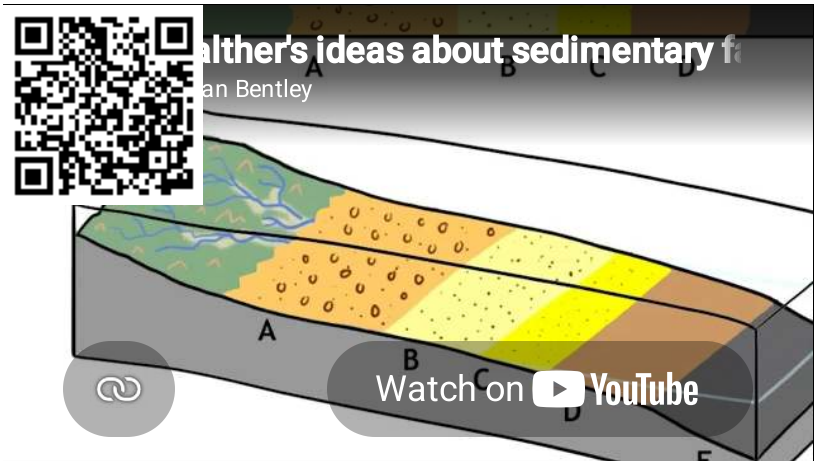
Formations and Walther's Principle

As mentioned above, when sediments are deposited into a basin, they are affected and also affect the basin. Where sediments ultimately come to rest is dependent upon the energy level of the system and how it is able to handle sediment grain size. Generally,

the energy level of water moving sediment decreases as gradient shallows, from mountains to the sea, gradually leaving behind successively smaller particles. Beaches, locations many of us are very familiar with, are full of sand because the energy levels at the seashore are well-suited to that grain size. In deeper water, grain sizes diminish even more, until all you find are marine muds well offshore. These successive sedimentological changes, along with the flora and fauna adapted to living in them and the unique structures created are how we determine local facies. The facies are the characteristics of a rock, formed under its unique depositional environment. Along a seashore, sedimentary facies will display a gradual decrease in grain size from areas well above the tidal zone (supratidal) to areas well below wave base (deep marine).

In these seashore locations, one of the key variables that affects how strata develop through facies are changes in sea level. Sea level change is continuous. Sediment deposition along shorelines is also continuous during these changes in sea level. When sea level rises, we call it a transgression. When it falls, we call it a regression. As water level rises, or transgresses, something interesting happens. These facies migrate shoreward also! Likewise, as water level drops, or regresses, the facies migrate seaward. Named after German geologist Johannes Walther as "Walther's Principle", these ideas describe how changes in sea level lead to changes in stratigraphic "stacking" of strata, or sequences.

Please watch this video below for an excellent explanation.



As seen in the video above, Walther's Principle applies to both siliciclastic basins and carbonate basins. In siliciclastic basins, our shoreward to deep marine progression might begin with conglomerate and then continue offshore with sandstone, siltstone, and end with mudstone. If sea level rises, deep marine facies will be pushed shoreward and deposited on top of shallower water units in a fining-upward sequence, typically of a transgressive sequence. Likewise, if sea level regresses, shoreward facies will migrate toward the sea, being deposited on top of deeper water facies, creating a stratigraphic succession of coarsening-upward layers.

In carbonate systems, the shoreward to deep marine sequence runs from grainstone (mostly fossil material) to packstone, wackestone (more mud), and ends with a limey mudstone. The same transgressive and regressive sequencing applies with changes in sea level up and down in these environments.

Packages of stratigraphic sequences can be transgressive or regressive in nature. When a transgressive sequence lies below a regressive sequence, the boundary between them is referred to a as a "Maximum Flooding Surface". Likewise, the top of a regressive sequence is erosional and referred to as a "Sequence Boundary".

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12.5: Pattern Matching and Stratigraphic Correlation

Millions of years of weathering and erosion shape our landscapes. Over that time, the records of deposition in ancient basins are washed away, often leaving only sparse remnants of these strata. Measured stratigraphic sections, separated by great distances due to erosion, provide the record necessary to reconstruct these larger ancient environments. The process of connecting layers across these distances is called correlation. Such correlation typically focuses on thicker strata, but can also be done in some cases with microstratigraphic strata, such as minute beds of fossils, volcanic ash beds, impact deposits, *etc*. Stratigraphic correlation is about making connections. It is a set of scientific activities that seek to match rock patterns across distances.

There are a wide variety of ways that stratigraphic patterns are identified and matched, and quite commonly, the patterns do not match up perfectly. Bedding thickness varies over distances. Beds “pinch out” between outcrops. Sometimes, in some places, a new bed is introduced. Physical or chemical characteristics may be missing. The reasons for these variations are multiple. Simple way to consider these variations is to imagine walking along the coast of eastern North America, and observing the changes you see. While there are many commonalities between the beaches of Florida, Virginia, and New Jersey, changes in the shape of the shoreline, development of large bays, variations in water chemistry, the presence or absence of barrier islands, subtle climate variations, inflowing river channels, the slope of the continental shelf, and a great many other features along a coast will cause shoreline conditions to change. As a result of these natural physical variations, populations of organisms and even entire ecosystems will vary, too. Post-lithification processes can also erase physical and chemical characteristics of the rocks through compaction and diagenesis. All of this natural variation can make correlation quite tricky! Stratigraphers, the geoscientists who study patterns within sedimentary rock layers, have developed a variety of complementary methods that aid in correlation between sites. Ultimately, these tools help establish relative age connections between distant units sometimes separated by millions of years of erosion, tectonic uplift and change, and other geological processes.

Dual GigaPans of Tonoloway Formation along Corridor H. You are seeing two sides of the highway at the same location. Click on the title of the Gigapan and then make it full screen so that you can play with the images. How would you go about correlating these layers from one side to the other? For what distinctive features might you look? List some in the quiz below.



? Exercise 12.5.1

Identify a feature contained within each of these outcrops that you could use to correlate across the road at this location. They are the same rock layers, separated by well over 100 ft of road (Corridor H, West Virginia, Tonoloway Formation).

a. Tracing out a distinctive bed.

b. Road Signs

c. Vegetation

d. Composition of the talus (loose material at the base of the outcrop)

Answer

a. Tracing out a distinctive bed.

Patterns that emerge from measured stratigraphic sections represent not only environmental changes on various scales, but also record the local interactions with the biosphere, atmosphere, hydrosphere, and even with the exosphere, the space environment. If stratigraphy is a record of events in a basin over time, then it will be recorded in complex ways due to these interactions between systems. There will be a record of climate change and large-scale weather events, biosphere events, evolutionary change, changes in the Earth's magnetic field, and astronomical events that leave an impact (pun intended!). Such changes will often be global and regional in nature and occur across different environments, requiring an analysis focused on events in time across a region. Over the years, stratigraphers have developed various subfields to help provide focused study in these areas. These subfields of stratigraphy include chronostratigraphy, lithostratigraphy, biostratigraphy, and magnetostratigraphy. Together, these fields provide tools that allow the piecing together of the geological past.

Lithostratigraphy

The name, lithostratigraphy, comes from the Greek words *lithos*, meaning rock, and *strata*, meaning layers. So lithostratigraphy is the study of rock layers. In lithostratigraphy, we are very interested in the sediment grains that make up these layers. We want to know their mineral composition. We want to examine their shape, whether they are rounded or not. We also want to know something about their sorting, whether they contain a diverse mix of grain sizes or just one grain size. These pieces of evidence describe, in part, the "facies" of the rock,which allows scientists to determine how the rock was formed and where it might have originated. In a stratigraphic section, stacked strata will often come vary from bottom to top, indicating a detailed record of environmental change. This record provides a history of the sediment deposition in that basin, changes in sea level, and can also provide important insights into climate fluctuations, all at a variety of time scales.

In lithostratigraphy, rock strata are placed in a hierarchy based upon several factors. Key among these is their lateral regional extent. The basic lithostratigraphic unit is called the **formation**. In order to be given this designation, a layer of rock must be mappable across an area. It must bear the very similar characteristics when seen in locations across that area,and have similar, distinct boundaries, or contacts, with layers above and below. Formations are the units used by field geologists when creating geologic maps. Formations also have a "type section" identified that represents the outcrop for which the standard description of the unit was created. They also have contacts with formations above and below, which are also mappable. Within formations, samller subunits called **members** can be identified. A member is a lithologically distinct bed within the larger formation. In stratigraphic columns below, the Kope Formation is divided into several members. And, within these members, beds are often discernable. The type Cincinnatian is an ideal section for the study of individual beds, as they are often laterally extensive and therefore traceable across the region. A single limestone bed, such as those numbered as 25-31 in column C below, mark distinct moments in time across the region. If enough stratigraphic sections exist from across a region, further patterns will emerge. Roadcuts, cliff faces, and other locations are ideal study areas. Individual rock layers may or may not be present in all of these. They may vary in thickness and content. But this information can help reconstruct the history of not just a single location, as represented by one section, but an entire region.

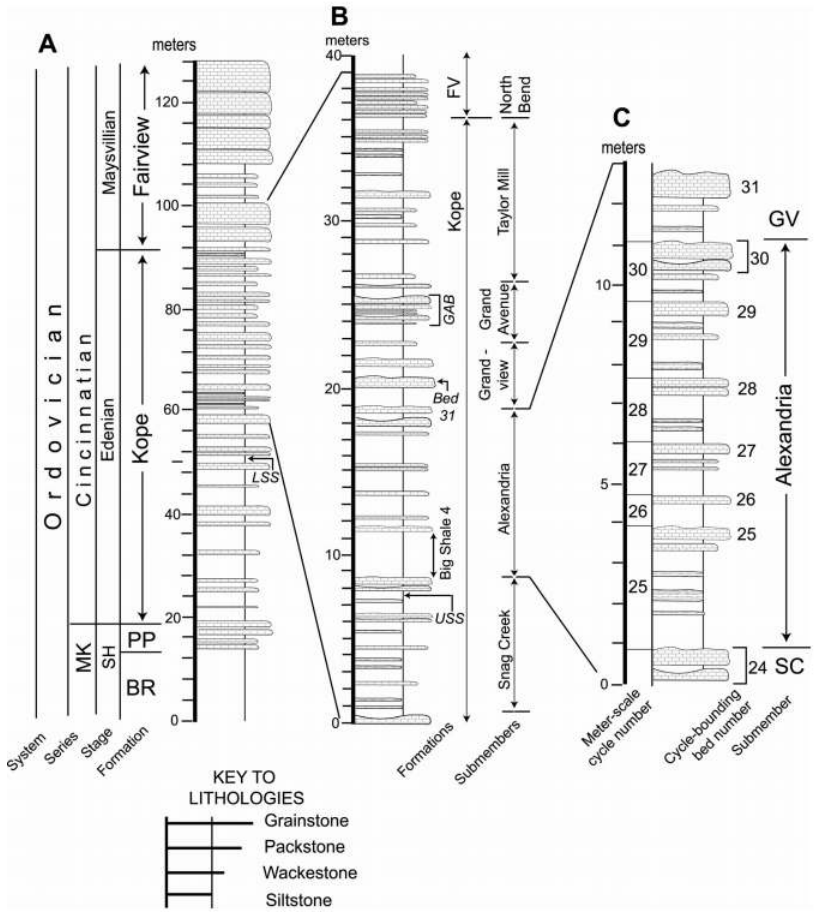


Figure 12.5.1: Upper Ordovician stratigraphy of the Cincinnati, Ohio region. A) Generalized stratigraphy of uppermost Mohawkian and lower Cincinnatian Series. BR Bromley Shale; LSS lower strophic Sowerbyella rugosa interval; MK Mohawkian Series; PP Point Pleasant Formation; SH Shermanian Stage. B) Stratigraphy of middle and upper Kope and lower Fairview Formations, showing Kope submembers . C) Detailed stratigraphy of the Alexandria submember and the upper Snag Creek (SC) and lower Grandview (GV) submembers. Numbers immediately to left of column are meter-scale cycle designations; numbers to right are designations of cycle-bounding limestone beds (Adapted and used with permission: Kirchner and Brett, 2008).

Normal Fault in Chalk – Correlation of both sides of the fault. Try to match the two sides up be moving the gigapan images around. Can you get ride of one of the faults? If so, you just correlated layers!



Lithostratigraphic Hierarchy

Flow (if volcanic materials are included) → Bed → Member → Formation (Basic Lithostratigraphic Unit) → Group → Supergroup

Packages of formations can be put together into **groups.** Packages of groups can be assembled into **supergroups**. Such higher designations are not always necessary or ideal, but in situations where the rock formations are closely related, their use makes good sense. The Grand Canyon Supergroup and the Belt Supergroup are two well known American examples of this largest designation.

Fossils as Stratigraphic Tools: Biostratigraphy

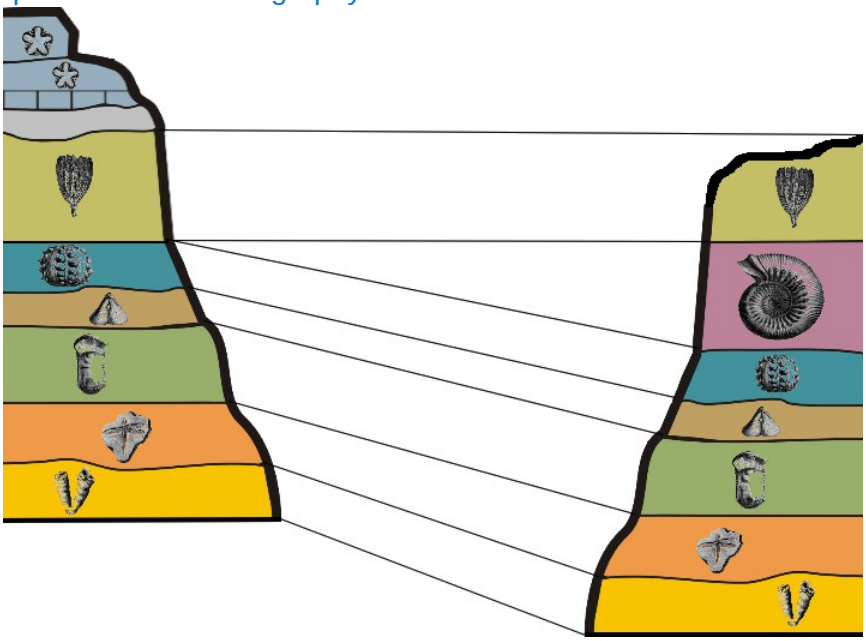


Figure 12.5.2: A simple example of biostratigraphic correlation (CC BY-SA 3.0; By אני יצרתי – דקי, https://commons.wikimedia.org/w/index.php?curid=17935309)

Biostratigraphy is a set of tools used to correlate patterns and beds using fossils. Fossils come in a variety of forms, either as body fossils, such as skeletal parts or impressions, or trace fossils, evidence of movement or activity. Usually, this information is collected in addition to the lithostratigraphic information. Biostratigraphic designations are based upon the existence of a particular diagnostic feature within strata specific to fossil material. Sometimes, such fossil material can be traced over great distances, much like lithostratigraphic layers (Kohrs and Brett, 2008). Biostratigraphic units can vary across a location and so such tracing of fossil beds is not necessarily a basic goal of biostratigraphy. Communities within an ecosystem vary across an environment, depending upon physical and biological conditions. Water depth, temperature, salinity and currents, along with sedimentation rates and nutrient levels can cause the distribution of particular organisms to be quite spotty. However, some fossils are more ubiquitous, for instance due to planktonic life habits, and so are found more commonly over larger areas as they are spread by currents. Because of all of this variation, while biostratigraphic designations can be used to correlate beds, they are also very important tools for understanding the paleoecological history of a location.

Index Fossils

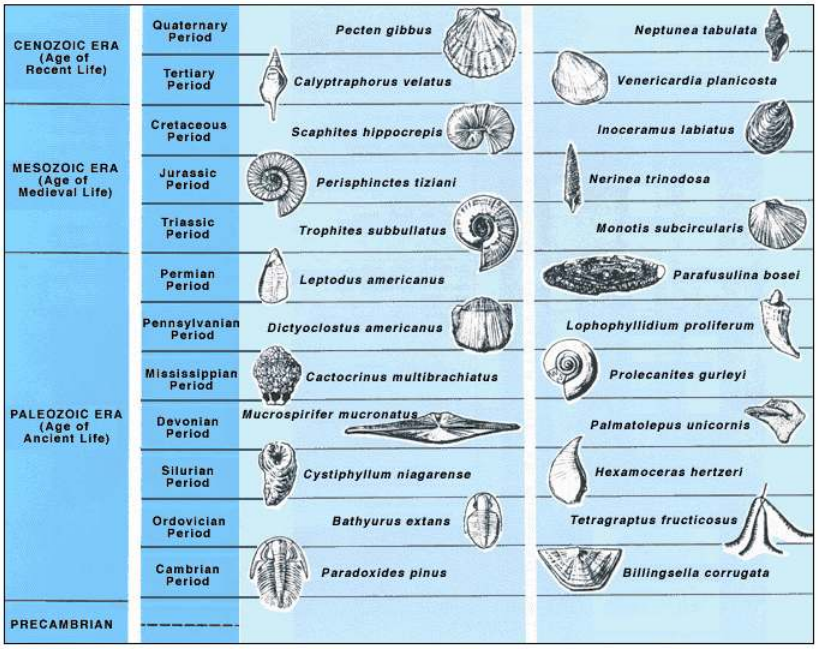


Figure 12.5.3: Some common index fossils (Source: USGS)

Across basins or even within a basin, fossil assemblages can vary a great deal. Consider a modern oceanic shoreline, even one with a barrier island system like the Outer Banks of North Carolina. In such areas, there are species that live across facies, and species whose lives are spent entirely within more restricted zones. Species with a broader distribution, that are plentiful, and that fossilize well can serve very well as markers for important intervals. Ideally, the organism's existence is short-lived. If all of these criteria are met, then that fossil can serve as a marker in strata, an index fossil. Some excellent examples of index fossils can be found here.

Biozones

Biostratigraphy comes with its own unique set of terminology. Unlike chronostratigraphy and lithostratigraphy, there is no hierarchical system of designations. The biozone is the basic element. Biozones are defined by the basic characteristics of their fossil taxa. There are five kinds of biozones. These include range zones, interval zones, assemblage zones, abundance zones, and lineage zones. Due to the lack of hierarchy, and the variety of ways biostratigraphic designations can be made (by single fossils or multiple fossils present), it is possible to have overlapping biozones within a single rock unit.

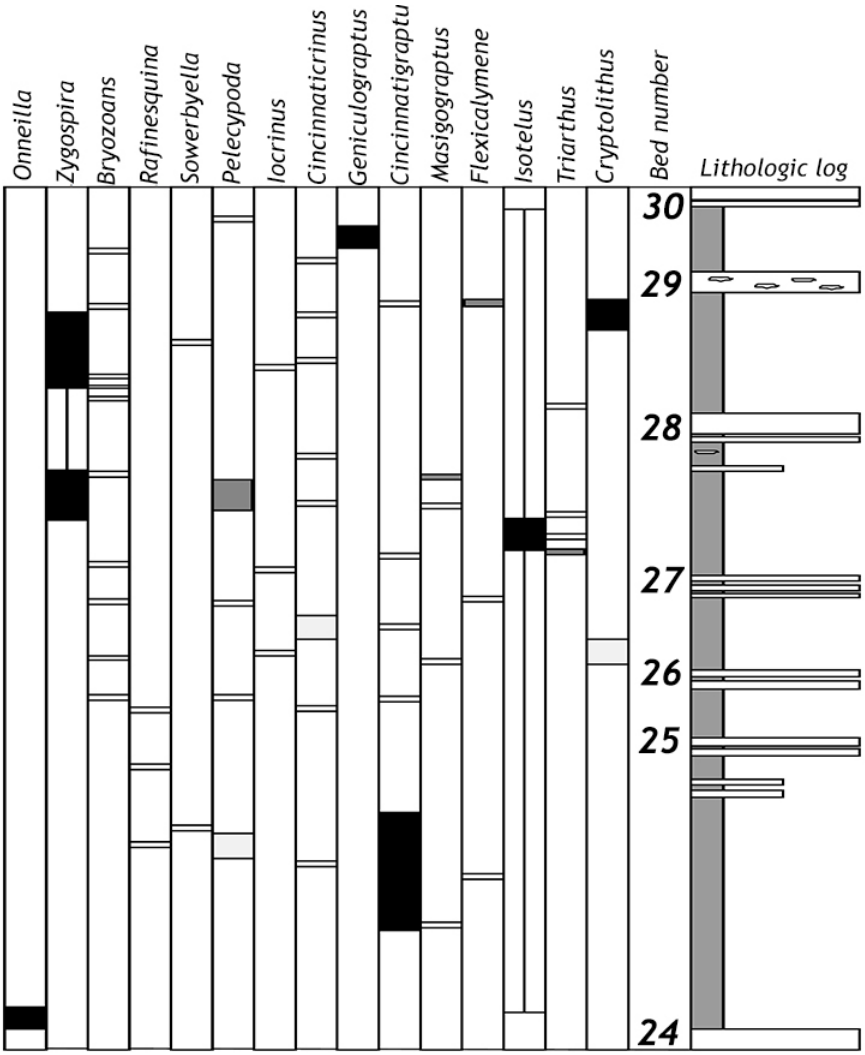


Figure 12.5.4: Fossil occurrences, stratigraphically, within the Alexandria Submember of the Kope Formation. This data was compiled from studies at six sites. Fossil occurrences at 5-6 sites are indicated with black bars. Gray bars identify fossils found at 3-4 sites. Light gray bars indicate fossils found at 1-2 sites. Dark gray to black bars could be considered fossils zones (range zones) useful for biostratigraphic correlation (Kohrs, 2003).

Range zones come in two varieties. The first, a taxon-range zone, is defined as a body of strata with a geographically and temporally defined occurrence of a single fossil taxon. The geographic range can be narrow or very broad, but is defined by the entire known set of occurrences of that taxon across the basin or set of basins where it has been identified. The second, concurrent-range zones, are only different from the taxon-range zone in that they consist of overlapping ranges of at least two taxa. Range zones can be used for stratigraphic correlation across locations, due to their geographic extent.

Interval zones are the fossiliferous strata that exist between two defined stratigraphic horizons. These are sometimes referred to as range zones. For instance, you will not find fossils of modern humans going back more than at most 350,000 years. The interval zone of our species only represents the last 350,000 years. Interval zones are bounded by the occurrence of the taxon in question, so do not necessarily have a correlative power. However, because they are defined by lower and upper horizons, these horizons can be used as stratigraphic correlation markers.

Lineage zones are defined by their evolutionary importance. They are zones that contain a particular evolutionary branch of a taxon. These very specific biohorizons, again defined by their lower and upper limits of occurrence, are very reliable means of relative time correlation in biostratigraphy. It is one thing to be able to correlate across a region the existence of a particular family or genus of a brachiopod, but quite another to be able to define a correlatable time horizon using a specific species.

Assemblage zones are defined by the lower and upper boundaries of strata that are defined by the presence of three or more fossil taxa. These taxa may represent members of a community within the paleoenvironment defined by the strata in which they are

contained. Across a region, not all members of the assemblage need be present to be defined in this way, but most of them should. Individual taxa can also have more extended ranges above and below the assemblage zone.

Abundance zones represent a body of strata where the abundance of a particular taxa is significantly greater than usual. The *Triarthrus eatoni* zones in the figure below from Kohrs (2008) can be defined as abundance zones within the context of a single location, though such zones are potentially also correlative tools. This trilobite can be found here and there within the Kope Formation, but at each one of these localities, could be found in unusual abundance and preservation as a fossil lagerstätte. As correlative markers, they mark a particular useful biohorizon.

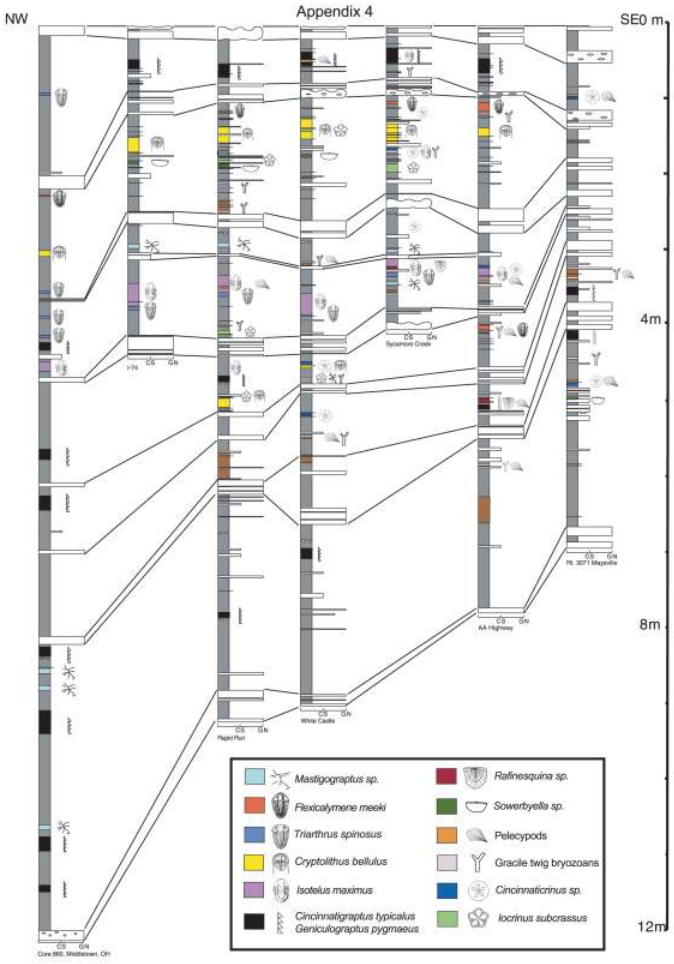


Figure 12.5.5: A lithostratigraphic and biostratigraphic interpretation of mudstone (gray) and limestone (white) layers within the Alexandria Submember of the Kope Formation in the Cincinnati/northern Kentucky region. Each column represents a different location across a transect covering over 50 miles of distance. These columns all represent sediments deposited in the same basin and during the same interval of time. Limestones and fossils can be correlated, or matched by patterns, across the region (Kohrs and Brett, 2008).



Figure 12.5.6: An exquisite speciment of Cryptolithus bellulus from the Alexandria Submember of the Kope Formation, Cincinnati, OH (Kohrs and Brett, 2008).

Some fossils are exceptional in how they are preserved. Exceptionally rich fossil deposits are given the name ***lagerstätten***, a German term that literally means "storage place" because the preservation is so exquisite as to represent what the original organism was like in appearance. Such fossils can sometimes be found in beds that are traceable as units across a region, their exception preservation being caused by the rapid deposition of sediment downslope during a storm event or by some other means. As such, they can be used as important biostratigraphic marker beds (Brett and Baird, 1997).



Figure 12.5.7: Fossil specimens, molts of the trilobite Triarthrus eatoni, from the Alexandria Submember of the Kope Formation, Cincinnati, OH. An exquisite fossil lagerstatten, these particular specimens could be indicating behavior. As these are molts, there are one of two options for how they arrived to be buried here. A large concentration of these molts managed to concentrate here from further upshore. Or, these molts indicate potential mating behavior, as the arthropod shed its exoskeleton (Kohrs and Brett, 2008).

In general, beds that are stratigraphically traceable across a region may contain fossils that appear in one place but not in another. Fossils are suddenly very abundant in bed or formation may represent what is referred to as an **epibole**, or a sudden flourishing of a species or community in the stratigraphic record. These can at times be important event beds for correlation across a region. Likewise, the opposite of this is an **outage**, which represents the disappearance of a fossil or community. This could even be a localized extinction event (*e.g.*, Brett and Baird, 1997). Again, such outages can be important biostratigraphic event be markers.

Principle of Faunal Succession

Because of the way evolution has influenced life on Earth, there is a distinct sequence of appearances of organisms through time. This sequence of life begins four billion years ago, with the appearance of single-celled prokaryotic microbes. Since then, origination and extinction have formed and killed off numerous species, always giving way to new species with novel adaptations based upon shared and derived characteristics. Important patterns develop across the fossil record that allow paleontologists to identify rock units as being of a certain age or environment. It is known, for instance, that humans never lived with dinosaurs. Human fossils have never been found with, among, or in units with dinosaur fossils, and these groups of fossils are separated by at least 61 million years of time, sediment, and events. If a rock contains a dinosaur fossil, we can thus tell that it is at least 65 million years old but not more than 243 Ma (*Nyasasaurus parringtoni*). If it contains a human fossil, our species specifically, it is likely less than 500,000 years old.

Chronostratigraphy

Some sediments are deposited at the same time within a basin, but within different depositional environments. Likewise, at any given point in geologic time, multiple basins around the planet accumulate sediments depending on regional tectonic and climatic conditions. Chronostratigraphy, or stratigraphy based upon time, provides a way to correlate units that would otherwise seem to be unrelated. Chronostratigraphy uses aspects of biostratigraphy and numerical dating to constrain the time of deposition within basins using notable fossil beds and other unique features, like volcanic ash deposits.

Chronostratigraphy places strata in a hierarchy based upon radiometric dates calculated from sampling minerals in a rock, rather than relying solely on physical features in the rock. Common examples of such markers are volcanic ash layers, magmatic intrusions, and zircon crystal inclusions. Globally, the International Commission on Stratigraphy (ICS) has created a time chart based upon an agreed global chronostratigraphy. Each boundary on this time chart is marked by a GSSP, or Global Stratotype Section and Point. A GSSP is a chronostratigraphic “type locality” where that boundary has been marked as the example for entire geological community. Global chronostratigraphic names can be different than regional names for these periods of time. There is a long tradition of applying regional names to rock and time units in the geosciences. Many of the names on the geologic time scale, such as Cambrian and Devonian are derived from locations in Europe. Sometimes, old regional names fall away in favor of others. Among scientists, more regionalized names become a part of the common geological vernacular, where there may be much less familiarity with global designations. For example, the Cincinnatian Series is the local name in the eastern United States for the global Upper Ordovician Series.

Chronostratigraphic designations are roughly equivalent to their lithological counterparts. What follows is a table showing the various lithologic units with their chronostratigraphic equivalents.

Comparison of Chronostratigraphic and Lithostratigraphic Designations

How to Name Our Moment in Geologic Time: Lithostratigraphically and Chronostratigraphically	Lithostratigraphic Designation	Chronostratigraphic Designation
Phanerozoic	Phanerozoic Eon	Phanerozoic Eonothem
Cenozoic	Cenozoic Era	Cenozoic Erathem
Neogene	Neogene Period	Neogene System
Holocene	Holocene Epoch	Holocene Series
Meghalayan	Meghalayan Age	Meghalayan Stage

In lithostratigraphy, we refer to stratigraphic units as being deposited within a particular Eon, Era, Period, Epoch, or Age. Associated names, like "Pennsylvanian", can be very regional. In Chronostratigraphy, these basic units have similar names, such as Eonothem, Erathem, etc. However, the names given that accompany these designations are agreed upon global time horizons.

Chronostratigraphic designations and boundaries are marked with radiometric dates. For the GSSPs that do exist, the dates are accompanied by a feature in the rock, such as a fossil zone, magnetic horizon, or other distinctive feature found widely at that time. This combination of factors can be hard to find. Because of this, not all boundaries on the ICS timescale have agreed upon GSSPs. Sometimes, the term “Golden Spike” is used to refer to GSSPs.

“Golden Spikes”

On May 10th, 1869, politician Leland Stanford drove a 17.7 karat copper-alloyed “golden spike” into a wooden railroad tie at Promontory Summit, Utah. At this location, the Union Pacific and Central Pacific railroad lines connected their respective ends of

the transcontinental railroad, ending many years of labor and connecting a nation just beginning the long recovery from a civil war. The term “golden spikes” is now used to mark moments of geological importance where significant changes, or shifts, occurred.

Geologically, a “golden spike” is used to mark a boundary on the geologic time scale (note the “golden spikes” on the ICS timetable linked above). “Golden Spike” is a colloquial term used for the more technical Global Boundary Stratotype Section and Point, or GSSPs, that are used to define most boundaries on the geologic time scale. There are rules for marking these, and they are as follows:

- A GSSP has to define the lower boundary of a geologic stage (body of rock deposited during an Age).
  The lower boundary has to be defined using a primary marker (usually first appearance datum of a fossil species).
- There should also be secondary markers (other fossils, chemical isotope signatures, geomagnetic reversal).
- The horizon in which the marker appears should have minerals that can be radiometrically dated.
- The marker has to have regional and global correlation in outcrops of the same age
- The marker should be independent of facies.
- The outcrop has to have an adequate thickness
- Sedimentation has to be continuous without any changes in facies
- The outcrop should be unaffected by tectonic and sedimentary movements, and metamorphism
- The outcrop has to be accessible to research and free to access. This includes that the outcrop has to be located where it can be visited quickly (International airport and good roads), has to be kept in good condition (Ideally a national reserve), in accessible terrain, extensive enough to allow repeated sampling and open to researchers of all nationalities.

GSSP markers provide geoscientists with a common language globally. They also provide common markers that are used to mark boundaries on the geologic time scale. In science, it is important to have a common language that everyone can use for efficient and clear communication. In the story of the rock record, it is important to identify important highlights. While the determination and description of a GSSP is not a flashy act, these markers are used throughout the discipline in a variety of important ways. “End-Permian” is used to discuss a very large mass extinction. “Frasnian-Fammenian”, two stages at the end of the Devonian, is used to discuss that extinction.

Magnetostratigraphy

Like other stratigraphic tools, magnetostratigraphy relies on special properties in the rocks to correlate them across distances. In the case of magnetostratigraphy, the record examined is the recorded inclination and declination of the Earth’s magnetic field at the time of deposition, also known as normal and reverse polarity. This includes correlating times of magnetic reversal, or times when the Earth’s magnetic poles have reversed polarity, with “Geographic South” becoming “Magnetic North”, and vice versa. Operating irrespective of the lithology or fossil content, magnetostratigraphy relies on minerals in the sediment at the time of deposition or in volcanic ash deposited during eruptions, allowing a different set of correlations to be applied.

Magnetostratigraphic “zones”, or magnetozones, are portions of the stratigraphic section that share polarity from the same time period and, within these zones, “magnetic chrons” can sometimes be identified that allow numerical dating of specific moments within such zones. Such zones and chrons, when traceable over distances, contribute unique information about a time period in that location and region, but also complement the tools offered by chronostratigraphy, lithostratigraphy, and biostratigraphy in characterizing the rock record or by filling in knowledge gaps left by them.

Many important GSSPs are marked by magnetozone boundaries. All of the important ratified GSSP boundary markers for the Pleistocene Series of the Quaternary System (1.8ma and 2.58ma) and the Pliocene Series of the Neogene System (3.6 Ma and 5.33 Ma) are marked by magnetic polarity changes in the stratigraphic record. These boundaries mark global changes in the Earth’s magnetic field polarity as seen in iron-bearing minerals. For example, the GSSP for the Gelasian Stage of the Pleistocene marks the end of the Pliocene and the start of the Pleistocene System. This magnetic boundary is described at an outcrop in Calabria, Italy, located just off of the main road near an old farm. In this bucolic setting and contained within the limey muds of the unit, lies a magnetic anomaly that marks an important reversal of the Earth’s magnetic field. You could sample this rock and never see it, as such a boundary is not visible to the eye, but only through a magnetometer. Still, this reversal should be recorded globally and represents an important boundary between two stratigraphic units.

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12.6: Measuring Stratigraphic Sections

In order to use stratigraphic tools provided by lithostratigraphy, biostratigraphy, etc., sections of strata have to be measured. This critical work is done primarily in the field. The stratigrapher plans a field outing, beginning with researching relevant background information collected from geologic maps of an area academic and state geologic survey publications. Direct exploration of brand new stratigraphic sections created by human or natural processes, such as cuts in a hillside for new roads or new stream cuts caused by flooding, are also excellent means to examine strata. Once the preliminary library or fieldwork is done, detailed fieldwork can begin. This includes measuring sections, photography, sample collection, and sketching. Sometimes, sections are measured at a meter-scale, sometimes at a centimeter-scale or finer, depending on the research questions being considered. Data is recorded at intervals, relative to a known datum and a detailed stratigraphy will emerge through the drawing of stratigraphic columns. However the measurements are taken, using consistent techniques from one site to the next is critical toward providing a solid stratigraphic model that is useful for drawing conclusions about the data and for generating predictions to be tested.

Measuring sections on multiple scales can be very important. At the time of deposition, sedimentary units are much thicker than they are once millions of years of compaction and lithification have taken place. Sedimentation rates can vary over time and compaction is one secondary effect that begins to remove some of the original information from a particular page in the stratigraphic record. While the original thickness of the unit can be a matter of some speculation, it is usually possible to produce a good estimate. This can give us an idea of the rate of erosion, climate, and other factors working at the time of deposition. Because compaction also leads to burial and, with burial, heating, we can obtain information about these secondary processes from fossils and other sedimentological evidence.

The analysis of fine layering within a unit of rock or entire stratigraphic column can be described as microstratigraphy. At these scales of analysis, from sub-millimeter to perhaps up to a meter of scale, it is possible to describe environmental events and processes occurring within rather small time scales. The Tonoloway Formation of West Virginia preserves packages of alternating gypsum crusts resulting from periodic evaporation events. While we cannot be absolutely certain about the timescales involved with each layer of limestone and gypsum crust, we can surmise that it is possible that such deposition represents time intervals of days to years, or perhaps seasonal changes over the course of a single year. Other microstratigraphic evidence could help constrain this further.

Microstratigraphic record of Silurian tidal flat carbonates, including salt casts, in the Tonoloway Formation, West Virginia. What can we learn from such fine layering? What can we read from the packages of sediment here that are no more than 2-3mm thick after compaction and lithification?



Studying stratigraphic layering is most often done at a larger scale, by describing strata on the order of decimeters on up to perhaps thousands of meters. These larger patterns of deposition can represent a few million to tens of millions of years. Providing a broader, larger, view of events occurring within a basin over larger spans of time allows for a deeper understanding of the behavior of the basin as a whole. These behaviors would include being able to study the tectonic processes occurring or major changes of sea level at a variety of scales. As space in a basin fluctuates and, eventually, a basin infills, this large-scale record provides the story of how deposition took place. Just as important, this story provides direct evidence for developing models of tectonic change for mountain belts, changes in paleogeography, or changes in ancient climates.

A larger stratigraphic record of the Devonian Foreknobs Formation in West Virginia, along Corridor H. This roadcut, created by the human activity of building a road, documents just one section of this formation and its alternating sequences of sandstone, siltstone, and shale.



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12.7: Stratigraphy is not just an Earth Thing...

Even on Mars, robots take pictures and measure crater stratigraphy! As human and robotic exploration push further out into our planet's solar system, we are increasingly finding application for stratigraphic principles outside of Earth. On Mars, one of the places in our solar system where might likely be found one day, the search for water, past or present, has always been important. In recent years, evidence of water has been found in phyllosilicate (clay mineral) layers in crater walls in places like Gale Crater. While the mere presence of water does not in itself indicate that life was ever present, it does help round out the requirements. From a stratigraphic point of view, the presence of water means that sediment was washed from highlands into a basin via streams.

"Shaler" outcrop at Gale Crater on Mars, taken on Sol 0309, Mars Science Laboratory (Curiosity). This outcrop contains shale deposits in a very typical layered strata, possibly representing an ancient lake bed. These can be measured on different scales. Near the top of the outcrop, a darker colored rock appears, perhaps basalt from a lava flow.



There is little doubt that the principles of stratigraphy tha we have found so useful and informative on Earth will cease to continue to reveal new insights into our own planet's past. But, as we dive more deeply into the geologic history of places like Mars, what will stratigraphy's insights have to say that help tell the unique story of that planet and various locations on it?

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12.8: Further reading

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CHAPTER OVERVIEW

13: Using Sedimentary Structures to Interpret Ancient Environments

Using primary sedimentary structures to interpret ancient environments

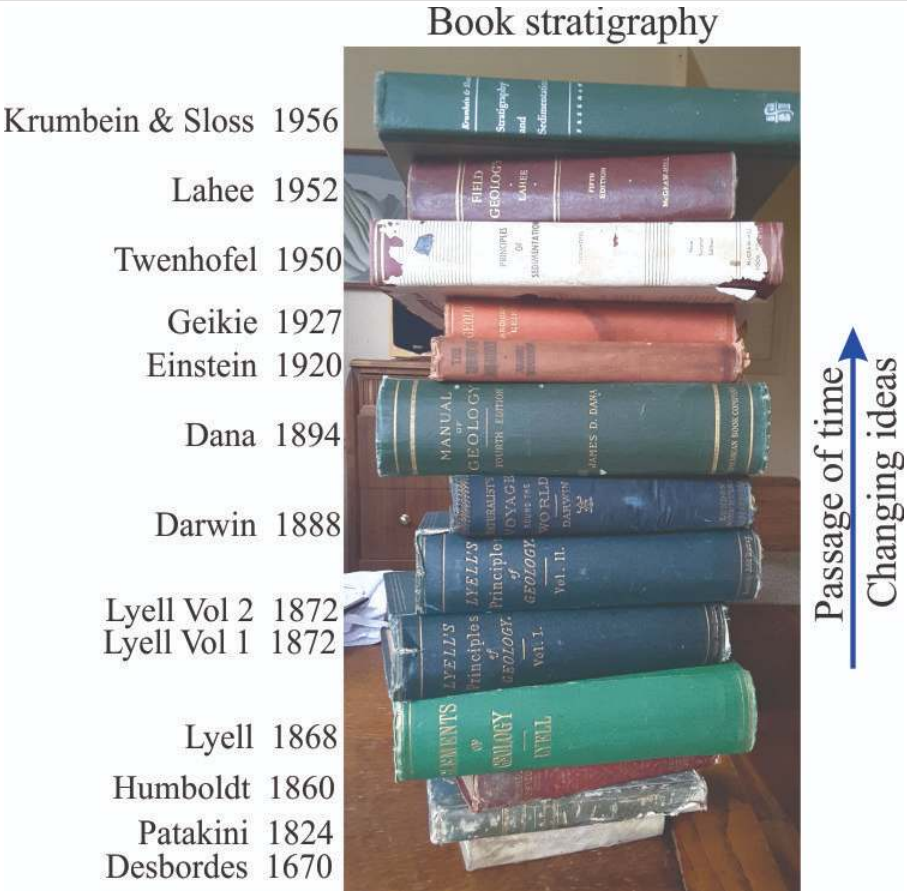
13.1: Preamble

13.2: Tools at a geologist's disposal

13.3: Some common sedimentary structures

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13.1: Preamble



Rocks to a geologist are like books and ancient manuscripts are to an historian. Rocks contain a record of past events and places. The rock record is incomplete. So part of the puzzle is to also try and figure out what is missing. Is it time? Is it other rocks that have since disappeared or perhaps been moved to another location? These are the scientific puzzles that motivate geologists.

The formation of sediment and sedimentary rock involves many physical, chemical and biological processes, sometimes operating separately but more commonly in concert. The journey from loose sediment to hammer-ringing rock is one of the marvels of the geological world. Deciphering this journey requires us to delve into the rock record.

Imagine that a local geologist tells you that the rocks in your backyard were originally deposited as sand and mud in shallow seas, where beaches and broad coastal tidal flats passed seawards to deeper waters, and landwards to marshes and scrubby coastal plains across which rivers and streams coursed. How did our geologist figure this out? What is it that geologists see in the rocks that helps them paint this picture of a world that existed so many millions of years ago?

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13.2: Tools at a geologist's disposal

Sedimentary rocks, whether terrigenous, carbonate, chemical, or volcaniclastic, contain a wealth of information that relate how they formed at or close to Earth's surface, and the various physical and chemical processes that affected them as they were buried. This article deals with the first part of the journey:

1. How sediment moves across a substrate,

2. The language of bedforms,

3. The hydrodynamic significance of bedforms, and

4. An atlas of common sedimentary structures

TRANSPORT OF SEDIMENT; BEDLOAD AND SUSPENSION LOAD



Figure 13.2.1: Surface sea waves meet standing waves on a New Zealand west coast river. (Brian Ricketts photo.)

Movement of sediment creates beds, structures within beds (e.g., laminations, crossbedding), and entire depositional systems like deltas and submarine fans. The processes by which sediment moves determine what the deposit will look like: a train of ripples, turbidites, a layer of mud, or Martian sand dunes.

In this section we will examine four of these processes:

- The case where a fluid, (water, air) interacts with a bed of loose sand – *i.e.*, **bedload transport**;
- Where sediment accumulates from a **suspension** of water and fine particles, and;
- Movement of sediment by surface waves;
- Bedforms in relation to fluid flow – **the flow regime**.

Bedload

As the name suggests, this element of sediment movement consists of loose, granular particles at the sediment-water interface (such as a stream bed or tidal flat): in other words, at the top of a bed. Air or water that moves across the bed will begin to move grains if the flow velocity is great enough to overcome the force of gravity and any resistance at grain contacts. This is the **threshold velocity**.

The bedload contains two main components:

- the **traction load**, or traction "carpet," and
- the **saltation load**.

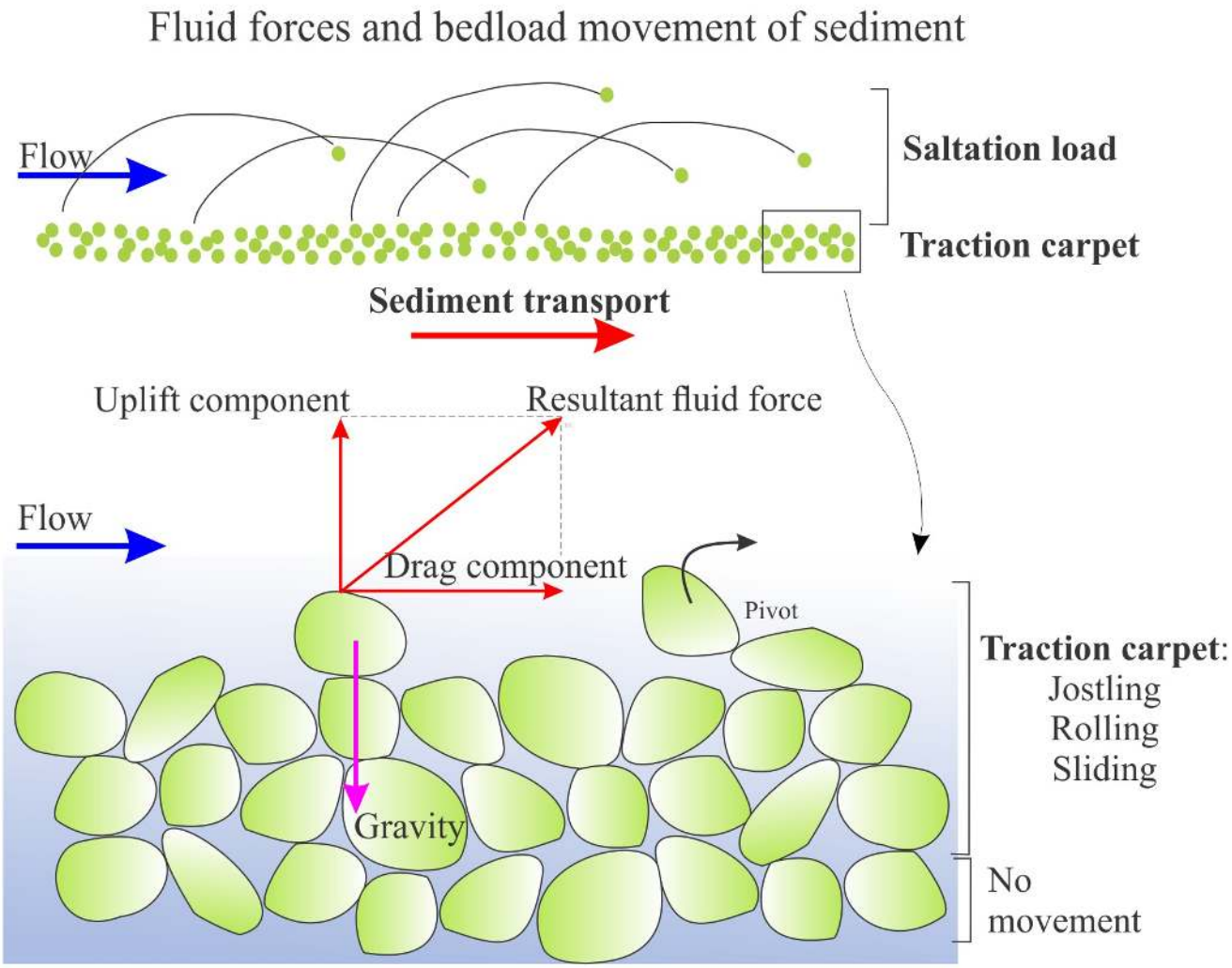


Figure 13.2.2: Diagram showing the components of bedload transport of loose sediment. (Brian Ricketts diagram.)

The various components of force involved in initiation of grain movement are shown here. Fluid flowing over a sediment bed produces shear stresses that can be resolved into a component of drag (parallel to the bed) and a lift component (perpendicular to the bed). At the threshold velocity when the resultant fluid force on grains becomes greater than gravity, grains begin to roll, slide and jostle along the bed like a moving carpet – the **traction carpet**.

The short video is of wind ripples where sand moves in a traction carpet (viewed best in full screen).

https://opengeology.org/historicalge...y/pss/#Bedload

As flow velocities increase, so too will the lift component of fluid forces and grains may temporarily leave the traction carpet, bouncing along with the flow; this process is called saltation. The saltation load is included in bedload because grain excursions

into the fluid do not last long. Saltating grains tend to travel much farther in air than water because the contrast in density between the grain and the fluid is much greater. Stand on any sandy beach on a blustery day and you will witness first-hand the effects of grain saltation. Saltation produces many collisions, and not just against your bare legs. Saltation collisions also result in grain **abrasion**.

The short video shows a saltation load across a windy, **black sand beach** (sand consists of magnetite, ilmenite and pyroxenes). Saltation also tends to winnow the lighter pyroxenes, separating them from the denser iron oxides.

https://opengeology.org/historicalge...y/pss/#Bedload

The ease with which grains move also depends on their density and shape. Quartz grains will move more easily than heavier minerals of the same size. Because they are shaped like little hang-gliders, platy grains like micas, even though they are denser than quartz, will tend to respond hydraulically as lighter particles.

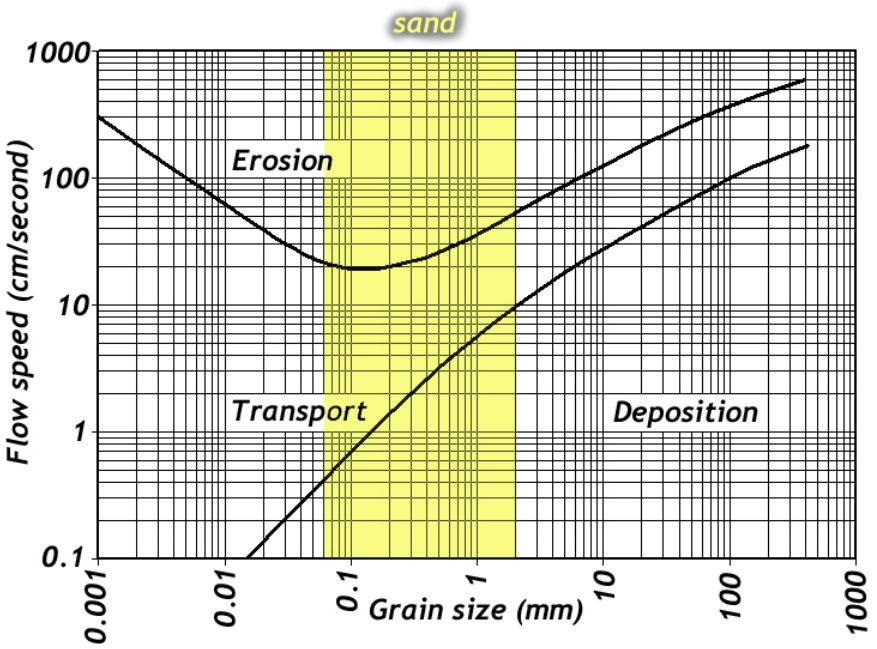


Figure 13.2.3: Hjulström's graph showing the relationship between grain size and flow velocity. Both scales are logarithmic. The graph defines three regions of response: erosion, transport, and deposition of loose sediment. The size of sand sized grains is highlighted in yellow.

The Hjulström curve plots two domains: the erosion curve shows the critical flow velocity required to move a sedimentary particle across a bed, and the deposition curve the flow velocity at which the particle will be deposited (cease to move). Both scales are logarithmic. For reference, the common size range for sand is 0.0625 mm to 2 mm, highlighted in yellow.

In considering these processes, we also neglect grain cohesion. For most sand and silt-sized particles, cohesive forces will be negligible. This is not the case in fine-grained sediment that contains clay. In this situation, clay particles tend to adhere to each other because of electrical charges on the surface of their crystal structures. Once clays have been deposited, it is difficult to remobilize them as individual particles because of electrically induced cohesion. When erosion does occur, it tends to produce lumps of clay, also referred to as rip-up clasts.

The suspension load

Most natural flows in rivers, shallow marine settings and air are turbulent. Even at low-flow velocities, the speed and trajectories of flow can vary considerably. Even seemingly tranquil streams show turbulent eddies and boils on their surface. Very fine particulate sediment (particularly particles of clay) can be kept in suspension for long periods by turbulence; the stresses generated by turbulent flow balance or overcome the gravitational force acting on the particles.

If turbulence decreases significantly, for example when a river empties into a lake, then most particles will gradually settle to the sediment bed. The rate at which a particle settles out of suspension is called the settling velocity, where the force of gravity

(downwards) exceeds the combined effects of upward-directed buoyancy forces acting on a grain and the drag on a particle caused by fluid (viscous) resistance. Thus, the rate of settling depends on the size, shape and density of particles, and the viscosity of the surrounding fluid. In general, settling through air is much more rapid than through water.

Both bedload and suspension load are important processes in the generation of sedimentary structures. In particular, bedload transport of loose sand is the critical process for growth of bedforms and their internal cross-stratification (crossbedding). The description of bedforms (crossbeds) and the flow conditions (flow regime) under which they form are described in following section.

Surface waves

Sea and lake surface waves are generated by wind. The wind provides the energy which is transferred to surface waters. As a general rule, the stronger the wind, the greater the wave amplitude, wavelength, and speed. Surface waves are basically pulses of energy. As such, a water mass does not move in concert with the wave. Instead, water particles beneath waves have a circular or elliptical motion (referred to as an orbital). The largest orbitals occur immediately below the wave crest, decreasing in size to a depth that equates to about half the wavelength. This depth where wave action ceases is called the wave base (or wavebase). If the wave base is deeper than the sea floor, then the waves will interact with the sediment there. Conversely, in deep water, the seafloor sediment is beyond the reach of the wave base, and so waves do not interact with the sea floor.

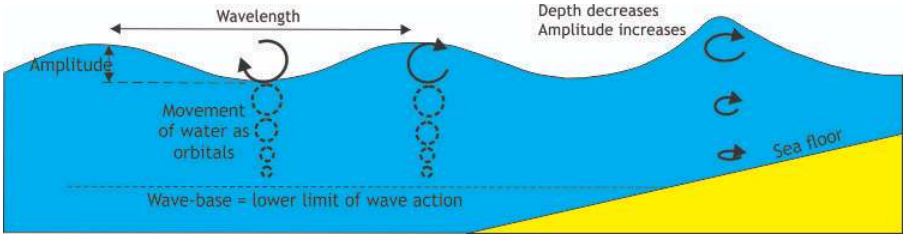


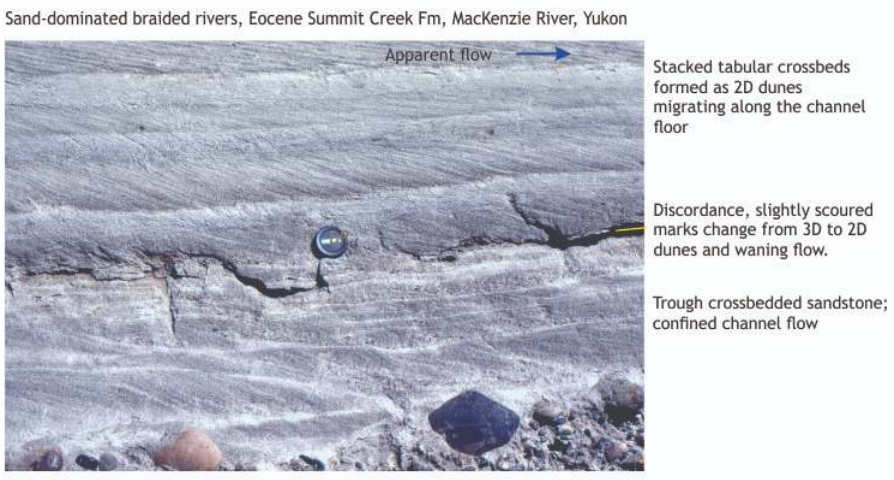
Figure 13.2.4: Diagram illustrating the orbital movement of water beneath surface waves. The water depth where wave orbitals interact with the sea floor is called wavebase – in this region waves can move sediment to form ripples and larger dunes.

In normal weather, the wave base is at a given depth, though that actual depth from place to place is variable, depending on wave amplitude, prevailing wind direction and speed, and the shape of the coastline. But during storms, there is more energy transferred from the wind to the sea surface, and so the waves are larger, and wave base is pushed deeper, where it may be able to interact with sediment that was previously “out of reach.”

Landward of wave base, the orbitals move sediment and generate bedforms, most commonly symmetric ripples where the lee and stoss faces have similar inclinations (because of the back-and-forth action of wave orbitals).

As waves approach a beach some of the energy is transferred to sediment on the sea floor, wave speed decreases (because of friction and loss of momentum),and wave amplitude increases. At a certain amplitude, waves become unstable and break; breaking waves transfer the water as a turbulent, aerated mass that rushes up the beach (swash) and then recedes (backwash), and in the process stirs surface sediment across the breaker zone and beach.

CROSSBEDDING – SOME COMMON TERMINOLOGY



Crossbeds are nearly ubiquitous in sedimentary rocks. They can be found on the deep ocean floor, the driest desert, and pretty well any depositional environment in between. They are most common in sandy deposits. They are less common —but no less important — in gravels (*e.g.*, low sinuosity settings such as braided rivers). Crossbeds form where air and water flow across a bed of loose sediment, so long as the individual sediment grains have low cohesion between the particles: in other words, they are not "sticky." Mud crossbeds are rare because individual clay particles tend to stick to one another (a result of the electrical charges on the surface of these tiny grains).

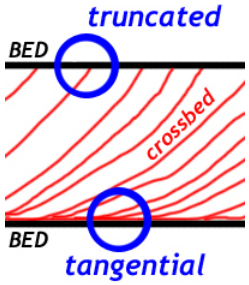


Figure 13.2.5: A cartoon cross-section showing the different crossbed/bed relationships at the upper and lower portion of the bed. (Image by Callan Bentley (2020).)

Crossbeds in the rock record are visible in bed cross-sections, or as exhumed 3D ripples and dunes on exposed bedding planes. The term **crossbed** refers to their internal structure; *i.e.*, laminations that are usually inclined in the down-flow, or down-stream direction. Crossbeds are therefore useful in interpreting paleocurrent flow direction. Because the laminations often show a **tangential** relationship to the main bed at the bottom of the main bed and a **truncated** relationship at the top, they are also useful as geopetal ("way-up") indicators.

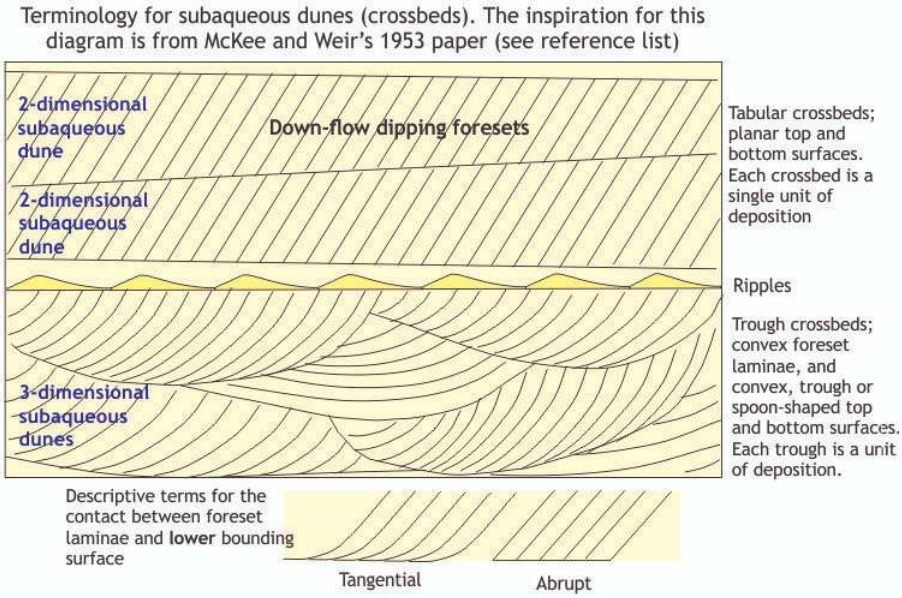


Figure 13.2.6: Terminology for common forms of subaqueous crossbeds (ripples, dunes). (The inspiration for this Brian Ricketts diagram is from McKee and Weir, 1953.)

The laminae are called foresets. In a 2D cross-section view, a single crossbed consists of any number of **foresets** bound above and below by flat or curved boundaries. The geometrical arrangement of foresets, their bounding surfaces and their size or amplitude gives us the information needed to decipher:

- the kind of crossbed,
- the hydraulic conditions under which the crossbed formed, and to some extent
- the paleoenvironment in which they formed; keep in mind that most crossbeds can be found in a range of paleoenvironments but used in conjunction with other criteria such as body and trace fossils, sediment composition and stratigraphic trends (e.g., fining upward) will help pin-point specific depositional settings.

Our interpretations can be advanced further if we are lucky enough to see exhumed structures on bedding, such that we can define:

- the shape of the ripple or dune crest line (is it straight or sinuous?),
- the wavelength between successive ripple or dunes, and
- a relatively unambiguous measure of ripple-dune migration across the bed (*i.e.*, paleocurrents).

Most of our knowledge about ripples and dunes (collectively referred to as **bedforms**) and how they form has been garnered from studies of modern environments. After all, if on your walks across a tidal flat or open-air dune field you see ripples that look identical to those preserved in rocks, it is quite reasonable to infer that the ancient bedforms developed in ways similar to the modern analogues (this is the Uniformitarian Principle at work).

In fact, they have also been seen and recorded forming in real time on Mars.

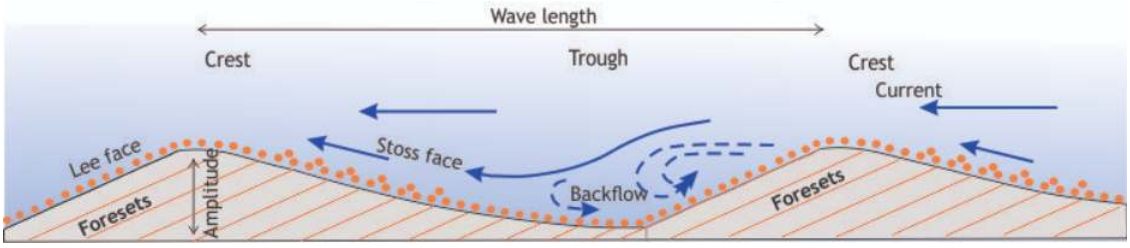


Figure 13.2.7: Ripple formation and terminology. The current moves sand grains up the stoss slope. At the crest of the ripple the grains enter a region of low velocity backflow where gravity takes over and the grains tumble or avalanche down the lee face. In this fashion the ripple moves down-stream with the current.

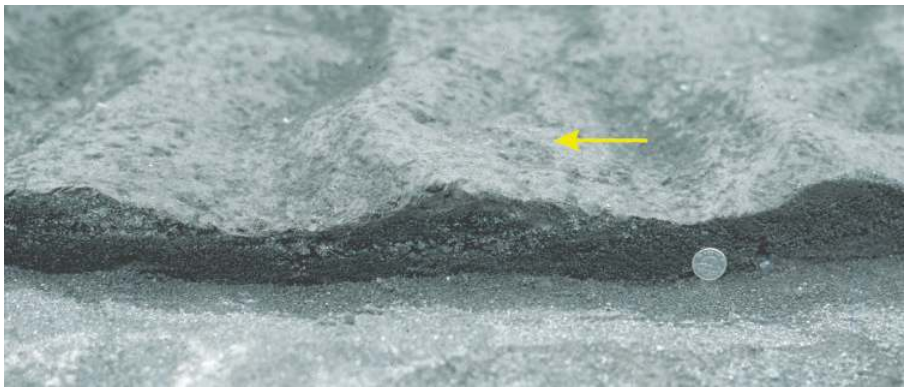


Figure 13.2.8: A slice through modern ripples in soft river sand. Current flow indicated. Compare the example with the diagram above for details of terminology and mechanism of formation. (Brian Ricketts photo)

THE HYDRODYNAMIC SIGNIFICANCE OF BEDFORMS – FLOW REGIME

In this section we introduce some basic hydraulics of sediment movement, bedforms and the concept of **Flow Regime**.

Ripples and dunes form when a fluid (usually water or air on Earth, but the same concepts apply to lava flows, crystal mushes in a magma chamber, or even the Martian atmosphere) flows across a sediment surface. Structures formed by air flow are called subaerial ripples or dunes; those in water have the qualifier subaqueous. These structures are given the general name **bedform**. The construction of bedforms requires certain conditions:

- The sand must be cohesionless (*i.e.*, the grains do not stick together).
- Flow across the sediment surface must overcome the forces of gravity and friction, and
- As noted above and illustrated in the Hjulström curve, there is a critical flow velocity at which grain movement will begin; this also depends on the mass of individual grains, and to some extent their shape. For example, flat, platy minerals like mica are easier to move than a grain of quartz with the same volume but a chunkier shape.

Ripples and dunes form under a relatively limited range of flow conditions. We can illustrate this in a graph of flow velocity against grain size, plotting areas on the graph that correspond to bedform growth. Most of the data for plots like this are derived from experimental flumes where flow conditions can be monitored closely. The plotted distribution shows that bedforms can be categorized according to flow and sediment conditions.

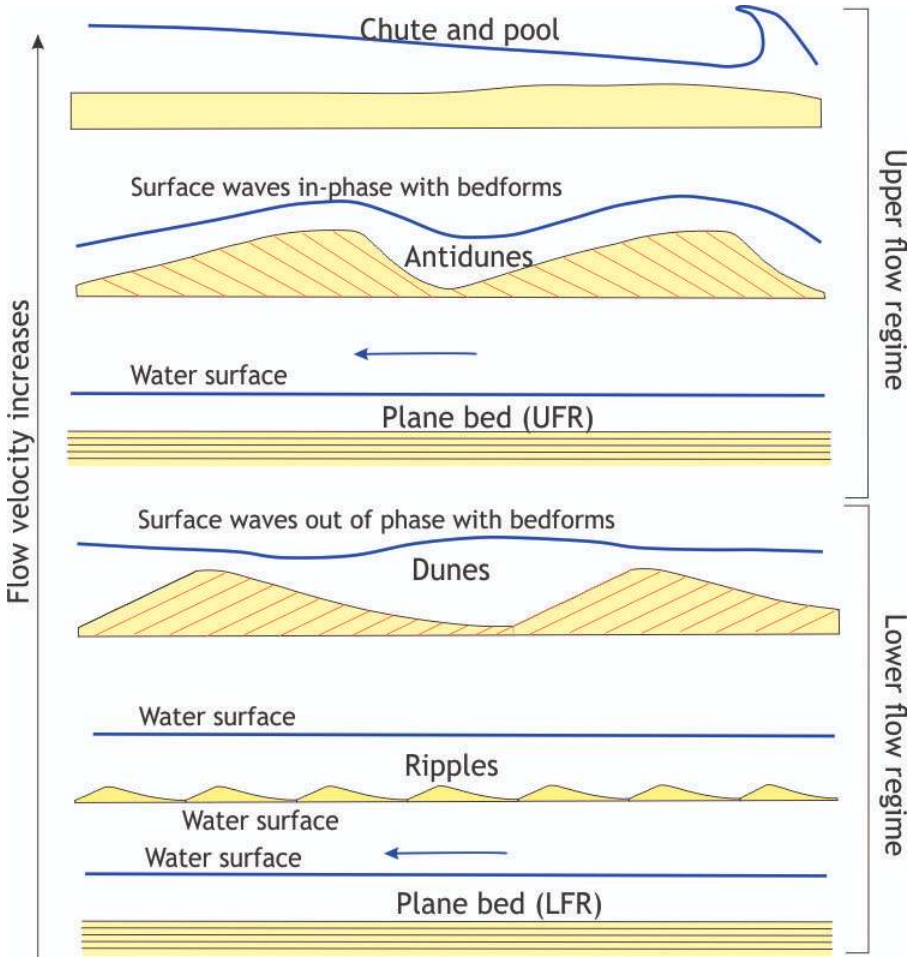


Figure 13.2.9: A schematic of bedform transitions and surface water configurations for Lower and Upper Flow Regime sedimentation – significantly modified from Harms and Fahnstock (1965), Plate 1. Note that the antidune foresets dip **up**stream. (Brian Ricketts diagram.)

This partitioning of bedforms was used to construct the Flow Regime hydraulic model, first published in the now classic 1965 paper by J.C. Harms and R.K. Fahnestock and used widely ever since.

The Flow Regime model considers three fundamentally different states of flow:

- No bed movement where there is too little energy in the system to initiate and maintain sand grain movement,
- A Lower Flow regime in which all common bedforms develop. Here, plane bed (basically parallel, planar laminae with no ripples) represents the lowest velocity, or energy conditions where sediment movement is initiated. It has been observed in flumes and in natural channels that the size of bedforms increases from ripples to large subaqueous dunes as flow velocity increases. Dune type also changes from two dimensional structures (straight crests and planar crossbed bounding surfaces), to three dimensional structures that have sinuous, arcuate and lunate outlines and spoon or scour-shaped bounding surfaces (commonly seen as trough crossbeds).
- An Upper Flow Regime where the power of stream flow washes out ripples and dunes, replacing them with plane bed (this kind of plane bed commonly has **parting lineations**), plus **antidunes**, and erosional chutes and pools.
- As stream flow increases the transition from Lower to Upper flow regime produces one of the more interesting bedforms – antidunes. They are mostly found in shallow channels (*e.g.*, fluvial and tidal channels). You can recognize that this transition has taken place when you see standing surface waves – watch closely and you will see the waves migrate **upstream**. **Antidunes** are the bedforms that develop immediately below standing waves (the two are **in-phase**). If high flow is maintained, the antidunes will also migrate upstream. However, once flow slackens they tend to wash out. This means that the **preservation potential** of antidunes is low.
- **Hydraulic jumps**: The transition from Lower to Upper flow regime passes with a change in bedform, in particular washing out of subaqueous dunes, but there is no sudden break in surface flow – the transition is reasonably smooth. This is not the case for

an Upper to Lower flow regime transition that is marked by an abrupt increase in water level and turbulence: a hydraulic jump. Hydraulic jumps can be thought of as standing waves. They are caused by a reduction in Upper Flow Regime velocity, a change in stream-bed gradient or water depth, or combinations of all three.



Figure 13.2.10: Standing surface waves will have corresponding antidunes on the sediment bed, in this small stream draining a beach at low tide. (Brian Ricketts photo.)

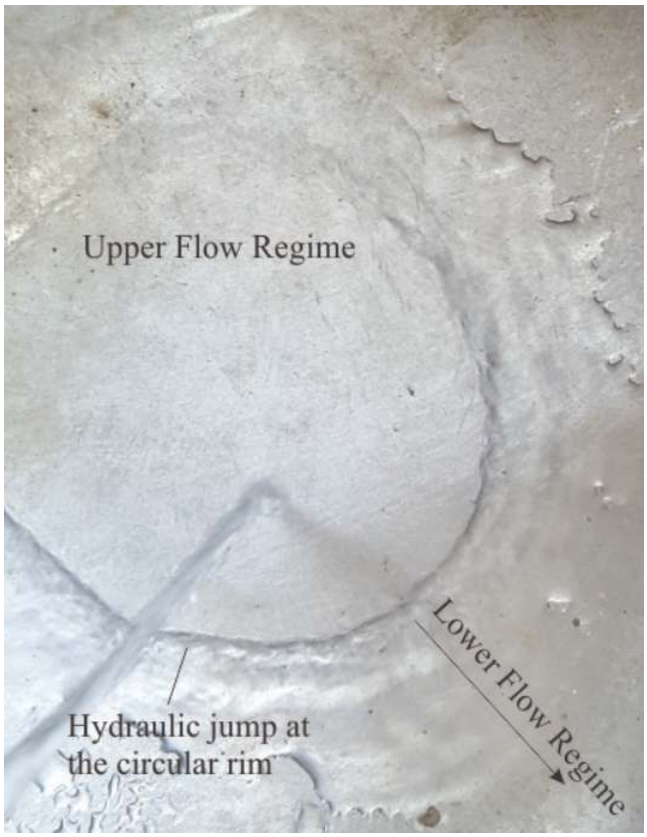


Figure 13.2.11: Formation of a hydraulic jump in the laundry sink. The transition across the circular rim marks the change from Upper to Lower Flow Regime. (Brian Ricketts photo.)

You can generate a hydraulic jump in your kitchen or laundry sink, so long as the sink floor is reasonably flat. Turn on the tap until you see a flow pattern like the one in the photograph at left. Flat laminar flow is generated by the downward force of tap water –

this is the plane bed. The hydraulic jump is located at the circular ridge where the flow changes abruptly. Beyond the jump is lower flow regime flow.

We can use the Flow Regime concept in the field as a qualitative indicator of changes in flow conditions with time (*i.e.*, stratigraphically) and space (laterally). For example, a succession of strata that contains a bed of ripples overlain by larger trough crossbeds indicates that flow velocities, and hence stream power increased abruptly. What kind of paleoenvironment could this have occurred in? This is one of the central questions for any sedimentological analysis.

Some common sedimentary structures, their hydraulic significance, and the environments in which they typically are found, are illustrated below.

A short movie of dune-ripple formation

The link here takes you to a short movie of subaqueous dunes forming in a flume (a **flume** is a narrow tank containing a sand bed, where the velocity of water flow that can be controlled – they are used in experiments to observe the formation of bedforms, and for modelling engineering problems such as the flow of water around bridge foundations). Flow in this experiment is Lower Flow Regime. The sequence begins by showing water flow in front of the migrating lee face; look carefully and you will notice (movie presented by Michael Calzi, SUNY Genseo Dept. of Geological Science):

- sand grains being carried along the stoss slope bed;
- when grains reach the dune crest, they avalanche down the lee face – forming crossbed foresets,
- water flow downstream of the lee face appears to flow backwards – this is the backflow shown schematically in the ripple formation diagram above.

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13.3: Some common sedimentary structures

Identifying sedimentary structures and deciphering how they formed are two of the most important tasks for any sedimentological study. They provide clues, in some situations the only clues, to interpreting ancient environments. The utility of sedimentary structures to unravel the past becomes even more powerful when used in conjunction with other rock properties such as fossil content and geochemistry.

Any examination of sedimentary rocks begins with descriptions of bed geometry, sediment texture, fabric, color and fossil content; sedimentary structures are part of this descriptive vocabulary. Some basic descriptors that are used for outcrop, core or hand sample are listed below.

Table 13.3.1: Descriptive elements

Bedding	Thickness
	Geometry (*e.g.*, parallel bedded, lenticular, lensoidal)
	Weathered attributes – is lithology resistant, recessive, cliff-forming?
Color	Note the color; is it uniform, variable?
Texture/Fabric	Grain size – The range of sizes (sorting)
	Grain size – Maximum clast size (*e.g.*, in conglomerates)
	Clast framework e.g. clast-supported, matrix-supported, variable
	Clast roundness
	Clast angularity
	Clast shape (*e.g.*, spherical, platy)
	Is there preferred clast orientation (*e.g.*, alignment, imbrication)?
	Is there preferred clast distribution (*e.g.*, graded, bimodal)
	Sediment/rock classification: *e.g.*, sandstone/grainstone, mudstone/lutite, conglomerate/breccia/rudite
Composition	Begin with the most basic classification: – is it a carbonate, siliciclastic, volcaniclastic?
	Main clast types: quartz, feldspar, lithics, bioclasts, ooids
	Cements: carbonate, silicic, iron oxides
	Induration: soft, hard; does it ring when you hit it with a hammer, or make a dull thud?
Fossils	Body fossils, trace fossils, casts or molds?
	Are there preferred faunal or sedimentary associations?
	Shells intact, broken
	Have shells been transported or are they in the same position when living?
	Relationship to bedding (*e.g.*, infaunal, epifaunal)
	Degree of preservation

In this section, we present examples of:

- Bedding

- Crossbedding (cross stratification)
- Structures formed at the soles of beds
- Structures formed by desiccation
- Structures associated with sediment texture and fabric
- Structures formed by sediment gravity flows (mainly turbidites and debris flows)
- Structures formed by deformation and mass movement soon after deposition
- Structures formed by organisms crawling, burrowing, or grazing through sediment in search of food, a place to live, or shelter from predators (these are **trace fossils**);
- Structures formed by chemical precipitation (evaporites, iron formations, phosphates);

For each structure, or set of structures, there is:

- An annotated image from outcrop describing the defining attributes;
- Where possible an annotated image of a modern analogue;
- A brief description of each structure and its formation;
- A list of common depositional environments in which it may be found.

BEDDING

Bedding is a sedimentary structure. Sediment moved by currents, or falling from suspension through water or air, will accumulate in a layer, or bed, a process that continues until the supply of sediment is terminated. The plane that defines the top of the bed (bedding plane) represents the termination of that depositional event. Renewed deposition will see the accumulation of a new bed. A bedding plane thus represents a period of time in which no, or very little sediment was deposited – the duration of the hiatus might be minutes, weeks, or centuries.

Common descriptive elements of bedding are listed in Table 13.3.1 above. Keep in mind that descriptive words like **tabular** bedding (where bedding planes are basically parallel), or **lenticular** (where beds are lens-shaped) depend on the scale of your observations. For example, beds that are tabular at the outcrop scale, may pinch out (as at the outer limits of a lens), or thicken in exposures farther afield. Note too, that outcrop views are commonly two-dimensional, so what appears to be a “tabular” bed may be lenticular in the third dimension (*i.e.*, deeper into the rock beyond the outcrop face).

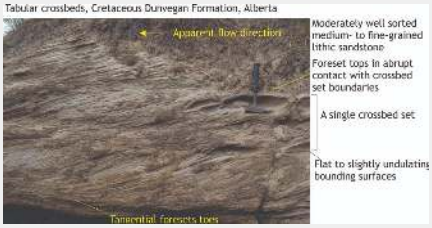
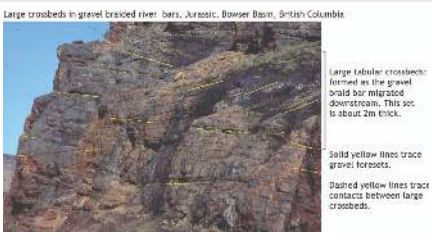
Table 13.3.2: Bedding

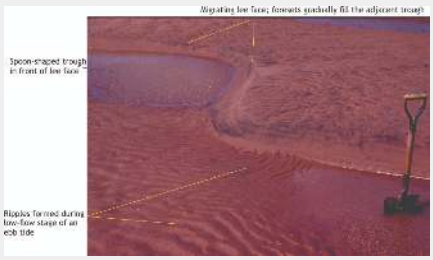
Sedimentary structure	Primary attributes	Common environments
	Tabular bedding can form in many environments. This example is from a turbidite succession. Beds occur in packages that are also tabular. Can form in almost any sediment type, from conglomerate to mudstone; siliciclastic, carbonate, volcaniclastic.	Almost any marine, lacustrine or terrestrial environment.
	Lenticular bedding is common in many environments. This example is from Carboniferous fluvial deposits.It can form in any sediment type, from conglomerate to mudstone; siliciclastic, carbonate, volcaniclastic.	Almost any marine, lacustrine or terrestrial environment.
	Channelized bedding downcutting the underlying lithology. This example is a fluvial channel that has eroded slightly older floodplain mudrocks. Common in coarse-grained lithologies (sandstone, conglomerate). The lower contact is erosional.	Many kinds of fluvial, tidal flat, estuarine, alluvial fan, delta, and submarine fan channels.

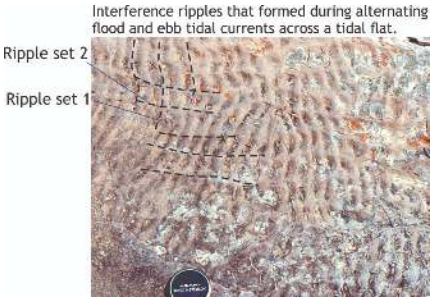
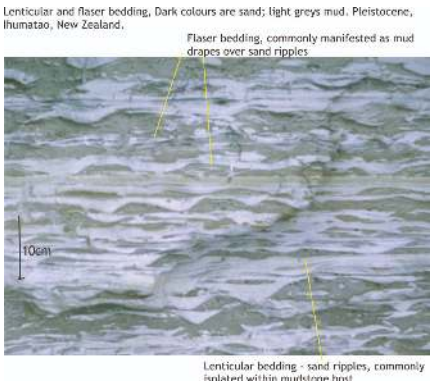
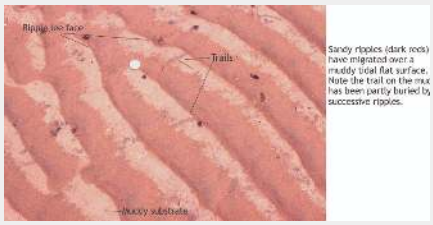
CROSSBEDDING

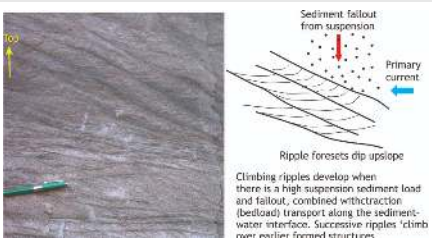
Refer to the text and diagrams above for an outline of fundamental processes of crossbed formation, and crossbed terminology.

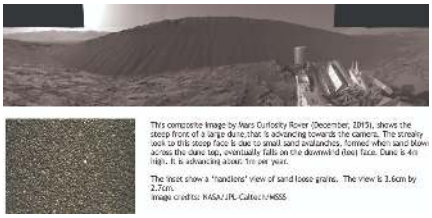
Table 13.3.3: Crossbedding

Sedimentary structure	Primary attributes	Common environments
	Tabular crossbeds in a fluvial channel. Each crossbed set has a planar basal contact with foresets tangential to that base. Paleoflow was to the right. Below the lens cap are several trough crossbeds. Each crossbed represents a migrating dune. See the modern examples below.	Common in many marine and terrestrial settings. Diagnostic of the energy of flow rather than any specific environment.
	Tabular crossbeds in a sandy fluvial channel; foreset contacts indicated. Flow was to the left. See the modern examples below.	Common in many marine and terrestrial settings. Diagnostic of the energy of flow rather than any specific environment.
	Modern dunes having straight crests are the bedform equivalent of tabular crossbeds. These examples grew during tidal flow on a sandy tidal flat. Small ripples have formed on the stoss slopes of each dune.	These examples formed during successive flood tides. The small ripples formed during subsequent ebb tides.
	Tabular crossbedding revealed in a slice through a dune on an exposed river channel bar.	This example formed during the flood stage in a mixed sand + gravel, braided river channel. Note, this kind of flow is quite different to the tidal flat examples shown above, but the crossbeds look similar.
	Tabular crossbeds up to 2m thick, in this gravelly, braided Jurassic river system. The crossbeds formed as part of large gravel bars within a river channel.	Tabular crossbeds are common components of sand-gravel bars in braided rivers. They commonly occur with trough crossbeds.

Sedimentary structure	Primary attributes	Common environments
 (Photograph by Brent Ward; reproduced with permission.)	A modern braided river, the Donjek, in the Yukon Territory of Canada. The channel bars consist of sand and gravel – tabular crossbedding is common here. It's also worth noting the gray-blue color of the water, indicating a through-going transport of suspended sediment.	Tabular crossbeds are common in meandering (high sinuosity) and braided (low sinuosity) river channels.
	In 2D exposures, trough crossbeds have convex-up basal contacts; foresets tend to be tangential. Troughs commonly cross-cut earlier-formed structures, resulting in a complex array of partly preserved bedforms.	Common in channelized (confined) flow; found in tidal, estuarine, fluvial, and alluvial fan channels. This example is from an ancient tidal channel.
	Recent dunes on an exposed tidal channel margin, are the 3D equivalent of trough crossbeds. Crest-lines are sinuous. The spoon-shaped troughs are scoured by tidal or stream flows (here covered by water) are filled by the advancing lee faces. Dune migration is to the left.	This example is from the margin of a tidal channel in Minas Basin, Fundy Bay, Nova Scotia.
	Straight crested ripples nicely exposed on bedding. Lee slopes on each ripple consistently face in the direction of the blue arrow. The regularity of ripples indicates consistent flow across this Paleocene tidal flat.	Ripples are one of the most common bedforms, and can form in environments ranging from sand dunes, rivers and alluvial systems, tidal flow on tidal flats, channels, estuaries and continental shelves, and deep ocean basins on submarine fans.

Sedimentary structure	Primary attributes	Common environments
	Modern straight- to slightly sinuous ripples on a tidal flat. Lee slopes consistently face left in the direction of tidal flow.	
Interference ripples that formed during alternating flood and ebb tidal currents across a tidal flat. 	**Interference ripples** contain two distinct sets of ripples at high angles to each other. One set is formed during flood tide and the other during ebb tide. However, distinguishing between flood and ebb is generally not possible unless there is independent evidence for the orientation of a paleoshoreline.	Common on tidal flats.
Modern interference ripples. 	**Interference ripples**. On modern tidal flats we can observe the two ripple sets forming at each stage of tidal flow.	
Lenticular and flaser bedding, Dark colours are sand; light greys mud. Pleistocene, Ihumatao, New Zealand. 	Lenticular and flaser crossbedding; Lenticular crossbedding refers to sand ripples that overlie or are draped by thin mud layers. Mud flasers drape sand ripples or occur in the troughs between ripples. Lenticular ripples dominate in this example from a Pleistocene tidal flat.	Lenticular and flaser crossbedding are good indicators of tidally influenced environments. Both structures indicate the reversal of tidal flows (ebb and flood) and the changing capacity to deposit sediment.
	A recent example of lenticular ripple crossbedding. Sand ripples (darker reds) overlie a mud layer (lighter colors). The ripples partly overlie invertebrate trails.	Tidal flats, lagoons, estuaries, and shallow subtidal environments.

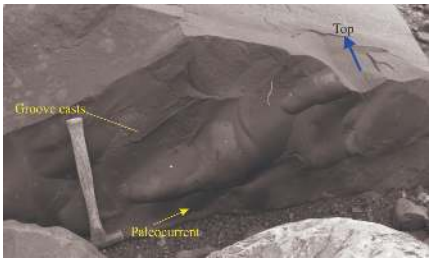
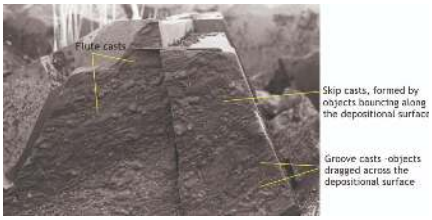
Sedimentary structure	Primary attributes	Common environments
	Climbing ripples (also called ripple drift); , Form when bedload transport acts simultaneously with the settling of fine sand from suspension. Successive ripples appear to climb over their predecessors.	Form when there is a high suspension load of fine sediment combined with sediment carried by flowing currents. Can form at the mouths of channels where flow velocities suddenly decrease, and in decelerating turbulent flows such as turbidites.
	**Hummocky cross-stratification (HCS)** forms at storm wave-base, when sea floor hugging turbulent flows interact with storm wave orbitals. Above storm wave base their preservation potential is low because of reworking by fair-weather wave orbitals.	HCS is common on open marine, sandy shelves and platforms.
 (Callan Bentley photo.)	Another example of hummocky cross-stratification, from Miocene volcaniclastic sandstones from western Costa Rica.	These convex-up scoops filled with half-“smiley faces” make excellent way-up indicators.
	Very large tabular crossbeds in aeolian dunes, this example from the Jurassic Navajo Sandstone, Navajo National Monument. Sets of dune bedding are indicated by black lines. Some sets are more than 10m thick. In most beds, foresets are tangential to the basal contacts.	Crossbeds are common in aeolian dune seas (ergs) and coastal sand dune complexes. There can be significant variability in crossbed geometry and size. Similar crossbedding is also seen on Mars.
	Multiple cross-cutting crossbed sets (mostly tabular) in a dissected Pleistocene, coastal dune complex. Prevailing wind was to the right (onshore). The lens of peat at bottom of image (about 50 cm thick) accumulated in a small interdune pond.	Coastal dune deposits will tend to show greater variation in crossbed thickness and lateral extent, and will be interbedded with interdune pond-lake deposits, and deposits from coastal storm incursions.

Sedimentary structure	Primary attributes	Common environments
 (Callan Bentley photo.)	Large-scale dune crossbedding exposed on a cliff of Jurassic Navajo Sandstone in Zion National Park, Utah. The image cycles between a raw photograph and an annotated view, where each of the ~10 principal beds are highlighted with different colors, and the traces of their internal crossbedding traced out with white.	Note the tangential attitude of the cross-beds at the bottom of each bed, and their truncation by the overlying bed. This relationship can serve as a useful geopetal indicator.
	This image from Curiosity Rover shows an advancing lee face of a large dune on **Mars**. The inset shows a typical dune sand as viewed by Curiosity's hand lens.	Aeolian processes are active and widespread on Mars. We can use our knowledge of Earth's aeolian processes to help us understand these Martian versions.

SOLE STRUCTURES

Sole structures are formed during deposition of sediment. The most common types are structures formed by scouring erosion of the substrate by a flowing current, by objects dragged across the sea or lake floor (e.g., bits of wood, shells, or pebbles), and by objects that bounce along the substrate. The depressions thus formed are filled by new sediment such that they are part of the basal contact of the overlying bed – i.e., they are casts of the depressions and are located on the bottom of the overlying bed. As the sole of a shoe is on its underside, sole structures are also formed on the underside of beds. They are sometimes called sole casts. Sole structures are most easily identified on exposed bedding. Flute, groove, and bounce casts are common in turbidites, but can form on shallow continental shelves and platforms, particularly during the passage of storms. Examples include gutter casts, narrow, elongate scours in the sea floor 15-20 cm deep, that form during coastal storm surges.

Table 13.3.4: Sole structures

Sedimentary structure	Primary attributes	Common environments
	Large flute casts at the base of a turbidite bed. Current flow (indicated) determined from flutes is generally unique. The groove casts parallel the flutes but inferred flow is ambiguous.	Flute casts are most commonly observed at the base of turbidites in submarine fan, submarine canyon, and prodelta deposits.
	Close-spaced flute casts in an Eocene shelf sandstone; flow direction indicated.	Flute casts are less common under normal conditions on a shelf or platform, but can form from offshore flow of storm surges.
	A nice collection of skip, groove and flute casts at the base of a turbidite bed. The flutes indicate flow to the right. There are also some load casts, but these formed after deposition was completed.	The utility of flute casts for determining paleoflow directions is nicely illustrated here.
	Groove casts at the base of a sandstone (shelf environment). Arrow indicates the ambiguity of paleoflow – possible flow directions are 180°apart.	
	Gutter casts (outlined) in Jurassic shelf sandstone; probably the result of a coastal storm surge. The same bed also has some pebbles and fossil debris that were transported offshore by the surge.	These structures known mainly from shelf and platform deposits, formed by bottom-hugging return flows generated by storm surges.

STRUCTURES FORMED BY DESICCATION

Desiccation means “drying out.” Desiccation of subaerially exposed surface sediment or soil drives off ambient moisture, reducing its volume, and resulting in shrinkage. Shrinkage cracks can extend many centimeters below the surface. Propagation of shrinkage cracks, or mud cracks, across a sediment surface commonly produce 5 and 6-sided polygons. If desiccation continues, the polygon margins will begin to curl upward. Mudcracks are common on river floodplains, the inactive parts of alluvial fans, and supratidal

environments that are exposed for long periods. Desiccated sediment is prone to reworking during river flooding, and periodic king tides or storm surges across intertidal and supratidal flats.

Mud cracks preserved in the rock record are excellent indicators of subaerial exposure and may indicate:

- falling sea levels,
- changes in the location of river channels and adjacent floodplains,
- arid climate

Table 13.3.5: Structures formed by desiccation

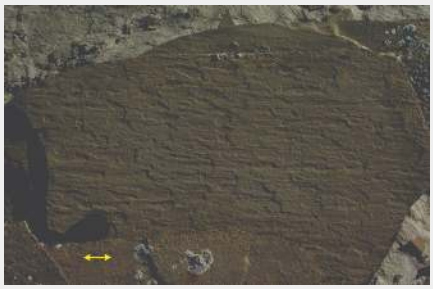
Sedimentary structure	Primary attributes	Common environments
	Desiccation of river overbank muds has produced 5- and 6-sided polygons, and shrinkage cracks 3-4 cm deep. Some polygon margins exhibit slight curling.	Common in muddy sediment and soils subjected to prolonged exposure: e.g., river floodplains, supratidal parts of tidal flats and lagoons, and terrestrial environments having fluctuating water tables.
	Desiccation polygons in a flat area adjacent to hydrothermal mud pots near the Krafla volcanic center in northeastern Iceland.	
	Desiccation polygons 10s of cm thick can form by repeated drying of successive mud layers, where the orientation of shrinkage deformation is guided by pre-existing mud cracks.	
	Intertidal to supratidal layers of carbonate mud were subjected to prolonged desiccation. Polygon margins are curled upward. Some desiccated layers were eroded by storms, producing abundant mud chips.	Desiccated sediment in supratidal regions is prone to erosion by king tides and storm surges.
 (Callan Bentley photo illustration.)	Halite casts can be preserved as 3D cube-shaped forms that protrude from the surface of the bed, as with this slab of Appekunny argillite (Purcell Supergroup, Alberta).	

Sedimentary structure	Primary attributes	Common environments
 (Callan Bentley photo.)	Gypsum is another common evaporite mineral whose form can be preserved in shape, even after dissolution and replacement. This photo shows gypsum casts in limestone, Silurian Tonoloway Formation of eastern West Virginia.	
 (Callan Bentley photo.)	Sometimes accompanying the desiccation process will be a brief spatter of rain. Individual raindrops can excavate very well-defined 'craters' upon impact with semi-cohesive damp mud. Cross-cutting relationships with mud cracks can help establish relative ages. Modern mud, central Virginia.	

STRUCTURES FORMED BY SEDIMENT TEXTURES AND FABRICS

This category of structures is based on textural properties of sediments, such as clast shape and orientation. We often think of sedimentary grains as approximating spheres. However, clasts commonly deviate from this ideal shape – they may be blocky, flattened or plate-like, rod-like, or tapered cones (fossil groups such as gastropods commonly conform to the latter shape). During transport in flowing currents (air, water), clasts like these will tend to be aligned according to their most stable hydrodynamic orientation. Identification of preferred clast orientations provides another useful tool for measuring paleoflow directions.

Table 13.3.6: Structures formed by sediment textures and fabrics

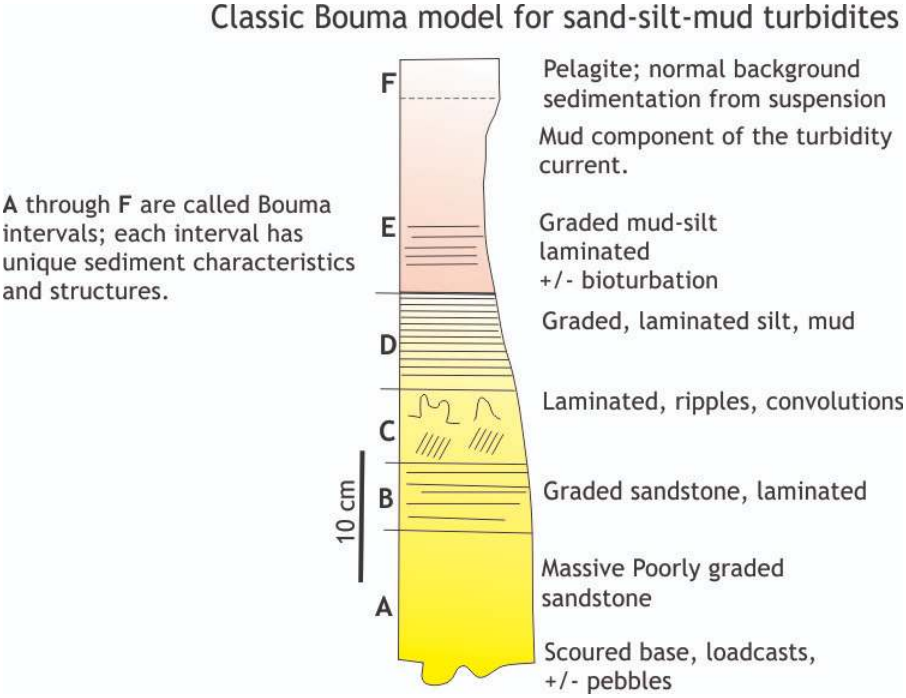
Sedimentary structure	Primary attributes	Common environments
	Imbrication (preferred alignment) of flat pebbles and cobbles in a recent river channel. The flat clasts are consistently inclined upstream; flow was to the right.	Common in gravelly rivers, particularly braided rivers, but can occur in any channel containing flattened clasts or shells.
 (Callan Bentley photo illustration.)	Imbrication of oblong or platy gravel clasts can serve as an indicator of paleo-current flow direction. This example comes from Death Valley, California.	A similar effect can be seen in leaves shed from deciduous trees, as they “shingle up” on on another when washed along by rainstorm currents.
	Current alignment of Permian fusulinids (foraminifera) on bedding. The sense of flow direction is ambiguous in this case, but might be confirmed by other structures in the rocks.	Current alignment of fossils is common in channels, or over broader substrates like the sea floor where strong tidal currents are present.
	Bedding view of parting lineation in laminated sandstone; each parting is one or two grains thick. This fabric develops under Upper Flow Regime Plane Bed conditions. Paleocurrent directions are ambiguous.	Most common in fluvial or tidal channels where high flow velocities are sustained.

STRUCTURES FORMED BY SEDIMENT GRAVITY FLOWS

This section contains images and brief descriptions of **turbidites** and **debris flows**.

Turbidites are mostly found in deep marine basins and lakes, but are also known to occur in shallow settings like continental shelves. They are deposited from turbulent, bottom-hugging flows of water carrying (mostly) sand and mud. Turbidity currents can travel many tens of kilometers across the sea floor. Turbidity current deposits, called “turbidites,” are the principal components of submarine fans.

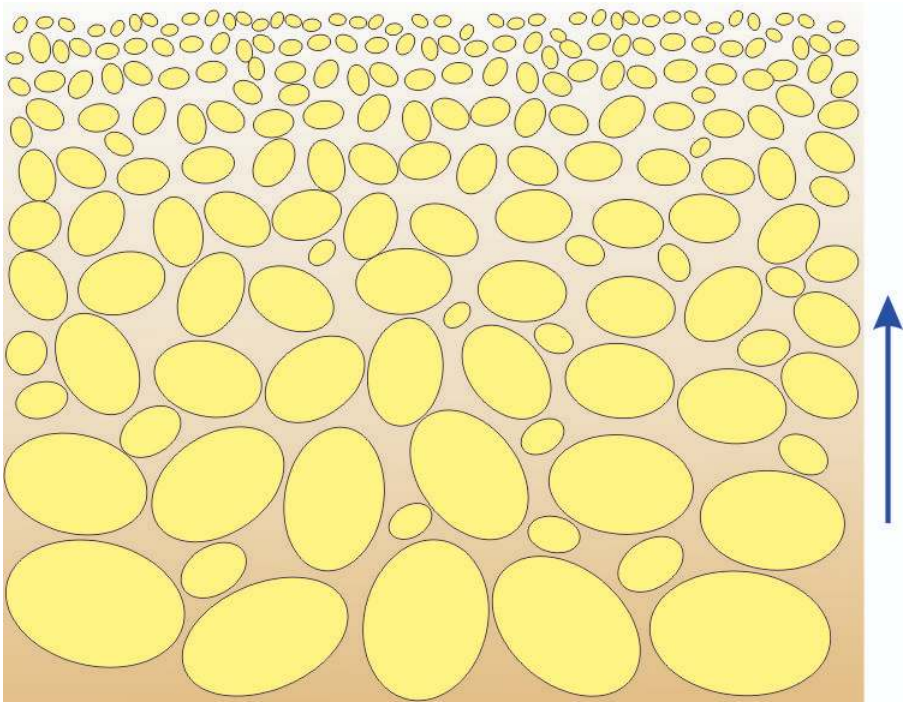
The classic descriptive model of a turbidite deposit was developed by Arnold Bouma, and is appropriately called a **Bouma Sequence**. The sedimentary components of a typical Bouma Sequence are shown below. A Bouma sequence represents a single turbulent flow.



A complete turbidite contains intervals A through E.

A Interval: This interval contains the coarsest sandstones, and may include pebbles or chunks of mudstone that were ripped from the sea floor during passage of the turbulent current – not surprisingly these chunks are called mudstone **rip-ups** *(or "rip-up clasts")*. Contact between the A interval and the underlying deposit is abrupt and may be scoured; the scours may be fluted and /or filled with pebbles.

B Interval: The B interval is also sandy, but is distinguished from the A interval by two common attributes: (1) the sandstones are graded (coarser framework grains at the bottom of the interval, becoming finer towards the top), and (2) the sandstones are commonly laminated.



C Interval: Mostly fine-grained muddy sandstone typically containing sandy ripples and climbing ripples, and ripple crossbedding that has been deformed into convoluted patterns (see the examples below).

D Interval: Mostly laminated, graded, muddy siltstone. This interval represents the late stages of deposition from the turbidite.

E Interval: This interval contains the mud component of a turbidity current, most of which falls out of suspension from the cloudy plume above the main flow. It is generally mixed with background suspended sediment from the water column above; it can be difficult to distinguish these two sediment components.

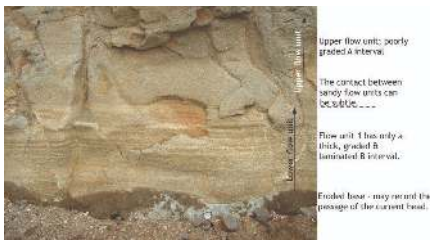
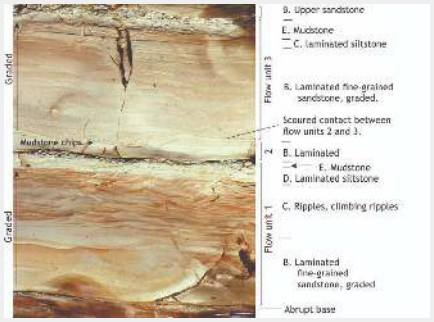
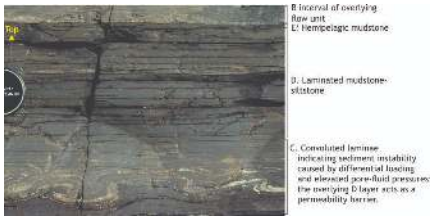
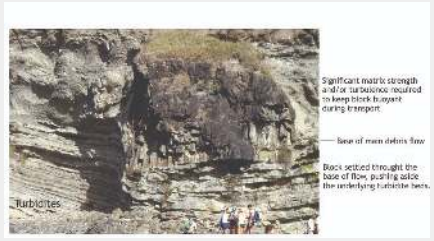
F Interval: This interval represents a return to 'normal' background sedimentation which in many cases is a mix of pelagic (*e.g.*, skeletal remains of micro-organisms) and suspended particles in the water column (the latter is also referred to as hemipelagic mud).

In thick turbidite successions, it is common to find individual flow units that are incomplete Bouma sequences; thus one flow unit may preserve only A and B interval sands, where others present B, C, D and E, or C, D, and E intervals. Variations like these generally reflect proximity to the source of the turbidity current, as well as the availability of sediment. For example, sandy turbidites containing thick A and B intervals are more likely to be deposited in the proximal parts of submarine fans (*i.e.*, closer to the sediment source and the head of the fan); finer-grained turbidites lacking thick A or B intervals will tend to accumulate in the more distal parts of submarine fans (*i.e.*, further to the sediment source and closer to the outer fringes of the fan).

Debris flows are also mixtures of mud, water, and coarse debris, but unlike turbidites they lack fluid turbulence during flow. The capacity of a debris flow to carry material, including house-sized blocks, lies in the viscosity and mechanical strength of its mud matrix. Debris flows are mobile, commonly destructive phenomena. In terrestrial settings, they can evolve from landslides, aided by high precipitation or snow melt. The equivalent phenomena in volcanic terrains are called **lahars** – debris flows consisting almost entirely of volcanic debris – that develop both during and after eruptions.

Subaqueous debris flows are commonly generated during slope failures (submarine landslides) that are triggered by seismic events or gravitational instability. Turbidites and debris flows are often found together in the rock record.

Table 13.3.7: Structures formed by sediment gravity flows

Sedimentary structure	Primary attributes	Common environments
	Two coarse-grained, sandy turbidite flow units – the low unit has a single, graded B interval; the upper unit begins with a poorly graded A interval (remainder of this unit out of the picture). Contact between the two units is planar.	Thick, coarse-grained turbidites tend to accumulate on the proximal parts of submarine fans.
	Three turbidite flow units: A lower B-C-D-E unit with excellent examples of ripples and climbing ripples in the C interval; a middle B unit, the top of which has been eroded by the upper B-C-D-E unit. Units 1 and 3 are graded. A lag of mudchips formed during scouring of the middle unit.	Probably mid-submarine fan
	A fine-grained, muddy turbidite with well developed C and D intervals. The convoluted bedding (small fold structures) formed after deposition during the early stage of compaction	Mud-dominated turbidites (lacking sand) tend to accumulate on the outer, more distal parts of submarine fans.
	A pebbly mudstone, or muddy debris flow, showing typical high mud content that supported the larger, denser clasts during flow – i.e., a matrix-supported texture, that indicates viscous flows (generally lacking the turbulence of a turbidite). The bed in this images dips to the left.	Commonly associated with turbidites on submarine fans. Can be initiated by collapse of submarine slopes or landslides.
	Good examples of matrix-supported debris flows in the San Onofre breccia. Most clasts here are angular indicating little or no mechanical abrasion during flow. Clasts include glaucophane schist derived from subducted rocks in a nearby metamorphic terrane.	Clasts in debris flows tend to be pebble through boulder size, and are extremely useful for identifying ancient sediment sources (i.e., provenance)
	An Early Miocene debris flow that carried megablocks – in this case, a large piece of jointed basalt. Clearly demonstrates the mechanical strength of the mud matrix and the mobility of such flows.	

Sedimentary structure	Primary attributes	Common environments
	A succession of submarine debris flows initiated on the submerged flanks of early Miocene volcanoes. Image right shows typical textural variations: very poor sorting, a range in clast shapes, and variable clast supporting characteristics ranging from matrix- to clast-supported.	The general category of debris flows includes lahars – flows generated on the subaerial or submerged flanks of volcanoes, and consisting of volcanic debris.

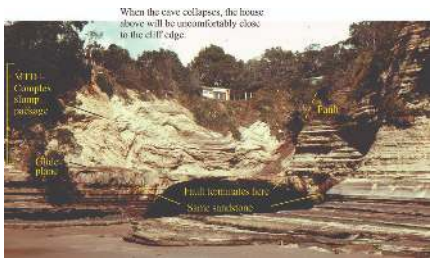
STRUCTURES FORMED BY MASS MOVEMENT (MTDS)

Mass Transport Deposits, or **MTDs**, is the term given to sedimentary slumps and slides, mostly generated on relatively high-angle slopes between the shelf or platform margin, and deep-water settings at the base-of-slope and beyond. The term is generally reserved for sediment packages that move and deform en masse under the influence of gravity.

There is a close association between MTDs and autochthonous (undisturbed) muddy slope deposits, and turbidites in submarine fans. MTD packages commonly overlie undisturbed turbidite assemblages, and in turn are overlain or draped by non-deformed strata. The slumps, and slides themselves generally consist of deformed turbidites and related deposits.

MTDs develop via a range of emplacement mechanisms and mechanical processes; most sediments will be ‘soft’, unconsolidated or only mildly so, and have high interstitial fluid contents (usually seawater). Sedimentary layers may bend and fold, or break like brittle materials. Liquefaction is also common where sediment becomes fluid-like. All these mechanisms may occur in the same structure. The deforming slump or slide package may also generate debris flows and turbidites.

Table 13.3.8: Structures formed by mass movement (MTDS)

Sedimentary structure	Primary attributes	Common environments
	Turbidite beds, while still relatively soft, have been folded and overturned into a recumbent anticline. This structure is part of a much larger package of deformed strata that moved down the ancient Miocene submarine slope	Most common in continental slope and deeper basin environments, associated with submarine fans.
	A smaller-scale folded and faulted sandstone bed within a succession of turbidites. Folding and sliding of soft sediment also formed small faults, and squishing of water-logged that liquefied the muddy sediment. Early Miocene, Auckland.	Deformation of soft sediment can involve brittle failure (small faults), plastic-like folding, and flow from liquefied sediment
	The slump package, or MTD, (Miocene, Auckland) is separated from undeformed turbidites by a relatively smooth glide plane (a plane of movement). The MTD is bound on the right by a fault – the fault terminates at the glide plane. The fault formed during the slumping and deformation event.	MTDs commonly overlie undeformed strata, separated by a planar surface that is also undeformed.
	Large-scale folds and faults in Pliocene turbidites, Ridge Basin, California. Some of this deformation may have been triggered by seismic activity on the ancestral San Andreas fault.	Slumping and sliding can be triggered by oversteepening of the sea floor as sediment accumulates (gravitational), by seismic events, or large storm surges.
	Small-scale folds and breakage of thin beds of carbonate mudstone (lutite). This example formed in lime mudstones on a slope adjacent to a Paleoproterozoic carbonate platform (Belcher Islands).	Sedimentary slumps and slides can occur in siliciclastic, carbonate, and volcaniclastic deposits

STRUCTURES FORMED BY SEDIMENT COMPACTION AND DEWATERING

The term **syn-sedimentary deformation** tends to be used rather loosely, as deformation that takes place during or soon after deposition; the ‘soon’ is the loose part of this broad definition. Sediment begins to compact almost immediately following deposition, where framework grains move closer together. As compaction progresses, interstitial water (i.e. the water between grains at the time of deposition) is expelled, and this process in itself can deform the sediment. Water that is expelled from sediment can escape to the surface (e.g. the sea or lake floor), or it can be prevented from flowing by low permeability layers –

particularly those containing mud. When this happens, local pore pressures increase, and this, in turn, reduces sediment shear strength, promoting deformation.

Note that this kind of deformation is an integral part of the Mass Transport Deposits pictured above, but it also takes place in beds where there is no wholesale slumping or sliding. Soft sediment deformation is common in conditions of rapid deposition where interstitial water is trapped, as in turbidity currents and sandy fluvial channels.

Some of the more common structures illustrated here include:

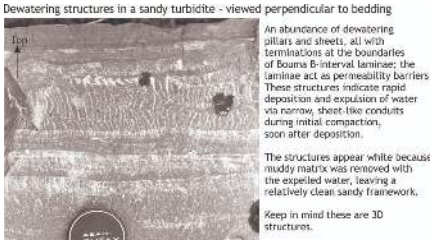
Load structures (also called **load casts,** or **ball-and-pillow structures**): Common in newly deposited beds where a higher density layer, like sand, overlies a lower density mud. Differential loading cause bulbous shaped sand bodies to project into the mud. Laminae within the load cast are also deformed. Load casts may detach completely and appear to float in the mud layer.

Flame structures: Closely spaced load casts will force relatively fluid mud upwards, forming tapered, flame-like mud wisps that protrude into the overlying sand.

Dewatering pillars and sheets: During compaction, interstitial water may escape along preferred pathways; these escape route may be tube-like, or form as vertical sheets. Dewatering pillars and sheets that extend to the sediment surface can carry suspended muddy sediment that will accumulate around the vent as mud blisters or small **mud volcanoes**.

Dish structures: Dewatering of saturated sands can truncate primary laminae, forming concave-upward, or dish-shaped structures.

Table 13.3.9: Structures formed by sediment compaction and dewatering

Sedimentary structure	Primary attributes	Common environments
	Detached load casts of volcanic ash appear to float in blue-grey mudstone. Laminae within the casts are intricately folded in the upper layer (lens cap). Load casts probably formed at about the same time as the dewatering pillars. Most of the primary bedding here was deformed soon after deposition. Paleoproterozoic, Belcher Islands.	Common in turbidites and interbedded lithologies where there are density and permeability contrasts. They indicate instability between the two layers during early compaction.
	Numerous load casts at the base of the top-most sandstone (coin for scale) have protruded into the underlying dark grey mudstone. Wispy mudstone flame structures are sandwiched between the casts. Paleoproterozoic, Belcher Islands.	Flame structures are commonly associated with load casts.
Dewatering structures in a sandy turbidite - viewed perpendicular to bedding 	Bed cross-section view of dewatering pillars in the laminated (Bouma) B interval of a sandy turbidite. The pillars are white because fine matrix was removed during water expulsion. Paleoproterozoic, Belcher Islands.	Indicative of rapid deposition of water saturated sediment. Sheets are layered because of permeability contrasts in the sandstone.
 Paleoproterozoic Omarolluk Fm, Belcher I.	Bedding exposure of small blisters, or mud volcanoes at the exits of dewatering pillars. Paleoproterozoic, Belcher Islands.	
	Dish structures in Rosario Group sandstone, San Diego. The dewatering pillars occur between adjacent dish-shaped laminae.	Indicative of rapid deposition of water saturated sediment.

TRACE FOSSILS



Trace fossils have the privilege of being two things at once: sedimentary structures, and fossils. They occur in sediment, are made of sediment, but represent the activities of creeping, crawling or burrowing critters, mostly at or immediately below the sediment water interface (marine, lacustrine, estuarine, swamp), or subaerial environments such as dune fields. As such, trace fossils represent the range of activities that critters are normally occupied with – grazing or foraging for food, home construction and house-keeping, predating or escaping predators, wandering aimlessly, or taking a nap after an exhausting day. Some critters like to rough it, preferring the tumble of waves or strong currents, while others like the peace and quiet of deeper realms. Lives are frequently interrupted by storms or violent, turbulent flows of sand and mud; their traces, or lack of them, also reflect these events.

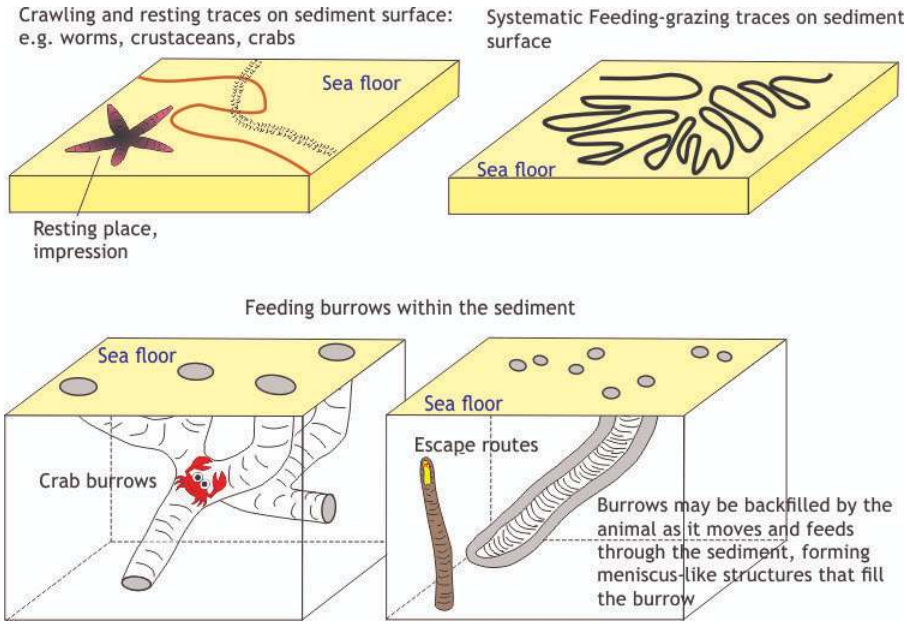
Most animals produce more than one kind of trace depending on what they are doing, which means that in most cases, traces reflect animal activity and biometrics, rather than the specific critter species. Most traces do not contain any remnants of the animal that made them (there are a few exceptions); finding a trilobite body fossil at the end of its scampering trail is pretty rare.

Trace fossils provide valuable information on benthic communities, environmental conditions such as wave or current energy, redox conditions (presence or absence of oxygen), rates of sedimentation, or periods of time when sedimentation slowed (e.g. hiatuses, disconformities, omission surfaces).

Intense bioturbation can also obliterate other kinds of sedimentary structures; for a geologist, this may be an annoyance or a happy circumstance. Most Precambrian successions are free of trace fossils and bioturbation; this changed during the Ediacaran, the period that appears to have been a kind of precursor to the Cambrian invertebrate explosion. Most Phanerozoic sedimentary successions (since 540 million years ago) have enjoyed the munching-burrowing efforts of a myriad nameless critters.

The images below are organized into six ethological (i.e. behavioral) structures:

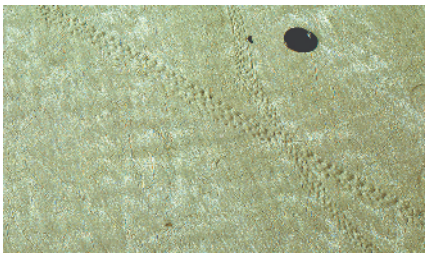
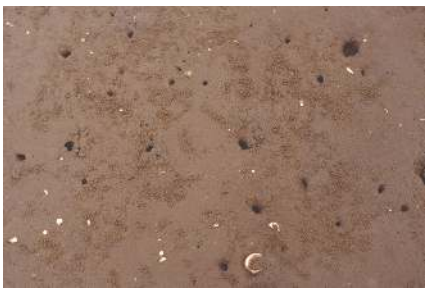
- *Resting traces:* the animal is taking a break or escaping a marauding predator.
- *Crawling traces:* basically moving from A to B, so the emphasis is on body and limb movement.
- *Grazing traces:* The search for food at the sediment surface or just beneath it – one interesting analogy is the similarity to strip mining.
- *Feeding structures:* Like grazing, but in this group the critters burrow (subsurface mining), forming as they go, simple temporary burrows or branched burrow complexes.
- *Dwelling structures:* These are more permanent burrows that the animals call home.
- *Escape structures:* These too are temporary burrows, made while attempting an escape from sediment burial, or from predators; hence the critter may move upward or downward in the sediment, depending on the circumstances.

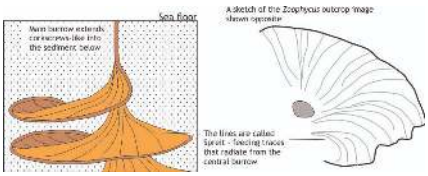
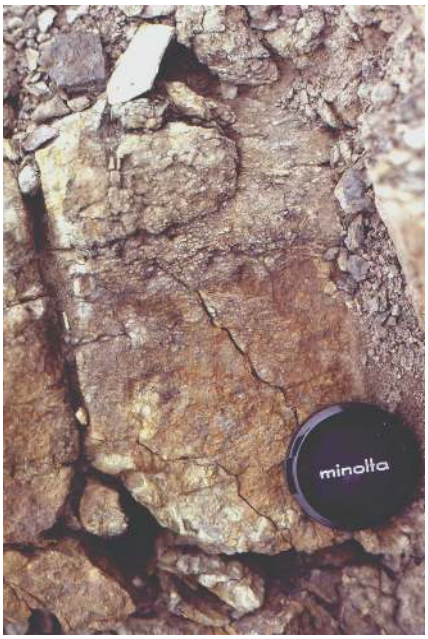


Trace fossils can be divided into groups according to the behavior of the animals that produced them. Here are four schematics of resting, crawling grazing, feeding, dwelling, and escape activities.

Table 13.3.10: Trace fossils

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Sedimentary structure	Primary attributes	Common environments
	**Resting trace:** A Permian starfish, or brittle star found its final resting place on this sand bed. The impression leaves a faithful record of the size of the starfish. Ellesmere Island.	Sandy substrates, intertidal to shallow subtidal.
	**Crawling trace:** Possibly a crustacean wending its way home, seemingly going in circles across the sea floor, or within the sediment just below it. The substrate was fine sand.	Common in many lower energy environments (away from direct wave action and strong currents)
	**Crawling & Grazing traces:** Recent, large crab traces with impressions from its appendages; this is a crawling trace. Look closely to see a myriad small, criss-crossing grazing trails made by small crustaceans. These examples from a sandy beach.	Intertidal to subtidal, soft substrates.
	**Dwelling structures:** The Ichnogenus *Ophiomorpha* – Simple burrows, vertical or obliques to bedding, typically lined with small muddy nodules, & uniform burrow fill. Commonly in sandstones that also contain evidence of shallow marine waters (*e.g.*, fossils, crossbeds).	Shallow subtidal, sandy substrates where sediment at the sea floor is moved by waves and tidal currents.
	**Dwelling structures:** Modern, vertical crab burrows in firm tidal flat sand. Note the small sand nodules at the entrance to each burrow; ejected by the animal as it excavates.	Intertidal, beach, to shallow subtidal, sandy substrates where sediment at the sea floor is moved by waves and tidal currents.

Sedimentary structure	Primary attributes	Common environments
	**Feeding traces:** The ichnogenus *Zoophycus* has a central burrow, or stem, about which the animal grazed into the sediment in a corkscrew fashion. Two specimens in this view.	Relatively quiet conditions where surface sediment is not continuously stirred. The deep, outer shelf, slope, and basin at bathyal depths.
	Schematic of basic corkscrew structures of *Zoophycus*. The sketch on the right outlines the central burrow and spreit (successive feeding traces) in the outcrop image above.	Feeding structures like these show systematic grazing patterns across the sediment surface and within sediment.
	**Escape structures:** Simple vertical structures that begin from the base of a bed, reflecting the critter's desire to escape to the sea floor	Can occur in almost any environment where rapid deposition of sediment takes place
	An example of rampant and pervasive bioturbation that has all but destroyed the primary layering (parallel laminations, crossbedding) within a bed.	Can potentially occur in many environments where benthic faunas proliferate.

Unless indicated otherwise, all the images used in this chapter are by Brian Ricketts.

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CHAPTER OVERVIEW

14: Facies



Figure 14.1: Limestone outcrop showing facies changes from finer-grained, deeper water light gray sediments at the base, to coarser, shallower water tan sediments at the top, Clays Ferry, Kentucky. (Photo by K. Layou.)

14.1: What are facies?

14.2: The CTSFAF model

14.3: Putting it all together - facies models

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14.1: What are facies?

Sedimentary rocks contain many characteristics at a variety of spatial scales, and basic observations at both the hand sample and outcrop level can aid in interpreting depositional environments. However, some of these hand sample and outcrop scale characteristics are not unique to one depositional environment–for example, well-sorted, rounded quartz sand can be found in both beaches and deserts. Therefore, other rock features like sedimentary structures, fossils, association, and form are needed to provide richer context. Together, these features are known as a facies, a body of rock (typically a sequence of several strata, or beds) that is distinct from adjacent strata based on observable characteristics.

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14.2: The CTSFAF model

One model for describing sedimentary facies in the lab or the field is the CTSFAF model. This model uses readily observable characteristics of the rocks and aids in the interpretation of depositional environments. The six aspects of CTSFAF are color/composition, texture, sedimentary structures, fossils, association, and form which are described in more detail below.

Table 14.2.1: CTSFAF

Characteristic	Description	Spatial scale of observation
C: color and composition	Grain colors and % mineral composition	Hand sample, microscope
T: texture	Grain size, shape, and sorting	Hand sample, microscope
S: sedimentary structures	Depositional/erosion features	Outcrop, hand sample
F: fossils	Fossil diversity and abundance	Outcrop, hand sample, microscope
A: association	Geologic contacts of beds	Outcrop
F: form	Bedding pattern	Outcrop

Color and composition

Color



Figure 14.2.1: Weathering rind showing difference in color on external surfaces versus inner portion of rock. (Photo by C. Bentley.)

The color of a sedimentary rock is typically derived from the minerals that make up the grains within the rock. For clastic sedimentary rocks, the cement holding the grains together may also significantly contribute to the color of the rock. For example, many of the gorgeous red and orange sandstones from the American Southwest are quartz sand cemented with hematite, a rust-colored iron oxide. Some sedimentary rocks can have a different color in a weathered state, thus when observing sedimentary rocks in the field, it is important to always conduct observations on a freshly broken surface of the rock.

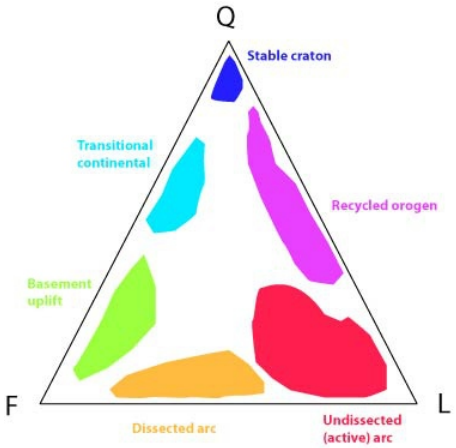


Figure 14.2.2: QFL ternary diagram for clastic sedimentary rocks.

Broadly, sedimentary rocks can be classified as clastic, biochemical, or chemical. In the context of facies, it is helpful to think about clastic and biochemical sedimentary rocks as either siliciclastic or carbonate. Siliciclastic rocks are those that are made from weathered grains of pre-existing rocks made of predominantly silicate minerals. For rocks with gravel or sand-sized grains (breccia, conglomerate, sandstones) noting percentages of quartz, feldspar, and lithic fragments (pieces of rock) in the rock (QFL), clues about tectonic history can be determined. High quartz percentages suggest tectonic stability or extensive reworking of sediments as the more chemically and less physically resistant minerals have eroded away. Lots of feldspar indicates immature sediments that have recently been produced via basement uplift of continental crust (*i.e.*, thick-skinned deformation including cratonic rifting), since otherwise feldspar quickly weathers to clay minerals. Lithic fragments are indicative of volcanic activity and/or shallow tectonic activity (*i.e.*, thin-skinned deformation) and also indicate immature sediment. QFL ternary diagrams can be used to classify these rocks. Siltstones, claystones, and shales are composed of the weathering products of silicate minerals.

Carbonate rocks are named such because they are made of minerals with a carbonate $(CO_3)^{-2}$ anion: usually calcite (which makes limestones) or dolomite (which makes dolostone). While there are some limestones that form in lake or hot spring environments on land, most limestones are associated with marine depositional environments, including epicontinental seas, continental shelves and slopes, and the abyssal plains. These limestones are made from clasts derived from skeletal elements (shells) or fecal pellets of organisms (peloids), chemically precipitated grains (ooids), or pieces of other carbonate material (rip-up clasts, clay-sized particles).

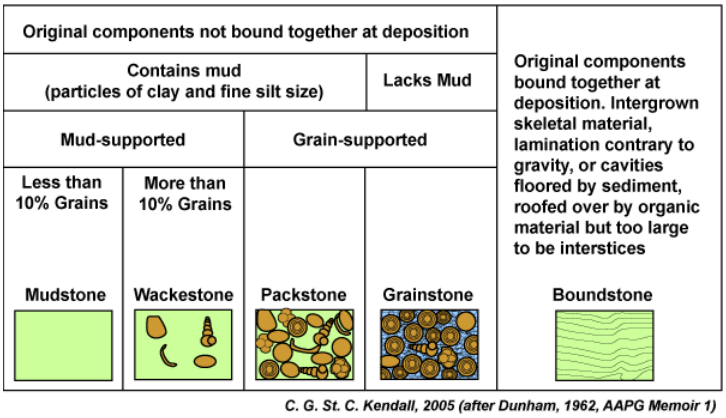


Figure 14.2.3: Dunham classification scheme for carbonate rocks

The Dunham classification system for carbonate rocks is useful to describe carbonates in the field and in hand samples. It compares the amount of clasts to the amount of clay-sized particles within a sample, and is analogous to the classification scheme for siliciclastic sedimentary rocks. Grain-supported grainstones and packstones have more clasts, while wackestones and mudstones have more fine grains.

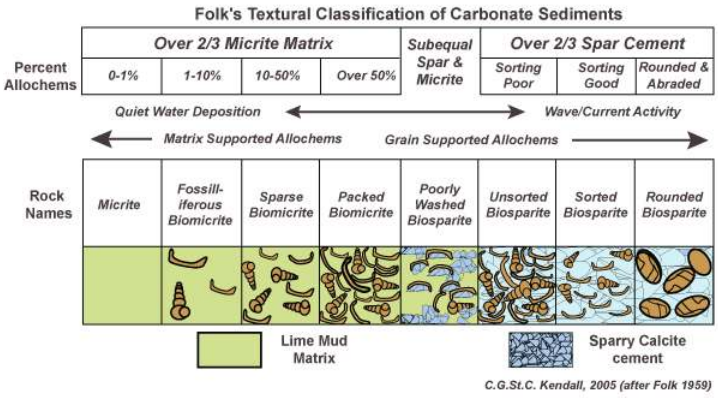


Figure 14.2.4: Folk classification scheme for carbonate rocks

Another classification system for carbonates, the Folk system, focuses on the type of clasts (allochems) and the character of binding (interstitial) material. This system is particularly useful when describing carbonates in microscopic thin section.

Texture

When describing sedimentary texture, three aspects must be detailed: grain size, shape, and sorting. All of these factors are dependent on the distance sediment has been transported from its source, where it was originally weathered from bedrock. As sediment is transported by currents from source to depositional environment, it becomes smaller, rounder, and more well-sorted.

	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 14.2.5: Classification of grain sizes. Silt and clay are considered fine-grained particles, sand is medium-grained, and particles larger than sand are considered coarse-grained. (CC BY 4.0; Karla Panchuk (2016))

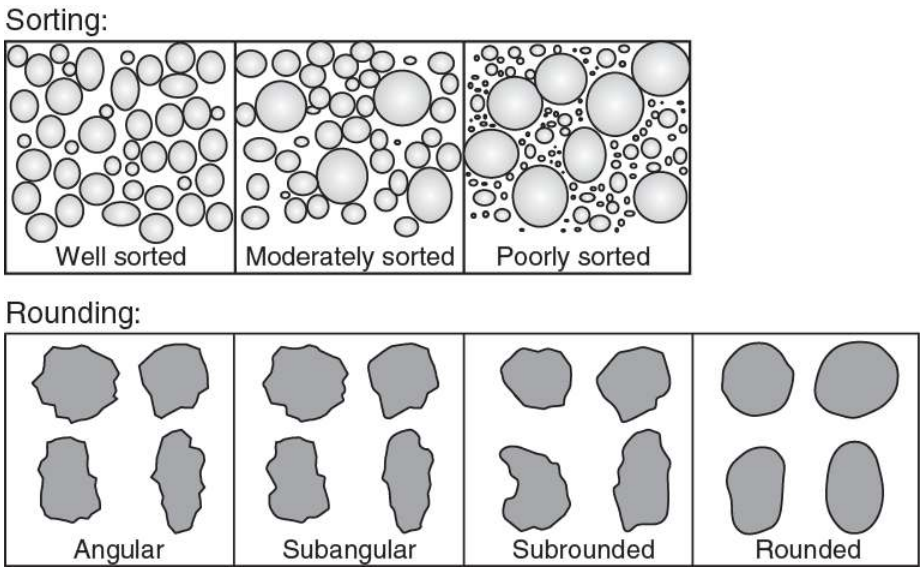


Figure 14.2.6: Top: Sorting of grains, ranging from well sorted where the grains are similar in size, to poorly sorted, where the grains vary greatly in size. Bottom: Rounding refers to how smooth or rough the edges of a clast are. Clasts with sharp edges and corners are angular. Clasts with smooth surfaces are rounded. Clasts that fall in between are sub-angular or sub-rounded. (CC BY 3.0; Reagan et al. (2015))

Sedimentary structures



Figure 14.2.7: Ripples in cross section, Corridor H, West Virginia. (Photo by K. Layou.)

Sedimentary structures are valuable indicators for determinations of current energy, direction of flow, intervals of wetting and drying and many other subtle aspects of depositional environments. Examples include mudcracks, ripples, cross-beds, tool marks, laminations, and more. See this page for detailed discussion and examples of sedimentary structures.



Figure 14.2.8: Fossiliferous layers of the Yorktown Formation at Cobham Wharf, Virginia. (Photo by K. Layou.)

The fossil content of a sedimentary bed can provide key information regarding the depositional environment in which the layer formed. As with the sediments, observations can be made at multiple spatial scales: microscopic, hand sample, and outcrop. Both diversity and abundance should be recorded. Diversity includes a list of all types of fossils found within the sample or outcrop. Abundances can record actual counts of fossil occurrences, or relative occurrences (*e.g.*, rare, common, abundant). Within a hand sample, qualities of fossil preservation (taphonomy) should be noted.

At an outcrop level, it is useful to consider the distribution of fossil content within individual beds to reveal any patchiness, or spatial variability, within the environment. Vertical stratigraphic changes in fossil content should also be recorded as the fossils can reveal both significant or subtle changes in water depth, temperature, salinity, and sedimentation rates.



Figure 14.2.9: Conformable Ordovician limestone layers near Carthage, Tennessee. (Photo by K. Layou.)

One important feature of strata at the outcrop level is how sedimentary layers, or strata, come into contact with one another. The place where one rock type touches another is known as geologic contact. Sedimentary layers exhibit depositional contacts. The contact is described as *conformable* when the layers pile atop one another in succession–they "conform" to expectation. Alternatively, if there is an unconformity present between layers then the contacts are *unconformable*. Remember an unconformity means there are missing layers due to nondeposition or erosion. Other types of geologic contacts include *fault* contacts where different rock layers touch on opposite sides of a fault, and *intrusive* contacts, where there is commonly contact metamorphism as an igneous intrusion of magma comes into contact with surrounding country rock.

Form



Figure 14.2.10: Lenticular bedding of a channel sandstone, note how the thickest light tan layer in the center of the image thins and pinches out into the foreground and background, Corridor H, West Virginia. (Photo by K. Layou.)

At an outcrop level, the form of a sedimentary bed can be observed. In some cases, individual layers may maintain their thickness across the entire outcrop, creating *tabular*, or parallel, beds. This type of bedding is common when sediment supplies are constant and the depositional environment is laterally pervasive. Other times, a sedimentary layer may pinch out, thinning to the point of disappearing from the stratigraphic section, creating a *discontinuous* bed. This is often seen in fan-shaped structures that develop in alluvial fans, deltas, and submarine fans. In some cases, a layer may pinch out in both directions, creating a *lenticular* bed. This is common for channel sandstones in river or deltaic systems.

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14.3: Putting it all together - facies models

Although many models exist, here we will discuss a range of commonly preserved sedimentary depositional environments, starting with terrestrial environments at the highest elevations on the continents, moving toward the coast, into marine environments of the continental shelf and onto the abyssal plains. The focus will be on the four facies model characteristics that are most observable in hand samples.

Terrestrial depositional environments

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Figure 14.3.1: Meltwater and lateral moraine till, Athasbasca Glacier, Banff National Park, Canada. (Photo by K. Layou.)

- Color: variable depending on landscape
- Composition: lithics
- Texture: gravel to clay, poorly sorted, angular to subrounded
- Sedimentary structures: may see some asymmetric cross-bedding in association with outwash deposits
- Fossils: rare
- Other notes: till, moraine deposits

Alluvial fan



Figure 14.3.2: Am Magh, the alluvial fan that divides Lochan na h-Earba into two lakes, Scotland. (CC BY-SA 2.0; Richard Webb)

- Color: tan to pink-red
- Composition: primarily feldspars and lithics
- Texture: gravel to clay, poorly sorted, angular to subrounded
- Sedimentary structures: generally absent
- Fossils: rare
- Other notes: massive beds common from one flow

Braided rivers



Figure 14.3.3: Braided river deposits, Banff National Park, Canada. (Photo by K. Layou.)

- Color: tan to gray
- Composition: primarily lithics, also quartz and feldspars
- Texture: gravel and sand, subangular to subrounded, poor to moderate sorting
- Sedimentary structures: asymmetric cross-beds
- Fossils: larger, robust fossils, commonly plants and vertebrates
- Other notes: gravel and sand bars (lenticular beds)



Figure 14.3.4: Loch an Eilean, lake near Aviemore, Scotland. (Photo by K. Layou.)

- Color: gray to black
- Composition: siliciclastic minerals (quartz, clay minerals from weathering)
- Texture: clay to silt, well-sorted
- Sedimentary structures: laminations
- Fossils: fish, insects, plants leaves/flowers, pollen
- Other notes: some carbonate sediment may be possible (tufas)



Figure 14.3.5: Elim Dune, Namibia. (Photo by Ron Bentley)

- Color: white to light pink or orange
- Composition: primarily quartz, hematite cements, evaporites
- Texture: sand, subrounded to rounded, well-sorted
- Sedimentary structures: large-scale asymmetrical ripples and cross-beds
- Fossils: vertebrates, some plants
- Other notes: sand grains may show some frosting from air abrasion

Meandering rivers



Figure 14.3.6: Meanders of Taskinas Creek, York River State Park, Virginia. (Photo by K. Layou.)

- Color: brown to reddish purple
- Composition: lithics and quartz
- Texture: sand to clay, subrounded, moderate sorting
- Sedimentary structures: unidirectional cross-beds, asymmetric ripples in channel; mudcracks, raindrop impressions on floodplain
- Fossils: terrestrial fish and other vertebrates, plants and roots
- Other notes: coarser sediment in the channel, finer grained material in the floodplain (coarse lenses in finer sediments in outcrop scale)

Transitional/coastal depositional environments

Deltas

- Color: brown to reddish purple
- Composition: lithics and quartz
- Texture:sand to clay, subrounded, moderate sorting
- Sedimentary structures: unidirectional cross-beds, asymmetric ripples in channel; mudcracks, raindrop impressions on floodplain
- Fossils: coastal plants (i.e. marsh grasses) and roots, marine fish, shells
- Other notes: coarser sediment in the channel, finer grained material in the floodplain, large foreset (incline beds) at outcrop scale

Beaches



Figure 14.3.7: Shoreline of beach at Cape Charles, Virginia. (Photo by K. Layou.)

- Color: white to tan
- Composition: primarily quartz, heavy minerals OR calcite
- Texture: moderate to well sorted sand, subrounded to rounded, moderate to well-sorted
- Sedimentary structures: symmetrical ripples, bidirectional cross-beds
- Fossils: shells
- Other notes: may see abundant shells and ooids in carbonate settings

Marine depositional environments

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Figure 14.3.8: Symmetrical ripples, tidal flat, Sapelo Island, Georgia. (Photo by K. Layou.)

- Color: light gray to tan
- Composition: quartz and clay minerals OR calcite
- Texture: silt to clay, subrounded, well-sorted
- Sedimentary structures: mudcracks, bidirectional ripples
- Fossils: stromatolites, snails, crustaceans and burrows
- Other notes: may see laminations if burrows not present

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Figure 14.3.9: Shallow carbonate marine environment, Bahamas. (Photo by K. Layou.)

- Color: gray to tan
- Composition: quartz and clay minerals OR calcite
- Texture: sand to clay, subrounded, well-sorted
- Sedimentary structures: hummocky beds, burrows
- Fossils: wide variety of invertebrates, fish, reefs in carbonate marine settings
- Other notes: may produce massive beds due to bioturbation; area of shelf affected by fairweather waves

Deep subtidal (lower shoreface)

- Color: dark gray to dark brown
- Composition: quartz and clay minerals OR calcite
- Texture: silt to clay
- Sedimentary structures:
- Fossils: wide variety of invertebrates, fish
- Other notes: area of shelf affected by storm waves may show evidence of storm deposits (more broken shells)

Offshore

- Color: dark gray
- Composition: quartz and clay minerals OR calcite
- Texture: silt to clay
- Sedimentary structures: laminations
- Fossils: planktonic microfossils, ammonites, vertebrates
- Other notes: area of shelf below storm wavebase

Submarine fan

- Color: dark gray
- Composition: quartz and clay minerals OR calcite
- Texture: silt to clay
- Sedimentary structures: Bouma sequences from turbidity currents
- Fossils: planktonic microfossils, ammonites, vertebrates
- Other notes: deposits will be thin as they are farther away (more distal) from sediment source

Abyssal plains

- Color: dark gray to black
- Composition: quartz (chert), calcite (chalk/micrite), clay minerals
- Texture: clay
- Sedimentary structures: laminations
- Fossils: planktonic microfossils, ammonites, vertebrates
- Other notes: deep sea sediments capping basalt of oceanic crust

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CHAPTER OVERVIEW

15: Paleoclimatology



Figure 15.1: Speleothem from Mawmluah Cave, Megahalaya, northeast India. The position of the 4.2ka event is shown. This specimen is 308mm long. (Source: Walker et al., 2018).

Key Concepts

After reading this chapter, students should be able to:

- Differentiate between climate and weather.
- Distinguish between climate and paleoclimate.
- Identify examples of extrinsic and intrinsic changes agents to the climate system.
- Describe a variety of paleoclimate proxies that are useful to scientists and that help us understand the history of our planet's climate through short and deep time.
- Identify several important climatic moments in our planet's past.

15.1: What is Climate?
15.2: Paleoclimatology- Historical Climate
15.3: Geological Sources of Important Elements for Life
15.4: Significant Paleoclimatological Events
15.5: Further Reading

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15.1: What is Climate?

Climate is the average of the weather for a region over time. There is a common confusion that arises when people use the word climate. Often, even if they think they are discussing climate, the conversation will begin with a discussion of the current weather. It might go something like this:

“It was really hot today! Must be global warming…”

And, indeed, it might be. However, what they are actually describing is the weather. Weather is the summation of current atmospheric conditions at any given moment. And, to measure it, we use things like thermometers, barometers, anemometers, and other devices. This information is collected and is used to develop models that can be used for predictive purposes. From these, you get your 1-10 day forecasts. These forecasts of the weather are entirely based on probabilities. A 20% chance of rain just means that, under current weather conditions (or those predicted), it tends to rain 20% of the time. These conditions are a part of a region’s climate story, but they are still defined as weather. While they use recent historical data, they are intended to describe the condition of the atmosphere at a moment in time…perhaps three days from now.

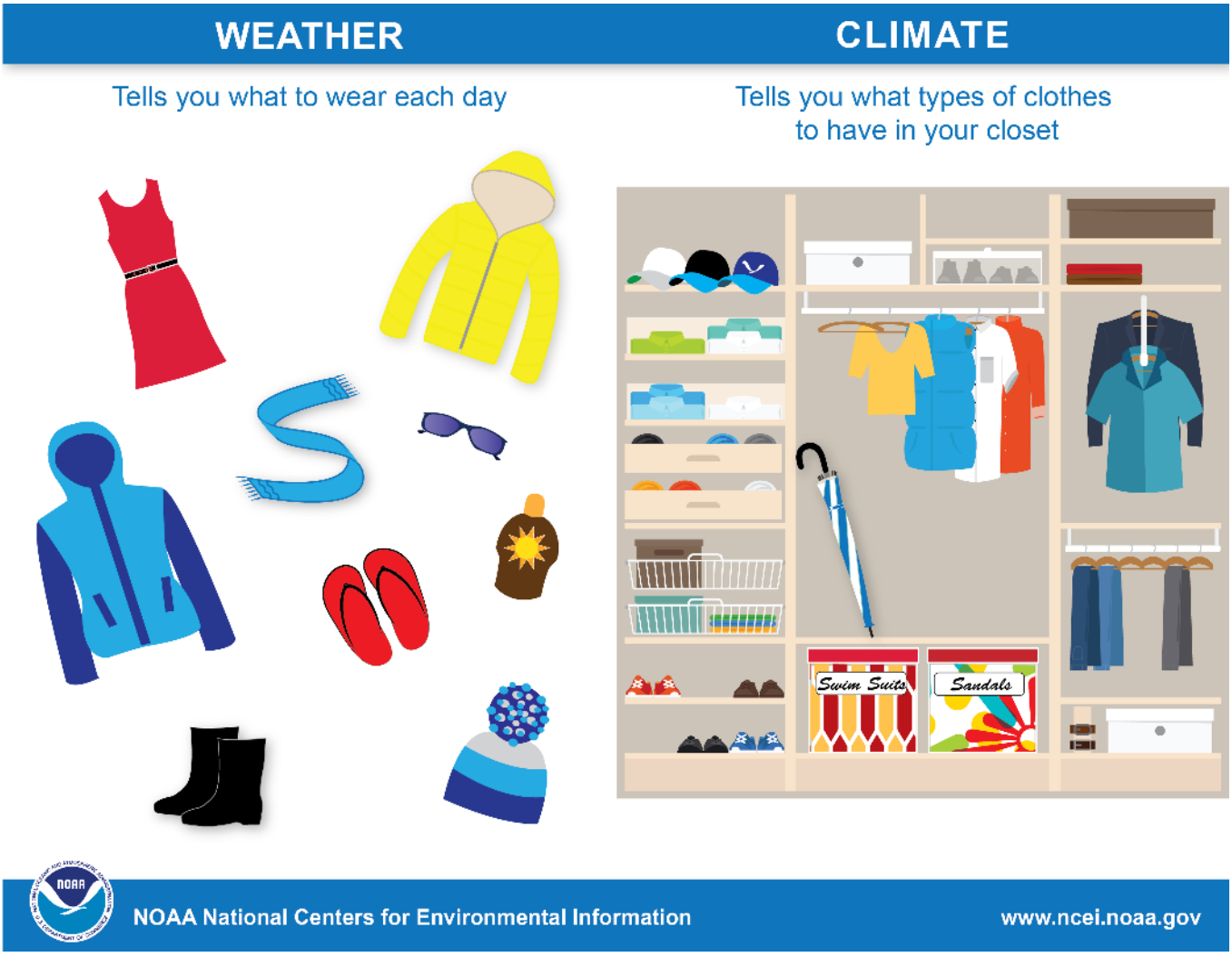


Figure 15.1.1: Illustration of weather versus climate. Do you own only shorts and sleeveless shirts? Or, do you wear a wide variety of clothing throughout the year? The composition of your wardrobe describes climate. What you pick out for the day describes the weather. (Source: NOAA, NCEI National Centers for Environmental Information)

Climate, on the other hand, is the average of these daily conditions over time. At its most basic and familiar level, climate can be described based upon seasons. Some places on Earth experience four of them (winter through autumn), others experience two (rainy and dry). These short term climatic conditions are based upon a large array of variables. These include latitude, topography,

proximity to the sea, time of year, and so forth. Looking at changes in climate over the course of the year at a given location on the planet provides a short-term picture unique to that location. Climate can also be described on a global average basis. Under these conditions, we can follow long-term changes in temperature and associated variables (greenhouse gases, ice loss, evapotranspiration, *etc.*) that provide a broader picture of global climate and how it is changing.

All of this is still be based upon instrumental records on the Earth's surface as well as satellite imagery and data from space. But what about the years before humans could measure these things? Is there any way of figuring out climatic conditions in the time before scientific measurement or human existence?

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15.2: Paleoclimatology- Historical Climate

Paleoclimatology is the study of the Earth's climate before instrumental records were available. The demarcation between paleoclimatology and climatology is analogous to paleobiology and biology. Both fields study climate, but one is very much focused on our current climatic situation. Paleoclimatologists must tease out of the Earth the story of its climatic past. And, that past is full of evidence of change.

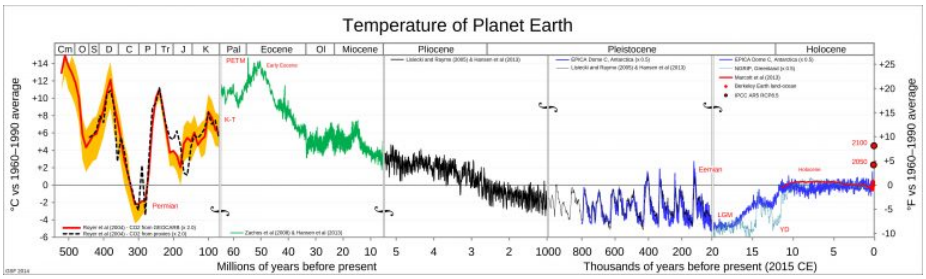


Figure 15.2.1: Global average temperatures for the past 540 million years, using graphs from various studies compressed together. This graph shows a great deal of time, but at a very low resolution (CC BY-SA 3.0; By Glen Fergus – Own work; data sources are cited below, https://commons.wikimedia.org/w/index.php?curid=31736468).

While we tend to focus on measuring our weather and climate using instrumentation on the ground and in space, the Earth has kept a record of its past climate in a wide variety of ways. In this chapter, we will explore a wide array of these kinds of paleoclimate evidence, including things like tree rings, layers of ice, and pollen. There are many more examples. Some of these record great detail over shorter periods of time while some record more coarse detail over much greater spans of time. In reality, this means that the Earth's data stores are really no different than the data stores we create in instrumentation. Measurement of climate is a matter of time and resolution. The usefulness of a particular data set is limited by time and resolution. So, like studies of modern climate, paleoclimatologists ask questions and pursue investigations that are guided by data limited by time and resolution.



Figure 15.2.2: Earth's various systems (Source: NASA).

There are some important assumptions accepted when studying paleoclimate. Like in all of the geosciences, these assumptions are uniformitarian. That is, we assume past processes generally operated the same way as in the present. That does not mean changes in the Earth's systems that affect climate always happen in the same way, but just that the processes that force or drive the changes do.

A key assumption is that the Earth's systems behaved throughout its history as they do today. And, we accept that the interactions among the Earth's spheres also have remained consistent. An event that forces movement within one system will have effects in other systems. These can lead to amplifying feedbacks that further upset the system's dynamic equilibrium and can lead to tipping points and a new set of system circumstances. Balancing feedbacks also still serve to work against such processes, as a natural means for the systems to right themselves and return to the former equilibrium.

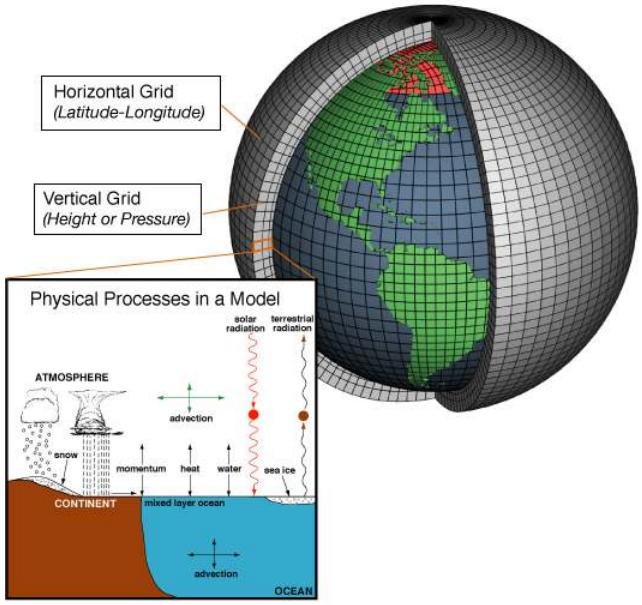


Figure 15.2.3: Climate models are made up of systems of equations that attempt to account for a wide variety of variables and their interactions. Known historical data is taken into account in creating those equations and physical processes, some of of which are listed in the inset here, are the focus. To run the model on a supercomputer, the globe and atmosphere are divided into a grid and the equations run (Source: NOAA).

The study of paleoclimate has lead to a wide variety of insights into our modern climate and how it works. It has helped to refine the models we use to study current problems associated with anthropogenic (human caused) warming. Studies of paleoclimate have also led to deeper knowledge of how our solar system works. While Galileo and others discovered sunspots, their periodicity and effect on climate at different scales and magnitudes would not be known if it were not for paleoclimatic research. Likewise, we would perhaps not know that the Earth, like other planets, sees gradual changes in its orbital parameters that affect its climate. Knowledge about these exospheric impacts informs our current research, not only because it helps us understand how our climate should be responding under only natural influences, but also because we can eliminate them as mechanisms behind the current changes we are causing. Paleoclimate research is critically important to understanding our future.

Intrinsic and Extrinsic Forcing Mechanisms

The Earth system is a part of the solar system. Extrinsic climate change agents must come from outside of the Earth system and result from the solar environment, or the exosphere. The solar system is powered by mostly by our Sun and also contains a variety of other objects, including planets, dwarf planets, and chunks of rock or ice hurtling through space in the form of asteroids or comets.



Figure 15.2.4: Fallen trees resulting from the Tunguska Event, Hushmo River, Siberia (Public Domain; By ru:Евгений Леонидович Кринов, member of the expedition to the Tunguska event in 1929. – [1] (original, black and white version of photo) / Vokrug Sveta, 1931 (current, color version of photo), https://commons.wikimedia.org/w/index.php?curid=200531).

Extrinsic forcing agents typically have a global effect, but in some cases effects can be more limited, such as the impact of a Tunguska-like asteroid versus a Chicxulub-like asteroid. The Tunguska asteroid impact, which happened over Siberia in 1908, had a wide variety of local effects, whereas the Chicxulub asteroid impact at 65 Ma devastated the planet's biosphere, killing off the dinosaurs and eventually giving mammals the evolutionary reins. The devastation of this impact led to massive global climate effects that contributed mightily to the extinction.

Likewise, intrinsic climate change agents come from within the Earth system. Overall, these forcing mechanisms can tend to be more limited in their impact, occurring on more local or regional scales rather than global. The Earth system consists of the atmosphere, hydrosphere, geosphere, cryosphere, and biosphere. Massive perturbation in any one of these could have climatic effects. From the geosphere, volcanic eruptions provide an excellent example of scale-dependent intrinsic climate forcing. In April 1815, Mt. Tambora, a stratovolcano in Indonesia, erupted. This eruption was very large and led to three years of significant climate change, particularly for the northern Hemisphere where Europe recorded the three coldest years in centuries and saw major crop failure and famine.

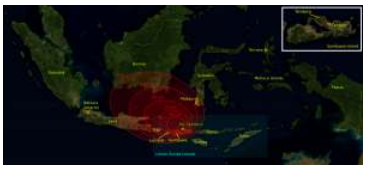


Figure 15.2.5: Projected ashfall resulting from the 1815 eruption of Mt. Tambora. While localized here, it was enough to force significant cooling over large parts of the globe for that year and into the next, leading to the "Year without a summer" (CC BY-SA 3.0; Wikimedia – The base map was taken from NASA picture Image:Indonesia_BMNG.png and the isopach maps were traced from Oppenheimer (2003).[1], https://commons.wikimedia.org/w/index.php?curid=1266774).

This calamity pales in comparison, however, to the major flood basalt eruptions of Earth's past. The Siberian Trap eruptions that took place at the Permian/Triassic boundary caused the largest mass extinction in the planet's history. This large igneous province in northern Siberia covered 7 million km^2 in lava and lasted for perhaps hundreds of thousands of years. CO_2 emissions from the lava led to runaway global warming and oceanic stagnation. Almost everything on Earth died.

EXTRINSIC CHANGE AGENTS

There are some key extrinsic change agents that affect Earth's climate from our exosphere. Below is a brief explanation of some of them. Ultimately, extrinsic factors do one of two things. They either adjust the amount of incoming solar radiation, like turning a handle on a spigot, or they cause direct perturbations to the Earth system. Most notably, changes in the hydrologic cycle and carbon cycle are good examples of systems that can be heavily affected by bolide impacts, for example. For this chapter, we will focus on three specific extrinsic change agents. These are Milankovitch Cycles, bolide impacts, and changes in solar energy output.

Milankovitch Cycles

Born in the rural Serbian village of Dalj in 1879, Milutin Milankovitch would make his mark on the study of climate in a big way. Through his research on insolation variation during Earth's seasons, he discovered a mathematical theory of climate that helped predict changes in Earth's climate due to changes in its orbit . His theory stated that as the Earth travels in its orbit around the sun, three cycles of variability will have an effect on changes in its climate. These are orbital eccentricity, axial obliquity, and axial precession.

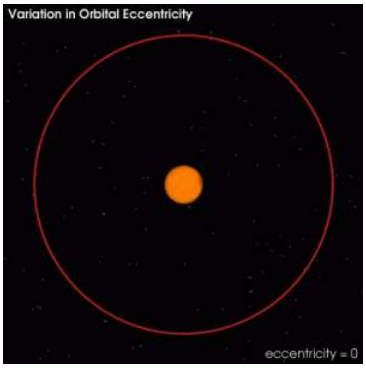


Figure 15.2.6: A perfectly circular orbit, which does not happen in planetary systems. Yet, it provides a model for what the other end of Milankovitch eccentricity is like. Note the position of the Sun at the center of the orbit. (Source: NASA)

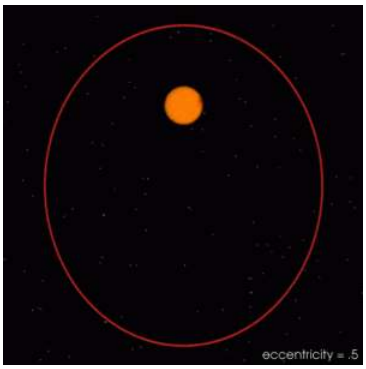


Figure 15.2.7: A highly elliptical orbit during an eccentricity cycle. Note the position of the Sun being located at one of the foci of the ellipse. (Source: NASA).

The longest of these cycles is eccentricity, lasting from 90,000 to 100,000 years per cycle. As this cycle progresses, the Earth's orbit stretches from more circular to more elliptical. The shift occurs because, while the Earth orbits the Sun, the giant planets Jupiter and Saturn over time exert gravitational forces on the Earth, which causes the Earth to shift over this period. As Johannes Kepler discovered well before Milankovitch, no planetary orbits are perfectly circular; all are elliptical (Kepler's First Law).

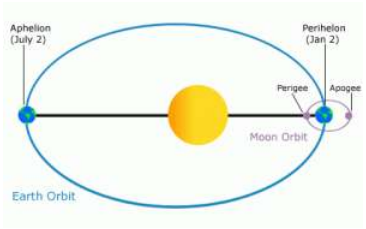


Figure 15.2.8: Image of Earth at aphelion and perihelion and the associated dates at present. Also shown here are positions of our Moon relative to its distance from Earth, apogee (farthest distance) and perigee (nearest distance). (Source: NOAA)

However, Milankovitch discovered that the shape of our orbit changes enough to have a significant effect on our climate. Currently, our orbit is in the phase of this cycle where it is more circular, leading to only a 6% increase in incoming solar radiation from January to July. In January, the Earth is at its closest approach to the Sun for the year (called *perihelion*) and in July, it is at its furthest approach (*aphelion*). When the orbit is more highly elliptical, the difference in incoming solar radiation at perihelion can be 20% to 30% greater than it is today. This would mean a much different climate than we have today. And, as you can see in the figure below, it is this 100,000-year cycle that governs cold and hot periods (glacials and interglacials) during the Pleistocene epoch.

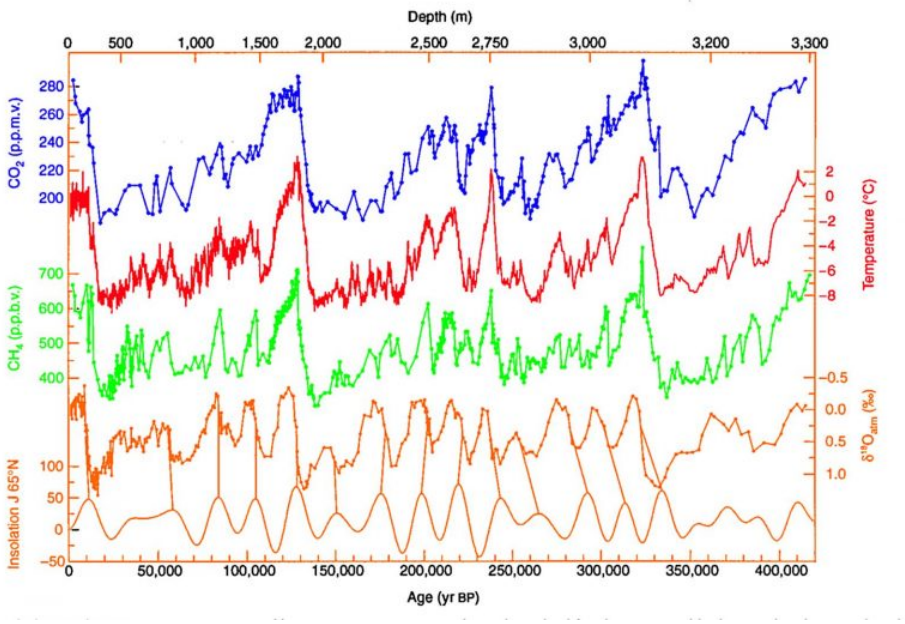


Figure 15.2.9: Insolation, methane, and carbon dioxide data extracted from the Vostok ice core, Antarctica. This ice core records in detail the last four glacial/interglacial cycles of the Pleistocene and the start of the Holocene Epochs. Note the approximately 100,000 year cyclicity that emerges in the data.

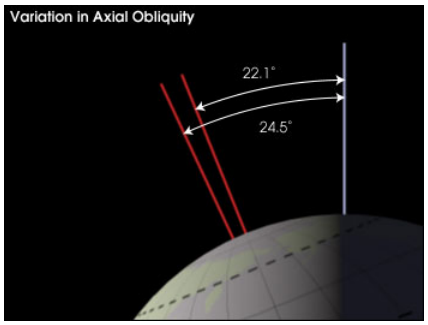


Figure 15.2.10: Changes in Earth's obliquity over ~41,000 years. (Source: NASA)

While the Earth's orbital path is changing, there are also at least two cycles of obliquity that will occur at about 41,000 years each. During one of these cycles, the Earth's tilt ranges from a minimum of 22.1° to a maximum of 24.5°. Currently, our orbital tilt is 23.5°, meaning that we are in the middle of one of these cycles rather than at one end or the other. 10,700 years ago, the planet was last at its maximum tilt. In 9,800 years, it will reach its minimum tilt. As the tilt of the Earth is the main factor that determines the behavior of our seasons, changes in this variable can also have significant effects on global climate. Greater tilt leads to more severe seasons while lesser tilt leads to milder changes between seasons. We think the greatest impact on climate occurs at times of greater tilt, as ice builds up and causes major changes in insolation albedo, leading to an amplifying feedback toward cooler climate. Higher reflectivity of the Earth's surface reflects more of the sun's radiation back to space. When this happens, ice sheets expand. As ice sheets continue to expand, even more sunlight is reflected back to space, cooling the planet.

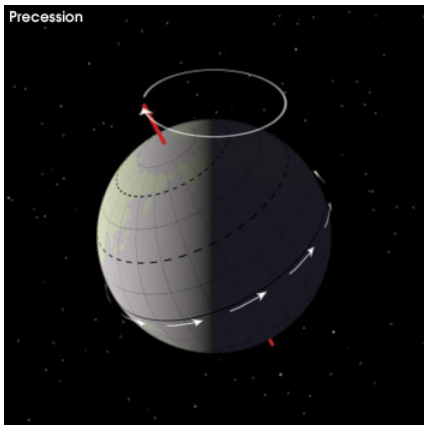


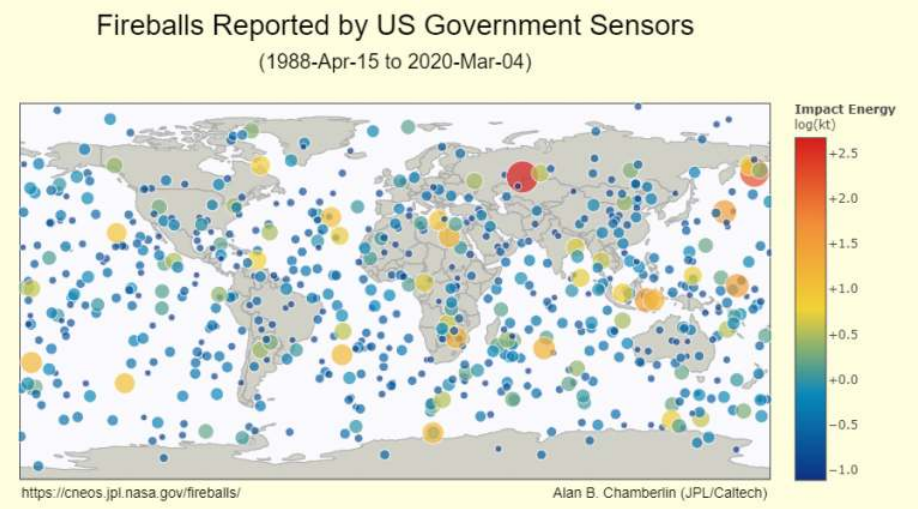
Figure 15.2.11: Changes in Earth's precession occur over the course of about 26,000 years. The Earth's axis "wobbles" like a top. (Source: NASA)

Finally, precession is the shortest of the Milankovitch cycles. Over the course of about 26,000 years, the Earth's axis (the imaginary line defining our geographic north and south poles) wobbles. As the axis does wobbles, it marks out a kind of circle on the celestial sphere over the course of that time. Currently, the axis is pointing toward the star Polaris, the North Star, making it the only star that stays in a "constant" position in the night sky. This is why it can be used for navigation. This has not always been the case over longer periods, however. At times, there is no north star, much like there is currently no "south star." At other times, other stars have that title, as the star Thuban did in 2000 BC when ancient Egyptians might used it to navigate through the desert. In terms of climate, the changes in precession alter the dates for aphelion and perihelion during the year, which affects how seasonal insolation affects the northern and southern hemisphere. As such, the effect of precession in global climate is significant, but more difficult to tease out of data as it is not nearly as powerful as obliquity and eccentricity.

Asteroid Impacts

Asteroid impacts are both exciting and terrifying. We hope that, should we ever be in danger of a massive bolide collision, NASA or some other space agency will see it coming. If they do, perhaps we will get a chance to do something about it. Until then – such imaginings make for excellent disaster B-movie plots!

However, these are very real hazards. Certainly, small impacts like the Tunguska event mentioned earlier are not likely to be seen by any NEA (Near Earth Asteroid) observatory, like the Center for NEO Studies, run by NASA through the Jet Propulsion Laboratory at Caltech. According to their data, there have been 822 recorded fireballs between April 15, 1988 and March 4th, 2020:



Most of these are small and cause only localized disturbances. To better quantify risks associated with impacts, the IAU (International Astronomical Union) in 1999 created the Torino Scale. This metric communicates to the public the hazard level of a particular NEA that is being watched by specialists. Displayed using colors from white (no hazard) to red (certain collisions), it is

simple and conveys important information on the type of resulting hazards associated at that level. Fortunately for Earth, none of the currently monitored NEA objects are listed as greater than 0 on the Torino Scale.

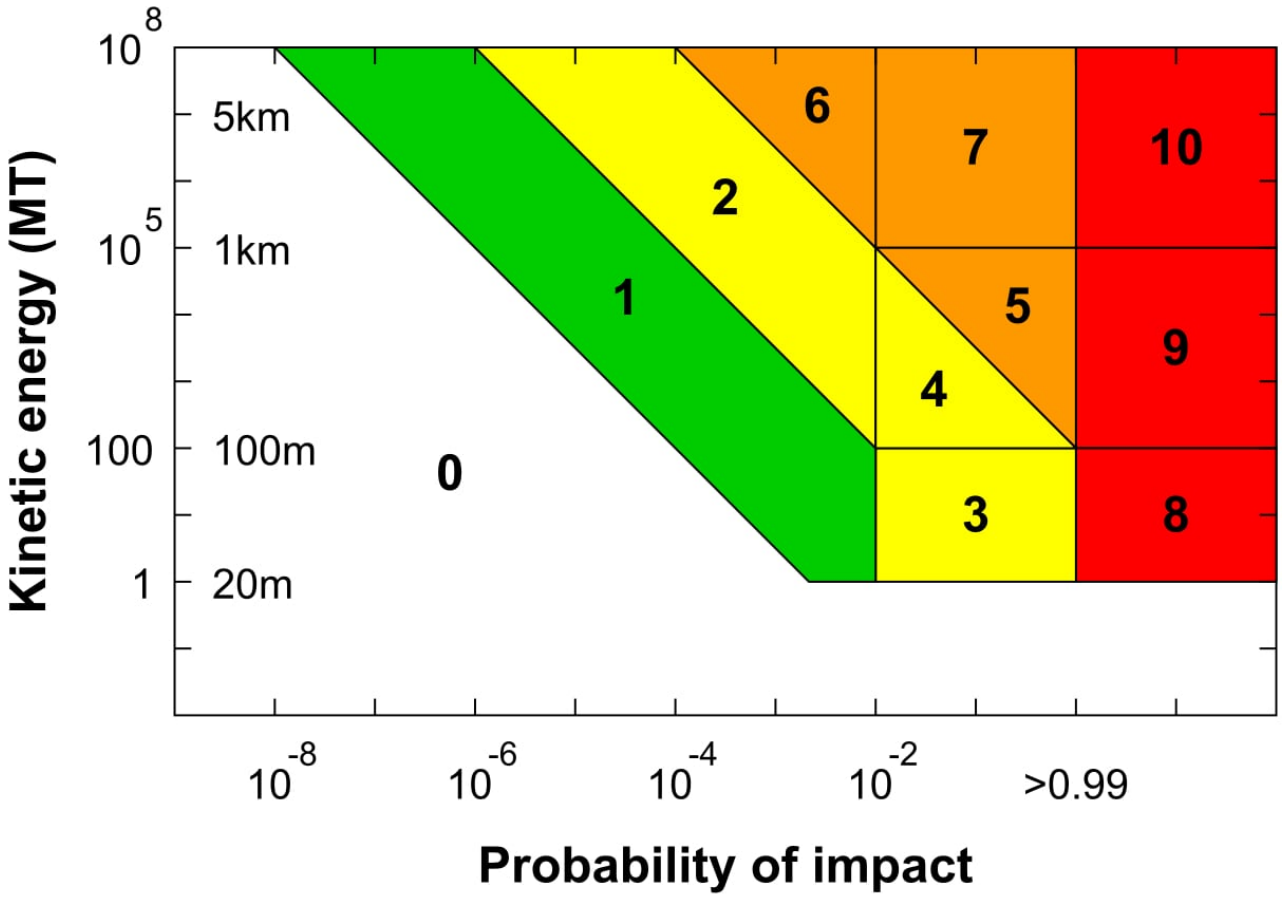


Figure 15.2.12: The Torino Scale measures the severity of an impact hazard based upon two variables, the probability of an impact and the kinetic energy (in Megatons) involved.

Asteroid impacts are important extrinsic climate change forcing mechanisms because of the effect they convey on the atmosphere, hydrosphere, geosphere, and biosphere. Impactors heat the atmosphere on their way in and then, as their ejecta return to Earth, heat the global atmosphere more evenly, possibly to temperatures as high as several thousand degrees. But this is a short-term effect of the impact. Post-impact, dust, ash, and other materials may block incoming solar radiation for many years to come, creating a long-term cooling effect. These are just two of the climate effects. As the oceans acidify from dissolved gases and rock debris, climate-regulating currents may be upended. New tectonic or volcanic activity from the geosphere can also contribute to climatic change.

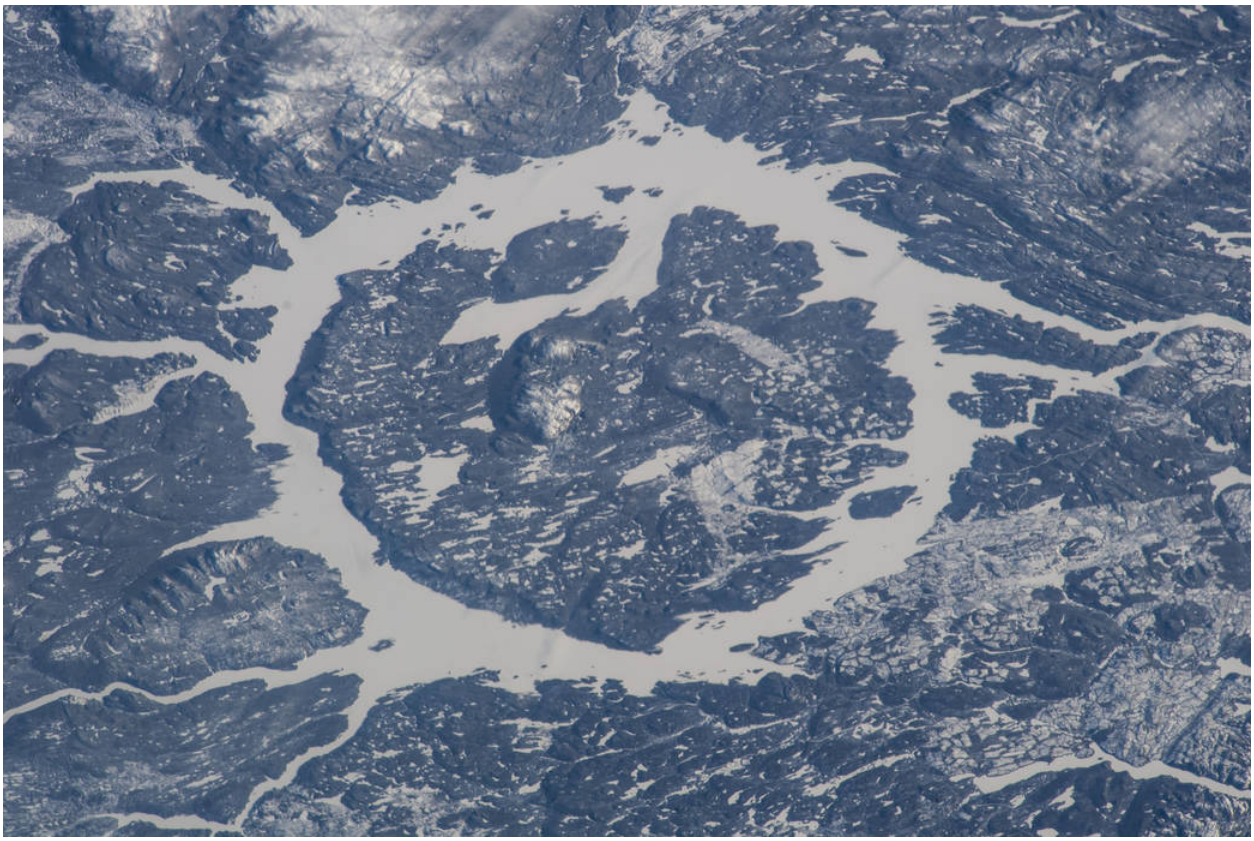


Figure 15.2.13: Manicouagan Crater Reservoir, northern Quebec, Canada. Photographed here by ISS Expedition 38 crewmembers, this crater is so visible from space that it is one of the most photographed. It was created during the Triassic Period, 215.5 Ma. (Source: NASA)

In the time periods where we study paleoclimate, none of these warning agencies existed. In the Earth's history, we know of many large impacts that have occurred, many of which have left lasting scars on the surface. One of the most photographed is of Manicouagan Crater in northern Quebec, which was created by a bolide that hit that part of Canada 215.5 Ma (during the Triassic period). The impactor is estimated to have been about 5 km in diameter (85km diameter crater), very close to the size of the Chicxulub impactor that caused the extinction of the dinosaurs at the K/Pg boundary (6km impactor and 150km crater).

Figure 15.2.14: The Earth Impact Database world map. Note that the colors indicate a rough time period of impact. These are also known impacts. There are very likely many, many more yet to be discovered. Hovering over a circle brings up more information about it.

Changes in solar insolation over time

Over the course of a year, the fraction of the energy Earth receives from the Sun is measured as insolation (the energy received), and varies by about 0.1%. While this fraction is small, it actually represents quite a bit of energy. Changes in short-term solar insolation are of great interest to climate modelers. Such changes result over time spans longer than a year due to the 11-year sunspot cycle, where the Sun's surface experiences a cycle running from cooler to hotter due to magnetic changes. These can affect the planet's climate over these time spans, though the evidence is clear that current climate change cannot be explained by this. To study changes in solar output, NASA in 2003 launched the Solar Radiation and Climate Experiment mission (SORCE).

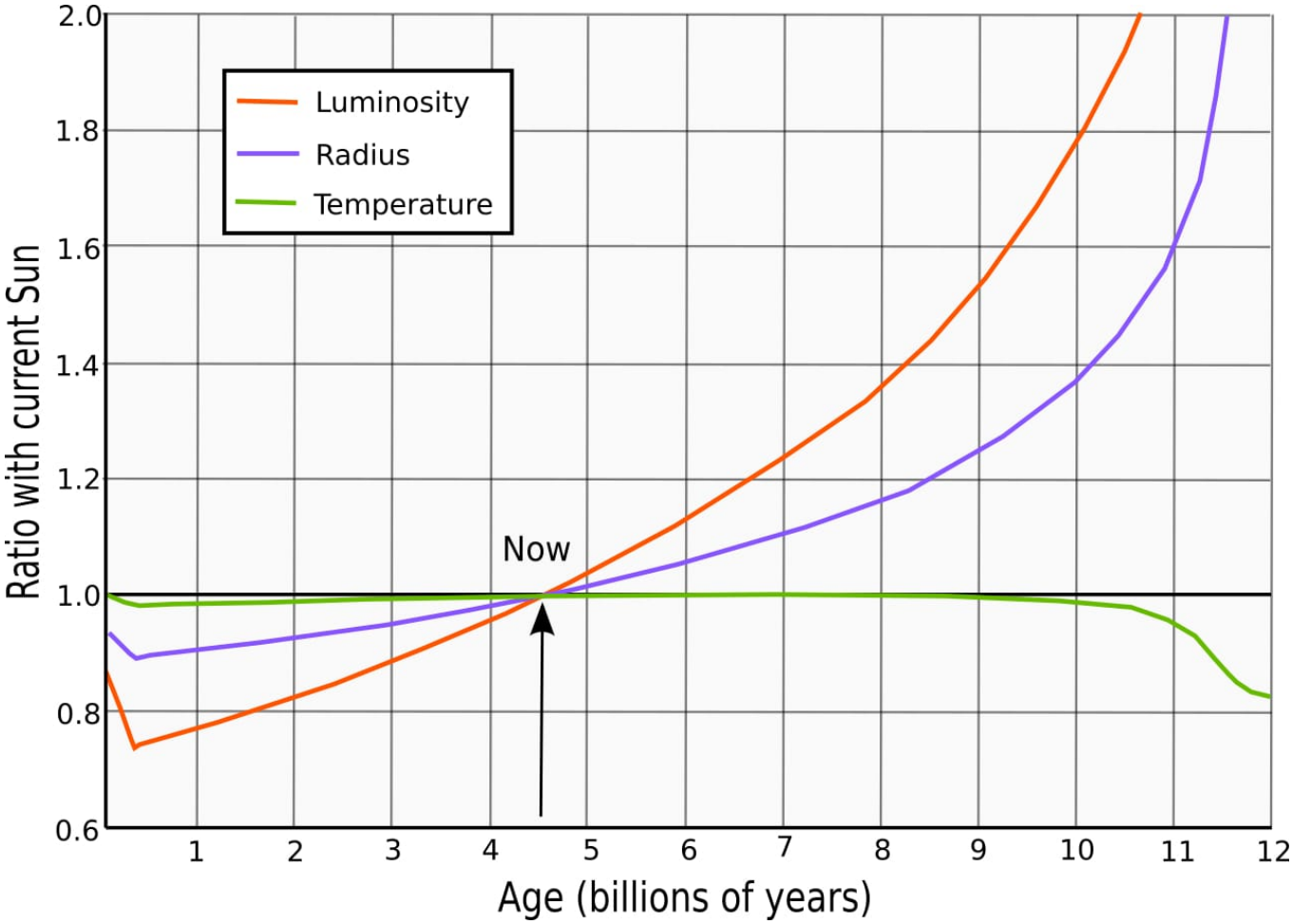


Figure 15.2.15: Variability in the Sun's luminosity (brightness), radius, and temperature over time, looking at past trends and future trends. The Sun will continue to increase its radius and continue to become more luminous over time as it moves toward the red giant phase of its life. Associated temperatures will rise some and affect climate over that time. (CC BY-SA 3.0; By RJHall – Own work, based on figure 1, Ribas, Ignasi (February 2010). "Solar and Stellar Variability: Impact on Earth and Planets, Proceedings of the International Astronomical Union, IAU Symposium". Error: journal= not stated 264: 3–18. DOI:10.1017/S1743921309992298., https://commons.wikimedia.org/w/index.php?curid=16799327)

The Sun itself changes over very long time spans also. Since its early days, it has been steadily warming and will continue to do so. The warming trend is significant, but does pale in comparison to the increases over time in the star's radius and luminosity. These two variables are related, as greater radius leads to greater surface area which leads to greater luminosity. In the figure to the left, it is predicted that the Sun will continue to grow larger. One misleading piece of data here is temperature. While the Sun's temperature may not increase as dramatically as its radius, as that radius increases, the surface of the Sun and its very hot atmosphere will inch ever closer to Earth. This effect on Earth will be very significant, very likely making within a billion years or less most life on Earth as we know it no longer possible. Ultimately, whether at this point the Earth is pushed outward be the Sun's lower mass or pulled inward by drag and enveloped is a matter of some lively scientific debate.

Lunar effects (stabilization, etc.)

What about forcing factors outside of the Sun and bolide impacts? Certainly, while there may be some climatic effects caused by phenomena such as supernova explosions and the passing of our solar system through dense interstellar clouds, these effects are tiny, if they exist.

What about the Moon?

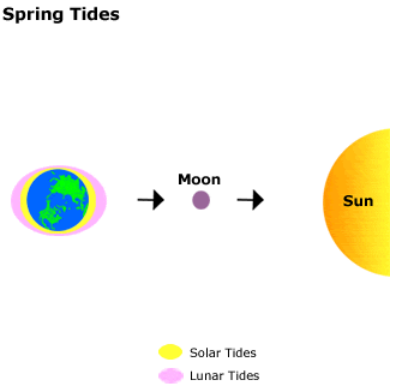


Figure 15.2.16: Animation of the lunar tidal effect. The combined pull of the Moon and Sun (the Moon's effect is much greater) during New and Full lunar phases create the largest tidal bulges and ranges and are called spring tides. (Source: NOAA)

The Earth's natural satellite (as that is really the proper term to use), the Moon, has two demonstrable effects, though small, on our climate. Both of these effects are related to the tidal gravitational forces that the Moon exerts on the Earth. When we think of tidal forces, we usually think of shorelines. This is correct, of course, but the effect on the shorelines is really a gravitational effect, as the Moon is pulling and releasing the water in the ocean as the Earth rotates and as a 29.5 day lunar cycle passes. As you may have learned, there are one to two diurnal high and low tides per day. During a lunar month, there are also times of greatest tidal range that occur during the New and Full phases of the Moon, the spring tides. The tidal range is the difference between the elevation of high and low tide. There are also what are called neap tides, when the tidal range is lowest during a lunar cycle, during the 1st and 3rd quarter phases (half moons).

These tidal effects actually affect climate on a small scale, by affecting the amount of precipitation that occurs. It does this not by pulling on Earth's water, however, but by exerting its gravitational pull on the atmosphere. When the Moon is overhead, a full moon in particular, air pressure is actually higher. This slightly depresses the chance for precipitation. The effect is tiny, but it does exist.

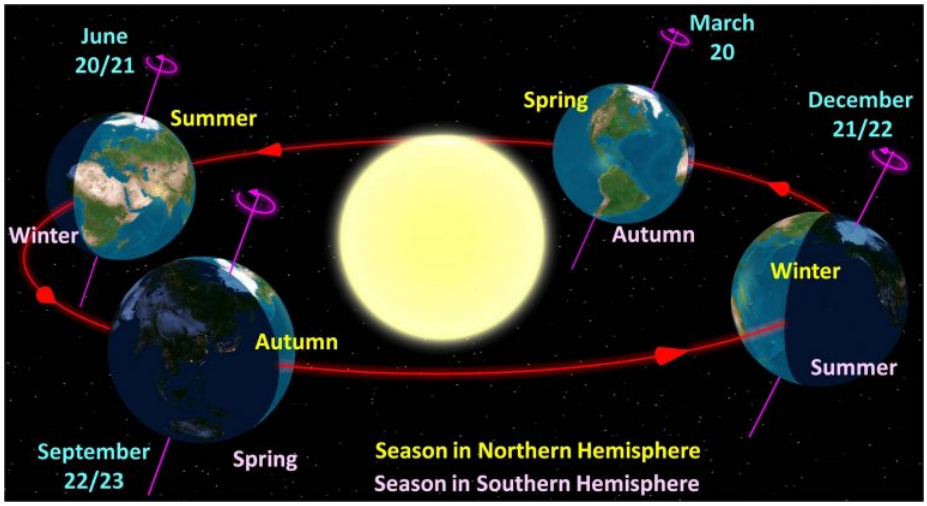


Figure 15.2.17: The Distribution of Earth's seasons relative to the Sun. The tilt of the Earth and its stability are what allow for the predictable cycle of seasonality we experience in temperate latitudes. At the time of the autumnal or vernal (spring) equinoxes, the equator faces most directly toward the Sun. At the time of summer solstice, the northern hemisphere experiences summer and at the time of winter solistice, the southern hemisphere is experiencing summer.

Perhaps a larger lunar impact is related to the Earth's obliquity cycle. Joutel and Robutel (1993) were able to demonstrate the Moon's gravity stabilizes our planet's tilt. In essence, it keeps our obliquity cycle stable, which prevents much more massive long term climatic changes. This also helps stabilize our seasons, which the Earth's tilt is critical in defining. In times when the tilt angle could be nearly 0° or much greater than 24.5° maximum our planet currently has, the seasons would be very different. In a situation where there is zero to little tilt angle, there would be no seasonality on the planet, or very little. In a situation where the tilt might

reach greater than 30°, the difference between summer and winter would be quite harsh in temperate latitudes and nearly impossible to bear in polar regions. As such, our Moon works to stabilize our climate.

The leading idea on the formation of the Moon is that it resulted from a massive impact, probably of an object about the size of Mars. In its early years then, the Moon was much, much closer to the Earth. Over time, it has gradually been moving farther and farther away. However, this also means that its gravitational tidal forces have also been lessening over time. One notable effect of this is the gradual change in the length of a day on Earth and, consequently, the number of days in a solar year. This is born out by fossil mollusks and corals from 80 Ma (and other time periods) that show that a year was 372 days at the time.

There are also potentially effects over time on other orbital parameters that affect climate. The affect of our natural satellite on our climate is still an active and exciting area of research.

INTRINSIC CHANGE AGENTS

Within the Earth system, there are a wide variety of factors that can affect the Earth's climate. These intrinsic factors all have one thing in common. In one way or another, they disrupt, or place in a state of disequilibrium, one or more biogeochemical cycles. There are many biogeochemical cycles and all of them interact with more than one of the Earth's subsystems, or spheres. Examples include the carbon cycle, nitrogen cycle, hydrologic cycle, phosphorus cycle, etc.

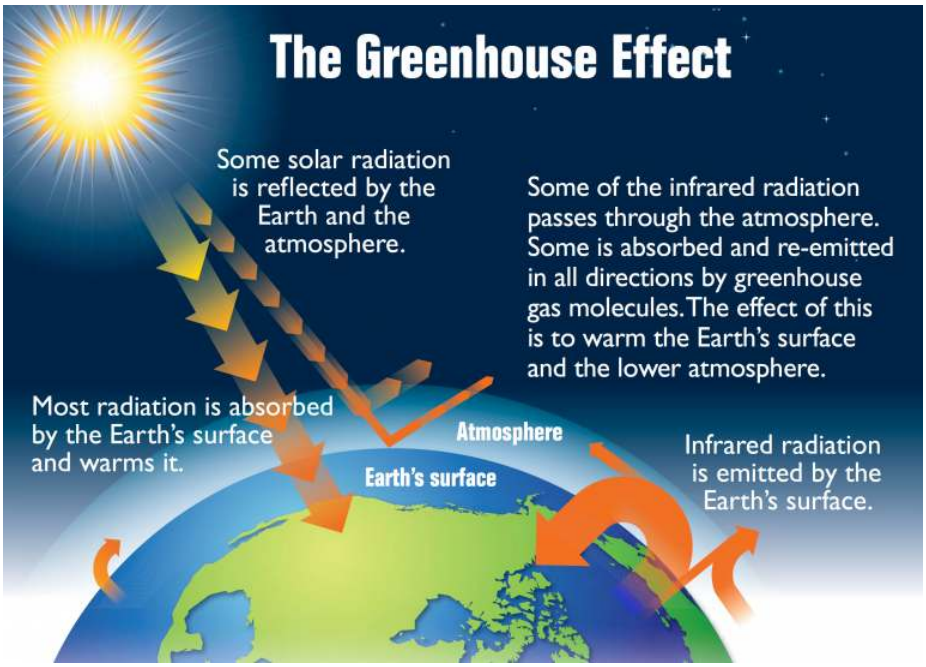


Figure 15.2.18: Earth's greenhouse effect (Source: EPA).

The Earth's climate is regulated by the greenhouse effect, or the retention of re-radiated longwave radiation through the absorption of this energy by greenhouse gases in our atmosphere. Earth is not the only planet that has a greenhouse effect. Venus has this also, in spades. It is what leads to surface temperatures on Venus exceeding 400 °C (~800 °F) in some instances. Mars has very little atmosphere and though what it does have it almost entirely carbon dioxide, it cannot retain much of the Sun's energy. Mars lacks a magnetic field and thus cannot hold onto its atmosphere. One other location in our solar system, Titan, a very large and gassy natural satellite of Saturn, has a moderate greenhouse effect, but due to the high concentration of methane in its atmosphere.

On Earth, the greenhouse effect is moderate, but absolutely critical for life. This is due in part to its distance from the Sun, but also because of its very different mix of atmospheric gases due to the presence of life. Photosynthesis has altered our atmosphere so that it contains 21% oxygen on average. 80% of the gas in our atmosphere is nitrogen, which also has biogenic sources. Life on Earth has, thus, reduced the carbon dioxide so prevalent in the atmospheres of Venus and Mars (from volcanism) to a mere 0.04% of the total gas content. As of this writing today, this amounts to a carbon dioxide concentration of about 416ppm. Common greenhouse gases include carbon dioxide, methane (another carbon compound), oxides of nitrogen (NO_x), HCFCs (anthropogenic chlorofluorocarbons), and water vapor.

Ultimately, all of the elements that make up compounds in our atmosphere share a primordial source from the Earth's interior. Biogeochemical cycling, aided by plate tectonics, makes these elements accessible. Biogeochemical cycles are cycles driven not only by geological processes (geo), but also biological processes (bio). As such, Earth's versions of these cycles are unique.

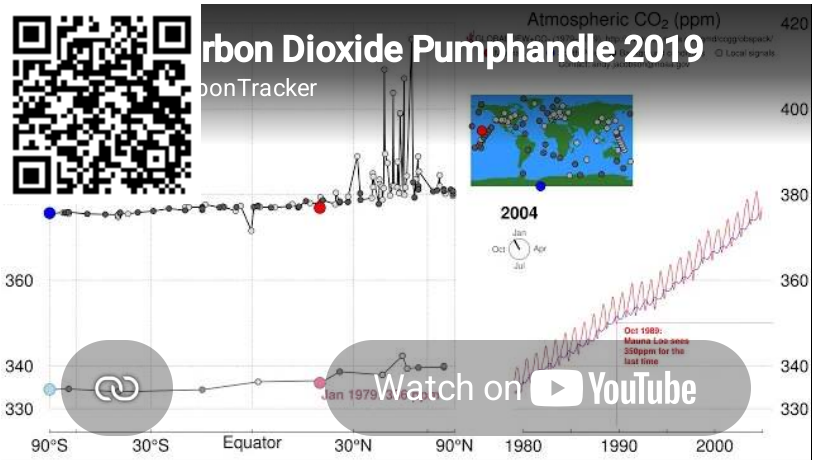
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15.3: Geological Sources of Important Elements for Life

Table 15.3.1: Table of important elements for life. All of these elements are ultimately sourced from within the Earth's interior. Plate tectonic processes, driven by Earth's interior heat, drive one end of the biogeochemical cycling of these nutrients while the Sun's incoming energy drives them from above.

Elemental Nutrient	Primordial Source on Earth	Important Life Processes
Carbon	Cosmogenetic, Mantle, Volcanism	Basic element of organic chemistry, important nutrient for photosynthesis of glucose
Hydrogen	Cosmogenetic, Mantle, Volcanism	Basic element of organic chemistry, important nutrient for photosynthesis of glucose
Nitrogen	Cosmogenetic ammonia, Earth's atmosphere	Important for the creation of amino acids, enzymes
Oxygen	Cosmogenetic, Volcanism, Carbon Dioxide dissociation, Photodissociation in Atmosphere of Water	By-product of photosynthesis, Necessary for respiration and associated oxidation
Phosphorus	Phosphides in Earth's core, Inorganic minerals such as apatite and fluorite, ocean-floor sediments	"Spine" of DNA molecule, important micronutrient that helps store energy in cells via ATP
Sulfur	Cosmogenetic, Stored in Earth's interior, Volcanism	Allows for the synthesis of a greater variety of amino acids, Important nutrient for chemosynthesis

Changes in greenhouse gas concentrations, particularly carbon dioxide, have led to changes in the Earth's climate today, but also in the geologic past. Today, we monitor carbon dioxide levels at NOAA's Global Monitoring Laboratory atop Mauna Loa, an extinct volcano in Hawaii. You can view carbon dioxide concentrations as measured instrumentally and then compared to data taken from ice cores from Antarctica (a source of proxy data for carbon dioxide) in the video below (Retrieved May, 2020):



Milkankovitch Forcing, Ice Cores, and Carbon Dioxide - Quiz

? Exercise 15.3.1

At 1:27 in the above video, the ice core record going back 800,000 years begins. As you watch, you will begin to see a pattern emerge in the carbon dioxide data. Compare this with the above section of the text on Milankovitch Cycles (Extrinsic factors). Which cycle do you see controlling the very large up and down swings in carbon dioxide?

a. Eccentricity, about 100,000 years

b. Annual, about 1 year

c. Precession, about 26,000 years

d. Obliquity, about 41,000 years

Answer

a. Eccentricity, about 100,000 years

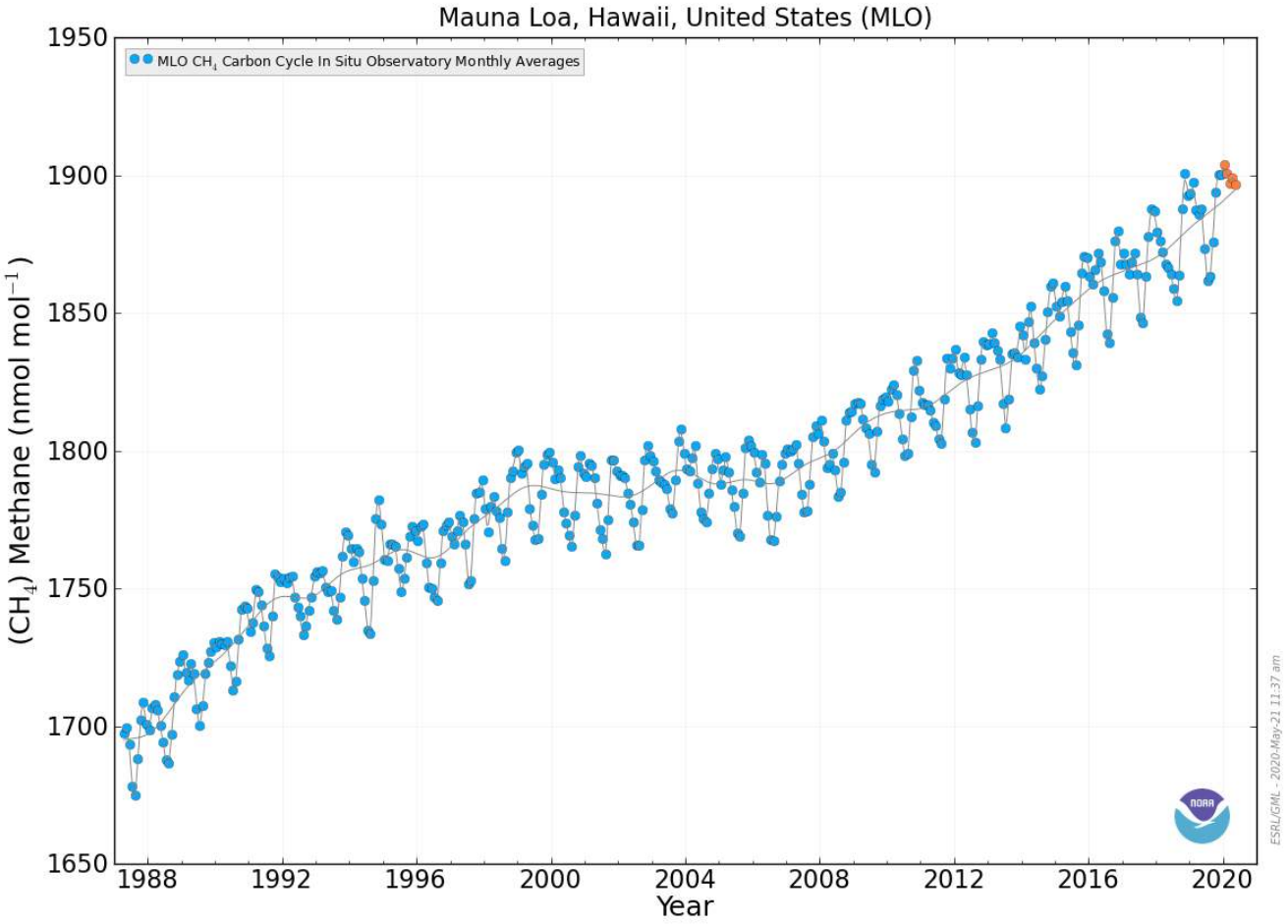


Figure 15.3.1: Atmospheric methane concentrations since 1988 (Source: NOAA Global Monitoring Laboratory, May 2020).

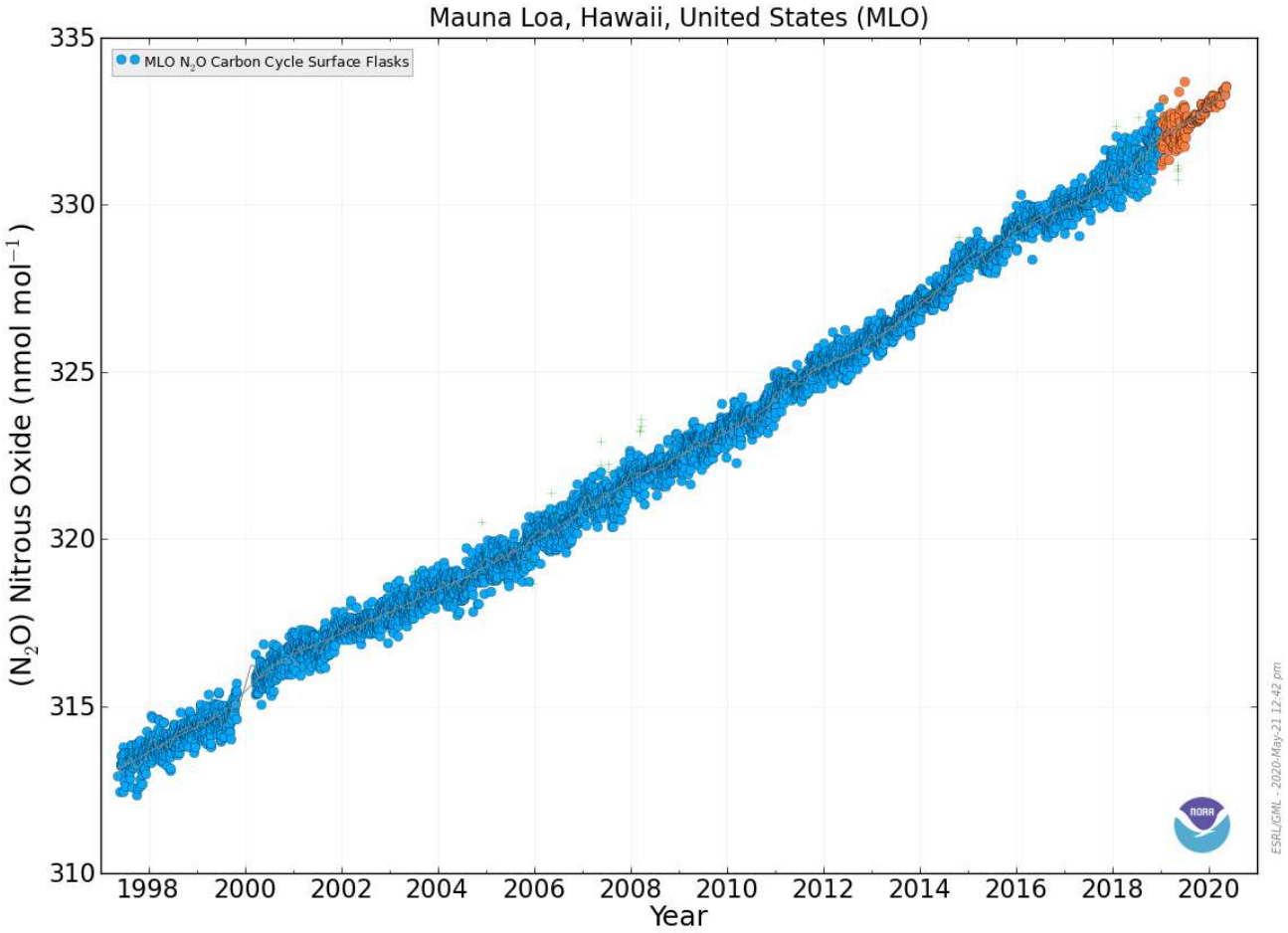


Figure 15.3.2: Atmosphere nitrous oxide concentrations, Mauna Loa, Hawaii, May 2020 (Source: NOAA Global Monitoring Laboratory).

Similarly, methane, another greenhouse gas, has seen a rapid increase in atmospheric concentration due to human activities since monitoring began. A nearly identical trend to carbon dioxide and methane can be seen in nitrous oxide gases. While nitrous oxide gases are also a major component of the nitrogen cycle, they are intimately intertwined within the carbon cycle because so many of the sources of carbon and nitrous oxides are the same processes. These include biomass burning, fossil fuel combustion, etc.

Carbon cycle gases are the real keys to understanding our changes in climate today. Human activities from the burning of fossil fuels to the production of fossil fuels, to deforestation, and other perturbations to the carbon cycle that lead to changes in carbon dioxide and methane concentrations are having a profound effect on our modern climate. The image below depicts natural and anthropogenic fluxes in the carbon cycle and their magnitude for a ten year period from 2000-2009. Note the magnitude of anthropogenic fluxes (red arrows and numbers).

Knowing the source of anthropogenic greenhouse gas emissions is one thing, pinning the increase and resultant climate effects is another. Apart from our ability to use carbon isotopic ratios to identify greenhouse gases as the course of new atmospheric carbon dioxide, stations monitoring atmospheric oxygen levels are seeing decreases across the globe as atmospheric oxygen is used in the combustion process of fossil fuels. Cape Grim, Tasmania, provides an excellent example.

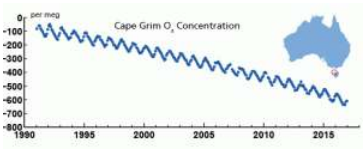


Figure 15.3.3: Declining atmospheric oxygen levels accompany increasing carbon dioxide levels due to combustion of greenhouse gases. While these decreases in oxygen are not enough to affect human health, they provide stark evidence for the effect of the burning of fossil fuels on the Earth's changing climate (Source: Scripps O_2 Program).

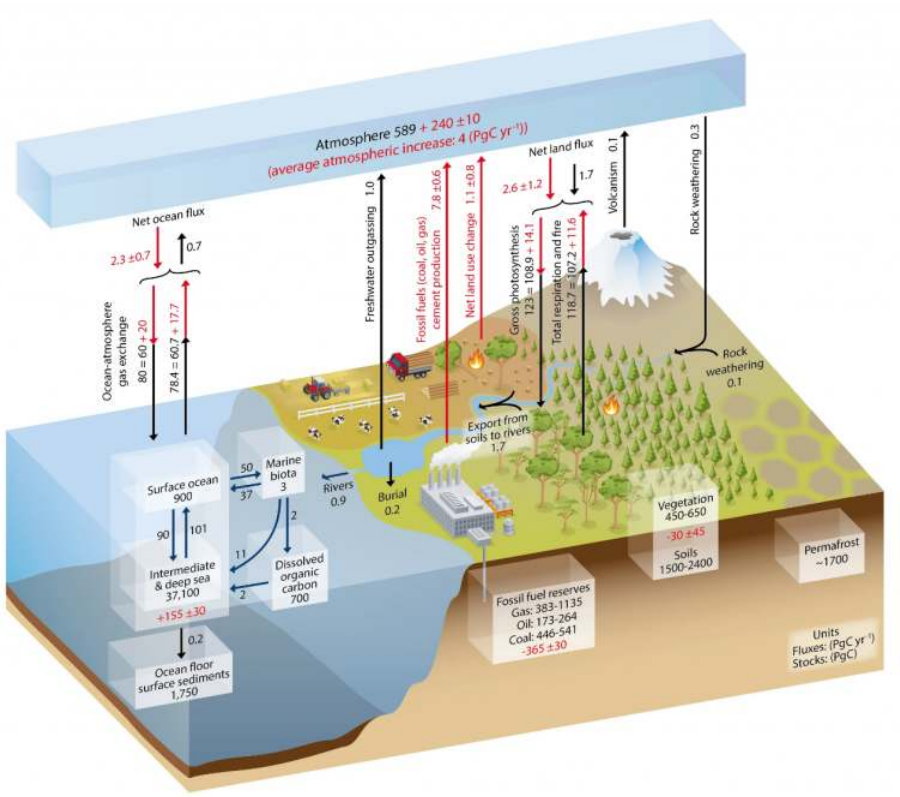


Figure 15.3.4: The carbon cycle, including anthropogenic inputs. Pg stands for petagrams of carbon per year (same as gigatonnes). Numbers in boxes represent the total mass of carbon in the reservoir (stock). Numbers next to arrows represent fluxes, or changes in carbon concentrations from 2000-2009. Black numbers are natural fluxes. Red numbers are anthropogenic fluxes. Both carbon dioxide and methane, among other carbon compounds, are included (Source: IPCC 5th Assessment Report).

The challenge of maintaining balance within the current carbon cycle and resulting climate is really no different today than in the past. The main difference today is the speed of changes. In Earth's past, changes in climate and the carbon cycle were more often much slower. To illustrate this, it is useful to think of the carbon cycle as having a slow component and a fast component, as illustrated below.

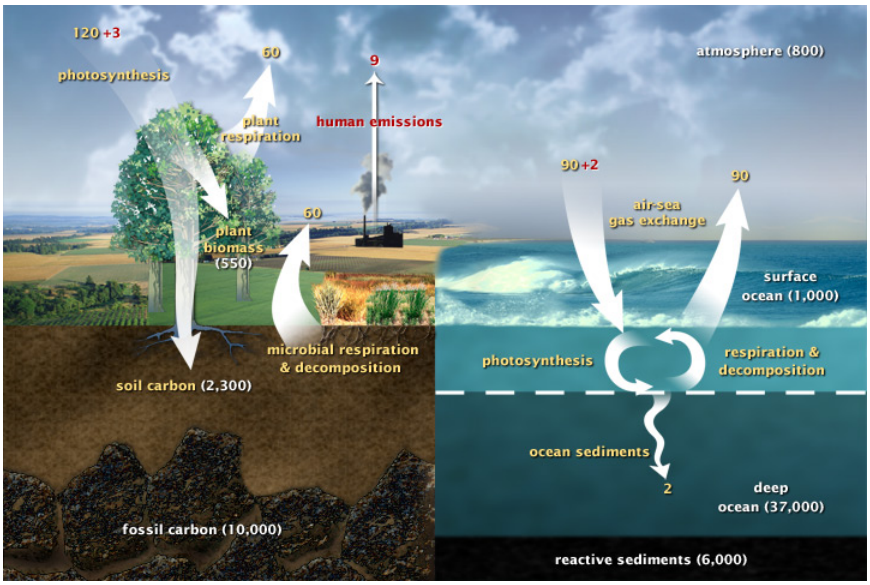


Figure 15.3.5: The carbon cycle, broken up into its two parts, the fast and slow carbon cycles. The fast carbon cycle moves carbon through the Earth's systems across short time periods, such as through photosynthesis and decomposition. The slow carbon cycle sees fluxes in carbon storage across much longer time spans, such as storage of carbon in limestones and fossil fuel deposits. Numbers indicate flux magnitudes and are unitless (Source: NASA Earth Observatory).

Today, human activity is pulling carbon out of reservoirs (storage) that have taken thousands to millions of years to accumulate (slow carbon cycle) and then adding these stored carbon to the fast carbon cycle. The speed at which the magnitude of these anthropogenic fluxes are changing is leading to rapid change in the Earth system, leading to a state of disequilibrium. At some point, tipping points will be reached. Such tipping points include releases of methane from Arctic permafrost, ocean acidification with the potential to trigger a marine mass extinction, *etc*.

Similar changes to the carbon cycle, or other biogeochemical cycles, also occurred in the past. Below, a few forcing mechanisms will be discussed that have led to changes in these systems, leading to large climate change, mass extinction, and other major geologic events.

Flood basalt eruptions

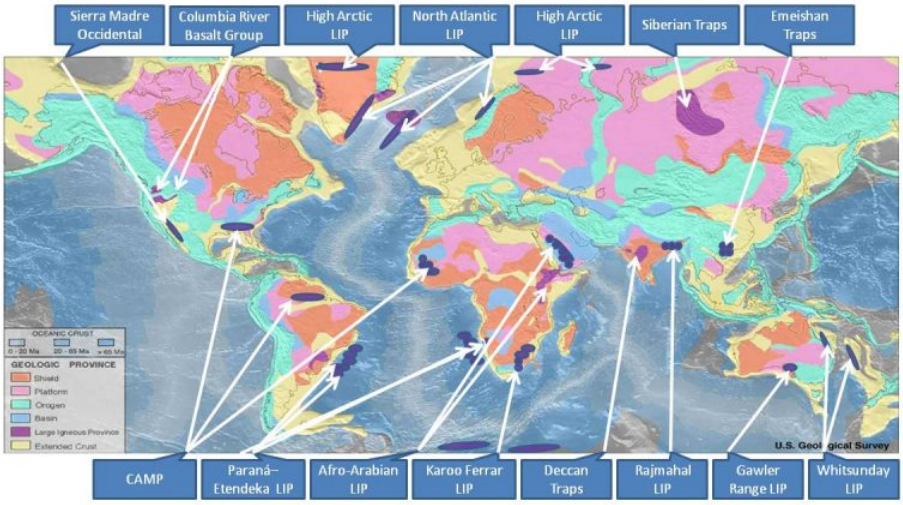


Figure 15.3.6: Large Igneous Provinces (LIPs), also known as flood basalt provinces, exist in a wide variety of locations all over the planet (Source: Williamborg after USGS via Wikimedia)

At various times throughout Earth's past, there have been major eruptions that have lasted for extended periods of time and that have affected large geographic areas. Called Large Igneous Provinces, or LIPs, these massive volcanic eruptions had a very unique

character, given the massive volumes of erupted lava, the very extensive nature of them, and their long duration.

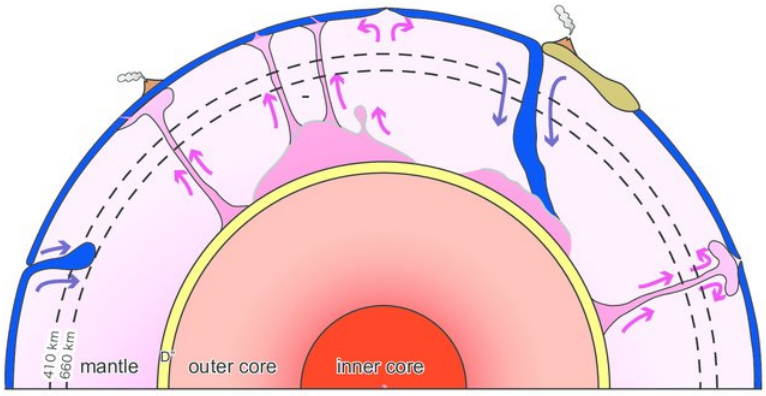


Figure 15.3.7: The fate of subducting lithosphere (blue) and the rising of hot spots and plumes that create LIP provinces. Transient LIPs, the kind that exist and then go away, likely originate from the core/mantle boundary from LLSPVs (Large Low Shear Velocity Provinces) (Used with permission: Coffin et al., 2006).

All of them share similar environmental effects also, most of which are related to climate. Some LIP events exhibited greater effects in one area than others. These include 1) global warming, 2) oceanic anoxia, 3) release of methane from gas hydrates, 4) oceanic calcification crises, 5) a period of global cooling, and 6) a significant extinction event. LIP eruptions may result from the pooling of upwelling plumes of warm rock from deep within the mantle, likely from the core/mantle boundary, sourced from large low shear velocity provinces (LLSPVs) that are probably partially molten.

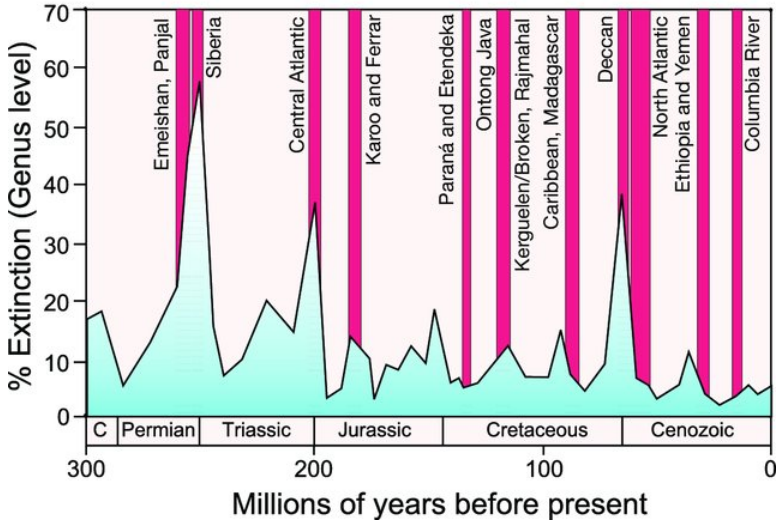


Figure 15.3.8: LIP eruptions superimposed upon Sepkoski (1996) extinction of marine genera over time. There is a close or direct correlation between LIP eruptions and a number of extinction events in Earth's history. Global climate changes are the main connection (Used with permission: Coffin et al., 2006).

The extinction connection is very significant. Within the last 258 million years, there have been five major LIP eruptive events. These include the Emeishan LIP (258 Ma) that correlates to the end-Guadalupian extinction within the Permian Period, the Siberian Traps (250 Ma) that likely caused the end-Permian "Great Dying" extinction, the Central Atlantic Magmatic Province (CAMP, 200 Ma) that likely caused the Triassic mass extinction, the Karoo-Ferrar Traps (180 Ma) that caused what is called either the Toarcian Turnover or the Toarcian Extinction during the Jurassic Period, and the Deccan Traps (65 Ma) that coincide with the extinction of the dinosaurs and the end of the Mesozoic Era.

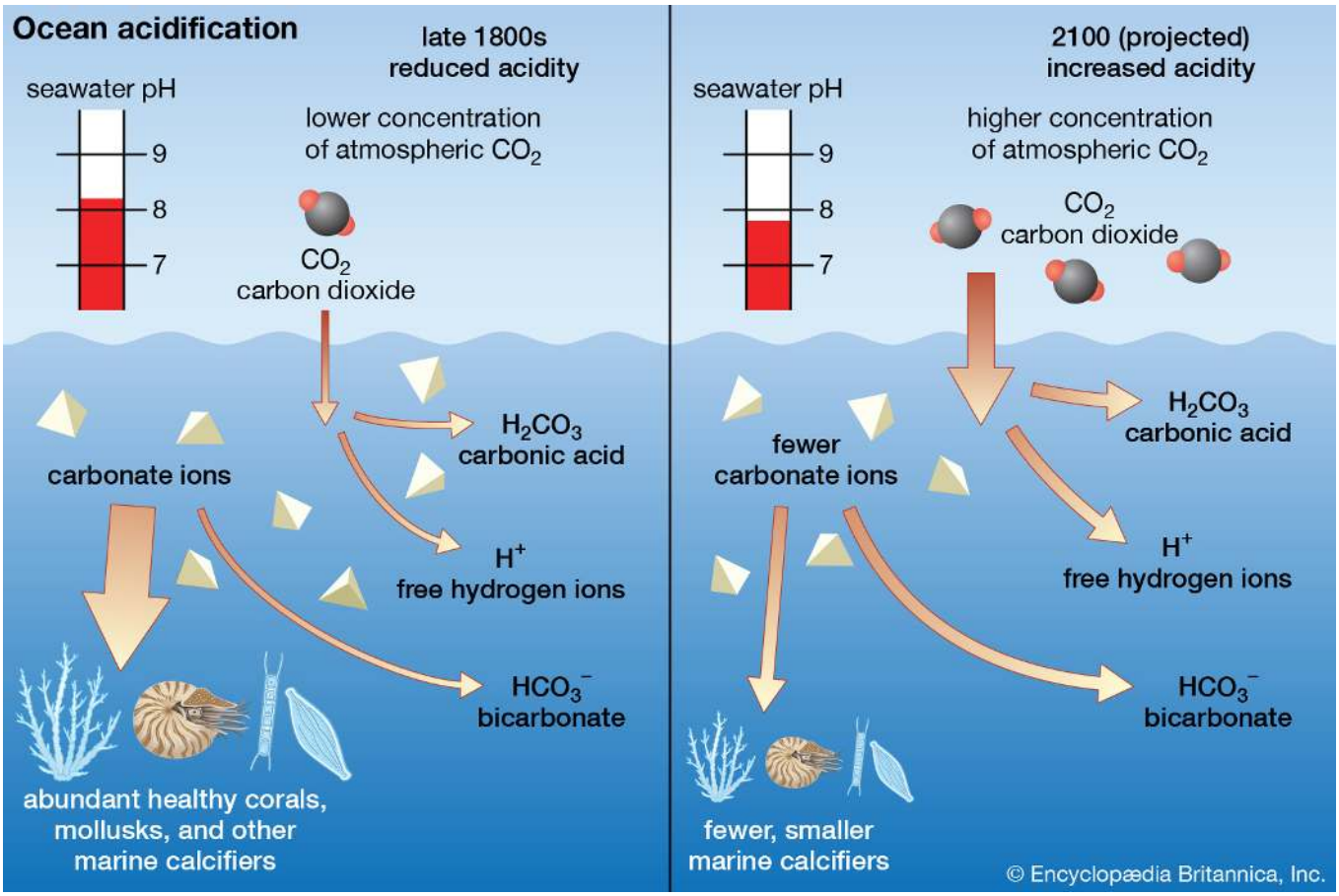


Figure 15.3.9: One effect of increasing carbon dioxide levels in the atmosphere is direct increase of carbon dioxide levels in the marine realm, as carbon dioxide emitted is concentrated first in the hydrosphere. Adding carbon dioxide to the ocean leads to a lowering of pH and dissociation of Ca^+ and bicarbonate, making shell production more difficult to impossible, in extreme scenarios (Source: Brittanica Online)

Such extinctions likely begin with the massive releases of carbon dioxide from the eruptions. These releases warm the atmosphere through the greenhouse effect, as is being done today by humans (though at a faster rate). Following this, the ocean waters warm also. Warmer waters hold lower concentrations of dissolved oxygen and, because the warmer air disrupts the ocean circulation system (at times disrupted also because of the positions of the continents), the water is able to hold very little oxygen. This is the source of oceanic anoxia.



Figure 15.3.10: Burning ice – methane hydrates (Source: USGS).

Also like today, the increased carbon dioxide in the atmosphere translates to increased carbon dioxide in the oceans, leading to the "calcification crises" due to ocean acidification. Animals that secrete calcite are competing with dissolved carbon dioxide for the available bicarbonate. Warming waters also releases accumulations of methane gas hydrates, which are common in offshore environments. This added methane acts as an amplifying feedback to warming, ramping up temperatures even more, thus amplifying also the warming of ocean waters and so on.

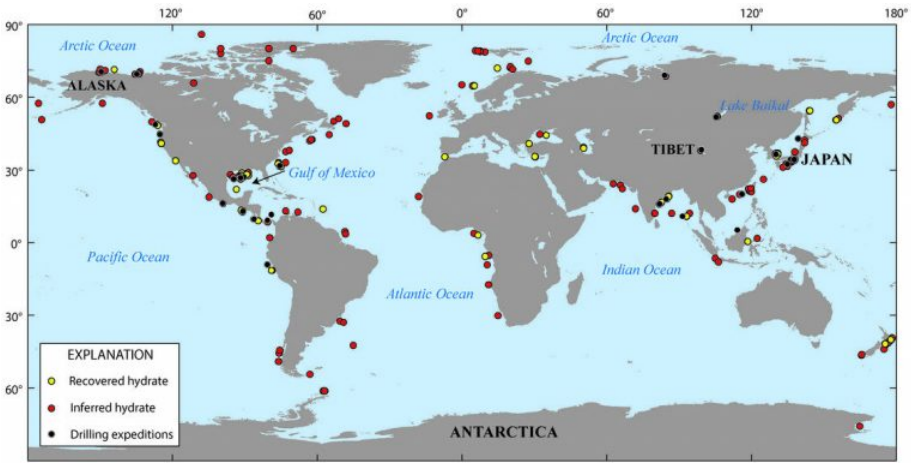


Figure 15.3.11: Locations of methane gas hydrates, recovered and inferred, today (Source: USGS).

The Biosphere

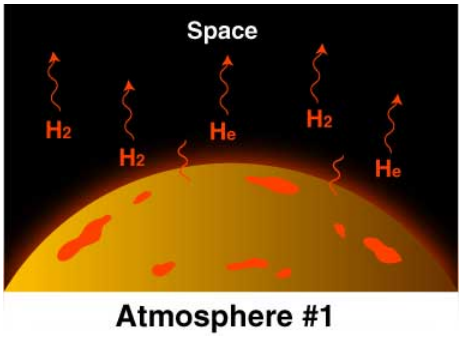


Figure 15.3.12: Earth's earliest atmosphere was made up of gases outgassing from the planet's early formation (Source: NOAA).

Like other terrestrial planets such as Venus and Mars, carbon dioxide was a very prevalent gas in our planet's early atmosphere. Over time, this changed. Earth's earliest atmosphere was composed of gases such as helium that did not stay put. Rather, Earth's mass is not great enough to retain helium and so it floated off into space, as it does today when you release it from a balloon. The planet's magnetic field had also not developed yet, so there was no way to prevent the solar wind from blowing away whatever atmosphere existed. The fact that Earth's liquid outer core is so substantial (and fluid enough to convect, generating a magnetic geodynamo) is surely one of the planet's "tricks for success" that have led to its stability over the geological long term.

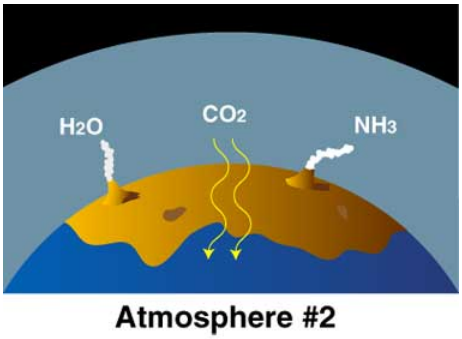


Figure 15.3.13: Earth's second atmosphere was primarily volcanic. By this time, the internal dynamo was creating the magnetic field that prevented the solar wind from flowing this atmosphere away. About 4.2 Ga (Source: NOAA).

Eventually, the formation of a magnetic field and lithosphere led to an atmosphere composed of accumulated volcanic gases, notably carbon dioxide. While it is hard to imagine what the climate of this early Earth was like, it was likely very hot. Life began, and started to evolve. It is likely that the first microbial organisms got their energy chemosynthetically, through the breakdown of compounds like hydrogen sulfide, which is common at deep sea volcanic vents. Genetic evidence also suggests that both Bacteria and Archaea had their origins in a very hot environment: the least derived members of both groups are hyperthermophilic extremophiles. In our solar system, the closest we get to examples of such an atmosphere are on Venus and Mars.

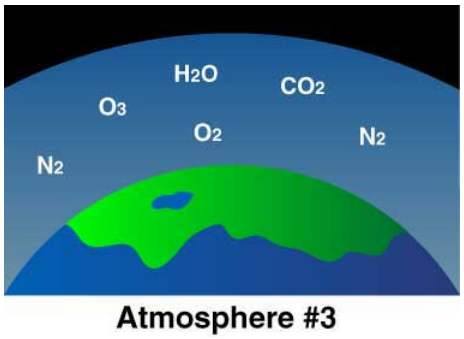


Figure 15.3.14: Earth's third atmosphere becomes our modern atmosphere. It is dominated by gases resulting from plant and animal respiration, but mostly planet photosynthesis. Begins to form about 3.8 Ga (Source: NOAA).

Earth's final atmosphere, the one we have today, began to form around 3.8 Ga with the innovation and evolution of photosynthesis. Most free oxygen produced through this process was absorbed and retained by the oceans, which kept atmospheric levels low. However, once saturated, oxygen was able to effervesce (exsolve) into the atmosphere and accumulate there. The presence of this oxygen would eventually lead to massive cooling events, as the prior carbon dioxide-rich equilibrium was upset. Several "Snowball Earth" events would follow throughout the remaining Precambrian time.

This last atmosphere is unique to Earth. No other planet in our solar system has an atmosphere like our planet. The biosphere's ability to alter its environment led to not just local changes, but global change. As evidenced by climate change today, we know this is still possible, even by a single species. Massive releases of a photosynthetic waste product, oxygen, would force major changes in the hydrosphere, followed by the atmosphere. Eventually, these would also be reflected in the geosphere as banded iron formations formed in the oceans and later by massive iron-rich red beds in continental environments. The iron so prevalent in the early Earth's geology was oxidized (chemically weathered) by all of this free oxygen.

Silicate Weathering

As mountains rise, they are also weathered and eroded. At different periods in Earth's history, there was more orogenic activity (mountain-building) going on than at other times. Orogenies occur at plate boundaries, the largest of which occur at convergent plate boundaries. As two continents plow into one another, as exemplified by India converging with Asia about 45 Ma, they form large mountain chains along massive collisional belts. In the case of the Himalayas, the largest mountains above sea level exist there rise to above 8,800m (~29,000 ft) of elevation, nearly to the stratosphere.

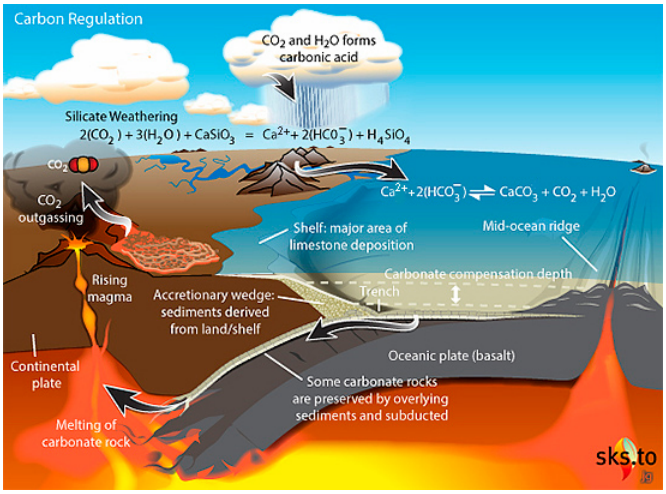


Figure 15.3.15: The silicate weathering cycle.

As mountains rise up, they are subject to more and more chemical weathering. This is because the troposphere's carbon dioxide combines with the clouds and precipitation occurring as a part of the hydrologic cycle, forming a weak carbonic acid. This acid over time reacts with the feldspar minerals so common in igneous and metamorphic rocks that form the cores of such mountain ranges, allowing them to be weathered downward. Mundane as it seems, this process not only leads to salty oceans, produces a

great deal of free bicarbonate and calcite for shell-making organisms, but also pulls carbon dioxide out of the atmosphere, leading to a cooler climate. As you can see in the image above, there is a chemical equation that describes this process:

$$\mathbf{2}(CO_2) + \mathbf{3}(H_2O) + CaSiO_3 \rightarrow Ca^{2+} + \mathbf{2}(HCO_3^-) + H_4SiO_4$$

Thus, the rise of mountains leads to global cooling. It also leads to the expansion of carbonate environments.

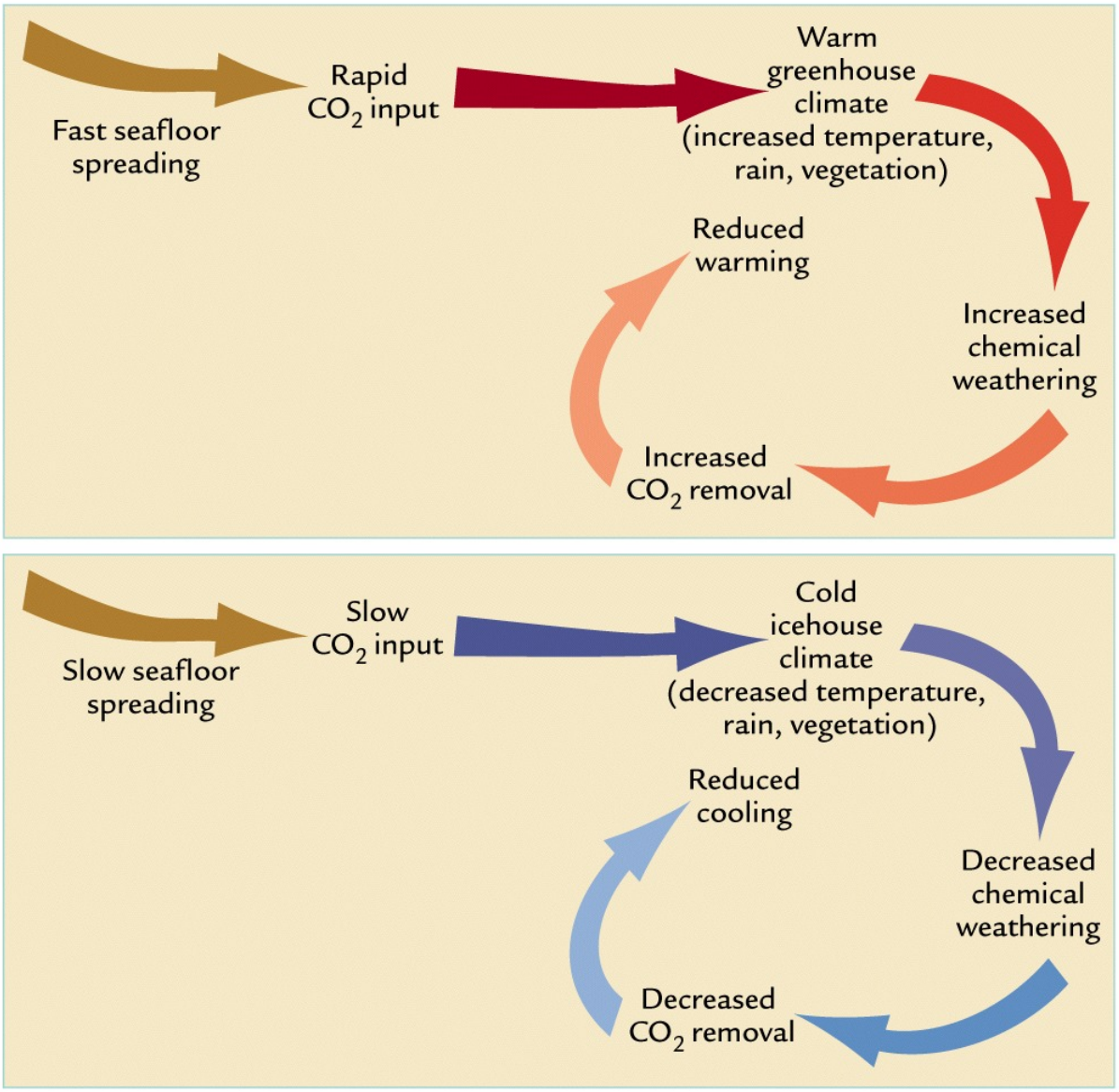


Figure 15.3.16: Tectonics drives climate on the scale of millions of years of change. A rise in atmospheric carbon dioxide can lead to a warmer climate. When coupled with rising mountains, increased carbon dioxide leads to increased chemical weathering, carbon dioxide removal, and then reduced warming. (Source: Public Domain)

An excellent example of the power of silicate weathering lies in its role in moderating the climate system. If tectonics tunes the climate over millions of years, weathering of silicates is a key process that serves to reduce or amplify warming or cooling, depending upon the initial inputs. In the image of feedback loops to the left, times of fast seafloor spreading (that also drive subduction and orogeny) lead to carbon dioxide input from volcanism which warms the climate. As resulting mountains intrude into the atmosphere, chemical weathering increases, carbon dioxide is removed, and warming is reduced. Likewise, when seafloor spreading slows, the reverse series of processes occurs, eventually moderates the resulting cooling trend.

By the end of the Pliocene Epoch, the initial conditions were very much like the top loop. There was a great deal of global tectonic activity and seafloor spreading that had been going on for nearly 200 Ma since Pangaea broke up. This led to accumulations of

large concentrations of carbon dioxide in the atmosphere and, with the exception of some smaller orogenic events such as those in the American west (Sevier Orogeny, Antler Orogeny), there were no really major orogenic uplifts until about 45 Ma, when India begin to slam into Asia, as mentioned above. Prior to this, the Earth reached a maximum warm period, referred to as the Paleocene-Eocene Thermal Maximum (PETM).

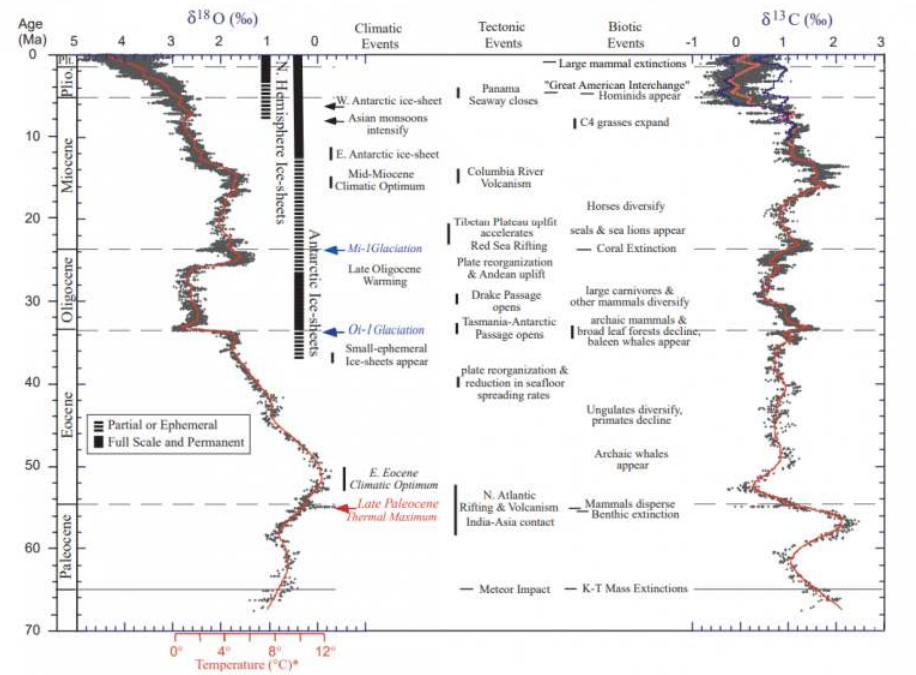


Figure 15.3.17: Major temperature fluctuations across the Paleocene-Eocene Thermal Maximum (PETM), as recorded by oxygen and carbon isotope excursions. Global temperatures reached a maximum at about 55 Ma, after which cooling began. Much of this cooling was very likely due to the collision of India with Asia at 45 Ma, a reduction in seafloor spreading rates at 39 Ma, and the rise of the Oligocene phase of the Andean Mountain Orogeny at 28 Ma. Note how much of the first half of the Cenozoic likely saw ice free polar regions. (Figure from Zachos et al., 2001; reproduced with permission)

Measuring Paleoclimatic Change: Proxy Records

By definition, paleoclimate data had no one there to collect it. So, how do we know what we know? As it turns out, the Earth has its own methods for recording its temperature (among other variables). As we get better as speaking the language of Rock, we learn to tease out of the stratigraphic record evidence of changes in temperature. Not all of these changes are recorded in rock, however. Some of them are recorded in ice. Others are recorded in wood. Still others are recorded in ocean or lake sediments. In many cases then, it's not rock, but sediment or other carbon-rich materials where we seek data. Collectively, we call these "proxy records". This simply means that they are records that stand in the place of the instrumentation we usually use to measure things like temperature.

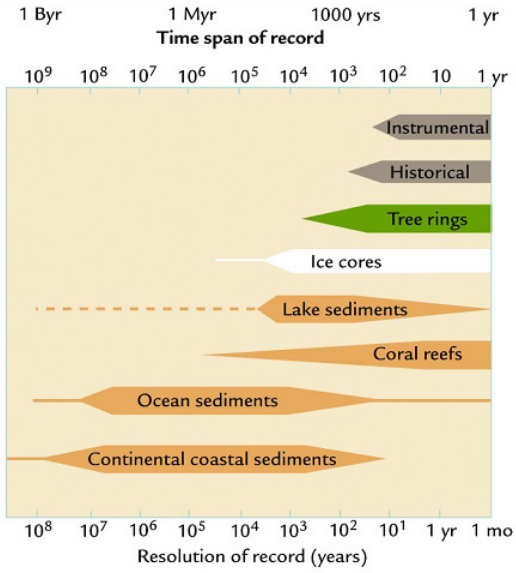


Figure 15.3.18: The time span of some examples of proxy records. (Sourc: NOAA).

Before we begin our brief overview of a variety of paleoclimate proxies, it is important to note that not all proxy records are created the same. Some record climate records over just the last few hundred to thousand years while others go back millions of years.

In the figure to the right, we can see the time limitations of historical and instrumental records. Beyond these time horizons, about 3-400 years, we refer to paleoclimate. Proxy records like tree rings can take our understanding of temperatures back to as much as 10,000 years ago and ice cores close to 800,000 years before present. Coral reef data extends to about the same time horizon as ice. From there, sediment records can give us climatic data going back as much as several tens of millions of years. At this point, the rock record and all of the data it contains become critical. For fossil shellfish, foraminifera, and other carbonate creatures, oxygen isotope records are very useful. Carbon isotope ratios extracted from black shales hundreds of millions of years old grant us insights into climate. There are numerous examples of proxy records.

You can find, explore, download, and freely use paleoclimate proxy data of great variety by using the Paleoclimatology Data Map maintained by the National Centers for Environmental Information.

Let's explore some key examples of proxy data sets commonly used by paleoclimatologists today.

Biological Proxy Records

Plant Fossils (Moderate to Long Timescales)

Plant fossils provide proxy records that vary in timeframes. Typically, macrofossils of plants, or remains large enough to be visible without a microscope, are the focus of such work. Using our knowledge of modern flora as they related to climate, including factors such as needle length, leaf shape, and more, we can use these macrofossils to describe ancient terrestrial environments. Some of these plant fossils are found in rock, other times they are found as a part of packrat middens. Fossil plants have more to say about really ancient environments while packrat middens might only describe environments of a few hundred to thousand years ago, and even then they are only applicable in arid regions.



Figure 15.3.19: Glossopteris sp. leaves from the Permian of Australia.

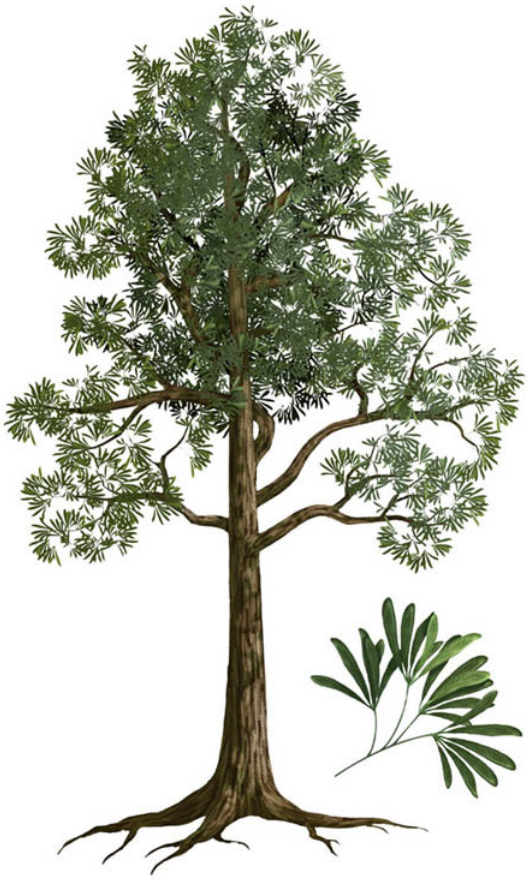


Figure 15.3.20: Artist's rendition of a Glossopteris tree (Public Domain).

When you think of plant-free harsh environments on Earth today, Antarctica might come to mind. But, a wealth of plant and animal fossils have been found there, dating back hundreds of millions of years to the Permian Period. At that time, Antarctica was a part of a very large land mass geologists call Gondwana. It was humid, what we might describe today as tropical. Antarctic land was carpeted with forests containing a wide diversity of plants. One of the well known fossil plant examples from this region is *Glossopteris*, woody plants with tongue-shaped leaves arranged in thick mats. They may have even been deciduous.

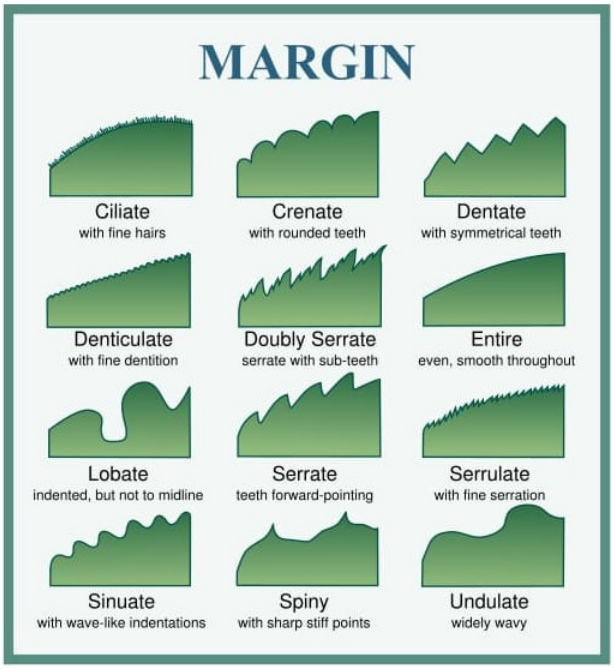


Figure 15.3.21: Leaf margin terms. “Entire” refers to leaves with smooth edges, no teeth, no lobes, etc. (Source: CC BY-SA 3.0; By derivative work: McSush (talk)Leaf_morphology_no_title.png: User: Debivort – Leaf_morphology_no_title.png, https://commons.wikimedia.org/w/index.php?curid=7681206)

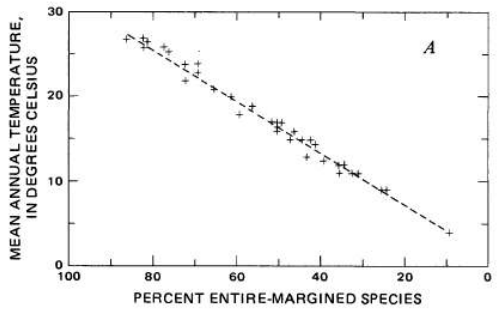


Figure 15.3.22: Percent Entire-Margined species versus temperature. There is a linear relationship between the two variables that is quantifiable. This relationship can be applied to fossil assemblages to help determine paleoclimatic conditions (Source: By Jack Wolfe – USGS, U.S. Govt. Printing Office, Public Domain, https://commons.wikimedia.org/w/index.php?curid=25994245).

Plants with leaves can be used to provide quantitative measurements of climate. Using uniformitarian principles, researchers evaluating modern environments have identified a linear relationship between the percentage of leaves in a location that have a smooth, “Entire” (toothless, non-lobate) margin and temperature. This relationship in modern species can be applied to assemblages of fossil leaves from the same time period to help determine what the temperature was like in that area.

Leaf margins were used successfully in this way for determining what the Eocene climate was like in the Geodetic Hills of Axel-Heiberg island in northern Canada.

Pollen (Short to Long Timescales)

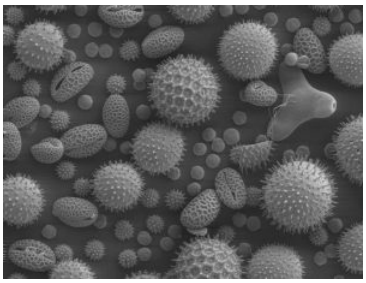


Figure 15.3.23: Electron microscope image of pollen grains from various common plants. Examples here include sunflowers, hollyhocks, lilies, caster beans, morning glories, and primroses. (Source: Windows2Universe.org via Chuck Daghlian and Louisa Howard, Dartmouth Electron Microscope Facility).

The record of pollen is directly linked to the existence of seeds. The earliest known plants that produced seeds were seed ferns. The oldest fossil seed ferns hearken to the late Devonian Period. These ranged from the size of small trees to shrubs. Since that time, over the subsequent 300 million years, the record of pollen has been accumulating. Pollen is by nature airborne. Whether transmitted through the air by insects or other organisms or blown by the wind, it can be carried great distances and, as such, was a fabulous evolutionary innovation. This adaptation persists today in all seed-bearing plants. You plant them in your garden, walk through them in the forest, sneeze because of them, and pick them for a lover or a friend.

Pollen provides a record of the vegetation of a region. Much of the record we recover and use as proxies for paleoclimate comes from lake sediments. Drifting airborne pollen becomes stuck when it touches water and eventually settles to the bed of lakes. From the emerging vegetation profile, painstakingly amassed via sometimes tedious work at microscopes, a detailed view of the climate of a region at the time and place may emerge. Pollen records run from recent times back several hundred million years. So, they are useful paleoclimate proxies over their entire time span.

Fire History (Short Timescales)

Fire history data tends to record more recent paleoclimatic data. It is also more sporadic, recording not continuously but in seasonal moments of intense burning. The data is also multiproxy, meaning that it comes in more than one form. Some fire data may come from tree rings. Other fire history data may come from lake sediments. There is even a global charcoal database!



Fire histories give us insights into ancient climate and how it has changed in a region over time. They also contribute to modern wildfire knowledge and contribute to the creation of forest management models. These system models use the history of fires over time as one of many variables to help understand feedbacks and variables related to everything as diverse as drought frequency to the volume of tinder on the forest floor.

Faunal Data (Short to Moderate Timescales)

Faunal data is very similar to plant data, in that the types of faunal remains found in a location provide excellent insights into the climate of the area at that time. Faunal remains come in various forms. There is fossil faunal data, of course. This can include everything from marine shellfish to terrestrial mastodon remains.

One significant project that has relied heavily on faunal data has been the characterization of the changes in the Sahara Desert region of north Africa during the Holocene Epoch. Analysis of the organisms from this vast region has produced evidence of past humid environments, rather than today's arid landscape. Animals of various types used waterways that existed then as migration routes. Some of these animals were aquatic. These waterways consisted of linked lakes, rivers, and inland deltas. These environments existed during the last interglacial of the Pleistocene Epoch and also during the early portion of the Holocene.

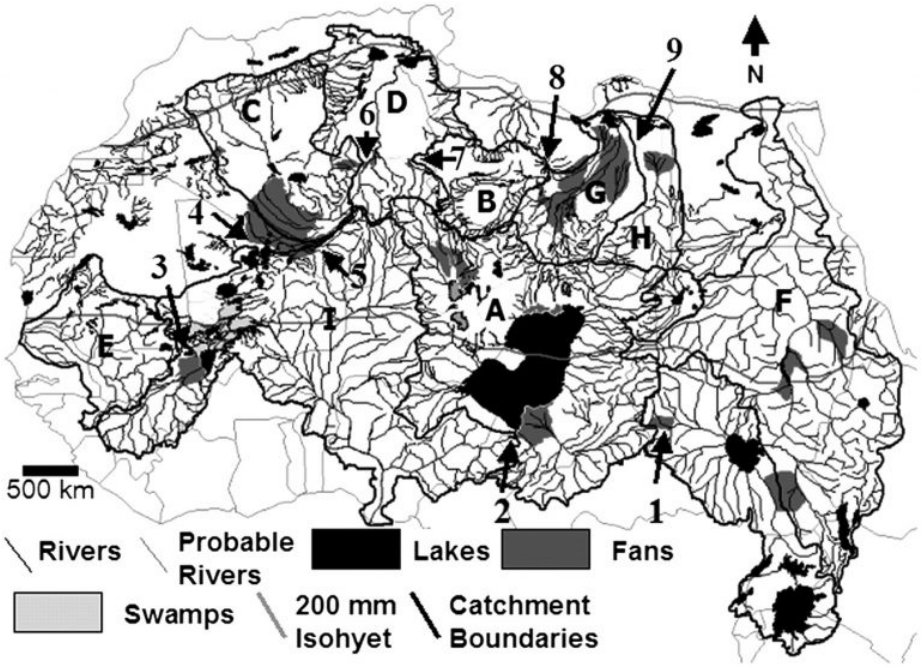


Figure 15.3.24: Reconstruction of late Pleistocene and early Holocene waterways of the Sahara, primarily from faunal data (Source: Drake et al., 2011).

The faunal evidence also includes human data. The migrations of humans were influenced by these waterways.

Physical Proxy Records

Sediments – Oceanic and Lake (Short to Long Timescales)

Sediments accumulate in basins after being eroded from highlands. As this occurs, they bring with them loads of evidence of their former lives and their journey to this new destination. Before becoming rock, these sediments contain a wealth of climatic clues, whether entombed in oceanic sediments or lacustrine.

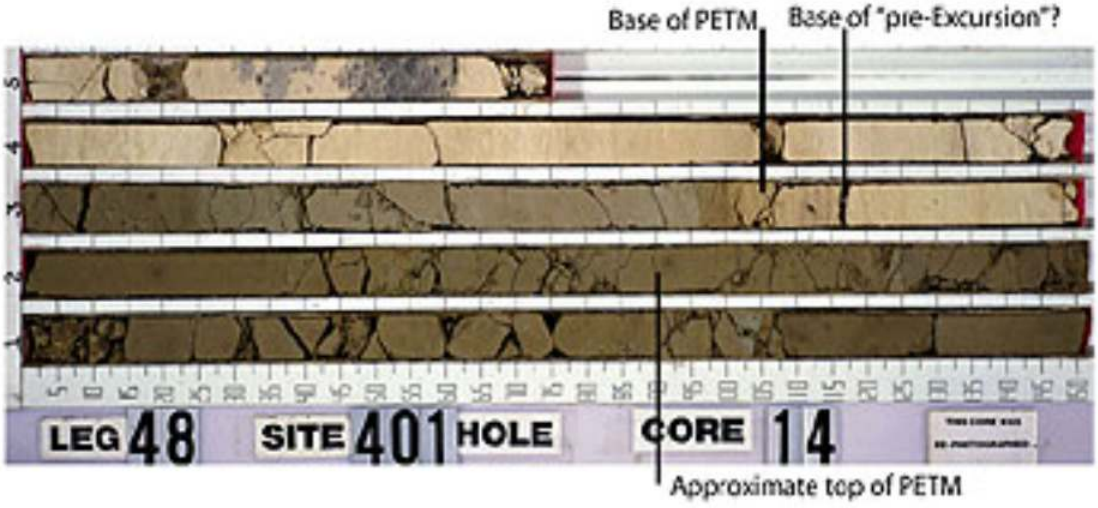


Figure 15.3.25: Ocean sediment core sample depicting, among other things, the base and top of the PETM period. Note the color change from lighter colored to darker colored during this time (Source: NSF, Scripps Institution of Oceanography).

Oceanic sediments are analyzed not only for their physical attributes, but also for chemical variables. These include isotopic data and trace metal analyses. They also contain a wealth of microfossil information. The ocean sediments themselves can give information on ocean health just by their color. Darker colored sediments may contain few shells and could provide evidence for a period of little to no oxygen (dysoxic to anoxic) . Light-colored sediments likely contain lots of shells and suggest a more healthy,

oxygen-rich ocean. Changes in ocean oxygenation are clues for past acidification events that are associated with periods of climate change, often associated with atmospheric changes such as a rise in carbon dioxide.

Lake sediments are used for similar studies, but also to analyze moisture profiles in a region. Because freshwater lakes do have so much depth variation that is directly related to seasonal and other hydrological changes, lake level is a useful measure of moisture in a region. As such, it becomes possible to describe the climate based upon this moisture. Is it humid or arid?

A great tool for exploring this is NOAA's "Lake Level Viewer" for the North American Great Lakes. Here, you can explore a wide variety of variables related to these lakes. In particular, predicted changes in shoreline can be explored using simple tools.



Figure 15.3.26: Lake Mead water level drop visualization, 2001-2015 (Source: USGS).

Apart from lake sediments, it is possible to find imagery of changes in lakes over time that also provide insights. While not specifically paleoclimatic, such imagery is very useful for documenting climatic change. Lake Mead, created by the Hoover Dam in southern Nevada, is a classic example of a lake experiencing arid conditions and, over time and through overuse, is drying out, leaving a very visible bathtub ring record.

Tree Rings (Short Timescales)

You were likely exposed to tree rings in your younger years. As children, exposure to tree rings is an introduction to a hidden world, one where curious information abounds at your fingertips. It is perhaps one of the most tangible and familiar proxy records. Major sources of tree ring data include the Southwest Paleoclimate CLIMAS portal, the Colorado River Basin Tree Ring Analysis, and the Living Blended Drought Prediction (LBDP).



Figure 15.3.27: Growth rings on a tree at the Bristol Zoo, England. Each ring contains lighter early wood and darker late wood and represent a single year of annual growth.

Rings record a good deal of information. Some rings are thin, some wide. All rings vary in thickness around the center of the tree. Why? Most rings have two different types of wood – so called "early wood" and "late wood." This couplet of two materials make up an annual growth ring. The early wood is produced in spring and feeds off of the starches stored in last summer's late wood. The thickness of early wood is a function of the quality of the late



Figure 15.3.28: Cross section of annual rings of Pinus Taeda. Early wood is lighter in color and late wood is darker in color (Source: CC BY-SA 3.0; By Pollinator, https://commons.wikimedia.org/w/index.php?curid=1674129)

wood from the year before, put on during the summer. Late wood is produced from the sugars created through photosynthesis in the leaves and then stored for the cycle to begin anew the next year. Late wood is also structurally stronger and more dense than early wood. Overall, most trees produce all of their new growth during about an eight week window from spring to early summer, until the hot weather arrives.

The thickness of a ring is a function of a variety of variables. Two of these are less important to the dendroclimatologist, the name of a specialist who studies climate through tree rings. Rings reflect the growth of the tree and its structure. A single ring will vary in thickness as you follow it up the tree and the tree narrows. It will also have a slightly variable structure depending upon how the tree grows–does it lean, split, or otherwise have diseased growths, etc. The one variable that a climatologist is interested in, moisture, has to be teased out using sophisticated statistical programs on a computer.



Figure 15.3.29: An increment borer, a drill with a hollow center, is used to extract core samples from trees without minimal damage. Samples can also be taken from structural wood (Source: CC BY-SA 2.5; By Hannes Grobe/AWI – Own work, https://commons.wikimedia.org/w/index.php?curid=1135047).

Tree rings are most often extracted from living trees or structural beams from buildings. This is done using an increment borer, which allows a small straw-like sample to be extracted from its core. This is mounted, polished, and the rings are meticulously measured along the length of the core. This data is input to a program, and compared to a database of other trees. All of this is meant to remove growth variables, leaving behind a record of moisture. From this, it is also possible to extract calendar dates for growth, particularly from buildings. This data is critical in understanding modern and paleodroughts. As a proxy record, tree rings only take researchers back several thousand years. This is enough for Quaternary geoscientists, who study the most recent geologic period, and archaeologists, who are interested in the material records of humanity, to learn how changes in climate have affected environments and human cultures. Today, governments use this and other data to monitor drought conditions. Other data are also useful, such as wood chemistry and isotopic fractionation.

Loess and Eolian Dust

Dust is something you breathe in daily. On the U.S. Gulf coast and in the Caribbean regions, some of this dust is blown in all the way from the Sahara desert. Over 100 tons of dust from the Sahara is blown westward annually. When silty sediment is windblown, it is called loess.



Figure 15.3.30: Dust blowing across the Atlantic Ocean between July 6-12, 2018. Data from Suomi/NPP satellite, NASA. (Source: NOAA, via climate.gov).

Dust is being blown all over the place, as long as the particles are small enough to be blown and wind energy great enough to keep the particles suspended. As such, ice cores are one of the most important repositories for the dust record, as much of it makes it over polar regions to find a resting place on ice caps.



Figure 15.3.31: The "Loess Hills" in the U.S. state of Iowa (Source: CC BY-SA 4.0; By BillyBlueJay – Own work, https://commons.wikimedia.org/w/index.php?curid=69537356).

What does it mean? Dust is sourced from dry land. Wet sediments do not blow away as easily as dry. So, from a paleoclimate perspective, large influxes of dust are an indication that some regions were experiencing periods of intense drought. This dust can be chemically analyzed too. Carbon dates, percentage of organic matter, and other measures can provide a wider variety of information about the dust source and depositional location.

One location where loess data has been very useful in understanding its paleoclimate is the Chinese Loess Plateau. Maher and Hu (2006) used magnetic susceptibility data and grain size measurements, for example, to create a detailed record of the southeast Asian monsoon during the



Figure 15.3.32: A canyon formed in the soft loess of Linxia County, Xihe Township, China (Source: CC BY-SA 3.0; By Vmenkov – Own work, https://commons.wikimedia.org/w/index.php?curid=7853712).

Holocene Epoch. Noteable periods of increased aridity, and dust, occurred at 12,500 BP (Before Present) and 11,500 BP, a time period also known as the "Younger Dryas", a time of sudden cooling that occurred after the start of the Holocene. Another noteworthy period of aridity began to occur after 5,000 BP, very close to what is now referred to as the start of the Meghalayan Age of the Holocene Epoch. These were likely periods of time when major shifts in the region's monsoons occurred.



Figure 15.3.33: Central China Loess Plateau region (Source: CC BY-SA 4.0; By Maintenance script – https://editors.eol.org/eoearth/wiki/Central_China_loess_plateau_mixed_forests, https://commons.wikimedia.org/w/index.php?curid=63963818).

Chemical Proxy Records

Chemical proxy records come primarily as isotopic data. You may recall from an early chemistry class that some elements exist in different forms. That is, there is a stable form and then there are versions of that element that have different numbers of neutrons, while retaining the same atomic number. Different isotopes of an element are named for their atomic mass, the total number protons and neutrons in their nucleus (for example, Oxygen-16 and Oxygen-18). Generally, in situations like oxygen and carbon, the stable versions of these elements, Carbon-12 and Oxygen-16, are found nearly everywhere. However, they do not make up 100% of the carbon or oxygen in the environment.

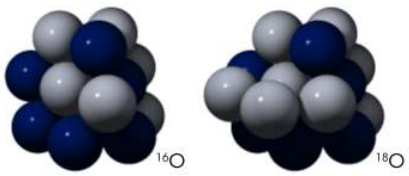


Figure 15.3.34: Stable oxygen atoms, Oxygen-16 and Oxygen-18. Both have eight protons, but Oxygen-18 has two extra neutrons (white balls), making it heavier by about 12.5% (Source: NASA Earth Observatory).

Let us explore oxygen, because the oxygen isotopic "thermometer" is such a critical one in paleoclimatic studies. In the environment, specifically the hydrologic cycle, there are two main forms of oxygen that exist, Oxygen-16 and Oxygen-18. There is some Oxygen-17 also, but its oxygen cousins are the big players. Oxygen-16 far outstrips Oxygen-18 in terms of prevalence, making up an average of 99.762% of all oxygen in the environment. Oxygen-18, by contrast, makes up an average of only 0.2% of all environmental oxygen. In both cases, these isotopes are taken up into organisms, clouds, precipitation, etc. and made into molecules. But, because Oxygen-16 and Oxygen-18 are treated differently than one another in these situations, depending upon the climate in a region, there can be very important variations in the ratio between these two isotopes. In some situations, such as in Arctic ice, Oxygen-18 is less common than average. In other situations, such as water vapor in clouds in the sub-tropics, Oxygen-18 would be more common than on average.

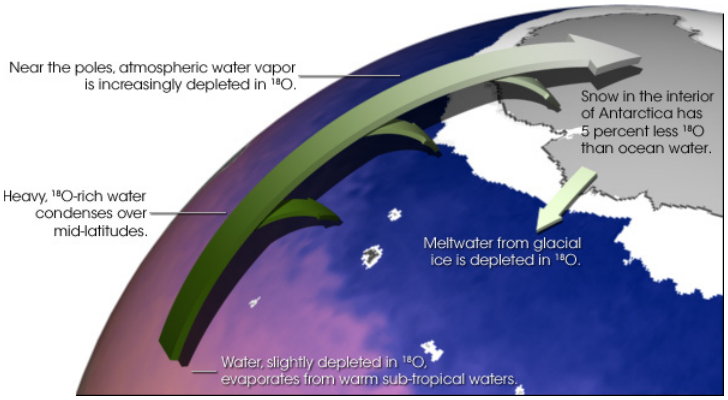
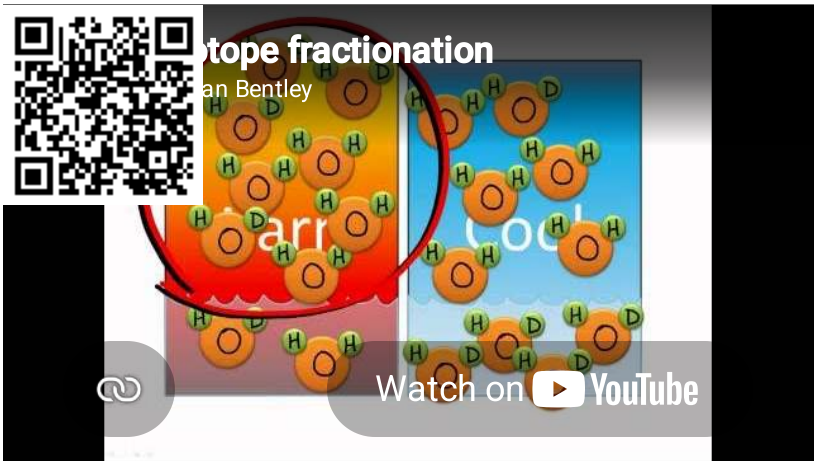


Figure 15.3.35: A version of the hydrologic cycle, focused on oxygen isotope fractionation. "Heavy", Oxygen-18 rich water condenses over lower latitudes and rains out, leaving "lighter" Oxygen-16 enriched water to head to the poles, depleted of Oxygen-18 (Source: NASA Earth Observatory).

An excellent and in depth explanation of isotopic fractionation can be had from watching this video:



$$\delta^{18}\mathrm{O} = \left(\frac{\left(\frac{^{18}\mathrm{O}}{^{16}\mathrm{O}}\right)_{\text{sample}}}{\left(\frac{^{18}\mathrm{O}}{^{16}\mathrm{O}}\right)_{\text{standard}}} - 1 \right) \times 1000‰$$

Figure 15.3.36: To calculate the Oxygen-18 and 16 ratio of a sample, the ratio of a the sample (obtained from spectroscopy) is divided by a known standard ratio, such as SMOW (Standard Mean Ocean Water), and then converted to “per mil” notation, or parts per thousand (Source: Wikipedia).

These variations from their average abundance provide useful metrics. In the case of some isotopes, the ratio between the two isotopes can be used as a proxy thermometer. In order to calculate the ratio and obtain a temperature, the Oxygen-16 to Oxygen-18 ratio from a sample is obtained using mass spectroscopy and then compared to a known standard. A commonly used known standard for coral reef data is SMOW (Standard Mean Ocean Water). The ratio is converted to “per mil” notation, or parts per thousand. The calculated ratio is represented using the Greek lower case letter delta, δ. So, the Oxygen-18 ratio is notated $\delta^{18}\mathrm{O}$.

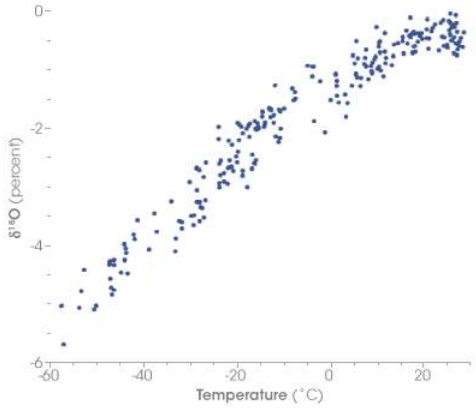


Figure 15.3.37: The oxygen isotope proxy thermometer. Concentration of Oxygen-18 decreases with temperature. The coldest sites in this data have about 5% less Oxygen-18 than ocean water (Source: NASA Earth Observatory).

The name of this field of study is stable isotope thermometry. And, oxygen isotopes are not the only example that fractionates in the environment in such a useful way. An isotope of hydrogen, deuterium, is also incorporated into water. Deuterium, $\delta^{2}\mathrm{H}$, is hydrogen with a neutron and is similarly as rare as Oxygen-18. It fractionates in much the same way in the environment when compared to its more common counterpart, plain old hydrogen.

Among the many other stable isotope thermometry examples are carbon isotopes. Carbon isotopes, notated as $\delta^{13}\mathrm{C}$, are reflective of different environmental variables. The carbon cycle governs the fractionation of carbon isotopes. Carbon-12 is most common, while Carbon-13 and Carbon-14 less so. For thermometry, we use Carbon-13 to Carbon-12 ratios. The primary reason for this is that photosynthesis discriminates against Carbon-13 (and 14) in favor of Carbon-12. The atom is smaller and is more easily dealt with by plants. Vegetation is then a reservoir of carbon depleted in Carbon-13.

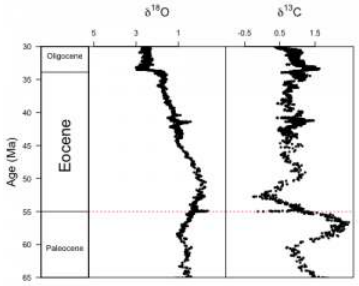


Figure 15.3.38: Depleted ratios of Oxygen-18 and Carbon-13 at the PETM boundary. Whatever caused the warming at the PETM may have been also responsible for the release of fossil carbon from decayed plant sources (After Zachos et al., 2001).

When the terrestrial biosphere is expanded, carbon with low $\delta^{13}C$ values is sequestered. When the biosphere is more restricted (such as after extinction events), atmospheric carbon reservoirs are more enriched in $\delta^{13}C$. Changes in this ratio tell us something about the health of the biosphere, which can be directly related to climate. Used in conjunction with $\delta^{18}O$ data, $\delta^{13}C$ provides deeper insights into environmental changes associated with climate and the biosphere. One example of this is a $\delta^{13}C$ excursion (big wiggle) that occurred right at the Paleocene-Eocene Thermal Maximum, which was discussed briefly earlier in this chapter. What caused a sudden and brief depletion of ^{13}C? One hypothesis is that warming led to release of methane from coastal methane hydrate deposits, decreasing the $\delta^{13}C$ ratio and contributing to a spike in warming and a small extinction episode. Methane from decomposing plant material would be much more depleted in $\delta^{13}C$.

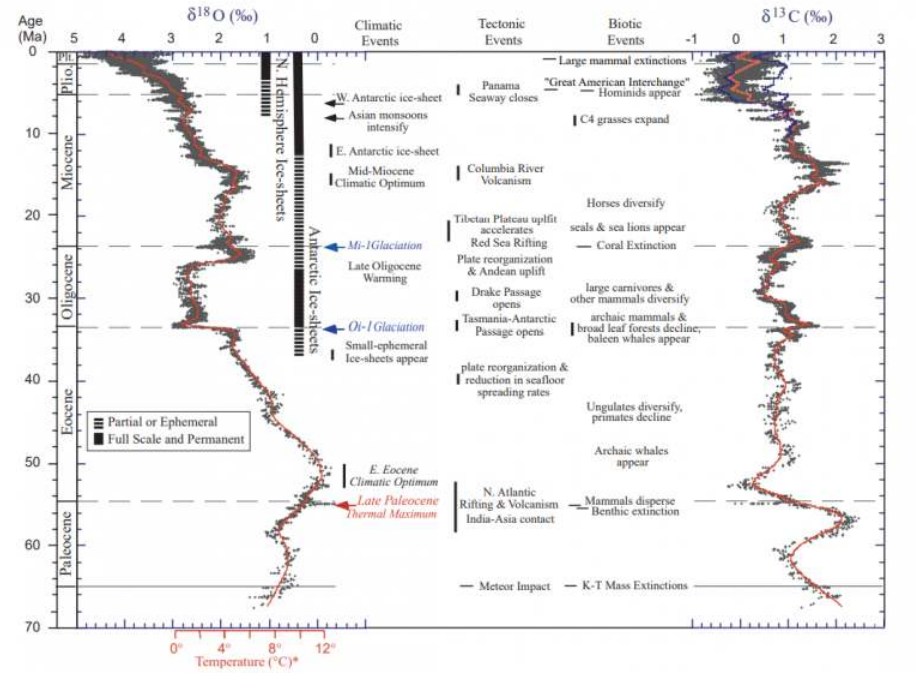


Figure 15.3.39: Major temperature fluctuations across the Paleocene-Eocene Thermal Maximum (PETM), as recorded by oxygen can carbon isotope excursions. Global temperatures reached a maximum at about 55 Ma, after which cooling began. Much of this cooling was very likely due to the collision of India with Asia at 45 Ma, a reduction in seafloor spreading rates at 39 Ma, and the rise of the Andes Mountains at 28 Ma. Note how much of the first half of the Cenozoic likely saw ice free polar regions (Used with permission, Zachos et al., 2001).

Speleothems

There is a cave in the U.S. state of West Virginia called Buckeye Creek Cave. Buckeye Creek Cave is similar to any other cave formed in karst regions. Regions underlain by limestone that chemically weathers to form certain features, such as sinkholes, caverns, *etc*., are called karst. Karst is very common around the world and is a landscape that is particularly environmentally sensitive, given how easily water tends to flow through the environment. Like other caverns, Buckeye Creek Cave is located in a limestone, the Greenbriar Limestone. It contains speleothems, cave formations like stalactites and stalagmites — again like many other limestone caves.

Importantly and distinctively though, scientists have found 7,000 years of climate data in these Buckeye Creek Cave speleothems.

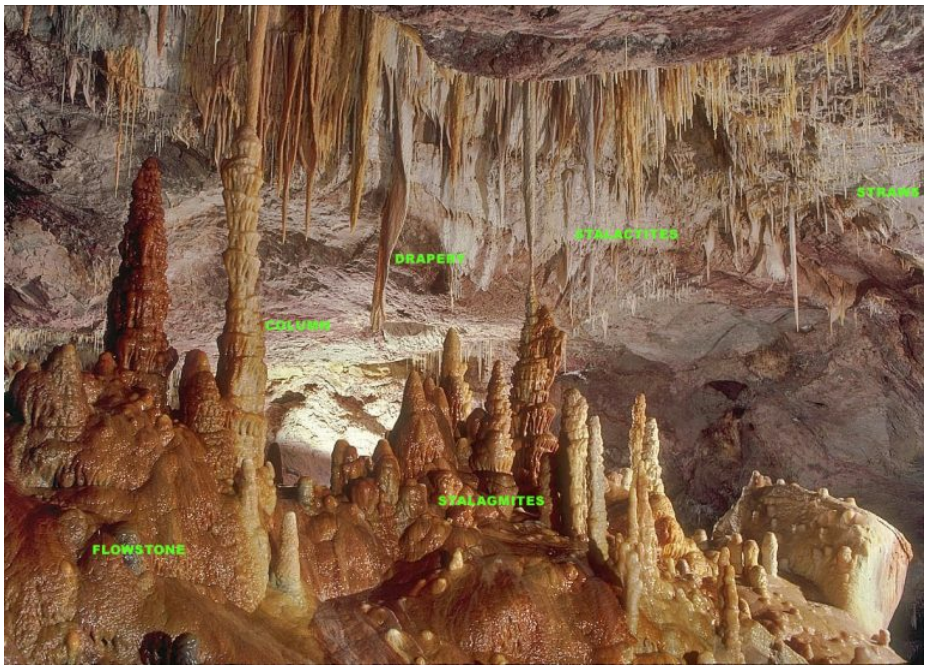


Figure 15.3.40: Cavern interior with six common speleothems labeled (Source: CC BY-SA 2.5; By Dave Bunnell / Under Earth Images – Own work, https://commons.wikimedia.org/w/index.php?curid=22613190).

Caves form while below the water table. Once the water table drops, or the cave is elevated above the water table for other reasons, water typically begins dripping through joints and bedding in the bedrock above. Drips of water, over time, leave behind calcium carbonate speleothem deposits, the ones tour guides refer to variously as “cave bacon” or “gnomes” or “bridal veils.” But how does this happen?

The atmosphere contains carbon dioxide (CO_2). This carbon dioxide, when combined with precipitating water, forms a weak acid(carbonic acid, H_2CO_3) that then infiltrates into the soil and bedrock. That soil, often the site of decomposition of organic material, can contain a good deal of carbon dioxide also, which can lower the pH of infiltrating water even further. This weak acid is enough to dissolve the limestone bedrock as it moves through, which dissociates the Ca^+ ions from their former bicarbonate partner. Now an aqueous solution of carbon dioxide, bicarbonate, and Ca^+ ions while in the rock, the water continues its work. That is, until it encounters the open space of a cavern. The water will run and drip as usual, but the open space allows the carbon dioxide to leave the solution and the Ca^+ and bicarbonate then recombine. The entire process is called carbonation, and its responsible for those lovely karst features.

$$\text{Air} + \text{Water} \rightarrow \text{Carbonic Acid:}$$
$$CO_2 + H_2O \rightarrow H_2CO_3$$

$$\text{Limestone} + \text{Carbonic Acid} \rightarrow \text{Calcium Bicarbonate:}$$
$$Ca(CO_3) + H_2CO_3 \rightarrow Ca(HCO_3)_2$$

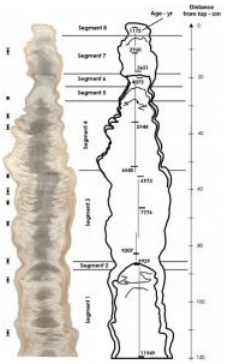


Figure 15.3.41: Cut face of a stalagmite from Jeita Cave, Lebanon (Source: EGU).

As speleothems form, they create layers. As rain can be very seasonal, these layers can take on an annual character and be directly attributed to particular years. Because of this annual accumulation, scientists can use speleothem layers much like they do coral growth layers, tree rings, and ice cores.

Buckeye Creek Cave has three speleothem formations that record 7,000 years of data. According to researchers (Hardt et al., 2010), years where calcite is enriched in $\delta^{18}O$ represent an increase in the summer precipitation in annual totals. Critically, the data also show a major change in $\delta^{18}O$ at 4.2 ka, which is now recognized as the start of the Meghalayan age, as documented by a speleothem from northern India. The presence of this data in both locations a globe away testifies not only to the usefulness of this data, but also to the global nature of the 4.2 ka climate event. Below is the 7,000 year record of $\delta^{18}O$ values.

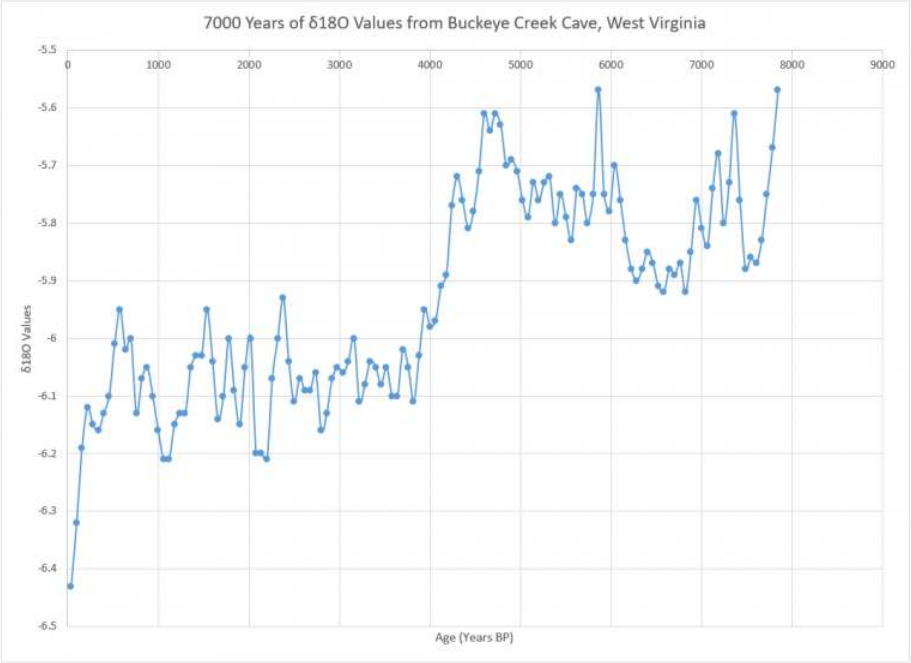


Figure 15.3.42: 7,000 years of $\delta^{18}O$ values from Buckeye Creek Cave, West Virgina. Note the change beginning around 4.2 ka and then again in the last few years. This kind of data from karst environments is a measure of moisture (Source: Hardt et al., 2010).

Coral and Sclerosponges



Figure 15.3.43: Coral reef outcrop, Great Barrier Reef (Source: CC BY-SA 3.0; By Toby Hudson – Own work, https://commons.wikimedia.org/w/index.php?curid=11137678).

Coral reefs are perhaps one of the more endangered groups of animals on this planet. Coral are marine animals that create massive skeletal reefs, adding to them annually, that create widespread habitat for myriad organisms in otherwise nutrient-poor tropical waters. Everything about coral reefs is about symbioses. It begins with the coral animals themselves, which cannot survive without the waste products, food to the coral, produced through photosynthesis practiced by the zooxanthellae algae that live with the coral polyps. Here is how it works.



Figure 15.3.44: Bleached Acropora coral, with unbleached (normal) coral in the background. Coral become bleached when they lose their zooxanthellae algae, typically due to high water temperatures (Source: Public Domain).

Coral polyps bring in dissolved minerals from the seawater and produce proteins. At this point, the coral polyp begins secreting calcium carbonate, using the water around it, in which the current $\delta^{18}O$ value will be recorded. The zooxanthellae algae provide the coral with oxygen, sugars, and fats through photosynthesis. In return, the coral polyp provides shelter, and nutrients needed by the algae for growth, such as nitrogen.



Figure 15.3.45: Montastrea cavernosa coral polyps. (Public Domain)

The coral polyps are filter feeders. This requires them to live near the surface of the water, where there is constant movement. This entire relationship requires water temperatures in a narrow range, stable sea levels, and the presence of other animals and plants in the community that contribute in other ways.

For paleoclimate researchers, it is the annual rings that are of particular interest. Because they record the $\delta^{18}O$ values of the seawater at that time and location, they are a window into the temperature, particularly the water temperature. Other geochemical analyses on these rings can add to the insights provided by oxygen isotopes. Since coral have been in around in one form another as far back as the Cambrian Period, they are very useful proxy records.

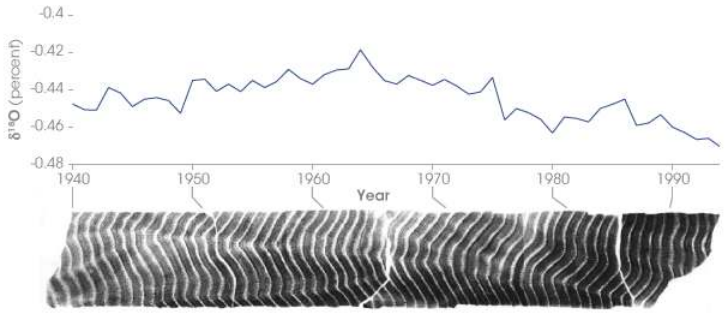


Figure 15.3.46: Coral growth rings are produced annually. The coral polyp will produce its calcite using oxygen available to it. In cooler times, there will be less Oxygen-18 than in warmer times. Because of this, both fossil and modern coral skeletons are excellent climate indicators (Source: NASA Earth Observatory).

Oceanic microplankton

Ocean microplankton form the very bottom of the food web and trophic pyramid. They are the primary producers of the marine realm, the survival of everything else hinging upon their continued happiness.

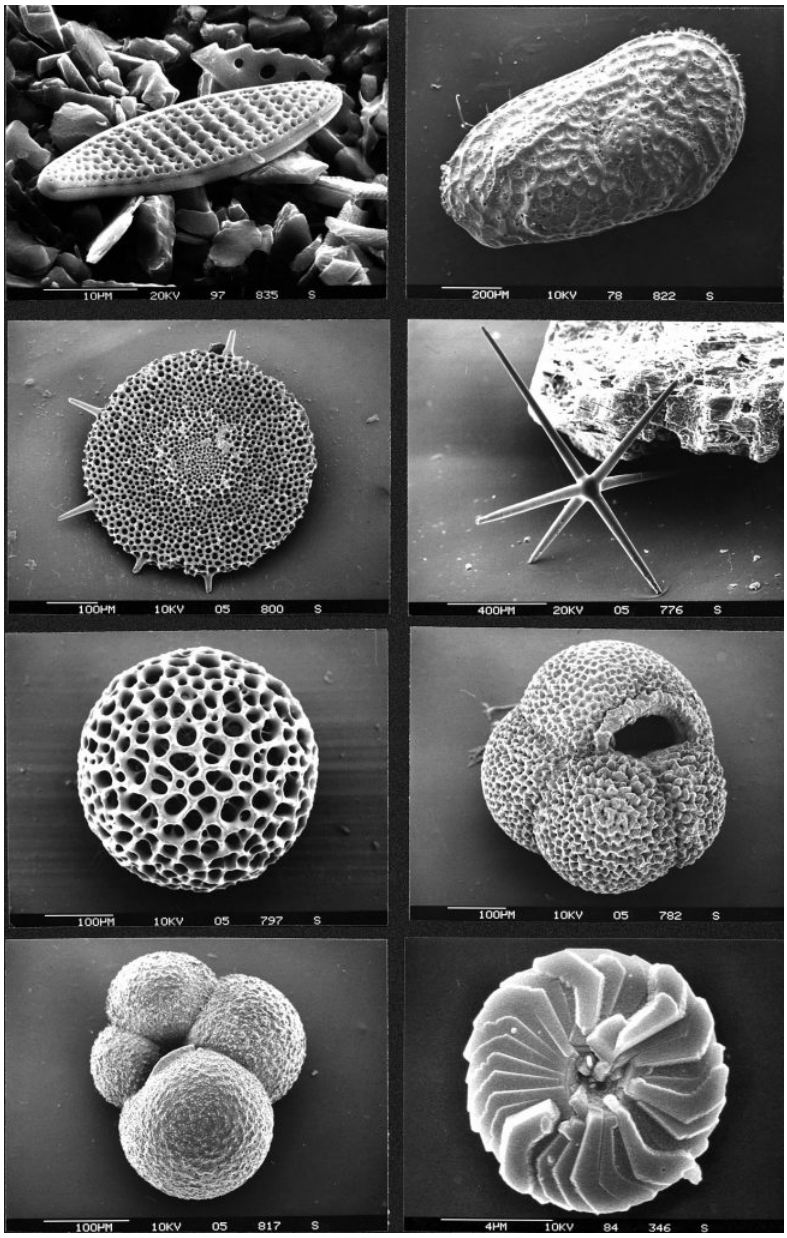


Figure 15.3.47: Various marine microfossils, inlcluding, from upper left to lower right, diatom, ostracod, radiolarian, sponge spicule, radiolarian, two planktonic foraminiferans, and a coccolith (Source: CC BY 3.0; By Hannes Grobe/AWI – Own work, https://commons.wikimedia.org/w/index.php?curid=8831075).

Some of these microfossils, foraminiferans, secrete calcium carbonate shells, or tests. Others secrete siliceous tests, made form silicon dioxide. The fossils will use the oxygen isotopes that are around and available, as well as the hydrogen, as they secrete these tests. Because they are so small and planktonic, which by definition means they float around, they are easily moved around by ocean currents. Once they die, their bodies rain down to the seafloor, where researchers can extract them from sediment core samples.

They are excellent paleoclimate indicators in a variety of ways. First, they do include $\delta^{18}O$ values that are useful temperature indicators of the surrounding seawater (higher $\delta^{18}O$ = lower temperature). This characteristic is not just unique to carbonate microfossils such as foraminiferans and coccoliths, but also to siliceous microfossils, like diatoms and radiolarians.

Some species are cold-tolerant. When this is known, finding them in particular areas is a really good indicator of a colder water temperature in that region at that time of their lives and deposition of their skeletal remains. Microfossils can also be indicators of upwelling and nutrient movement, which can tell scientists about the kinds of wind and weather patterns in that area at the time.

Ice Cores

Of all of the paleoclimate proxies, ice cores are perhaps the most glamorous. This is not just because they are much more familiar to people, but also because the wealth of information contained in an ice core is quite astounding. Ice cores are extracted from glaciers using a core drill. Ice is layered on annually, like tree rings, coral growth, and speleothems. Each year, the annual snowfall eventually compacts to firn and then to ice. While doing so, it traps dust and bubbles of gas associated with that moment in time with it. Then it sits. Conveniently for us, scientists can come back 800,000 years later and pull this data "out of the freezer."

Ice core data includes oxygen, deuterium, carbon, and often other stable isotopes. It also includes dust, atmospheric gas concentrations (such as carbon dioxide and methane), along with other geochemical data, including lead, sulfur, and other industrial and environmental parameters. Thin layers of volcanic ash help constrain the age of the ice. It's a rich treasure trove of information!

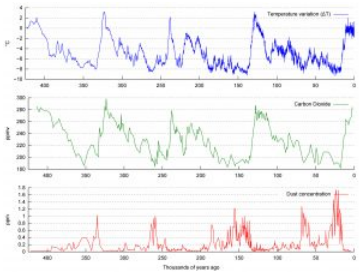
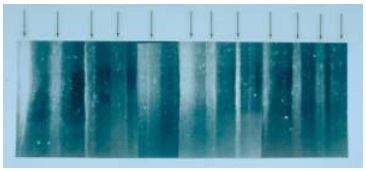


Figure 15.3.48: Dust, carbon dioxide, and temperature (from $d^{18}O$ values) for the last 420,000 years as taken from the Vostok Ice Core (Source: CC BY-SA 3.0; By Vostok-ice-core-petit.png: NOAAderivative work: Autopilot (talk) – Vostok-ice-core-petit.png, https://commons.wikimedia.org/w/index.php?curid=10684392).

There are numerous ice cores that have been extracted from mountain glaciers, ice sheets, and more. Perhaps the most famous is the Vostok ice core, extracted by the Soviet Union at Vostok Station in 1987. Formerly the oldest, but still one of the most detailed cores we have, goes back 800,000 years and is referred to as Antarctic Dome C, extracted by the EPICA project (European Project for Ice Coring in Antarctica) in 1996. The oldest ice core, containing ice dating as far back as 2.7 Ma, was extracted from the Allen Hills in Antarctica. by a team led by Princeton University in 2015.



Greenland has produced some important cores that contain long climate records also. The Greenland Ice Core Project Program (GRIP) and North Greenland Ice Core Project Program (NGRIP) have extracted multiple core samples from areas of Greenland where ice thickness exceeds several thousand feet. Like the Antarctic ice cores, these cores have been critical for understanding past global conditions, including temperatures, moisture, anthropogenic emissions, and other environmental factors.

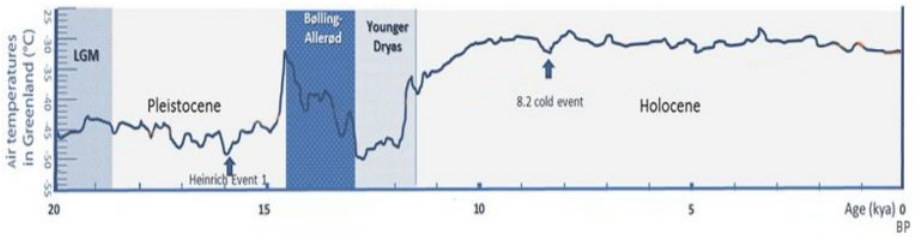


Figure 15.3.49: The evolution of temperature after the last major glacial advance and into the Holocene Epoch. Note the period of the Younger Dryas and the 8.2 ky event. This last event defines the boundary between the Northgrippian Age and the Greenlandian Age of the Holocene Epoch (Source: CC BY-SA 4.0; By Daniel E. Platt et al., https://commons.wikimedia.org/w/index.php?curid=78989262).

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15.4: Significant Paleoclimatological Events

Many significant paleoclimatological events from Earth's long history have been mentioned in the text above. However, should you wish to explore in more detail some of these compelling stories, below are a series of links you can pursue. All of these have in common several themes. The climate changes. It does so for a variety of reasons and at a variety of rates. As you read more about these specific events, you should see many common themes emerge, but you should also see proxy data at work helping us understand our past climate.

Great Oxygenation Event



Snowball Earth

Ordovician Extinction

PETM



Permian Extinction

Pleistocene Glaciations

The Younger Dryas

Defining Our Age: Holocene, Anthropocene, Meghalayan

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15.5: Further Reading

Coffin, Millard & Duncan, Robert & Eldholm, Olav & Fitton, J. & Frey, Fred & Larsen, Hans & Mahoney, John & Saunders, Andrew & Schlich, Roland & Wallace, Paul. (2006). Large Igneous Provinces and Scientific Ocean Drilling: Status Quo and A Look Ahead. *Oceanography*. 19. 150-160. 10.5670/oceanog.2006.13.

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CHAPTER OVERVIEW

16: (Case Study) Nebular theory and the formation of the solar system

16.1: In the beginning...

16.2: Nebular theory

16.3: A star is born

16.4: Age of the solar system

16.5: The implications of meteorites

16.6: Further reading

16.S: Summary

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16.1: In the beginning...

How and when does the story of Earth begin? A logical place to start is with the formation of the planet, but as you'll soon see, the formation of the planet is part of a larger story, and that story implies some backstory *before* the story, too. The purpose of this case study is to present our best scientific understanding of the formation of our solar system from a presolar nebula, and to put that nebula in context too.

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16.2: Nebular theory

The prevailing scientific explanation for the origin of the Earth does a good job of not only explaining the Earth's formation, but the Sun and all the other planets too. Really, it's not "the Earth's origin story" alone so much as it is the origin story of the ***whole*** solar system. Not only that, but our Sun is but one star among a hundred million in our galaxy, and our galaxy is one of perhaps a hundred million in the universe. So the lessons we learn by studying our own solar system can likely be applied more generally to the formation of other solar systems elsewhere, including those long ago, in galaxies far, far away. The vice versa is also true: Our understanding of our own solar system's origin story is being refined as we learn more about exoplanets, some of which defy what we see in our own system; "hot Jupiters" and "super-Earths," for instance, are features we see in other star systems but not our own.

When we use powerful telescopes to stare out into the galaxy, we observe plenty of other stars, but we observe other things too, including fuzzy looking features called nebulae. A nebula is a big cloud of gas and dust in space. It's not as bright as a star because it's not undergoing thermonuclear fusion, with the tremendous release of energy that accompanies that process. An example of a nebula that you are likely to be able to see is in the constellation Orion. Orion's "belt," three stars in a row, is a readily identifiable feature in the northern hemisphere's night sky in winter. A smaller trio of light spots "dangle" from the belt; this is Orion's sword scabbard. A cheap pair of binoculars will let you examine these objects for yourself; you will discover that the middle point of light in this smaller trio is not a star. It is a nebula called Messier 42.

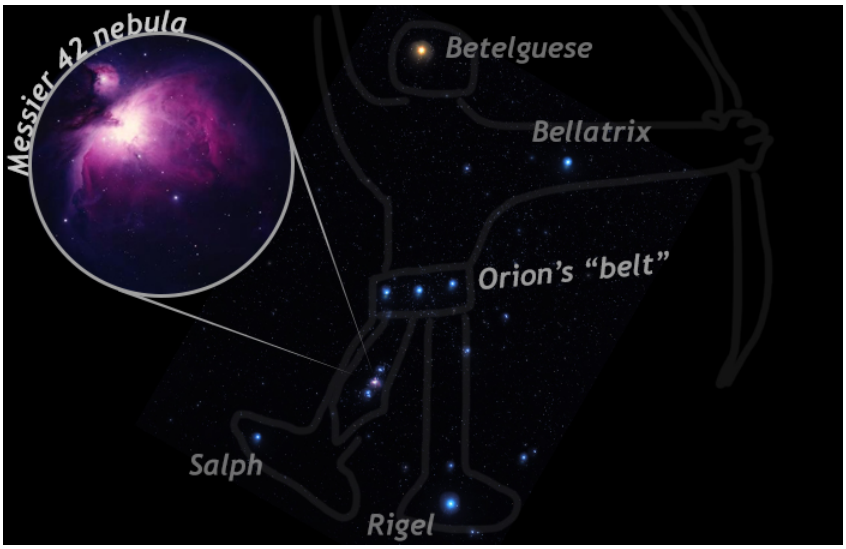


Figure 16.2.1: The Messier 42 nebula, shown in the context of the "scabbard" of the constellation Orion.

Nebulae like Messier 42 are common features of the galaxy, but not as common as stars. Nebulae appear to be short-lived features, as matter is often attracted to other matter. All that stuff distributed in that tremendous volume of space is not as stable as it would be if it were all to be drawn together into a few big clumps. Particles pull together with its neighboring particles under the influence of various forces, including "static cling" or electrostatic attraction. This is the same force that makes tiny dust motes clump up into dust bunnies under your couch!



Figure 16.2.2: Three dust bunnies under the couch: a household hassle, or a glimpse into the birth of the solar system?

Now, electrostatic force is quite strong for pulling together small particles over small distances, but if you want to make big things like planets and stars out from a nebula, you're going to need gravity to take over at some point. Gravity is a rather weak force. After all: every time you take a step, you've overcoming the gravitational pull of the entire Earth. But gravity can work very efficiently over distance, if the masses involved are large enough. So static cling was the initial organizer, until the "space dust bunnies" got large enough, then gravity was able to take over, attracting mass to mass. The net result is that the gajillions of tiny pieces of the nebula were drawn together, swirling into a denser and denser amalgamation. The nebula began to spin, flattening out from top to bottom, and flattening out into a spinning disk, something between a Frisbee and a fried egg in shape:

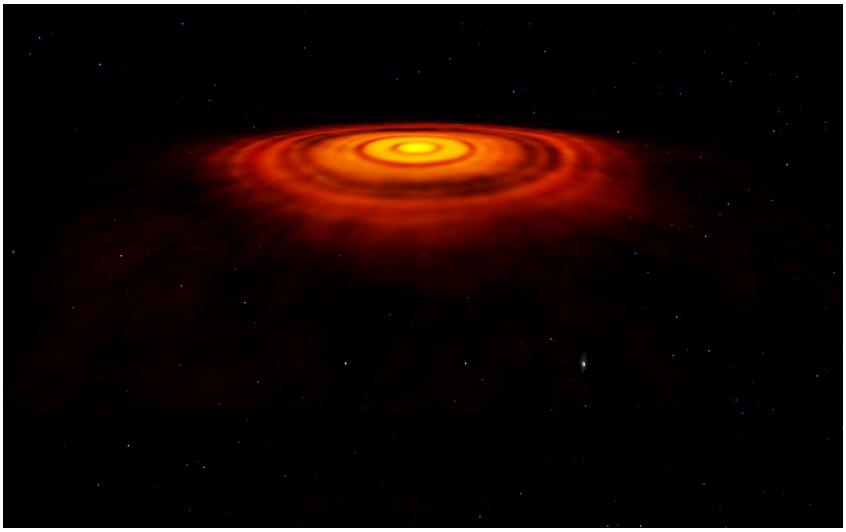


Figure 16.2.3: An artist's conception of an oblique view of a protoplanetary disk, using actual photoimagery of the HL Tauri disk, originally gathered by the European Southern Observatory.

Once a star forms in the center, astronomers call the ring of debris around it a protoplanetary disk. Two important processes that helped organize the protoplanetary disk further were condensation and accretion.

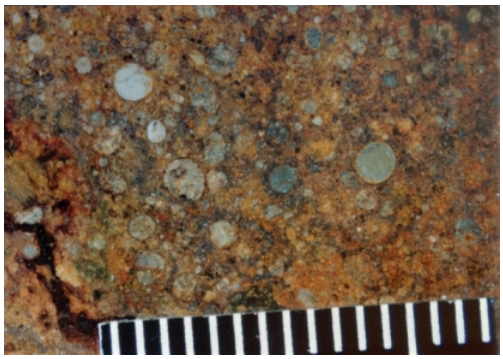


Figure 16.2.4: Chondrules in a meteorite, with a scale in mm.

Condensation is the process where gaseous matter sticks together to make liquid or solid matter. We have evidence of condensation in the form of small spherical objects with internal layering, kind of like “space hailstones.” These are chondrules, and they represent the earliest objects formed in our solar system. (Occasionally, we are lucky enough to find chondrules that have survived until the present day, entombed inside certain meteorites of the variety called chondrites.)

Chondrules glommed onto other chondrules, and stuck themselves together into primordial “rocks,” building up larger and larger objects. Eventually, these objects got to be big enough to pull their mass into an round shape, and we would be justified to dub them “planetesimals.” Planetesimals gobbled up nearby asteroids, and smashed into other planetesimals, merging and growing through time through the process of accretion. The kinetic force of these collisions heated the rocky and metallic material of the planetesimals, and their temperature also went up as radioactive decay heated them from within. Once warm, denser material could sink to their middles, and lighter-weight elements and compounds rose up to their surface. So not only were they maturing into spheroidal shapes, but they were also differentiating internally, separating into layers organized by density.

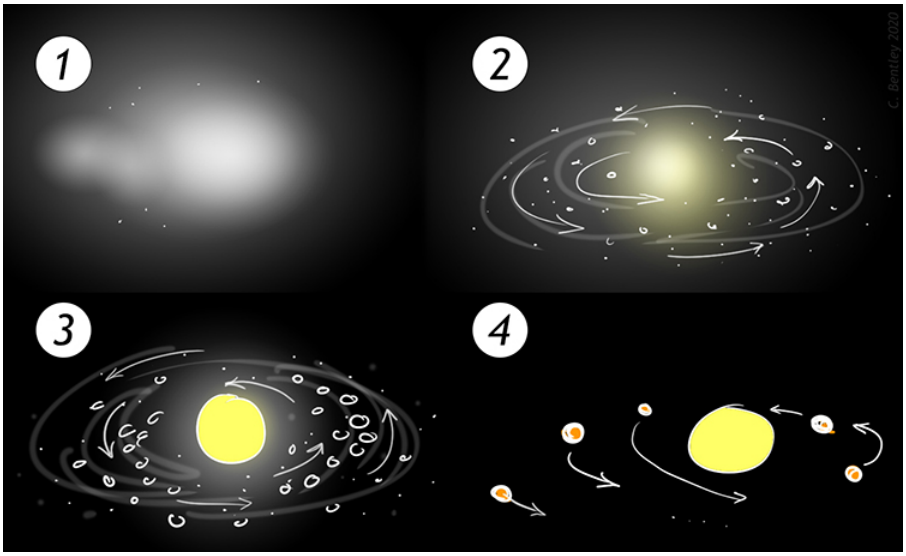
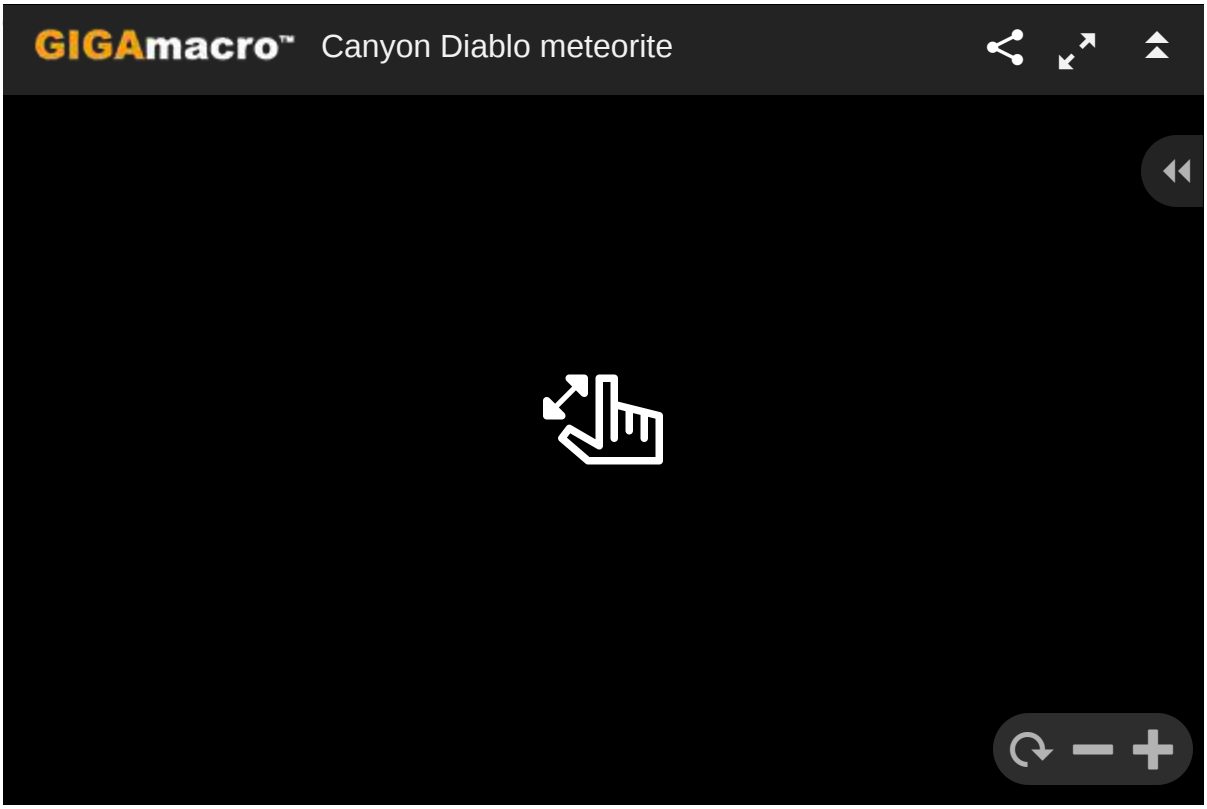


Figure 16.2.5: A cartoon model showing the evolution of our solar system from a pre-solar nebula.

Meteorites that show metallic compositions represent “core” material from these planetesimals; core material that we would never get to glimpse had not their surrounding rocky material been blasted off. Iron meteorites such as the Canyon Diablo meteorite below (responsible for Arizona’s celebrated Meteor Crater) therefore are evidence of differentiation of planetesimals into layered bodies, followed by disaggregation: a polite way of saying they were later violently ripped apart by energetic collisions.



If you were to somehow weigh the nebula before condensation and accretion, and again 4.6 billion years later, we'd find the mass to be the same. Rather than being dispersed in a diffuse cloud of uncountable atoms, the condensation and accretion of the nebula resulted in exactly the same amount of stuff, but organized into a smaller and smaller number of bigger and bigger objects. The biggest of these was the Sun, comprising about 99.86% of all the mass in the solar system. Four-fifths of the remaining 0.14% makes up the planet Jupiter. Saturn, Neptune, and Uranus are huge gas giants as well. The inner rocky planets (including Earth) make up a tiny, tiny fraction of the total mass of the whole solar system – but of course, just because they are relatively small, that doesn't mean they are unimportant!

The process of accretion continues into the present day, though at a slower pace than the earliest days of the solar system. One place you can observe this is in the asteroid belt, where there are certain asteroids that are basically nothing more than a big 3D pile of space rocks, held together under their own gravity. Consider the asteroid called Itokawa 25143, for instance:

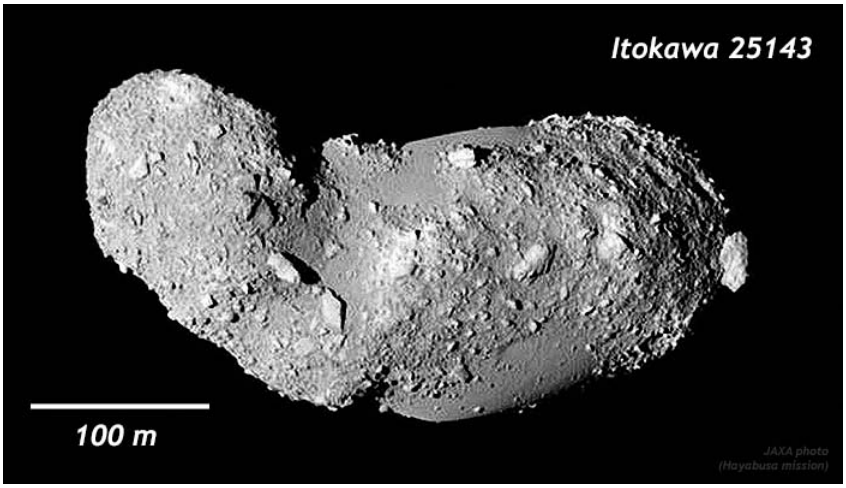


Figure 16.2.6: The asteroid 25143 Itokawa.

Only about half a kilometer long, and only a few hundred meters wide, Itokawa doesn't even have enough gravity to pull itself into a sphere. If you were to land on the surface of Itokawa and kick a soccer-ball-sized boulder, it would readily fly off into space, as the force of your kick would be much higher than the force of gravity causing it to stay put.

Another example of accretion continuing to this day is meteorite impacts. Every time a chunk of rock in space intersects the Earth, its mass is added to that of the planet. In that instant, the solar system gets a little bit cleaner (fewer leftover bits rattling around) and the planet gets a little more massive. A spectacular example of this occurred in 1994 with Comet Shoemaker-Levy 9, a comet which had only been discovered the previous year. Jupiter's immense gravity broke the comet into chunks, and then swallowed them up one after another. Astronomers on Earth watched with fascination as the comet chunks, some more than a kilometer across, slammed into Jupiter's atmosphere at 60 km/second (~134,000 mph), creating a 23,700 °C fireball and enormous impact scars that were as large as the entire Earth. These scars lasted for months.

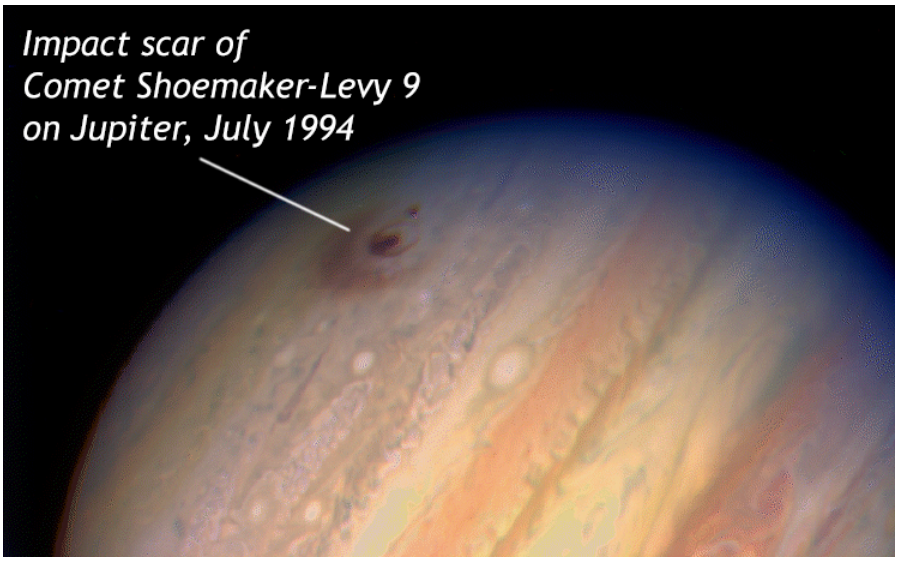


Figure 16.2.7: Accretion in action: scar from the July 1994 collision between a fragment of Comet Shoemaker-Levy 9 and the surface of the atmosphere of the planet Jupiter.

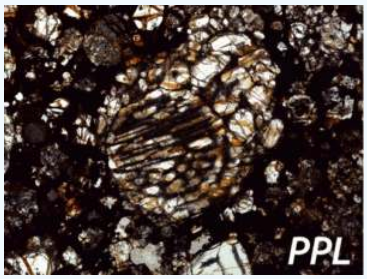
This incredibly dramatic event perhaps raises the hair on our necks, seeing the violence and power of cosmic collisions. It's a reminder that Earthlings are not safe from accretionary impacts even today – as the dinosaurs found out. For the purposes of our current discussion, though, bear in mind that the collision was really a merger between the masses of Comet Shoemaker-Levy 9 and the planet Jupiter, and after the dust settled, the solar system had one fewer object left off by itself, and Jupiter gained a bit more mass. This is the overall trend of the accretion of our solar system from the presolar nebula: under gravity's influence, the available mass becomes more and more concentrated through time.

Did I Get It? - Quiz

? Exercise 16.2.1

What feature is shown here, and when did it form?

(Microscopic thin section, field of view about 5 mm, shown in plane-polarized light (PPL) and cross-polarized light (XPL).



a. This is an asteroid. It formed last week.

b. This is a nebula. It formed in the aftermath of a old star's supernova explosion within a few million years of the Big Bang.

c. This is a granite. It formed during the growth of the continental crust.

d. This is a chondrule. One of the oldest objects known from our solar system, it formed during the earliest stages of condensation of the presolar nebula.

Answer

d. This is a chondrule. One of the oldest objects known from our solar system, it formed during the earliest stages of condensation of the presolar nebula.

? Exercise 16.2.2

What has happened to the amount of matter (*e.g.*, the mass) of the solar system between its early nebula days and today?

a. It has decreased. It weighs less now.

b. It hasn't changed at all - merely been organized into a smaller number of bigger bits.

c. It has increased. It weighs more now.

Answer

b. It hasn't changed at all - merely been organized into a smaller number of bigger bits.

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16.3: A star is born

Because the Sun is so massive, it is able to achieve tremendous pressures in its interior. These pressures are so high, they can actually *force two atoms into the same space*, overcoming their immense repulsion for one another, and causing their two nuclei to merge. As two atoms combine to make one more massive atom, energy is released. This process is thermonuclear fusion. Once it begins, stars begin to give off light.

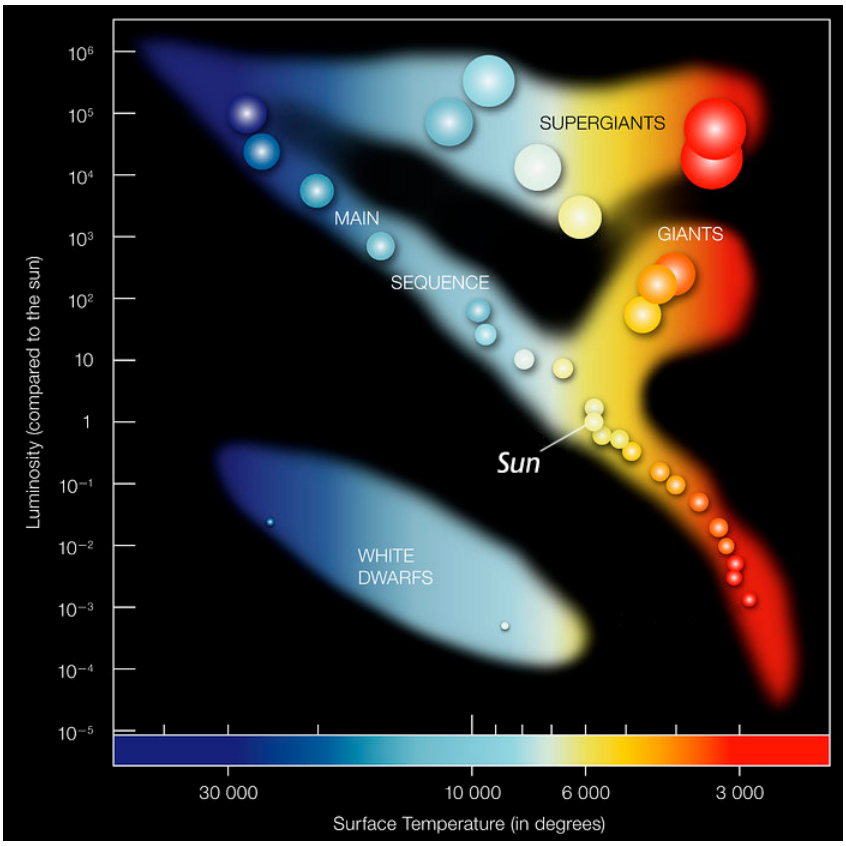


Figure 16.3.1: A Hertzprung-Russell diagram, plotting the temperatures of stars (how hot they are) are against their luminosities (how much light they emit). The position of a star in the diagram provides information about its present stage and its mass. Stars that burn hydrogen into helium lie on the diagonal branch, the so-called main sequence. Our Sun is an example of a main sequence star, about halfway through its "life" expectancy. Red dwarfs lie in the cool and faint lower right corner. When a star exhausts all the hydrogen, it leaves the main sequence and becomes a red giant or a supergiant, depending on its mass (upper right corner). Stars with the mass of the Sun which have burnt all their fuel finally develop into a white dwarf (lower left corner). Modified from an original by the European Southern Observatory.

The ability of stars to make big atoms from small ones is key to understanding the history of our solar system and our planet. Planet Earth is made of a wide variety of chemical elements, both lightweight and heavy. All of these elements must have been present in the nebula, in order for them to be included in Earth's "starting mixture." Elements formed in the Sun today stay in the Sun, fusing low-weight atoms into heavier atoms. So all the elements on Earth today came from a pre-Sun star. We can go outside on a spring day and enjoy the Sun's warmth, but the carbon that makes up the skin that basks in that warmth was forged in the heart of *another* star, a star that's gone now, a star that blew up.

This exploding star was the source of the nebula where we began this case study: it's the backstory that occurred before the opening scene. Our solar system is like a "haunted house," where billions of years ago, there was a vibrant, healthy main-sequence star right here, in this part of the galaxy. Perhaps it had planets orbiting it. Perhaps some of those planets harbored life. We'll never know: the explosion wiped the slate clean, and "reset" the solar system for the iteration in which we live. The ghostly remnants of this time before our own still linger, in the very stuff we're made from. This long-dead star fused hydrogen to build the carbon in our bodies, the iron in our blood, the oxygen we breathe, and the silicon in the rocks of our planet.

This is an incredible realization to embrace: everything you know, everything you trust, everything you *are*, is stardust.

Did I Get It? - Quiz

Exercise 16.3.1

What process do stars undergo?

a. Thermonuclear fusion

b. Accretion

c. Galactic hydrogenation

d. Fission

Answer

a. Thermonuclear fusion

Exercise 16.3.2

The fusion of smaller atoms (like hydrogen and helium) in stars results in _____ and _____.

a. emission of light; nebulae

b. emission of light; bigger atoms

c. nebulae; bigger atoms

Answer

b. emission of light; bigger atoms

Exercise 16.3.3

Everything you've ever touched is made of _____.

a. chondrules

b. radiation

c. helium

d. stardust

Answer

d. stardust

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16.4: Age of the solar system

So just when did all this happen? An estimate for the age of the solar system can be made using isotopes of the element lead (Pb). There are several isotopes of lead, but for the purposes of figuring out the age of the solar system, consider these four: ^{208}Pb, ^{207}Pb, ^{206}Pb, and ^{204}Pb.

^{208}Pb, ^{207}Pb, ^{206}Pb are all radiogenic: that is to say, they stable "daughter" isotopes that are produced from the radioactive "parent" isotopes. Each is produced from a different parent, at a different rate:

Table 16.4.1

Parent isotope	Stable daughter	Half-life
^{232}Th	^{208}Pb	14.0 b.y.
^{238}U	^{206}Pb	4.5 b.y.
^{235}U	^{207}Pb	0.70 b.y.

^{204}Pb is, as far as we know, non-radiogenic. It's relevant to this discussion because it can serve as a 'standard' that can allow us to compare the other lead isotopes to one another. Just as if we wanted to compare the currencies of Namibia, Indonesia, and Chile, we might reference all three to the U.S. dollar. The dollar would serve as a standard of comparison, allowing us to better see the value of the Namibian currency relative to the Indonesian currency and the Chilean currency. That's what ^{204}Pb is doing for us here.

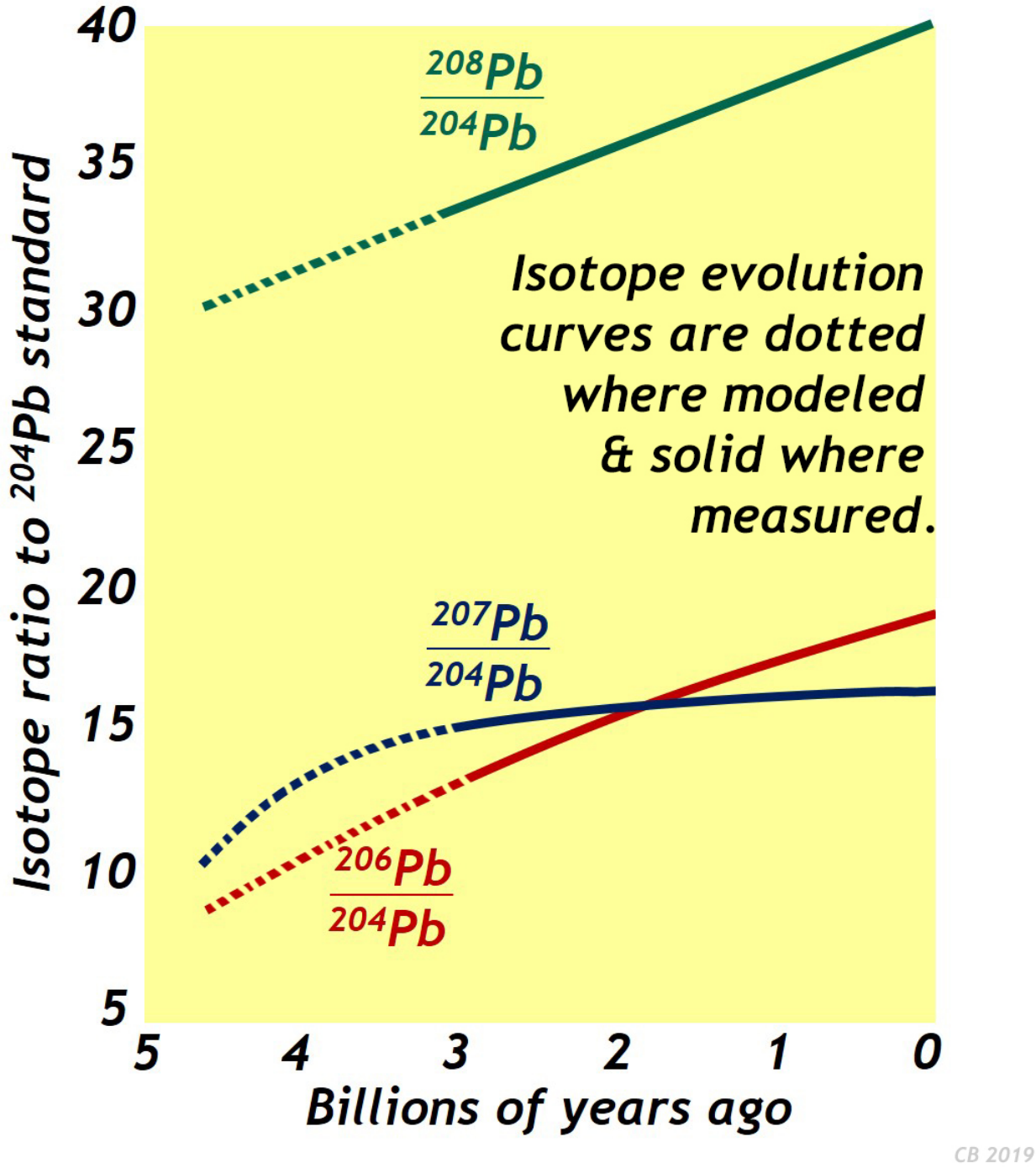
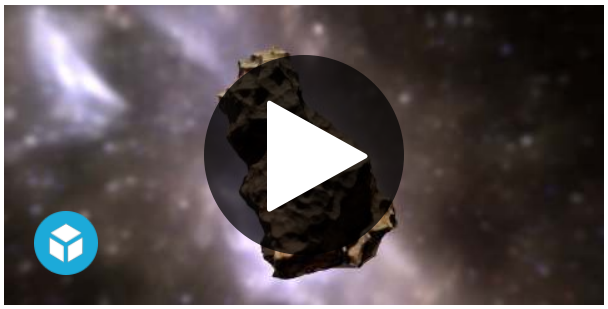


Figure 16.4.1: Lead (Pb) isotope ratio evolution over the past 5 billion years, including both terrestrial (Earth rock) measurements and projections of primordial evolution, though no Earth rocks of that age persist.

This is a plot showing the modeled evolution of our three radiogenic lead isotopes relative to ^{204}Pb. It is constrained by terrestrial lead samples at the young end, and projected back in time in accordance with our measurements of how quickly these three isotopes of lead are produced by their radioactive parents. Of course, if we go back far enough in time, we run out of samples to evaluate. The Earth's rock cycle has destroyed all its earliest rocks. They've been metamorphosed, or weathered, or melted – perhaps many times over! What would be really nice is to find some rocks from the early end of these curves – some samples that could verify these projections back in time are accurate.

Such samples ***do*** exist! But they are not from the Earth so much as "from the Earth's starting materials." If the nebular theory is correct, then a few leftover scraps of the planet's starting materials are found in the solar system's asteroids. Every now and again,

bits of these space rocks fall to earth, and if they survive their passage through the atmosphere, we may be lucky enough to collect them, and analyze them. We call these space rocks “meteors” as they streak through the atmosphere, heating through friction and oxidizing as they fall. Those that make it all the way to Earth’s surface are known as “meteorites.” They can be often be distinguished by their scalloped fusion crust, as with this sample:



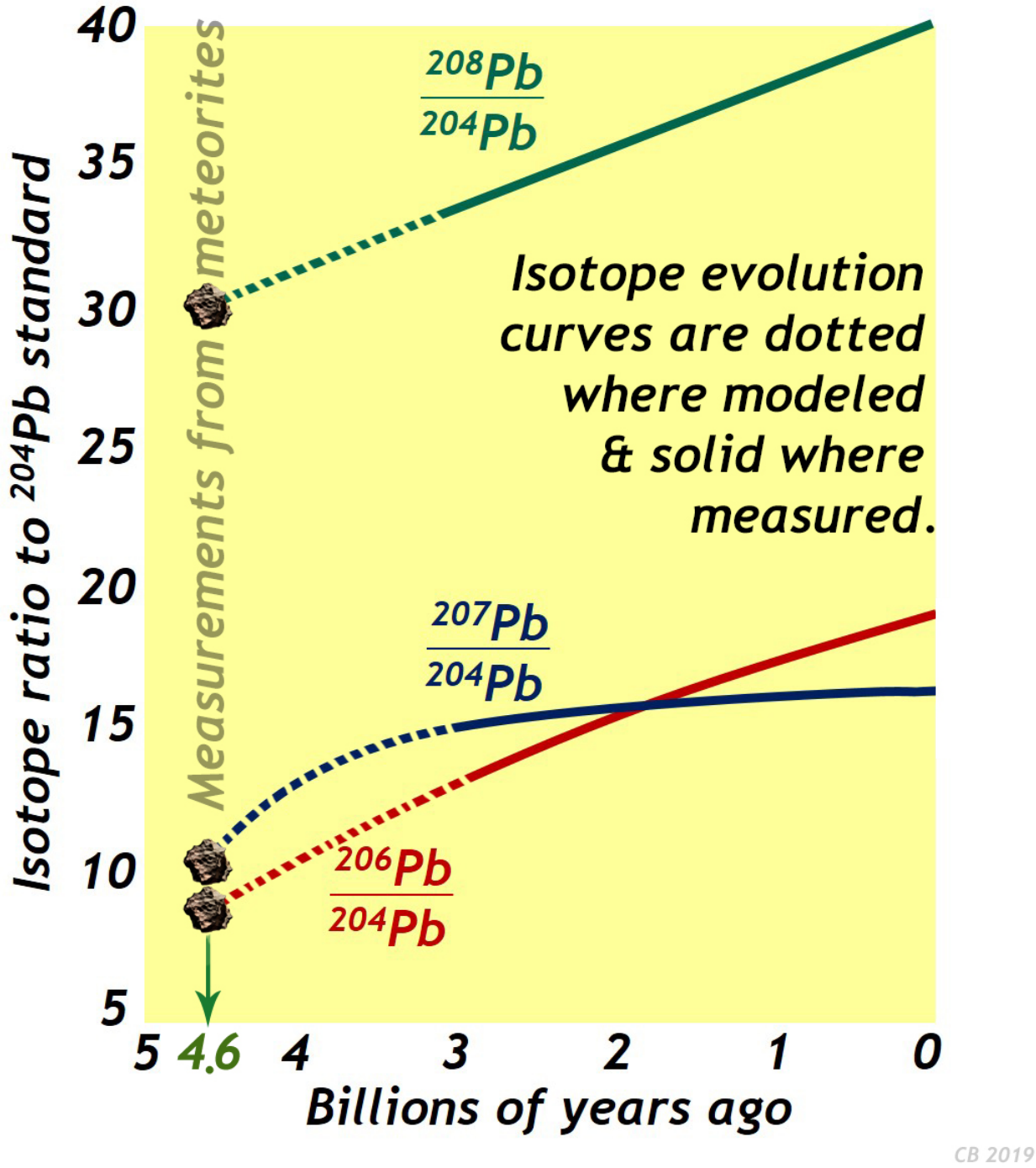


Figure 16.4.2: Lead (Pb) isotope ratio evolution over the past 5 billion years, including terrestrial (Earth rock) measurements and projections of primordial evolution, and values derived from measurement of meteorites.

Meteorites come in several varieties, including rocky and metallic versions. It is very satisfying that when measurements of these meteorites' lead isotopes are added to the plot above, they all fall exactly where our understanding of lead isotope production would have them: at the start of each of these model evolution curves. Each lead isotope system tells the same answer for the age of the Earth, acting like three independent witnesses corroborating one another's testimony. And the answer they all give is **4.6 billion years ago** (4.6 Ga). That's what ^{208}Pb says. That's what ^{207}Pb says. And that's what ^{206}Pb says. They all agree, and they agree with the predicted curves based on terrestrial (Earth rock) measurements. This agreement gives us great confidence in this number. The Earth, and meteorites (former asteroids), and the solar system of which they are all a part, began about 4.6 billion years ago...

Did I Get It? - Quiz

? Exercise 16.4.1

How old is the solar system?

a. 6.4 million years.

b. 6.4 billion years.

c. 4.6 million years.

d. 4.6 billion years.

e. 6000 years.

Answer

d. 4.6 billion years.

? Exercise 16.4.2

Isotopes of what element are used to calculate the age of Earth (and meteorites) ((and the rest of the solar system))?

a. Uranium

b. Carbon

c. Hydrogen

d. Lead

Answer

d. Lead

...But what came before that?

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16.5: The implications of meteorites

In 1969, a meteorite fell through Earth's atmosphere and broke up over Mexico. A great many pieces of this meteorite were recovered and made available for scientific analysis. It turned out to be a carbonaceous chondrite, the largest of its kind ever documented. It was named the Allende (*"eye-YEN-day"*) meteorite, for the tiny Chihuahuan village closest to the center of the area over which its fragments were scattered.

One of the materials making up Allende's chondrules was the calcium feldspar called anorthite. Anorthite is an extraordinarily common mineral in Earth's crust, but the Allende anorthite was different. For some reason, it has a large amount of magnesium in it. When geochemists determined what kind of magnesium this was, they were surprised to find that it was mostly $^{26}\mathrm{Mg}$, an uncommon isotope. The abundances of $^{25}\mathrm{Mg}$ and $^{24}\mathrm{Mg}$ were found to be about the same level as Earth rocks, but $^{26}\mathrm{Mg}$ was elevated by about 1.3%. And after all, magnesium doesn't even "belong" in a feldspar. The chemical formula of anorthite is $\mathrm{CaAl_2Si_2O_8}$ – there's no "Mg" spot in there. Why was this odd $^{26}\mathrm{Mg}$ in this chondritic anorthite?

One way to make $^{26}\mathrm{Mg}$ is the break-down of radioactive $^{26}\mathrm{Al}$. The problem with this idea is that there is no $^{26}\mathrm{Al}$ around *today*. It's an example of an extinct isotope: an atom of aluminum so unstable that it falls apart extremely rapidly. The half-life is only 717,000 years. But because these chondrules condensed in the earliest days of the solar system, there may well have been plenty of $^{26}\mathrm{Al}$ around *at that point* for them to incorporate. And Al, of course, is a key part of anorthite's $\mathrm{CaAl_2Si_2O_8}$ crystal structure.

So the idea is that weird extra $^{26}\mathrm{Mg}$ in the chondrule's anorthite could be explained by suggesting it wasn't always $^{26}\mathrm{Mg}$: Instead, it started off as $^{26}\mathrm{Al}$,and it belonged in that crystal's structure. However, over a short amount of time, it all fell apart, and that left the $^{26}\mathrm{Mg}$ behind to mark where it had once been. If this interpretation is true, it has shocking implications for the story of our solar system.

To understand why, we first need to ask, what came before the nebula? What was the 'pre-nebula' situation? Where did the nebula come from, anyhow?

It turns out that nebulae are generated when old stars of a certain size explode.

Figure 16.5.1: Light traveling outward after a nova-like explosion from the star V838 Monocerotis. (via GIPHY)

These explosions are called supernovae (the plural of supernova). The "nova" part of the name comes from the fact that they are very bright in the night sky – an indication of how energetic the explosion is. They look like "new" stars to the casual observer. Supernovae occur when a star has exhausted its supply of lightweight fuel, and it runs out of small atoms that can be fused together under normal conditions. The outward-directed force ceases, and gravitationally-driven inward-directed forces suddenly dominate, collapsing the star in upon itself. This jacks up the pressures to unbelievably high levels, and is responsible for the nuclear fusion of big atoms – every atom heavier than iron is made instantaneously in the fires of the supernova.

That suite of freshly-minted atoms included a bunch of unstable isotopes, including $^{26}\mathrm{Al}$.

And here's the kicker: If the ^{26}Al was made in a supernova, started decaying immediately, and yet enough was around that a significant portion of it could be woven into the Allende chondrules' anorthite, that implies a *very short amount of time* between the obliteration of our Sun's predecessor, and the first moments of our own. Specifically, the 717,000 year half-life of ^{26}Al suggests that this "transition between solar systems" played out in less than 5 million years, conceivably in only 2 million years.

That is very, very quickly.

Did I Get It? - Quiz

? Exercise 16.5.1

How much time passed between the supernova that ended the Sun's predecessor star (and solar system) and the condensation of chondrules in the beginning of our own solar system?

a. Probably about 1.2 billion years.

b. Probably less than 5 million years.

c. Probably about 10 years.

Answer

b. Probably less than 5 million years.

? Exercise 16.5.2

The isotope ^{26}Al is _____.

a. Produced through thermonuclear fusion in our Sun.

b. Extinct; it's so radioactive that all of the initial population has decayed.

c. Very massive and heavy; triggering gravitational attraction of the dust in the presolar nebula.

Answer

b. Extinct; it's so radioactive that all of the initial population has decayed.

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16.6: Further reading

Marcia Bjornerud's book *Reading the Rocks*. Basic Books, 2005: 226 pages.

Jennifer A. Johnson (2019), "Populating the periodic table: Nucleosynthesis of the elements," *Science*. 01 Feb 2019 : 474-478.

Lee, T., D. A. Papanastassiou, and G. J. Wasserburg (1976), Demonstration of ^{26}Mg excess in Allende and evidence for ^{26}Al, *Geophysical Research Letters*, 3(1), 41-44.

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16.S: Summary

In summary, the planet Earth is part of a solar system centered on the Sun. This solar system, with its star, its classical planets, its dwarf planets, and its “leftover” comets and asteroids, formed from a nebula full of elements in the form of gas and dust. Over time, these many very small pieces stuck together to make bigger concentrations of mass, eventually culminating in a star and a bunch of planets that orbit it. Asteroids (and asteroids that fall to Earth, called meteorites), are leftovers from this process. The starting nebula itself formed from the destruction of a previous star that had exploded in a supernova. The transition from the pre-Sun star to our solar system took place shockingly rapidly.

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CHAPTER OVERVIEW

17: (Case Study) Earth's oldest rocks

When did Earth really become Earth? We've learned about how the solar system formed and the earliest stages in Earth formation in the case study exploring the nebular theory and the formation of the solar system. Early Earth grew through the process of collision and accretion of nebular material that ranged in size from space dust to planetesimals, perhaps some as large as small planets. We have dated Earth at 4.567 billion years (Ga) based on lead isotope data from meteorites [1]. Meteorites are the "ingredients" that formed Earth and the other terrestrial planets. This "birth of Earth" date is assigned to the onset of accretion, not at a point in which Earth had largely attained its full, accreted mass [2], [3].

17.1: The Hadean

17.2: Events that shaped the Hadean

17.3: Earth's Oldest Rocks

17.4: Earth's oldest fossils?

17.5: References and Further Reading

17.S: Summary

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17.1: The Hadean

Earth's first eon of geologic time is appropriately named the "Hadean" with reference to Hades, the God of the Underworld in Greek mythology. The Hadean is often described as "Hell on Earth," the time when an extremely hostile environment existed with magma oceans boiling on the surface and noxious gases and steam enveloping the young planet. Perhaps the accompanying picture is an analogy. In this extreme environment, no life, as we know it, could possibly survive.

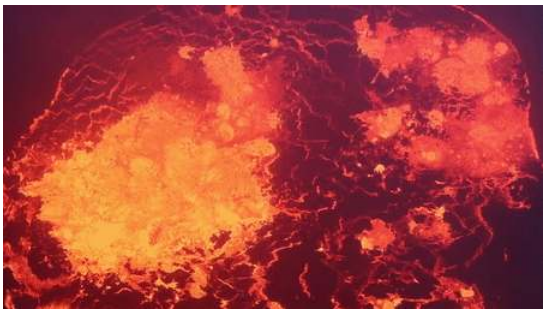


Figure 17.1.1: A lava lake in the Hawaiian volcano, Kilauea. This may be analogous to the magma ocean that existed on Earth's surface during the Hadean. (USGS public domain.)

So when can we say, "there you have it – Earth is now Earth!"? Perhaps if we could find the first rock formed on Earth, we could assign Earth's true age as the inception of a solid surface.

The geologic time span of Hadean contains no Eras or Periods because we know so little about what happened as only trifle bits of Earth material remain. The Hadean as it appears on the geological time scale below indicates that its beginning is approximate. The little clock located at the divisions of time on the right side of the scale above the Hadean indicate that the dates chosen are somewhat arbitrary as geologists continue to examine remnant mineral and rock evidence.

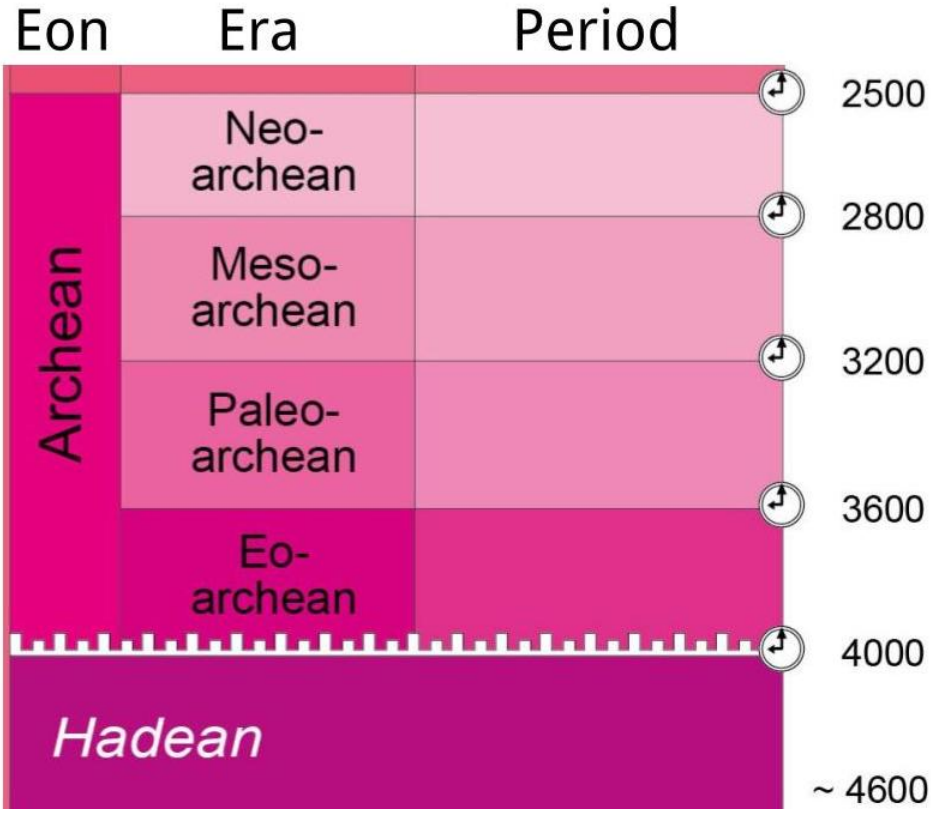


Figure 17.1.2: Portion of the Geologic Times Scale including the Hadean and Archean Eons of Geologic Time. (Permission granted for educational use. Cohen, K.M., Harper, D.A.T., Gibbard, P.L. 2020. ICS International Chronostratigraphic Chart 2020/01. International Commission on Stratigraphy, IUGS. www.stratigraphy.org.)

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17.2: Events that shaped the Hadean

Differentiation of Earth's Interior

What evidence do we have of Earth shaping events occurring in the Hadean? Planetary geologists theorize that toward the end of the accretionary stage in Earth formation, 4.5 – 4.6 Ga, Earth was pummeled by larger, planetesimal-sized space debris. Planetesimals were other "wanna-be" planets traveling in the same orbit as the developing Earth. They were massive in their own right, but the developing Earth was larger and able to withstand their impacts and remain intact. The energy of these massive collisions melted the surface of the young planet resulting in the formation of vast "magma oceans." Energetic radioactive decay of unstable elements added to the heat production from inside the early Earth. This double whammy of interior and exterior heat generation may have melted the entire planet, or turned it into a thick, slushy mass of highly convective molten rock material.[1], [2]

This allowed the Earth, and other developing planets in our solar system, to go through a process of differentiation where the heaviest, siderophile (iron associated) elements migrated toward the central core while the lighter lithophile (lithosphere concentrated) elements rose toward the surface. Differentiation happened very quickly with respect to the immensity of geologic time – over the span of several tens of millions of years. Differentiation was not a wholesale turnover of the entire interior of the planet but more like a percolation of siderophile elements through the slushy magma ocean (see image below).

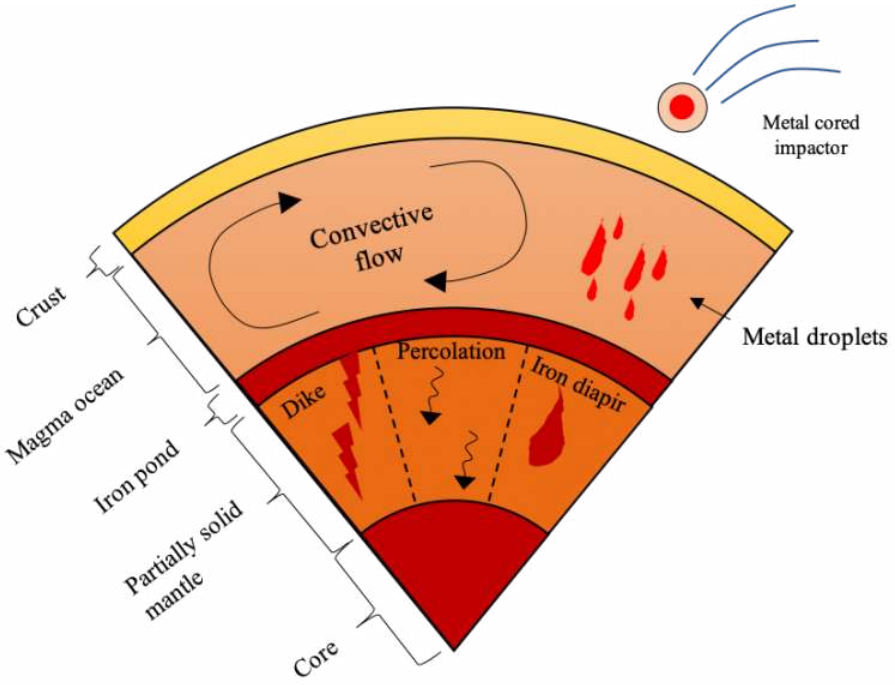


Figure 17.2.1: Hypothetical core-mantle differentiation processes: Percolation, diking, and diapirism. After Rubie et al. (2015). (CC BY-SA 4.0; By AlexInMetal – Own work, on Wikimedia.)

Analysis of seismic waves produced by earthquakes has helped us understand the density differences and relative composition of the materials that exist within the modern Earth at different depths. These seismic wave velocity analyses over almost the past 100 years have allowed us to see a very clear picture of Earth's layered interior [4].

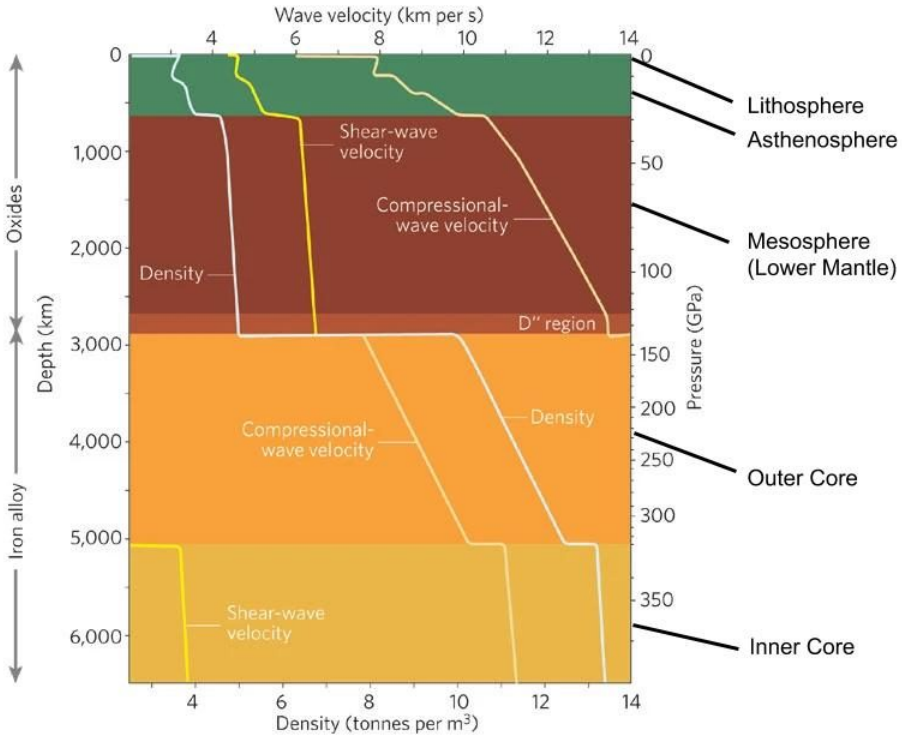


Figure 17.2.2: The radial structure of Earth's interior. The diagram details the concentric shells within Earth and their approximate composition. These were established over the first half of the twentieth century from measurements of the travel times of seismic waves refracted and reflected inside Earth. Each abrupt shift in seismic wave velocity a phase change (liquid/semi-solid, solid), and a compositional change. (Modified from [5].)

Earth's first crust

As differentiation proceeded, the light-element-enriched magma ocean at the surface was exposed to the chill of space and a thin, early crust began to form. Meteors, asteroids and comets continued their impact unabated, puncturing the earliest crust and allowing the magma, hidden just beneath, to flow again. The composition of this earliest crust was similar to very rare ultramafic rock komatiite, a volcanic rock largely composed of the mineral olivine. Komatiites are very rare in Earth's rock record, almost entirely restricted to rock of Archean age.[6] The komatiites were largely derived from the very hot mantle of the young Earth, and pooled on the surface as lava. As Earth continued to cool and the mantle solidified, generation of highly ultramafic magma that has made its way to the surface has been rare.[7]



A quick review of magma composition

Differentiation produced our iron-dominated inner and outer core and silicate-dominated rock material of the mantle and ultimately, the crust. Silicates are minerals that are largely composed of the elements of silicon and oxygen bound together with other lithophile elements including aluminum, calcium, potassium, sodium, and magnesium. Not all of the iron migrated to the core. In the early Hadean, the Earth was continually being bombarded by space debris that contained all of Earth's natural elements in varying percentages. We can think of the composition of the magma ocean as being silica (silicon and oxygen) dominated and

relatively homogeneous. The first crust, composed of komatiite, represented this composition. This convecting ocean of silicate mush brought diapirs (rising teardrop shaped blobs) of deep magma ocean material toward the surface where it pooled and cooled in thick komatiite masses. These heavy, dense komatiite masses dripped back into the magma ocean for recycling. This remelting process further differentiated the early crust. Eddys of convecting magma near the surface partially melted the komatiite. This "partial melt" was more silica enriched because the minerals in the komatiite with the most silica would be first to melt (see discussion on Bowen's reaction series to understand this process). This is referred to as an "evolved magma," not primary, but remelted and more silica enriched. Silica enriched magmas are lower in density and more buoyant and can be envisioned as rising to the surface like ice cream in a rootbeer float. This evolution of magma leads to the varying compositions we recognize today in the classification of igneous rock. Notice where komatiite appears on the diagram below. As magma evolution proceeds, the composition becomes more silica enriched, creating magma that becomes increasingly more "felsic" in composition.

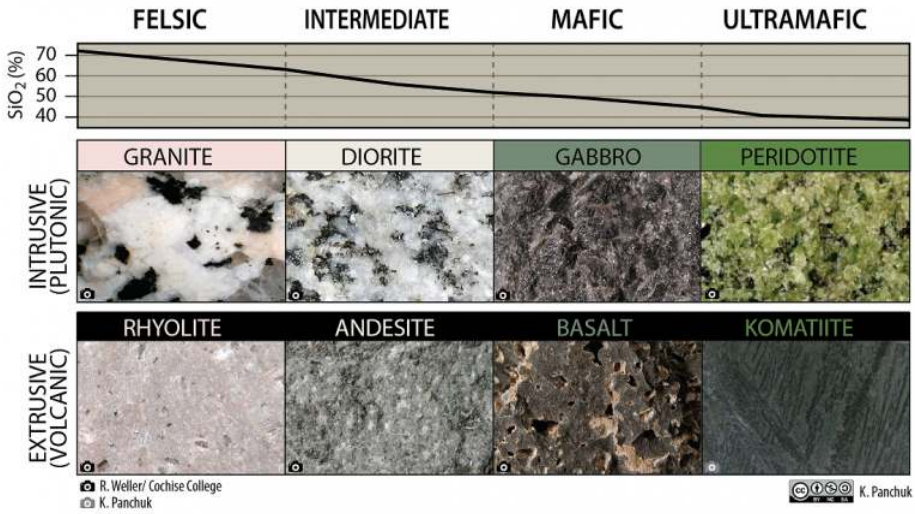


Figure 17.2.3: Igneous rock classification diagram.

Did I Get It? - The Hadean Quiz

? Exercise 17.2.1

The early Earth grew by adding space debris and planetesimals through the process known as

a. convection

b. differentiation

c. accretion

d. percolation

Answer

c. accretion

? Exercise 17.2.2

During the process of differentiation, the _____ elements migrated toward Earth's core while the _____ elements rose toward Earth's surface.

a. siderophile, lithophile

b. light, heavy

c. least dense, most dense

d. lithophile, siderophile

Answer

a. siderophile, lithophile

Exercise 17.2.3

Earth's first crust had the composition of which type of igneous rock?

a. granite

b. komatiite

c. basalt

d. felsic

Answer

b. komatiite

Exercise 17.2.4

What have scientists studied for more than 100 years to learn about the layered interior of Earth?

a. komatiites

b. rock from the core

c. images taken of Earth from space

d. seismic waves from earthquakes

Answer

d. seismic waves from earthquakes

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17.3: Earth's Oldest Rocks

The Jack Hills Zircon

What followed the formation of thin, perturbed, komatiite crust is highly debated and an area of fervent geological research. Insight into the evolution of the earliest crust was unknown until the discovery of Hadean age zircons from the Jack Hills area in southwest Australia (see map). Zircon is a small, durable mineral with the chemical formula $ZrSiO_4$. It is a common mineral in felsic rock with the composition similar to that of the continental crust, like granite. It is not a common mineral of the rock of the ocean crust, like basalt and does not occur in the earliest rock of the crust, komatiite.

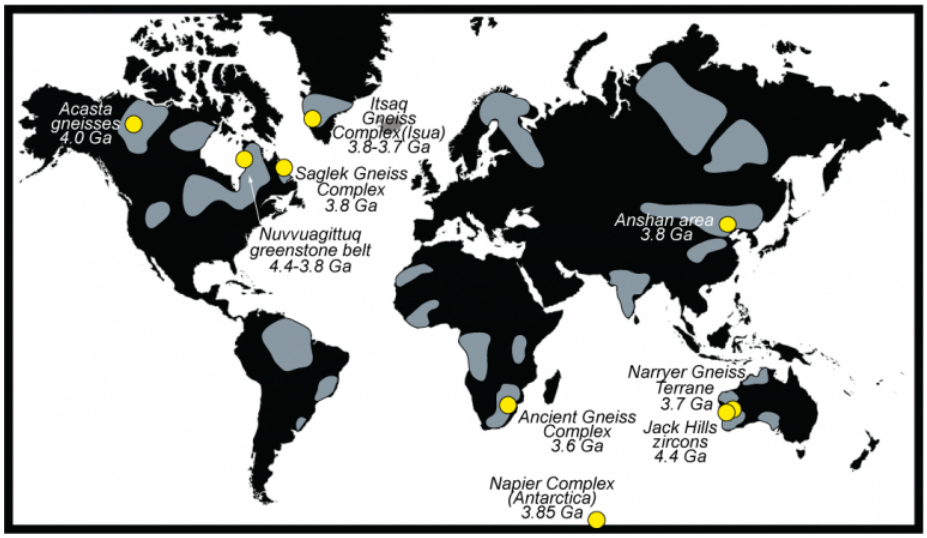


Figure 17.3.1: Archean Cratons (light grey) and locations of Earth's oldest rocks on a world map. By Jonathan O'Neil. (With permission from: UOttawa Early Earth Blog)



Figure 17.3.2: Quartz-pebble metaconglomerate (Jack Hills Quartzite, Archean, 2.65 to 3.05 Ga; Jack Hills, Western Australia) (Cranbrook Institute of Science collection, Bloomfield Hills, Michigan, USA) (CC BY 2.0; James St. John on Flickr)

In the mid-1980s field geologists sampled a metamorphosed sedimentary rock (metaconglomerate) from the Jack Hills. The metaconglomerate is dated as Archean, approximately 3.6 Ga. The detrital sedimentary particles found within the metaconglomerate are, of course, older than their deposition as gravel. The environment of deposition of the original conglomerate is thought to be an alluvial fan, where sediment derived from the weathering and erosion of mountains were moved by water and deposited in an adjacent valley. Small zircon crystals were extracted from the metaconglomerate for analysis. The zircons were originally part of the rock that composed these earlier mountains. Remnants of these mountains have not been discovered and have most likely been erased from Earth's surface by erosion.

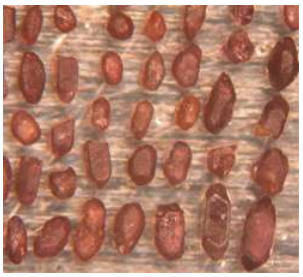


Figure 17.3.3: This image shows an array of Jack Hill zircons arranged on a piece of tape. Largest zircons ~ 1mm. (CC BY-NC-SA 3.0; Science Education Resource Center)

Zircon is a small but mighty mineral. It is one of Earth's little timekeepers. Zircon typically forms during the crystallization of magma where radioactive uranium can substitute for zirconium in the mineral lattice. Following crystallization, the radiometric clock starts ticking. The unstable radioactive uranium atoms break down through a process known as "decay." The atoms lose subatomic particles and emit energy. Particle loss includes a decrease in the number of protons which ultimately changes the uranium to lead. The rate of this decay is well known and allows scientists to very accurately date the zircon. Radiometric dating analysis of the Jack Hills detrital zircon grains yield dates as old as 4.404 Ga! This is the oldest Earth material discovered to date, formed merely ~150 Ma after the inception of Earth. That is amazing!!! This tells us that in those 150 million years, the entire Earth was largely formed, its interior differentiated, it cooled enough to have a solid crust of komatiite, and, remelting produced evolved magma where the zircon crystallized in solid rock similar to the composition of our continental crust today. WOW!!! (The method of dating the zircon is discussed more fully in the chapter on Geologic Time).

This discovery has significantly advanced our understanding of the Hadean environment and the evolution of the crust during the Hadean. The discovery of the zircon means that within the first several hundred million years of Earth's existence, crust of varying composition existed, some of which was more felsic in composition, very much like the continental crust that exists today. This provides significant insight into the processes at work in the Hadean as this type of rock is continuously being formed and reformed through plate tectonic processes today.

Analysis of oxygen isotopes data from the zircon have revealed even more incredible evidence about the environment that existed during the very early Hadean. Oxygen has several isotopes, ^{16}O is most abundant with 8 protons and 8 neutrons existing in the nucleus. ^{17}O and ^{18}O also exist in much less abundance. ^{18}O is concentrated in an aqueous environment, such as oceans, as the lighter ^{17}O and ^{16}O will preferentially evaporate. Analysis of the relative amounts of different isotopes of oxygen ^{16}O and ^{18}O (ratio denoted with the lowercase Greek delta $\delta^{18}O$) in the Jack Hills zircon are skewed toward "heavy" ^{18}O, as opposed to the more common "light" ^{16}O. This heavy oxygen signature in rock is an indication that it formed by cool, wet, sedimentary processes at the Earth's surface. Thus, the magma that eventually gave rise to the zircons is theorized to have been formed from what had once been sediments deposited on the floor of an ancient ocean [3]. So not only was the very young Earth capable of making felsic-composition crust, it also was cool enough to have liquid water in oceans. These are surprisingly familiar conclusions about a "hellish" young planet.



Figure 17.3.4: Hadean age zircon discovery locations on a world map. (CC BY 4.0 International; By karaclc on Wikimedia.)

Since the discovery of the Hadean age zircon in the Jack Hills area of Australia, other detrital zircons have also been discovered in Archean age rock in other parts of the world. See the map at right.

Did I Get It? - Jack Hills Zircon Quiz

? Exercise 17.3.1

The mineral zircon is commonly found in which type of igneous rock?

a. komatiite

b. ultramafic

c. basalt

d. granite

Answer

d. granite

? Exercise 17.3.2

How old are the zircon from the Jack Hills of southwest Australia?

a. 4.040 Ga

b. 4.040 Ma

c. 4.404 Ma

d. 4.404 Ga

Answer

d. 4.404 Ga

? Exercise 17.3.3

Which of the following statements is NOT true?

a. The Jack Hills zircon were dated through a process that uses the decay of radioactive elements.

b. The Jack Hills zircon crystals were extracted from sedimentary rock, not from the original rock in which they formed.

c. The Jack Hills zircon crystals originally crystallized in the Earth's first komatiite crust.

d. Oxygen isotope analysis of the Jack Hills zircon crystals indicate that water was present on Earth's surface at their time of formation.

Answer

c. The Jack Hills zircon crystals originally crystallized in the Earth's first komatiite crust.

The Acasta Gneiss Complex

While the Jack Hills detrital zircons tell us that rocks were definitely around at 4.404 Ga, actual rocks that old have not yet been found. In fact, some controversy exists in the scientific community as to the oldest rocks discovered to date. The Acasta Gneiss Complex located in the Slave Craton of Northwest Canada (see map of Archean Cratons, above) preserve the oldest unquestioned isotopically dated rock thus found anywhere on Earth. This complex includes a variety of highly deformed and metamorphosed tonalite-trondhjemite-granodiorite (TTG) rock. This type of rock is discussed in more detail in the Case Study on Greenstone Belts, Primordial Tectonics. TTGs are similar to granite with some chemical and mineralogical differences. They are typical igneous rocks produced as intrusive bodies along tectonically active continental margins today.



Figure 17.3.5: A fragment of the Acasta Gneiss, the oldest known rock on our planet. In exhibition at the Natural History Museum in Vienna. (CC BY-SA 3.0; By Pedroalexandrade on Wikimedia.)

The Acasta Gneiss has been dated using U/Pb isotopic dating techniques on zircon formed during the crystallization of these rocks in their original igneous environment. The presence of zircon in an igneous rock indicates that the magma has “evolved” – it was formed through the remelting of a pre-existing rock. The isotopic dates yield a record of several different intrusive magmatic events ranging in age between 2.94 Ga and 4.02 Ga. The oldest episode of igneous activity recorded in the gneiss complex occurred between 3.92 and 4.02 Ga which straddles the arbitrary Hadean/Eoarchean division on the geologic time scale (above).[8]



Figure 17.3.6: Acasta gneiss in exposure. Notice the tight folds from tectonic deformation. (CC BY NC ND 4.0 International; By: The Traveling Geologist)

Interpretation of the Acasta Gneiss zircon ages provides verification that continental crust existed in the Hadean. Extensive geochemical analysis of many different isotopic pairs and element/isotope concentration comparisons have led to the interpretation that the oldest rock of the Acasta Gneiss Complex was derived from the partial melting of mafic Hadean crust that was 4.3 billion years old [8] [9].

The Nuvvuagittuq Greenstone Belt

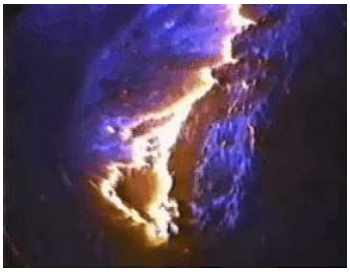


Figure 17.3.7: Modern pillow lava forming underwater.

Dating of the rock from the Nuvvuagittuq Greenstone Belt (NGB) has been controversial but, if accepted, may actually provide a glimpse of Earth's earliest crust. (Read more about Greenstone Belts here). The NGB is located in northern Quebec, on the eastern shore of the Hudson Bay (See Archean Craton map above). Metamorphosed mafic and ultramafic volcanic rock dominate the NGB. The fact that these are volcanic rocks erupted in an ancient ocean is documented by the presence of lava pillows. Lava that erupts underwater forms conspicuous pillow shapes as a crust solidifies instantly around the oozing lava as it spills out underwater.

Another rock type included in the NGB is banded iron formation. BIFs (as they are affectionately known) are sedimentary rocks that also indicate an ocean environment as these sedimentary iron minerals form and settle out of the water column. (Read more about the formation of BIFs here). Intrusive mafic and ultramafic dikes also occur within the NGB.



Figure 17.3.8: Exposure of the metamorphosed rocks of the Nuvvuagittuq Greenstone Belt from the Porpoise Cove locality, Quebec, Canada. Some of these rocks have Sm/Nd ages in excess of 4.0 Ga and may be the oldest rocks on Earth. (Public Domain; NASA.)

The NGB is bounded by a felsic intrusive rock known as a tonalite (see map below). Both the NGB and the tonalite are cut across by a similar type of felsic rock called a pegmatite. The tonalite and pegmatite intrusive bodies contain zircon needed for U/Pb radiometric dating which has provided an accurate date of 3.77 Ga. Due to the cross-cutting relationship of the intrusive rock to the NGB, this provides only a **minimum** age for the NGB of 3.77 Ga [10] [11]. Further investigation is necessary to understand the **maximum** age.

The greenstone rocks of the NGB are remnants of ancient ocean crust with a mafic to ultramafic composition (see composition chart above). Rocks of ultramafic composition do not contain zircon and mafic rocks rarely do. Zircon is a mineral that is typically found in igneous rock more felsic in composition, like granite. Uranium-lead (U/Pb) isotopic dating is the “gold standard” for accuracy, particularly in very old rock. A lack of zircon in the NGB will call into question any other method used to provide dates.



Figure 17.3.9: Exposure of the rock in question as the oldest rock found to date from the Nuvvuagittuq Greenstone Belt. It is described as a “faux” amphibolite – a basalt rock that has been subjected to medium grade metamorphism. (With permission by: Jonathan O’Neil from: Guest Post: Oldest Rocks Could Weigh A Man Down)

In 2008, Jonathan O’Neil, a young PhD student from McGill University in Quebec, Canada, challenged the assumed date for NGB by investigating an odd looking amphibolite patch in the NGB (picture). An amphibolite is a metamorphosed mafic rock therefore, zircon is not present. O’Neil used a dating technique and isotopic ratio comparisons of the rare but ubiquitous elements of samarium (Sm) and neodymium (Nd). Using this technique, O’Neil determined the actual age of the NGB to be 4.28 Ga [12]. Further analysis of additional samples since then has pushed the date back even further to 4.31 Ga [11]. If true, the NGB rock would represent the oldest preserved Hadean age rock found on the planet.

Not all geologists investigating these rocks concur with O’Neil’s findings. Their interpretation of the isotopic data and comparisons propose that the isotopic signatures represent only that the NGB was derived from pre-existing rock of Hadean age and that the NGB age of formation is Eoarchean (3.7 Ga). This controversy is sure to continue until zircon is found within the NGB. Are you intrigued by the search for the Earth’s oldest rocks? Consider pursuing graduate study in geology. Much of the NGB extent has yet to be sampled.

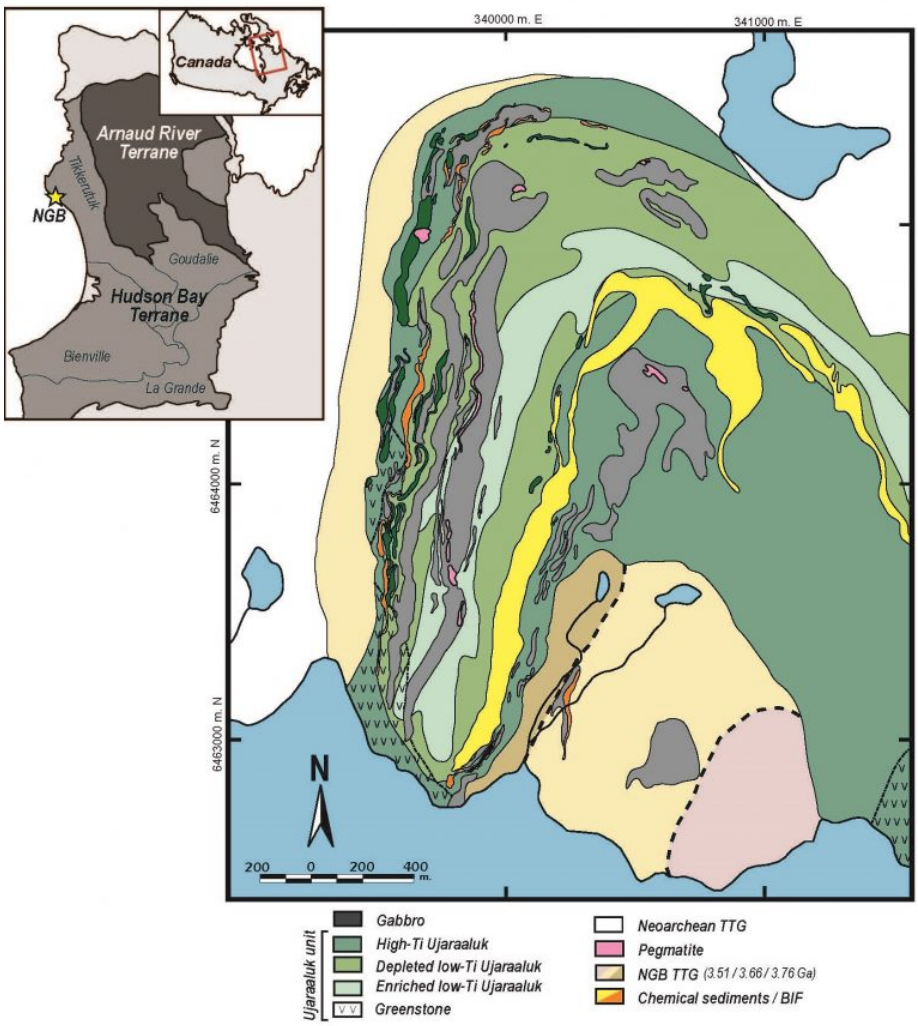


Figure 17.3.10: Geologic map of the Nuvvuagittuq Greenstone Belt (with permission by Jonathan O'Neil.)

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17.4: Earth's oldest fossils?

The Nuvvuagittuq Greenstone Belt continues to give. In 2017 a team of researchers presented their findings on what they believe to be traces of life found in the banded iron formation of the NGB [13]. Their microscopic investigation of the BIF revealed small tubes made of the iron mineral hematite. These tubes match the shape and size of those made by bacteria in modern hydrothermal vent environments. The bacteria use the iron for their metabolism. The researchers also discovered the mineral graphite in the BIF which is composed entirely of carbon. The graphite likely formed by metamorphism of organic material as it contains reduced levels of the heavy carbon isotope, carbon-13 (^{13}C). Lifeforms are preferential in selection of the lighter isotope of carbon, carbon-12 (^{12}C).

If this discovery holds true, this will be the oldest evidence of fossil life yet discovered. The NGB banded iron formation exists within the NGB and can be interpreted as having been deposited on the ancient sea floor represented by the greenstone. This NGB has been unquestionably dated by cross-cutting igneous intrusions at 3.77 Ga. This would pre-date any fossil evidence previously discovered by several hundred million years (see In the Beginning). If O'Neil's age for the NGB holds up, then the fossils may perhaps be as old as 4.31 Ga.



Did I Get It? - Acasta Gneiss and Nuvvuagittuq Greenstone Belt Quiz

? Exercise 17.4.1

Is the Acasta Gneiss Complex Hadean in age?

a. No, none of the samples from Acasta Gneiss Complex date to the Hadean, they are 3.92 Ga which is Eoarchean in age.

b. Yes, some of the samples of the Acasta Gneiss Complex have been dated to 4.04 Ga which is in the Hadean Eon of geologic time.

Answer

b. Yes, some of the samples of the Acasta Gneiss Complex have been dated to 4.04 Ga which is in the Hadean Eon of geologic time.

? Exercise 17.4.2

How do we know that the Acasta Gneiss Complex is Hadean in age?

a. The Acasta Gneiss rocks have been isotopically dated using zircon crystals.

b. The Acasta Gneiss is composed of TTG rock which is from Earth's first crust.

c. The Acasta Gneiss contains fragments of the Earth's first komadiite crust.

Answer

a. The Acasta Gneiss rocks have been isotopically dated using zircon crystals.

Exercise 17.4.3

How do we know that the greenstone from the Nuvvuagittuq Greenstone Belt is rock from the ocean crust?

a. The greenstone contains zircon crystals which are common in rock of the ocean crust.

b. The greenstone contains pillow lava structures which indicate the eruption of lava underwater.

c. The greenstone is composed of TTG rock which is the most common rock of the ocean crust.

Answer

b. The greenstone contains pillow lava structures which indicate the eruption of lava underwater.

Exercise 17.4.4

Why is the age of the Nuvvuagittuq Greenstone Belt (NGB) questioned?

a. The location of the NGB is so remote that other scientists have not been able to investigate the rock and validate the claim.

b. Zircon has not been found in the greenstone rock of the NGB therefore U/Pb isotopic dating cannot be used which is the "gold standard" in dating very old rock.

c. The team of NGB geologists have not collected enough samples with zircon therefore the isotopic dates provided are not reliable.

Answer

b. Zircon has not been found in the greenstone rock of the NGB therefore U/Pb isotopic dating cannot be used which is the "gold standard" in dating very old rock.

Exercise 17.4.5

Which rock type of the NGB contains fossil structures?

a. greenstone

b. TTG

c. banded iron formation

d. amphibolite

Answer

c. banded iron formation

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17.S: Summary

The oldest minerals from Earth's crust yet discovered are the zircons found in Archean metamorphosed sedimentary rock from the Jack Hills of southwestern Australia. Analysis of the zircon consistently provides dates over 4.0 Ga with the oldest being 4.4 Ga. The rock in which the zircon originally formed would have been some of the very oldest continental crust to exist; unfortunately, the original rocks that supplied these zircons are most likely long gone. The oldest intact rock found on Earth to date is from the Acasta Gneiss Complex of northwest Canada. U/Pb dates of zircon from the gneiss reach into the Hadean Eon at 4.02 Ga. Rock from the Nuvvuagittuq Greenstone Belt in northern Quebec, Canada does not contain zircon as it is mafic to ultramafic in composition and represents ancient ocean crust. Alternate isotopic dating techniques yield a much older reach back into the early Hadean of 4.31 Ga. Controversy within the scientific community does exist concerning this date and investigation continues.

Geologists continue to search Archean cratons for slivers of Hadean crust. More discoveries are assured which will continue to advance our understanding and knowledge of the environment of the earliest time of Earth.

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CHAPTER OVERVIEW

18: (Case Study) Greenstone belts - primordial tectonics

18.1: Plate tectonics - what governs Earth's surface dynamics today

18.2: It was different in the Archean

18.3: Rocks found in greenstone belts

18.4: Structure of greenstone belts

18.5: Further reading

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18.1: Plate tectonics - what governs Earth's surface dynamics today

As we have covered elsewhere, the way the world works today is largely governed by plate tectonics: the differential movement of large plate-like slabs of the planet's lithosphere, mainly horizontally, but with significant vertical motion too at sites of subduction and continental collision. The fathomless deep sea trench or a towering Himalayan-style mountain range are ephemeral topographic features, however. While they allow us to spot plate boundaries at a glance today, they don't last over geologic time: the trench gets filled in with sediment; the mountains are eroded to their roots.



Figure 18.1.1: A key feature of plate tectonics is the subduction of cold oceanic crust, leading to a high-pressure & low-temperature metamorphic regime. Blueschist is evidence of those conditions, which we would not expect in the absence of plate tectonics. (Callan Bentley photo)

So how can we figure out when plate tectonics began on this planet? We must look to other aspects of the geologic record. This is a vibrant area of research in geology now, and there is genuine scientific controversy about how best to interpret a patchy record. This much is agreed upon, however: Plate tectonics results in several features that are decisive "signatures" of its operation, and are potentially "durable" enough to persist in the geologic record through time. First is the production of **blueschist** as an example of a high-pressure but low-temperature metamorphic rock, only results when you have the subduction of old, cold, dense, oceanic crust. The oldest known blueschist on Earth is merely Neoproterozoic in age (Condie and Kröner, 2008, and references therein).

Second, blueschist is usually associated with other subduction metamorphic rocks, such as eclogite. Distinctive minerals, including glaucophane (a blue amphibole mineral that makes blueschist **blue**) and lawsonite mark these rocks. But the belt of "subduction metamorphic rocks" is paralleled by a second belt of metamorphic rocks, with a higher temperature signature, the rocks forming in association with the accompanying volcanic arc. These include classic Barrovian metamorphic minerals (sillimanite, staurolite, garnet, etc.) in rocks such as phyllite, schist, and gneiss.

The set of **two parallel belts with different temperature/pressure regimes** is the key observation: that's how it works today, with plate tectonics. And unlike the deep trench or the towering volcanic arc, the paired belts of rock with distinct metamorphic signatures stand a good chance of being preserved in the geologic record.

But as we go further and further back in the geologic record, we see fewer and fewer blueschists, and fewer and fewer paired metamorphic belts, forcing us to conclude that plate tectonics was not always the way the Earth's outer rocky layer behaved.

Did I Get It? - Quiz

? Exercise 18.1.1

Though some manifestations of plate tectonics are topographic (*e.g.*, trenches, collisional mountain belts), others are signatures more likely to be preserved through geologic time. Which of the following is considered a durable signal of plate tectonics we might expect to "read" from the ancient geologic record?

a. Blueschist, a high-pressure, low-temperature metamorphic rock

b. Granite, a coarse-grained, felsic-composition igneous rock

c. Basalt, a fine-grained, mafic-composition igneous rock

d. Hornfels, a high-temperature, low-pressure metamorphic rock

Answer

a. Blueschist, a high-pressure, low-temperature metamorphic rock

? Exercise 18.1.2

Why are paired metamorphic belts considered significant?

a. Different settings in plate tectonics produce different combinations of temperature and pressure conditions. These result in different suites of metamorphic rocks forming at the same time, in adjacent settings. In plate tectonics, we would expect a high-pressure, low-temperature belt associated with subduction to be paired with (not too far away) a parallel higher-temperature belt associated with a volcanic arc.

b. Because over time different belts of metamorphic rock tend to pile up together, starting off in wildly different locations at different times, and plate tectonics eventually brings them together in pairs.

Answer

a. Different settings in plate tectonics produce different combinations of temperature and pressure conditions. These result in different suites of metamorphic rocks forming at the same time, in adjacent settings. In plate tectonics, we would expect a high-pressure, low-temperature belt associated with subduction to be paired with (not too far away) a parallel higher-temperature belt associated with a volcanic arc.

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18.2: It was different in the Archean

We don't find blueschist in Archean rocks. We don't find paired metamorphic belts in Archean rocks. It appears, therefore, that plate tectonics was not operating on Earth in that early time. Instead, our oldest rocks appear distinctively different, in both rock type and structure, and they therefore imply a different tectonic regime for Archean Earth.

So what was that "style" of tectonics, and what is the evidence for it? The oldest coherent features we see that can answer this question are called **greenstone belts**. These are linear or branching zones of distinctive low-grade metasedimentary and metavolcanic rocks. Notable greenstone belts on Earth include examples from the oldest parts of the cratons of Australia, South Africa, Europe, Siberia, Brazil, Antarctica, India, Greenland, China, and northcentral North America. All are included in otherwise fairly homogenous felsic crust of similarly ancient age. We call each of these "continental nuclei" by the term craton. Where the craton's rock is exposed at Earth's surface, we call it a "shield." Areas where the craton is present, but buried beneath a sedimentary blanket, are called "platforms."



Figure 18.2.1: Map showing the distribution of Precambrian (Archean + Proterozoic) crust on Earth. (Callan Bentley graphic)

Examples of cratons include the Pilbara Craton of Western Australia, the Kaapvaal Craton of South Africa, and the Superior Craton of the U.S. and Canada. The greenstone belts are typically sandwiched between large bodies of crustal rocks such as granite and gneiss. Not all greenstone belts are Archean; some are also known from the Proterozoic. But the Proterozoic's tectonic record is more diverse, so for the sake of this case study, we will focus on the Archean.

The greenstone belts are interpreted as the remains of ancient ocean basins, which were squished between ancient proto-continental terranes, in the same sense as how in modern plate tectonics, ophiolites mark the position of former ocean basins. The process of partial melting had repeatedly "distilled" the silicate crust of the planet into small mobile blobs, roughly the size of modern U.S. counties or states, and once solidified, these resisted subduction due to their low density. They drifted about, approaching other similar granitic terranes, with the intervening ocean basin sagging out of the way, straight down below. The more sagging occurred, the more room there was for sediments to accumulate atop the subsiding region, until the neighboring granitic blobs got close enough to compress the rocks between them. Once they collided, they often stayed stuck together, resulting in a granite/greenstone belt/granite horizontal sequence.

Let's now examine them in more detail, starting with the roster of rocks we find in greenstone belts, and progressing to an overview of their large-scale structure.

Did I Get It? - Quiz

? Exercise 18.2.1

What word describes linear or branching zones of distinctive low-grade metasedimentary and metavolcanic rocks that are found sandwiched between granite "proto-continents" in Archean aged rocks?

a. paired metamorphic belts

b. greenstone belts

c. mountain belts

Answer

b. greenstone belts

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18.3: Rocks found in greenstone belts

Rock types in greenstone belts are of both igneous and sedimentary varieties. In general, both tend to be lightly metamorphosed (greenschist facies). Here we will examine the igneous ones first, both volcanic and plutonic, and then move on to the sedimentary ones.

IGNEOUS

Basalt and komatiite



Figure 18.3.1: Archean pillow "basalts" near Ely, Minnesota, showing a greenschist overprint (including a prominent tectonic cleavage). It's a palimpsest rock! (Callan Bentley photo)

Volcanism and plutonism have been themes in Earth history from very early times. In the fact that they show evidence of the production of magma and lava, Archean greenstone belts are not that different from modern oceanic ridge systems.

Basalt was a common rock then as it is today. Greenstone belts get their very name from this prominent rock, which served as protolith for the low-grade metamorphic rock called greenstone. Then, as now, basalt erupted as mafic lava onto or near Earth's surface, cooling rapidly to make a fine-grained igneous rock. We know there must have been at least some water present, because greenstone belt basalts include **pillow structures**, features that form when basaltic lava erupts underwater. Other primary volcanic structures include vesicles, the "fossil gas bubbles" that formed as exsolved gases were trapped in solidifying lava.

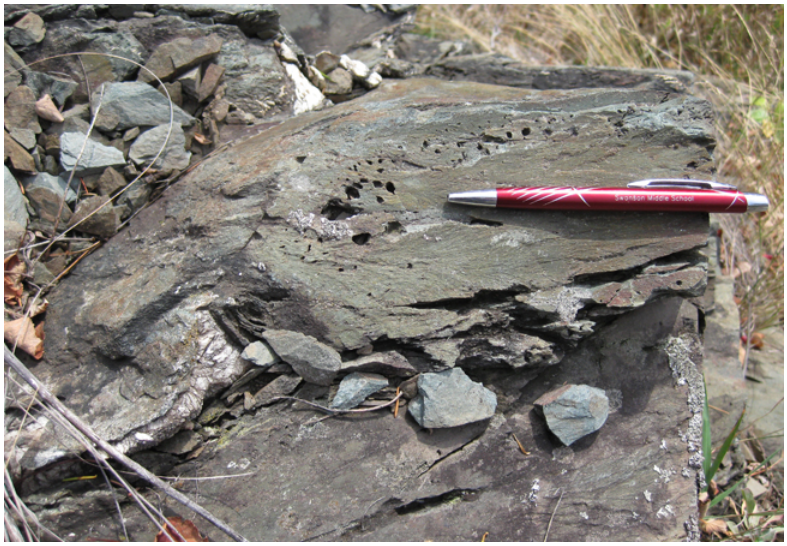


Figure 18.3.2: A vesicle "train" within a pillow structure from greenstone (meta-basalt) from the Superior Craton, near Ely, Minnesota. Pen for scale. (Callan Bentley photo)

While basalts in greenstone belts looks more or less like basalts forming today, that's not true for komatiite, an ultramafic volcanic rock. It is named for the Komati Valley of South Africa, a river valley in the area of the Barberton Greenstone Belt. Komatiite is often cited as an example of an "extinct rock," since it hardly ever forms on Earth any more. The reason for its formation in the Archean but not today is temperature. Komatiite is dominated by the mineral olivine, and olivine has a very high crystallization temperature, around 1500 °C. Lavas that hot don't erupt on our planet today; a more typical temperature for a modern mafic lava (like the Kilauea volcano in Hawaii) would be around 1100-1200 °C.

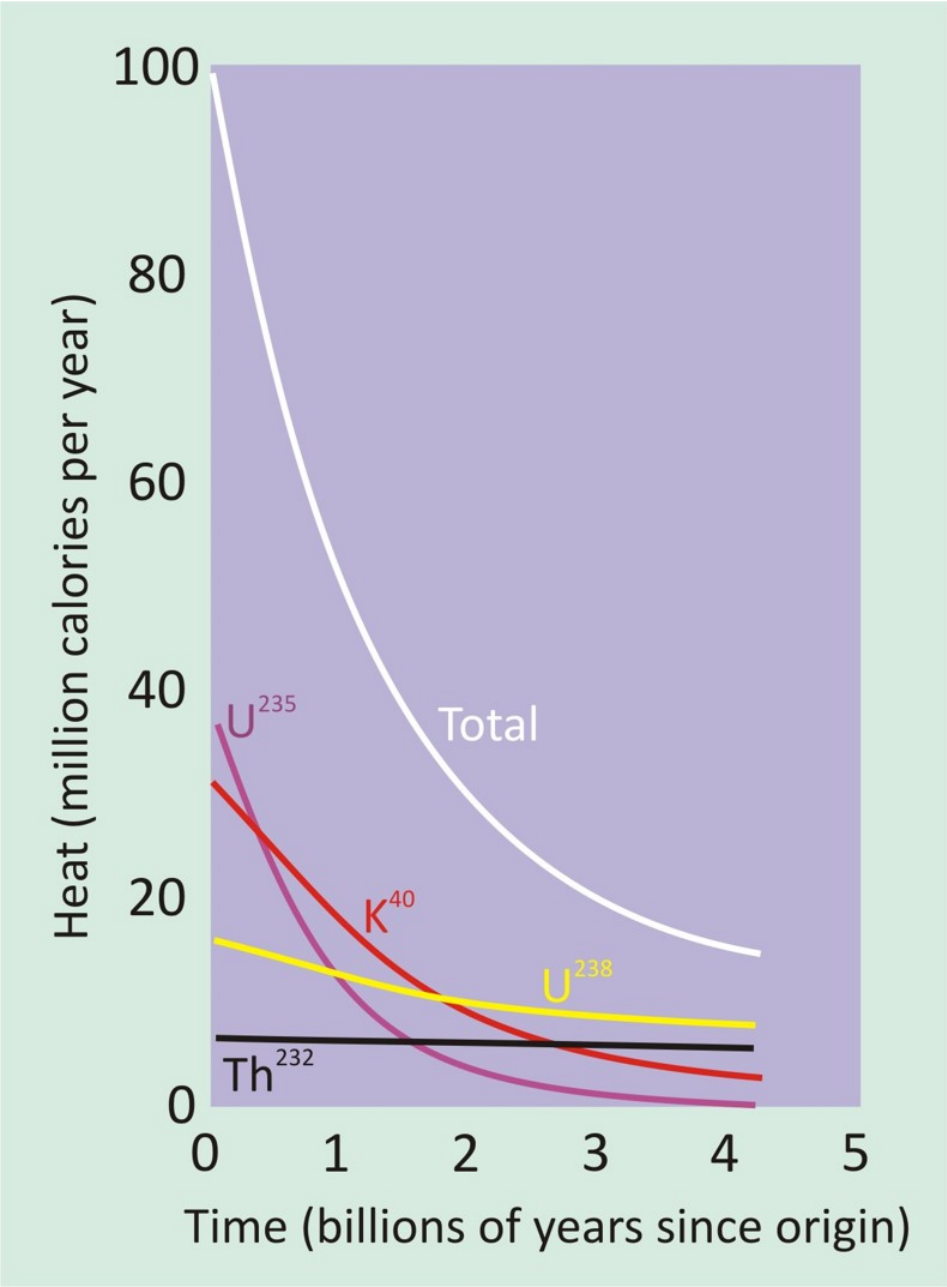


Figure 18.3.3: Heat production was much higher during the Hadean and Archean eons than it is today. There were many more unstable radioactive isotopes then, decaying and releasing heat energy into the Earth system. (Callan Bentley graphic)

The fact that we see komatiites in Archean-aged deposits is strong evidence that the early Earth was hot. At least internally, it was hotter than today. Why was this? A few reasons: (1) First, less time had passed since the planet was assembled from an unthinkable number of meteorite impacts, events that instantaneously transformed a very high kinetic energy into heat energy. So it was "still hot," like a pie recently removed from the oven. It may be cool on the outside, but you'll burn your mouth if you take a bite! (2) Second, the rate of *new* heat production was much higher in early Earth history. Less than a billion years after the nebula-forming

supernova of Earth's predecessor star, there were a lot more unstable (radioactive) isotopes around. Remember that [link to numerical dating discussion here] half of a population of radioactive atoms fall apart in the first half-life, followed by a quarter of them in the second half-life, followed by an eighth in the half-life after that, and so on. Totaling all known radioactive isotope sources suggests something like five times as many radioactive decay events in the Archean as today. So there was a much higher level of heat production in the first eons of Earth history than there was in subsequent eons.

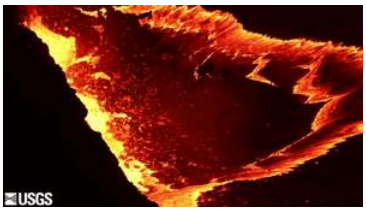


Figure 18.3.4: A modern lava lake, or the Hadean Earth? Uniformitarianism says, "Yes." (From a video by the USGS.)

In summary, not only was the planet hot from its initial smash-up mode of formation, but radiation also kept it "on a hotplate" for its initial stages. Only gradually did that heat naturally get "turned down" over time, as there were fewer and fewer unstable isotopes left to decay and make new heat. The result was that the mantle, and the crust it produced, were both much hotter than they are today. Things were hot enough that ultramafic magma stayed molten all the way to Earth's surface, and was erupted hot to flow as lava.

Examine this pair of komatiite samples and note two features: (1) the prominent brownish-orange weathering rind, indicating the high iron content of the rock, and (2) the long needle like crystals of olivine, a feature called "spinifex texture."



Callan Bentley GigaPan

Though komatiite is a volcanic rock (and thus we would expect a mostly fine-grained aphanitic texture), these large crystals of spinifex olivine are a common sight in komatiite outcrops, particularly near the top of the lava flows. Apparently, as the ultramafic lava lost heat, it began to crystallize most rapidly from the top down. Crystals of olivine nucleated on the chilly upper surface of the flow and then grew downward into its molten interior, hanging into the molten rock like chandeliers on the ceiling. Meanwhile,

chunkier olivine crystals nucleated within the flow, free floating and untethered. These settled to the floor of the komatiite flow, building up a cumulate layer there.

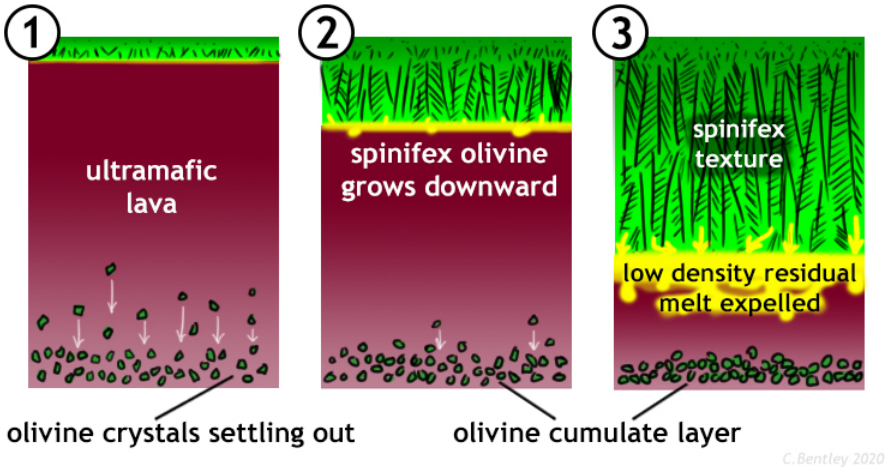


Figure 18.3.5: Spinifex-shaped olivines nucleate on the underside of a solidified "crust" on the top of an ultramafic lava flow, growing downward into the flow interior, expelling a residual melt. This diagram also shows the development of a cumulate layer of chunkier-shaped olivines at the base of the lava flow. (Callan Bentley cartoon)

Did I Get It? - Quiz

? Exercise 18.3.1

Komatiites are an indication of what aspect of the Archean Earth?

a. The Archean Earth must have been much hotter (internally) than today.

b. The Archean Earth's atmosphere was devoid of free oxygen, unlike today.

c. The Archean Earth was host to many more meteorite impacts than today.

Answer

a. The Archean Earth must have been much hotter (internally) than today.

? Exercise 18.3.2

What is spinifex structure, and how does it form?

a. It's long needle-like growths of olivine that formed on the "ceiling" of komatiitic lava flows.

b. It's fossilized spinifex grass, a common spiky grass that still exists on the modern Earth.

c. It's a description of the overall structure of greenstone belts, with thin "keel" like projections of sedimentary and volcanic rocks between dome-like bodies of granite-like plutonic rocks.

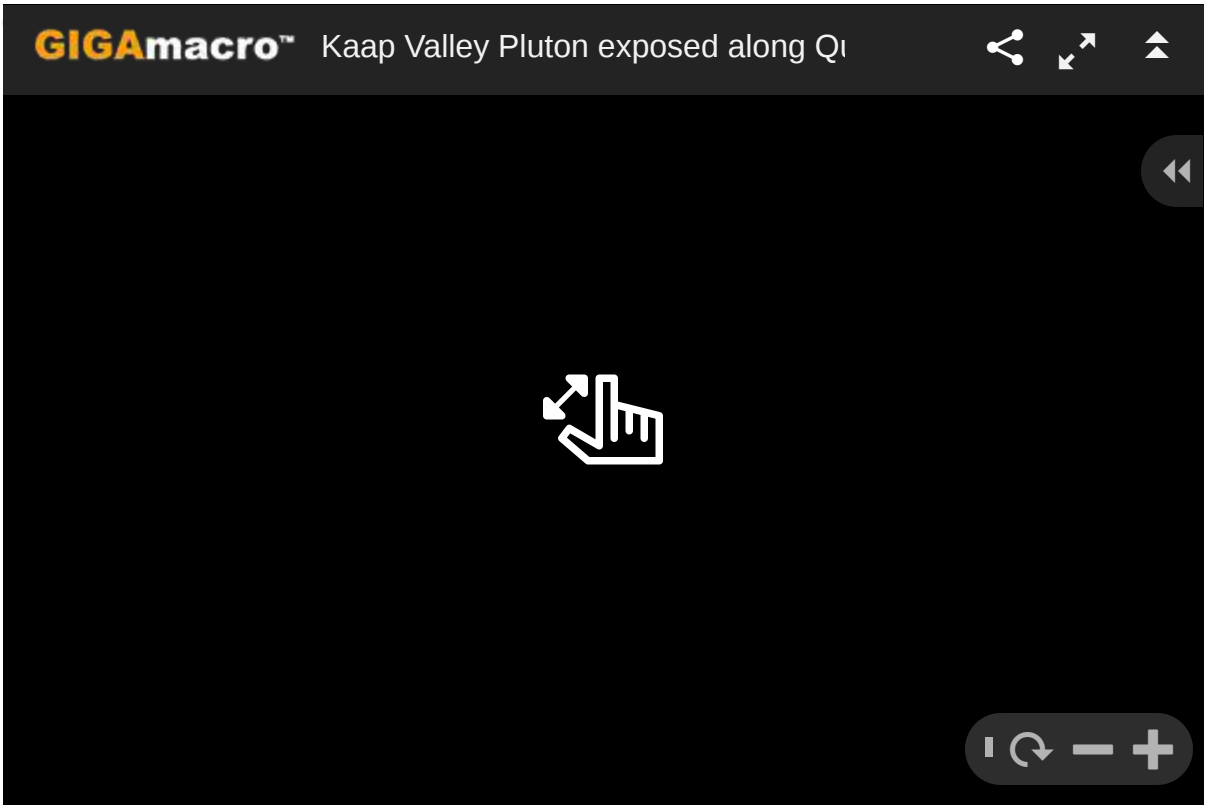
Answer

a. It's long needle-like growths of olivine that formed on the "ceiling" of komatiitic lava flows.

Tonalite, trondjhemite, granodiorite suites

Granite is a coarse-grained, felsic igneous rock. Its coarse grained (phaneritic) texture indicates it cooled slowly, which implies it cooled underground – that is to say, it is intrusive. Granite is a common rock of the modern continental crust. There are granite-like rocks in Archean cratons, too. Technically, these are not precisely granite, but instead a suite of three similar plutonic rocks: tonalite, trondjhemite, and granodiorite. Collectively, this trio of rocks are often called the "TTG" suite. The distinction between TTGs and a true granite is the absence of the mineral potassium feldspar in the TTGs.

Here is an example of one such unit: an outcrop of the Kaap Valley Tonalite, a 3.2 Ga pluton from South Africa's Kaapvaal Craton. Explore it and and count how many different textural or compositional aspects to the pluton you can find:



Callan Bentley GigaPan

Overall, we think that the TTG plutons originated the way modern granitic plutons originate: through the partial melting of other rocks, "sweating out" a magma of the easiest-to-melt minerals (those at the bottom of Bowen's reaction series). Those minerals are nonferromagnesian and silica-rich, and the result is a felsic magma. Then as now, those magmas cooled slowly underground, developing a coarse-grained texture.

Did I Get It? - Quiz

? Exercise 18.3.1

What rock type indicates the Archean Earth was **hotter** than the modern Earth?

a. Tonalite

b. Komatiite

c. Basalt

d. Blueschist

Answer

b. Komatiite

? Exercise 18.3.2

Which rock suite is similar to granite, but lacks potassium feldspar?

a. Banded iron formation

b. Blueschist

c. Komatiite and basalt

d. Tonalite, trondjhemite, and granodiorite

Answer

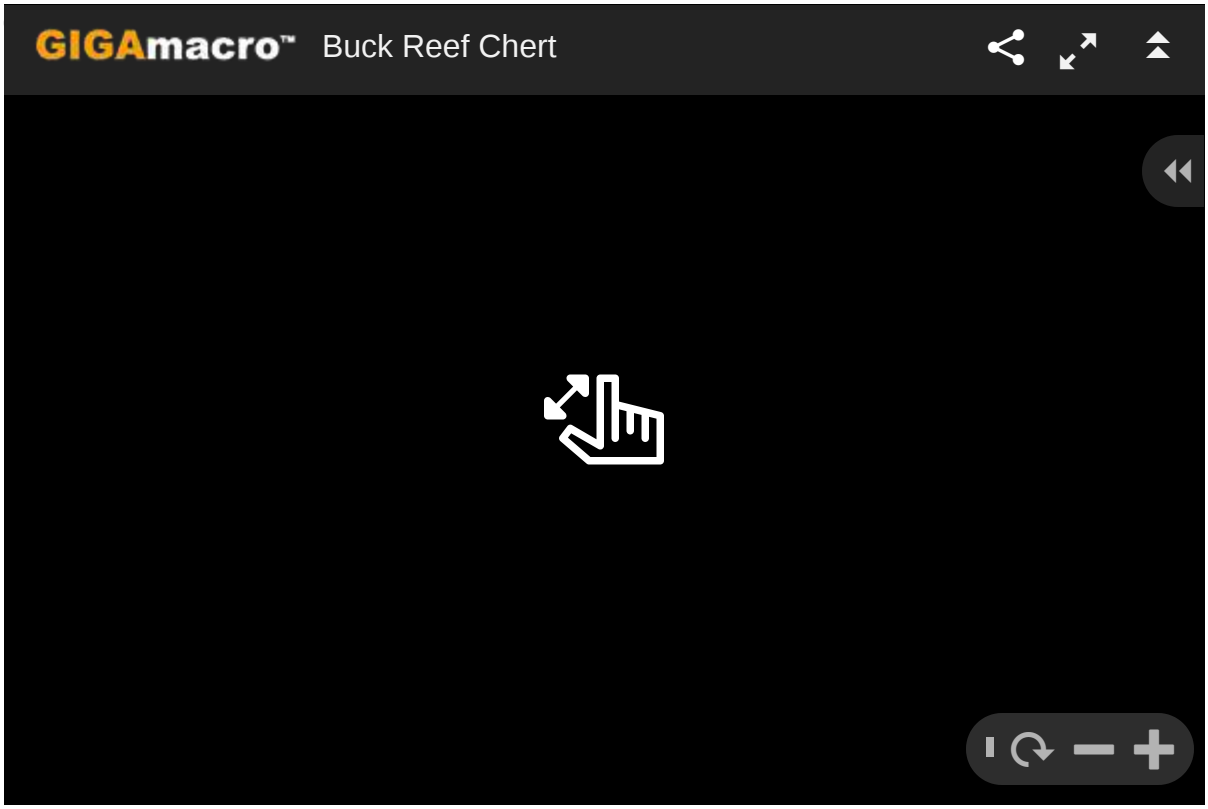
d. Tonalite, trondjhemite, and granodiorite

SEDIMENTARY

Chert

Interestingly, the oldest Archean sedimentary rocks are *not* silicliclastic. There are no sandstones. There are no conglomerates. There are no shales. This implies, quite strongly, that **there was no land** exposed above sea level – no rock exposures available for weathering and erosion. But underwater, there were still chemical sedimentary rocks forming due to mineral precipitation from the ocean. Among the oldest Archean sedimentary rocks is chert.

The Barberton Greenstone Belt's Buck Reef Chert is one of the most extensive chert deposits in the Archean geologic record. It is thick and covers a wide area, and records an extended period of chemical precipitation of silica from seawater, sometime around 3.4 billion years ago. Examine this boulder of the Buck Reef Chert, and consider its alternating black and white layers, subsequent faulting, veining, and pressure solution, and a modern surface veneer of lichens.



Callan Bentley GigaPan

Did I Get It? - Quiz

? Exercise 18.3.1

What does the (a) presence of chemical chert and (b) the absence of clastic sedimentary rocks imply about conditions during the Archean?

a. There was no land.

b. There were frequent meteorite impacts.

c. Water was shallow.

d. It was hot.

Answer

a. There was no land.

? Exercise 18.3.2

Which of the following features are *primary* aspects of the Buck Reef Chert in the Barberton Greenstone Belt?

a. bedding

b. faults

c. lichens

d. veins

Answer

a. bedding

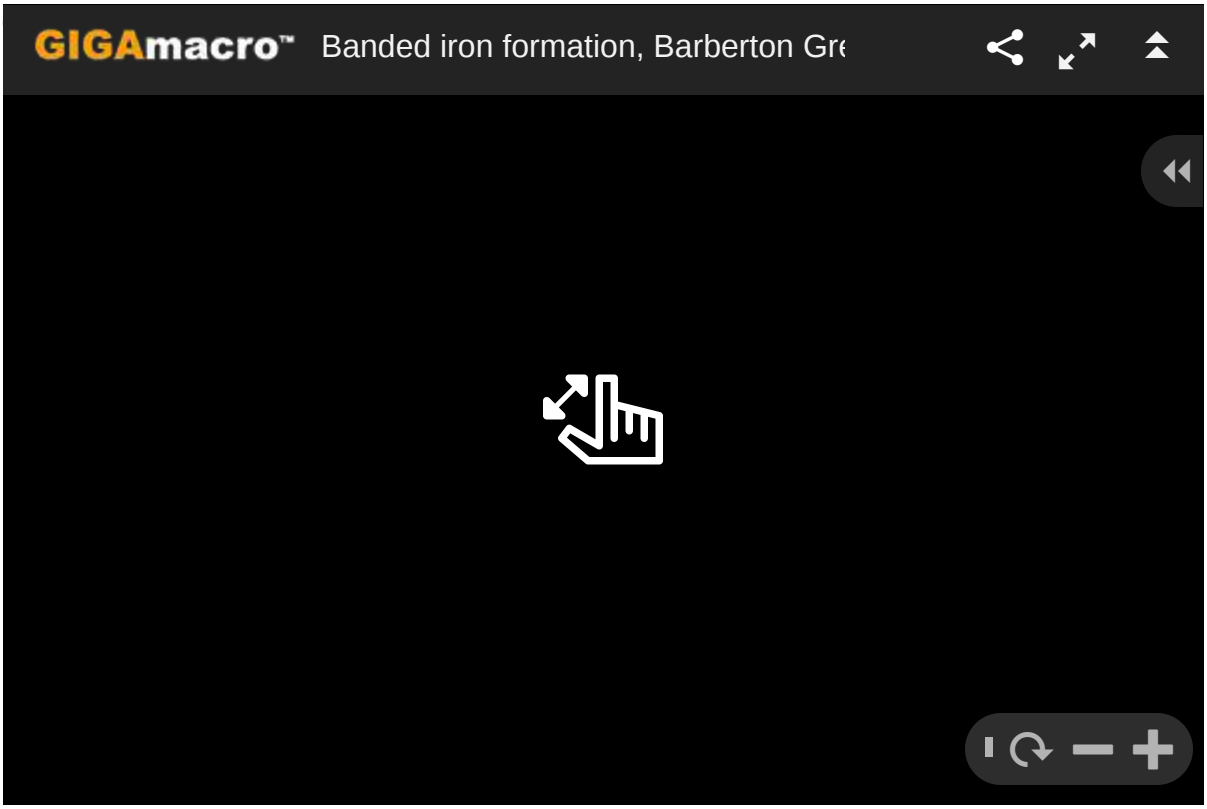
Banded iron formation



Figure 18.3.6: Deformed banded iron formation, Soudan, Minnesota (Callan Bentley photo)

Banded iron formations ('BIF's) are another distinctive kind of chemical sedimentary rocks, known only* from the Archean Eon and subsequent Paleoproterozoic Era. They are chemical sedimentary rocks composed of alternating layers of iron-rich oxides (usually hematite or magnetite) and silica precipitates (chert or jasper). Banded iron formations are the main ore of metallic iron for human use, but in the context of Historical Geology, the thing that's really interesting about BIFs is that they are a signature of radically different ocean chemistry, and therefore also imply a radically different atmospheric chemistry.

Specifically, BIFs imply a world without widespread free oxygen. This is hard for us to imagine, considering that oxygen makes up around 21% of Earth's modern atmosphere. But oxygen is a highly reactive element, and it likes to form bonds with other atoms. After all, about half of Earth's crust is oxygen, it's just bonded to silicon and other elements to make up minerals.



Callan Bentley GigaPan

As to the rhythmic nature of the alteration in BIF's internal composition, several hypotheses have been proposed. One possibility, the straightforward one, is that these are primary sedimentary layers: bedding, in other words. In that case, a rhythmic alteration in sediment source is required. This sedimentary "rhythm" is potentially tidal, seasonal, or represents fluctuations in the position of the chemocline (a boundary within the water column separating anoxic, iron-rich water below from oxygenated, iron-free water above). However, another possibility is that the original sedimentation was more or less homogeneous, and only later during diagenesis did the silica-rich and iron-rich components separate out from each other into layers.

Regardless of the details, BIF's mode of formation must have been a chemical compromise between the upwelling waters loaded with dissolved iron and shallow surface waters, pumped full of oxygen by the photosynthetic action of microbes, including those microbial mats we today call stromatolites.

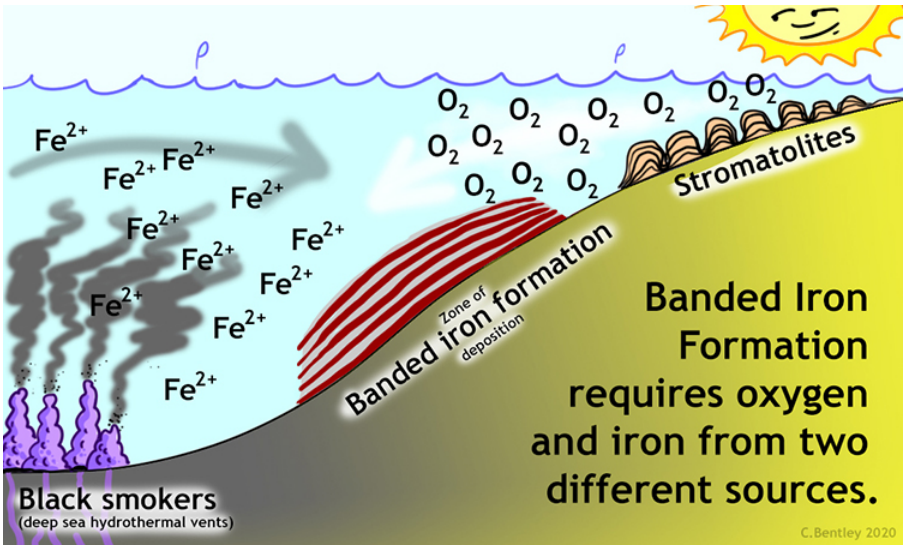


Figure 18.3.7: A simple model for the formation of iron oxide chemical sedimentary deposits (BIF) in a largely anoxic ocean, given inputs of dissolved oxygen as a waste product of photosynthesis in the sunlit shallows. (Callan Bentley cartoon)

Did I Get It? - Quiz

? Exercise 18.3.1

Why are banded iron formations important in geologic history?

a. They imply very cold conditions.

b. They imply a very hot early Earth.

c. They imply an early atmosphere and ocean devoid of free oxygen.

d. They imply violent meteorite impacts.

Answer

c. They imply an early atmosphere and ocean devoid of free oxygen.

Turbidites of graywacke and shale



Figure 18.3.8: Graded bedding in Lake Vermilion Formation graywacke, Minnesota (Callan Bentley photo)

Later in the Archean, enough land had risen above sea level that significant quantities of siliciclastic sediments were produced: mud, sand, and pebbles. Turbidity currents flowed into deep-water marine basins of the Archean, adjacent to significant sources of clastic sediment such as proto-continental terranes. Then, as now, these turbidity currents transported a slurry of sediments of different sizes. As the current slowed, its capacity dropped, and sedimentary grains settled out, in order of their weight. The biggest ones tended to be the heaviest ones, and settled out first, followed by progressively more and more lightweight grains (which tended to be smaller). The result was a graded bed: coarse at the bottom, fine at the top. Using this pattern, can you tell which way is stratigraphic "up" in the graded bedding photo here?

Shallow-water sandstone with microbially-induced sedimentary structures

In shallower water, in river deltas and nearshore settings, there were better-sorted, more mature sand deposits, dominated by quartz sand. These sand deposits record evidence of widespread microbial mats, preserved as thin, sediment-binding dark layers called "microbially-induced sedimentary structures" (MISS). These are essentially the siliciclastic version of stromatolites. Here's an example from Barberton, seen in cross-section:



Callan Bentley GigaPan

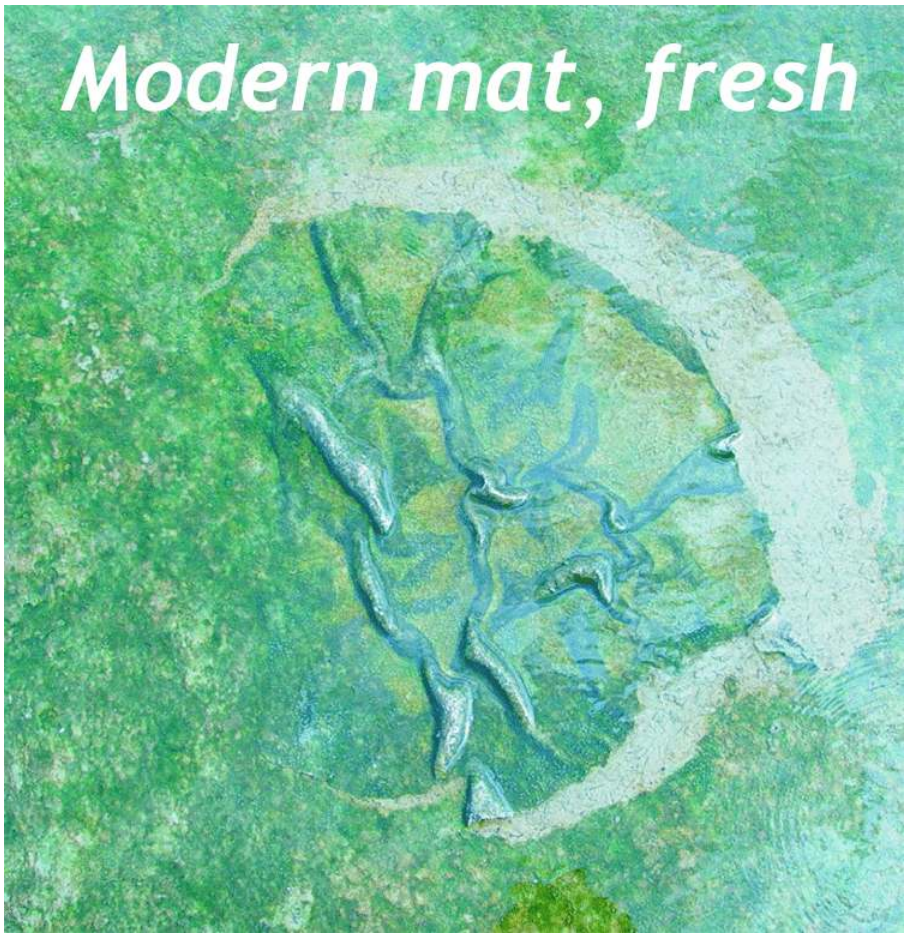


Figure 18.3.9: Taphonomy of a microbially-induced sedimentary structure: a 3 frame sequence showing (1) a modern mat, freshly ripped, rotated a bit, and wrinkled, then (2) the same structure a week later after desiccation, and finally (3) a fossil example from the Mesoarchean Pongola Supergroup. All three examples are about 40 cm in diameter. (Photographs by Nora Noffke; reproduced with permission.)

Next, consider an example from the Pongola Supergroup in Africa, as viewed on the bedding plane, and compared with a modern analogue. Many MISS show soft-sediment (and soft-mat!) deformational features such as rips and tears, wrinkles, and “elephant skin texture.” Sometimes they are rolled up, or folded back on themselves.

MISS show us that microbial mats were a key feature in Archean sedimentary systems, helping to bind sediment in place. They also show us that at least one aspect of the biology on Earth has been consistent for billions of years.

Did I Get It? - Quiz

? Exercise 18.3.1

The appearance of clastic sedimentary rocks in younger Archean strata implies which of the following?

a. The presence of sufficient land exposing rocks that can be weathered.

b. Frequent, violent meteorite impacts.

c. Very low levels of free oxygen in the early Earth atmosphere.

d. High internal temperatures for the early Earth.

Answer

a. The presence of sufficient land exposing rocks that can be weathered.

? Exercise 18.3.2

Are there any fossils in Archean clastic sedimentary rocks?

a. Yes, there are dinosaur fossils in Archean sedimentary rocks.

b. Yes: microbially-induced sedimentary structures show microbial mats binding sedimentary grains together.

c. No, there are no fossils in Archean rocks at all.

Answer

b. Yes: microbially-induced sedimentary structures show microbial mats binding sedimentary grains together.

Conglomerates



Figure 18.3.10: Strained metaconglomerate of the Seine Group, Ontario. Note the different styles of deformation exhibited by different clasts in this smeared-out rock. (Callan Bentley photo)

Some Archean sediments are very similar to modern day conglomerates – full of cobbles and pebbles and sand, that were deposited by fast-moving river currents. Based on their size, sorting, and the rounding of the cobbles in them, we think they accumulated in modestly-sized wrench basins formed in releasing bends along the edges of transpressional embryonic orogens. The clasts that make up these conglomerates tend to be volcanic in origin, as one might expect for a hot, young, immature planet. As small proto-continental masses were colliding with one another along what we today call greenstone belts, dilational wrench basins opened up, ready to receive the sloughed-off clastic detritus of the primordial mountains.

Some Archean conglomerates contain sand grains and pebbles of pyrite. These **detrital pyrites** are primary sedimentary grains, not some later manifestation of metamorphism. While pyrite exists as a mineral in the modern world, we never see it as a major constituent of sediment, because it rapidly rusts (oxidizes) in the planet's current oxygen-rich atmosphere. The fact that pyrite grains could tumble along, getting physically rounded but not chemically weathered is a surefire signal that the atmosphere of the Archean Earth was decidedly oxygen-poor. In this way, the "testimony" of the detrital pyrites corroborates the testimony of the banded iron formations we discussed earlier.



Figure 18.3.11: Detrital pyrite in conglomerate of the Steyn Reef Placer, Witswatersrand Supergroup of South Africa. (Photo by Frances Deegan; reproduced with permission.)

Meteorite impact strata

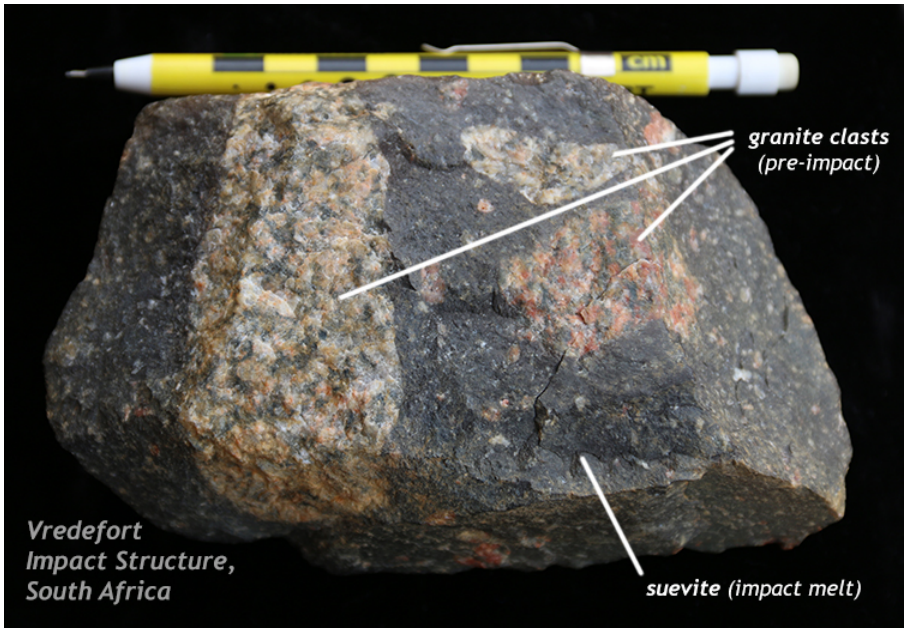


Figure 18.3.12: Meteorite impact melt hosting inclusions of pre-melt granite, from the 2.02 Ga Vredefort Impact Structure of South Africa. (Callan Bentley photo)

Another unusual aspect of really old sedimentary rocks is that they record a fair number of violent meteorite impact related rocks. Theoretically, we would expect more meteorite impacts in early Earth history than today, since then the Archean was much closer in time to the planet's accretionary formation. Since then, we've had more than three billion additional years of the planets' gravity cleaning stray chunks of space rock out of our solar system. in other words, we expect the number of meteorite impacts to decrease through time.

But during the early Archean, even though the Late Heavy Bombardment had concluded, impacts were still fairly frequent events. Fortunately, we can now begin to read the story of their impact because unlike the Hadean, a rock record still persists from the Archean. The record of impacts is recorded in rocks that come in several varieties: ejecta and fallout as well as jumbled and disturbed sedimentary layers deposited by tsunamis, rocks that record perturbations in ocean chemistry due to the impact, and also rock melted directly by the collision, called suevite.

A famous example of suevite can be found in the Vredefort Impact Structure in South Africa, and and even older example occurs in the Sudbury Impact Structure of southern Ontario in Canada. In both cases, clasts of pre-impact basement rocks (granites, mostly) are found ripped up and "floating" in the glassy suevite.

Events energetic enough to melt rock also blasted rock high into the sky, generating energetic clouds of suspended ash and water vapor, violently twirling around (generating static electricity). Clumps of ashy particles formed, and grew layer upon layer in the manner that hailstones do. We call these odd little sphere-shaped structures "spherules" and they are present in sedimentary layers of the time. Violent volcanic eruptions (of ultramafic komatiitic composition) also produces little "volcanic hailstones" in the same fashion, though it didn't take a meteorite impact to make them happen. Their similar-shaped (and sized) structures are called

"accretionary lapilli" rather than spherules, but they are essentially the same thing. Compare the two (both Archean, from the Barberton Greenstone Belt) below:



Callan Bentley GigaPans

Meteorite impacts, being violent events on an Archean planet without true continents, were very likely to happen in the oceans, which generated tsunamis. These high-energy events ripped up pre-existing strata and mixed them with spherules, redepositing the mix as energy conditions abated and the waters calmed, as seen in the virtual sample below. See if you can find a foliated chunk of basement rock, an angular chunk of chert, and a spherule...



Callan Bentley GigaPan

Another consquence of crust-piercing, ocean-churning impacts was to alter the chemistry of the oceans radically. One interesting feature associated with impact-origin strata are odd chemical sedimentary deposits of barite, a mineral rich in the element barium. Note that these strata show bedding-perpendicular oriented crystal forms that are interpreted as nucleating on the ocean floor and growing ~vertically upward into the water column. Ocean chemistry must have been seriously weird to permit such odd layers:



Callan Bentley GigaPan

Did I Get It? - Quiz

? Exercise 18.3.1

Which of the following is cited as evidence for frequent large meteorite impacts during the Archean?

a. Barite layers

b. Tsunami deposits with angular chert chunks, sand, and spherules

c. Spherules (accretionary lapilli)

d. All of these answers are correct

e. Circular- or ring- shaped deposits of suevite including chunks of basement rocks

Answer

d. All of these answers are correct

METAMORPHIC

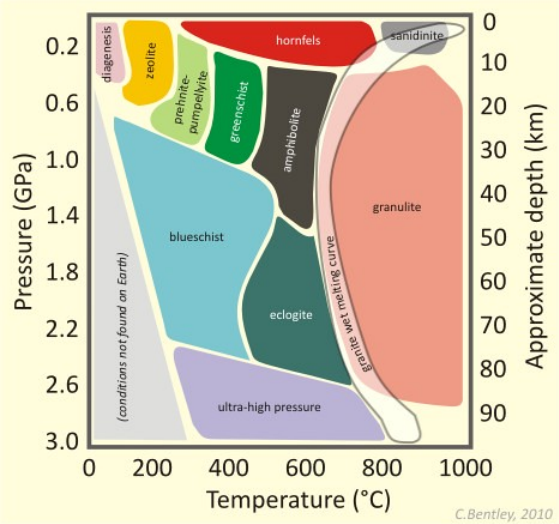


Figure 18.3.13: A plot showing the various metamorphic facies that result from different combinations of temperature and pressure. Note that blueschist and eclogite result from relatively low-temperature but relatively high-pressure metamorphism: conditions we would expect in a subduction zone. Greenschist reflects relatively low-temperature/low-pressure conditions. (Callan Bentley graphic)

As their name implies, greenstone belts have seen ***relatively*** light levels of metamorphism: merely "greenschist facies," which is sufficiently hot and high-pressure to transform basalt to greenstone (or its foliated equivalent, greenschist) but this allows certain primary features such as vesicles, pillows, and cooling columns to persist. There are no high temperature+pressure metamorphic rocks (*e.g.*, amphibolite, granulite), nor subduction related blueschists (high pressure and relatively low temperature).

Because the volcanic and sedimentary packages of greenstone belts were protected by the surrounding buffers of "granitic" material (see next section), they were never subjected to proper modern-style mountain-building or subduction, and this is a good thing from the perspective of the historical geologist, as it has allowed subtle, delicate primary features to survive, relatively undisturbed, for billions of years! If this hadn't been the case, we would today find completely recrystallized metamorphic rocks, and instead we can see primary features such as these pillow basalts in Michigan:



Ron Schott GigaPan

Did I Get It? - Quiz

? Exercise 18.3.1

True or false: "Greenstone belts are sites of high-grade metamorphism."

a. True

b. False

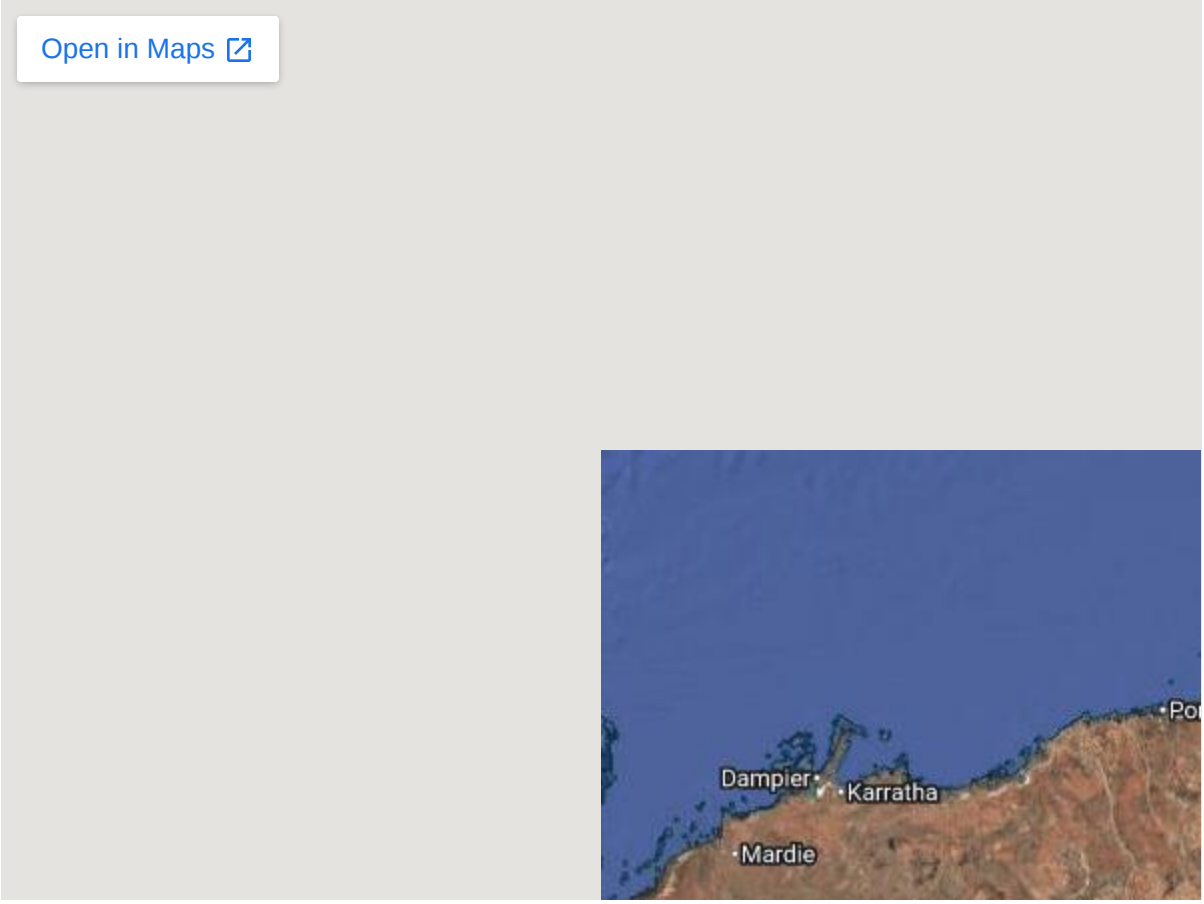
Answer

b. False

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18.4: Structure of greenstone belts

If you examine a map view of the Pilbara Craton in Western Australia, you'll see a distinctive pattern of shapes... Ignore the roads and towns, and focus on the colors and shapes in the unvegetated landscape instead:



Notice there how there are broad oval shaped light colored features, separated by more linear dark colored rocks? This is a fairly typical arrangement of rocks in Archean crust. The big light colored features are 'granite' *(really: tonalite, trondjehmite, and granodiorite)* and the dark colored lines separating them are steeply dipping layers of lightly metamorphosed volcanic and sedimentary rock. In the third dimension (depth), we think the metavolcanic and metasedimentary cover sequence pinches out, making a thin septum between blob-like granite domes. We call this geometry "dome and keel," and it's thought to represent a kind of vertical tectonics that dominated during the Archean. Here's a diagram showing the relative motion of the plutonic domes and the surface strata folded into the thin keels:

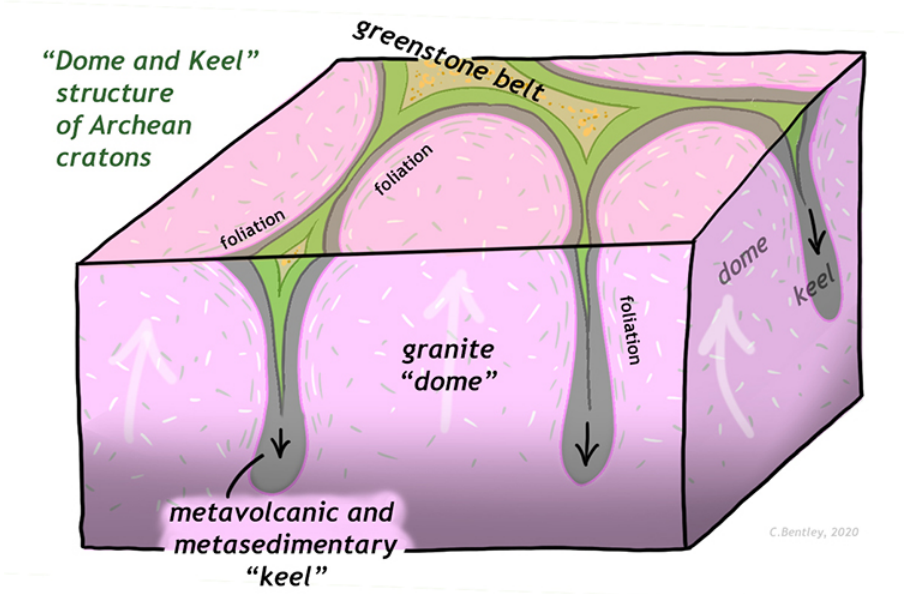


Figure 18.4.1: Block diagram showing the structure of a typical Archean craton, such as the Pilbara. Arrows show relative motion of the granite "domes" relative to the metavolcanic & metasedimentary "keels." (Callan Bentley cartoon)

The idea here is that the crust hadn't yet organized itself into thick, laterally-continuous slabs like plates. Instead, on a much smaller scale, the primordial crust was forming through both plutonic processes (pushing up from below) and surface processes (eruption of lava and production and deposition of sediment). The rising domes of granite moved vertically, and to accommodate this relative motion, the surface rocks sagged downward, perhaps even foundering (akin to a sinking ship) and "dripping" into the dynamic mantle.

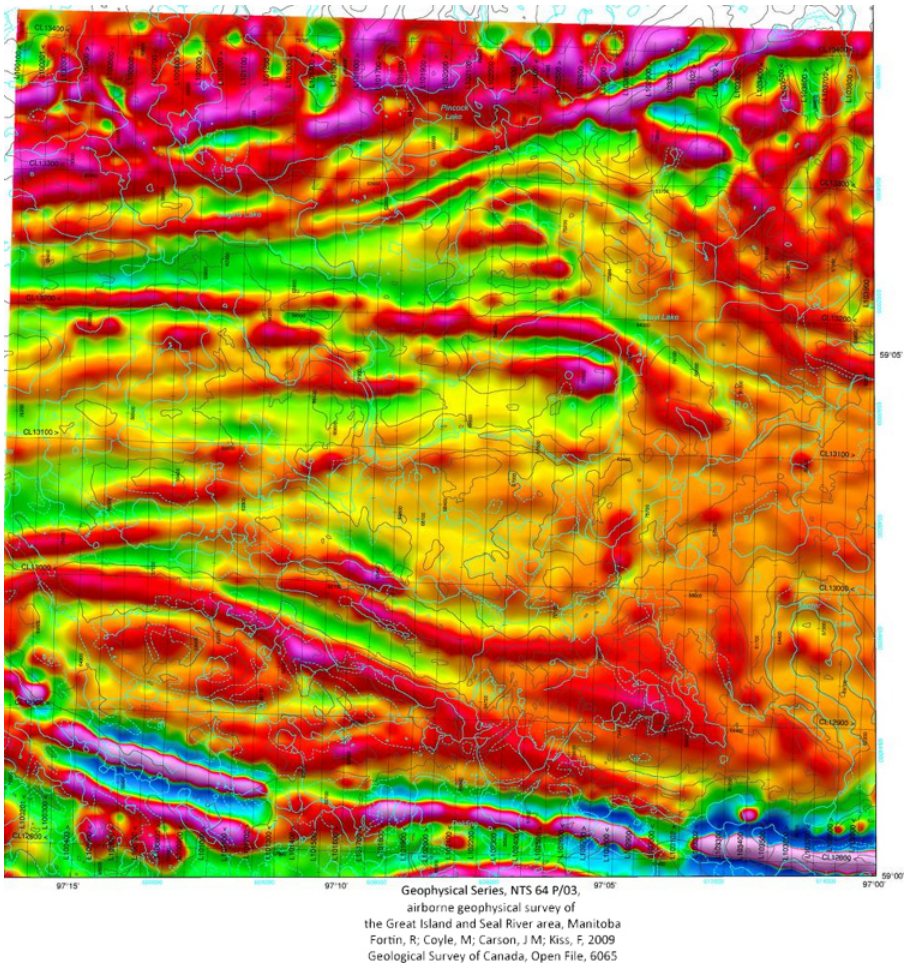


Figure 18.4.2: Dome and keel structure seen in a Canadian greenstone belt. (Geological Survey of Canada map)

Though Archean cratons and greenstone belts tend to be in relatively placid corners of the modern Earth, where we can't get a significant view into their three-dimensional structure, airborne magnetometry and gravitotometry can produce some insights into their structure, as this magnetic map of a greenstone belt in Manitoba, Canada, shows.

Overall, the dome-and-keel structure suggests that during the Archean, vertically-oriented, small-scale tectonics dominated, forming stable cratons of continental crust that have fortunately persisted until the present day, where they form the nuclei of most continents.

Did I Get It? - Quiz

? Exercise 18.4.1

What is the large-scale structure of greenstone belts?

a. Greenstone belts are found wrapping around dome-shaped bodies of "granite" in Archean cratons, thinning downward into blade-like "keels."

b. Broad rift-like features that show normal faulting and evaporite sedimentation.

c. Parallel metamorphic belts, indicating subduction and parallel magmatic arcs.

Answer

a. Greenstone belts are found wrapping around dome-shaped bodies of "granite" in Archean cratons, thinning downward into blade-like "keels."

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18.5: Further reading

Condie, K. C., & Kröner, A. (2008). When did plate tectonics begin? Evidence from the geologic record. In *When did plate tectonics begin on planet Earth* (Vol. 440, pp. 281-294). Geological Society of America Special Papers.

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CHAPTER OVERVIEW

19: (Case Study) Starting tectonics and making the first continents

Starting Tectonics and Making Continents

Learning Objectives

After reading this chapter, students should be able to:

- Describe the general idea related to the formation of continental crust and modern tectonics
- Differentiate different ideas about the differences in early tectonics models and how they are different than modern tectonics
- Explain why some rocks were formed in the Archean but are not still being formed today

19.1: The Beginning of Plate Tectonics and the Beginning of Continents - How One Caused the Other

19.2: The Chicken and Egg Problem of Making Continents

19.3: A Very Different Archean

19.4: Conclusion

19.5: References

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19.1: The Beginning of Plate Tectonics and the Beginning of Continents - How One Caused the Other

Introduction

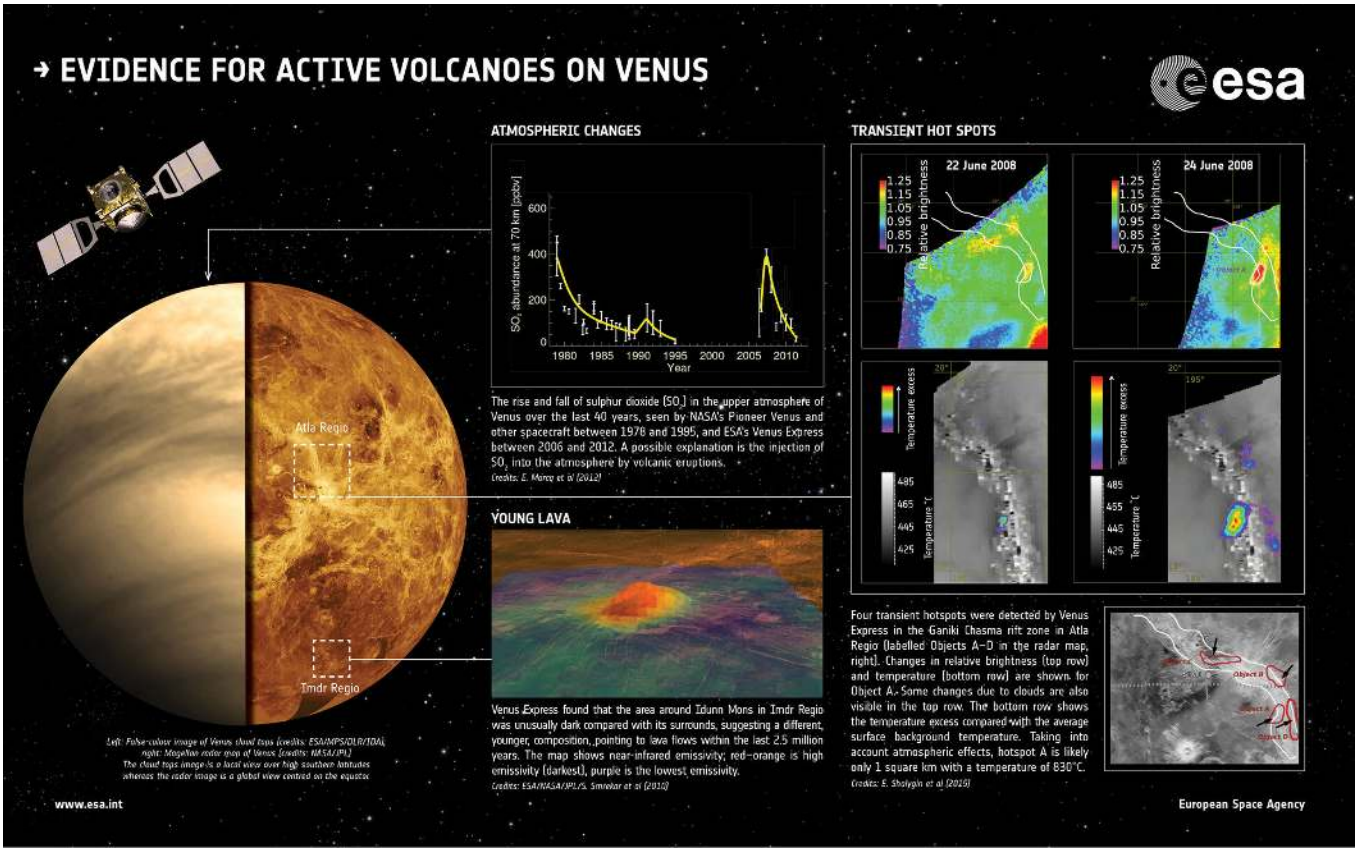


Figure 19.1.1: Diagram summarizing evidence for active volcanism on Venus. (ESA graphic)

In some ways, the first rocks that existed on Earth are fairly straightforward, even though no direct evidence of their existence has ever been found. After coalescing via impacts, the material that would become the Earth was basically a hot ball of magma. The densest (metallic) material sunk down to form the core, and lighter (silicate) material floated upwards, eventually cooling enough to form the first rocks: most likely a very mafic veneer on a magma ocean. At this time, Earth may have more resembled modern Venus, with no significant tectonic movements and volcanism as the driver which shapes the surface. So, how does an Earth-like tectonic system with subduction start on a planet with no plates? Also, since continental material today is made via processes like subduction, how does continental crust evolve on a planet without tectonics? Even on Earth today, the process of how a subduction zone initiates is a bit of an enigma. How do we solve the seemingly unanswerable question of how subduction started on Earth billions of years ago? For details on these beginning steps, please review the case study on Earth's Oldest Rocks.

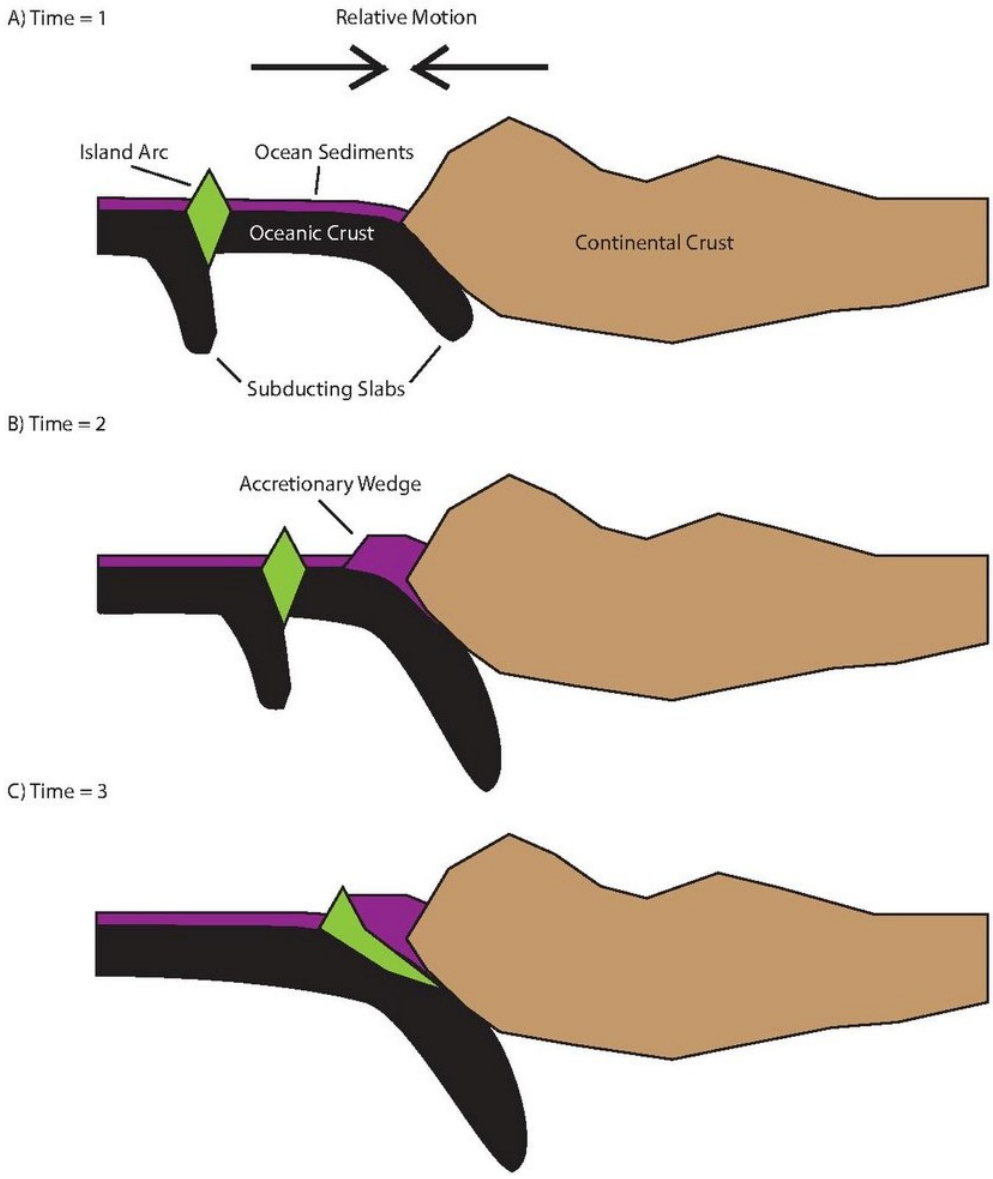


Figure 19.1.2: Cartoon showing the general process of accretion in growing continents. (Graphic: BZenith via Wikimedia.)

Why is the start of subduction so important for understanding the start of plate tectonics? There are two main reasons: First, subduction is an important driver of plate movement today. In fact, some would argue it is the ***main*** driver. Once subduction begins, other processes, like the building of continents through collisions and the creation of new seafloor at oceanic ridge systems can also begin. Secondly, some models of early Earth focus on subduction, and use it as an assembly process for Earth's first continental cores. Basically, subduction would create an island arc, and island arcs would accrete to make the first continents. Though this model has fallen somewhat out of favor (e.g. Bedard and Harris, 2014) in place of models discussed below, it is still a possible mechanism for Earth's early history, and should be considered. The only rocks we have from the Hadean and Archean eons are found in buoyant continental cratons, and thus, are the only rocks you can use to directly surmise the start of plate tectonics. While subduction is a driver of felsic magmatism today (the makings of the continental crust), is it proper to assume that subduction is the main process that made it in the past? Lastly, subduction also causes the initiation of many geochemical cycles (deep water cycle, many chemical and isotopic heterogeneities) which are used to understand the mantle. If the mantle has evolved over time via tectonic movement or other methods, it is important to have a time constraint on this mixing process.

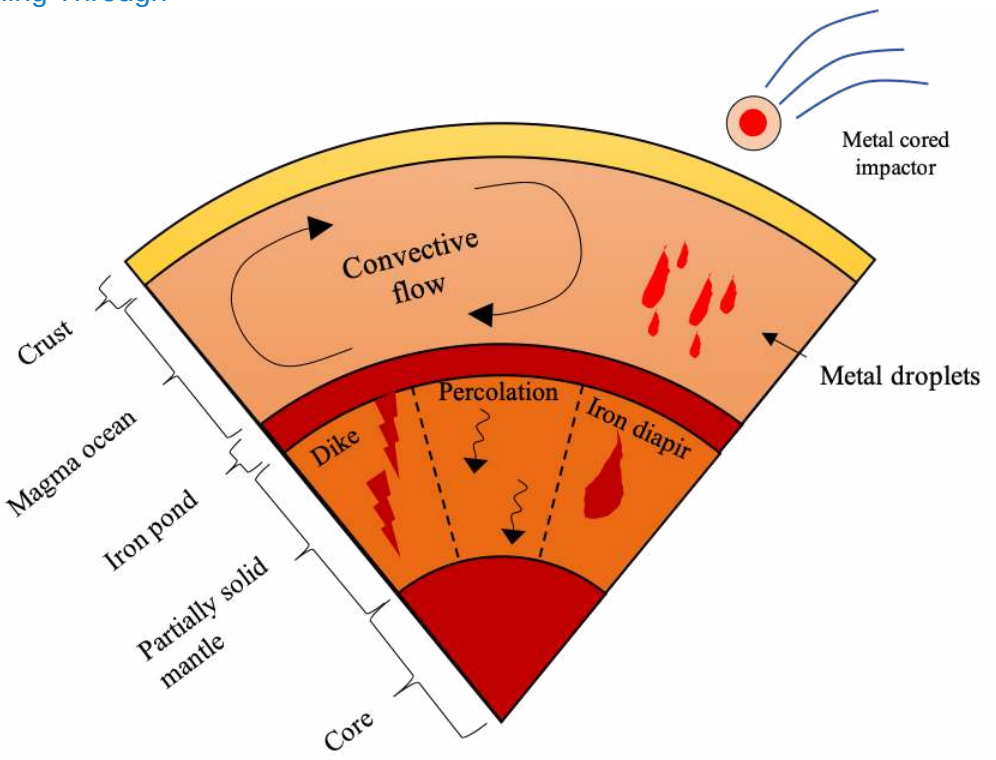


Figure 19.1.3: Hypothetical core-mantle differentiation processes: Percolation, diking, and diapirism. After Rubie et al. (2015). (CC BY-SA 4.0; By AlexInMetal – Own work, https://commons.wikimedia.org/w/inde...curid=76607390)

One of the biggest issues with modern tectonics and subduction is the heat the early Earth possessed. It is speculated that the internal heat of the early Earth was much higher than today (as much as 300 degrees Celsius), suggesting both a higher heat flow out of the planet, and thus a warmer lithosphere. Modelling has shown that this heat would have made the lithosphere weaker, more prone to breaking. A trigger, like the extra heat that comes with a mantle plume (Greya *et al.*, 2015) or even an extraterrestrial impact (*e.g.*, O'Neil et al, 2019) may have weakened the lithosphere enough to start the subduction process.

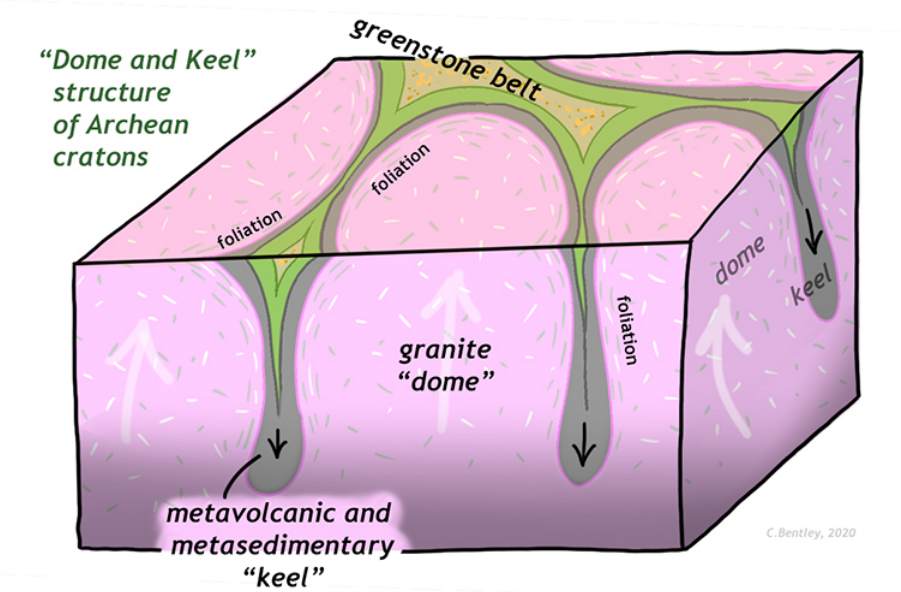


Figure 19.1.4: Block diagram showing the structure of a typical Archean craton, such as the Pilbara. Arrows show relative motion of the granite "domes" relative to the metavolcanic & metasedimentary "keels." (Callan Bentley cartoon)

Early models for the movement of material within Earth, *i.e.* a proto-tectonics, have focused on vertical tectonics, found within these greenstone belts: upward moving plumes and diapirs and downward-moving delamination or “drip” tectonics. The extra heat in the mantle could have driven a more-active plume system all over the planet. Each plume could have been a place where the lithosphere was heated and weakened. This is known as plume-lid tectonics–the rising material from plumes was somewhat capped by the young, hot, but continuous lithosphere. Another part of this plume-lid model has to do with a rock known as ecologite. Models predict a thicker oceanic crust during this time; much thicker than today’s. At about 40 km, basalt metamorphoses to eclogite, which is much denser type of rock. This dense lower part of the crust would “peel off” in a process known as delamination,or ‘drip’ downward. This is a possible way material can cycle before the beginning of plate tectonics.

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19.2: The Chicken and Egg Problem of Making Continents

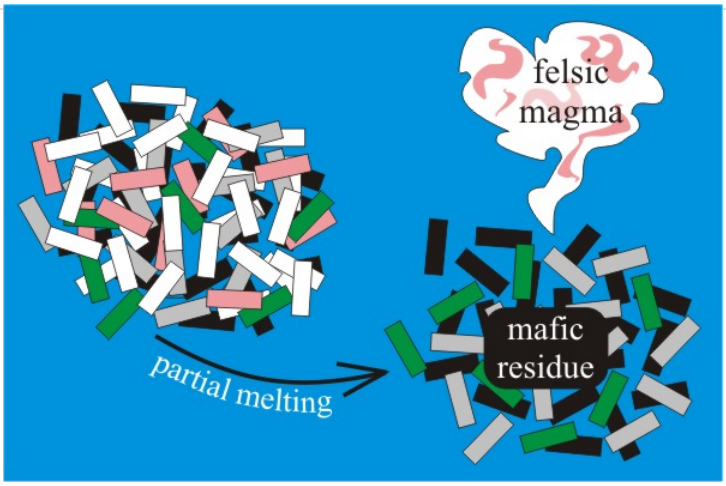


Figure 19.2.1: Partial melting makes magmas that are more felsic than their source rocks, and leaves behind solid rocks that are more mafic. (Cartoon by Callan Bentley.)

The process of making continental material on modern Earth with plate tectonics has one major requirement: continental material. So, how can you make a continent before you have a continent? With modern tectonics, the making of felsic continental rock that composes the thicker continental plates, involves the differentiation of partially melted mantle material through the subduction process and partial melting. This results in diapirs (magma chambers) of more felsic magma rising through the pre-existing crust, becoming increasingly more felsic as the buoyant magma melts surrounding rock in its journey toward the surface. Models that predict a thicker mafic crust early in Earth's history can allow for more magmatic differentiation as magma rises in a plume (or otherwise) than would happen within oceanic plates on Earth today. Another factor is felsic input directly into magmas, a major part of today's formation of continents. Through partial melting of host rock, which pulls more felsic components out of the host than mafic components, and assimilation, in which felsic material can enter a melt and drive the overall composition of it towards being more felsic (less mafic), large portions of recent continental material has been made. The problem is both of these methods require felsic host rocks present in the beginning of the process to work properly in large quantities. Even without plate tectonics as we know it today, there certainly is a propensity for making some felsic material, or at least material that is more felsic than the average crust at the time, in a variety of ways. And perhaps that is all that is needed; a slight increase in felsic material in a section of the crust might have been enough at the time to make it buoyant and start the more complex processes described above and make full-fledged continents. It seems that if this is the case, plume-lid volcanism or early subduction or both may have seeded the first continents. One final piece of the modern subduction story may come down to something as simple as sediment (Sobolev and Brown, 2019). It turns out the sediment is a great lubricator of slab movement, and the amount of sediment even affects modern slab descent. The problem is this: without a continent to supply sediment to lubricate subduction, it may be slow or inhibited. Sobolev and Brown suggest only after continents form and provide sediment as lubrication (such as with the post-Archean, 2.4-2.1 billion year old Huronian Glaciation), does subduction really take off.



Figure 19.2.2: A fragment of the Acasta Gneiss, the oldest known rock on our planet. In exhibition at the Natural History Museum in Vienna.
(CC BY-SA 3.0; By Pedroalexandrade on Wikimedia.)

Archean continental rocks may be hinting at a different process in their formation. This is due to a high concentration of rocks which are only rarely produced in tectonic settings today: the tonalite-trondhjemite-granodiorite (TTG) suite of intrusive igneous rocks and their metamorphosed descendants/equivalents, like the Acasta Gneiss. TTGs are superficially similar to normal crustal granites, but typically have a much lower amount of potassium feldspar from a lower overall elemental potassium content. According to modeling, TTGs are only formed via partial melting of metamorphically altered mafic rocks, which seems to be less common today than in the past. One thing TTGs cannot tell us is the source of this melting metamorphic mafic rock. It could come from a melting subducting slab, or it could come from lower crust delamination. It is helpful to remember that slabs today do not really melt to produce magma, but rather introduce volatiles to produce flux melting in the mantle wedge. If slabs were melting in the past this would be a clear difference between early Earth and modern tectonics. Water is also an important factor in TTG formation according to geochemical models, which does lend some credence to an early but not fully modern subduction-like process. Since water is needed to form TTGs, some process like subduction should have brought water down into the zone of formation of TTG rocks. Trace elements also support the argument that mantle mixing (with water also helping mantle convection by making it more ductile) happened with the formation of TTGs (Hastie et al, 2016). Again, a plume could also be creating weaknesses, inducing subduction (Beas et al, 2020), but either way, it appears material was moving from the surface to depth.

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19.3: A Very Different Archean



Figure 19.3.1: Blueschist cobble (with white vein) on Kayak Beach, Angel Island, California. (Callan Bentley photo)

There are several things that occurred in the Archean that did not occur afterwards. Perhaps the most famous example is komatiite, the ultramafic lava that does not erupt out of today's volcanoes. Another example, TTG rocks are only in high abundance in cratonic rocks dated from the Archean and are rarely formed today. A third example has to do with certain metamorphic rocks that are not present in Archean rocks. These include UHPM (ultra-high pressure metamorphic rocks) and blueschists. Both of these are only formed with the lower temperature/higher pressure conditions in subduction. Two conditions must be true to form these rocks: buoyant continents and strong plates, which could explain the lack of blueschists and UHPM rocks before the late Archean is the evidence used, according to Brown and Johnson (2018). Additionally, as mentioned earlier, these melting broken slabs may be the source for the TTG suite of rocks. Finally, various geochemical isotope signatures (Dhuime *et al.*, 2012, Tush *et al.*, 2020) show that looking at rocks before 3 billion years ago and after 3 billion years ago, processes were just different. Isotopes of hafnium, oxygen, and tungsten show dramatic changes occurred in mantle mixing processes around 3 billion years ago. Oceanic basalts also show concentrations of trace elements that shift around this time (Condie 2015). These are significant enough to convince many scientists that the true start of modern tectonics occurred at this time. At a recent conference on the origins and evolution of plate tectonics, a majority of the researchers put the start of plate tectonics at 3 billion years ago. Some are still convinced that it could have started as early as 4.2 billion years ago, others as recent as 1 billion years ago. It should be noted that there are qualifiers, where some scientists say a different tectonic paradigm existed earlier, whereas only modern plate tectonics, like we know, started later. Evidence for this includes kimberlites, rocks that bring up diamonds that may owe their origin to modern tectonics. While some are old, many more are found within the last billion years of the rock record (Stern *et al.*, 2016). Also, since the oldest blueschist is 780 million years old in China, the oldest UHPM rock is 620 million years old in Mali, it could mean a much more recent start to modern tectonics. Ophiolites are similar in their record, and are a clear indicator of modern tectonics. There are a few ophiolites that are much older, close to 2 billion years old, but the vast majority are less than 1 billion years old. Of course, the incompleteness of the rock record, especially as one ventures further into the past, should be considered. It is certainly possible that the lack of these rocks going back in time is due to destruction via the rock cycle and tectonics.

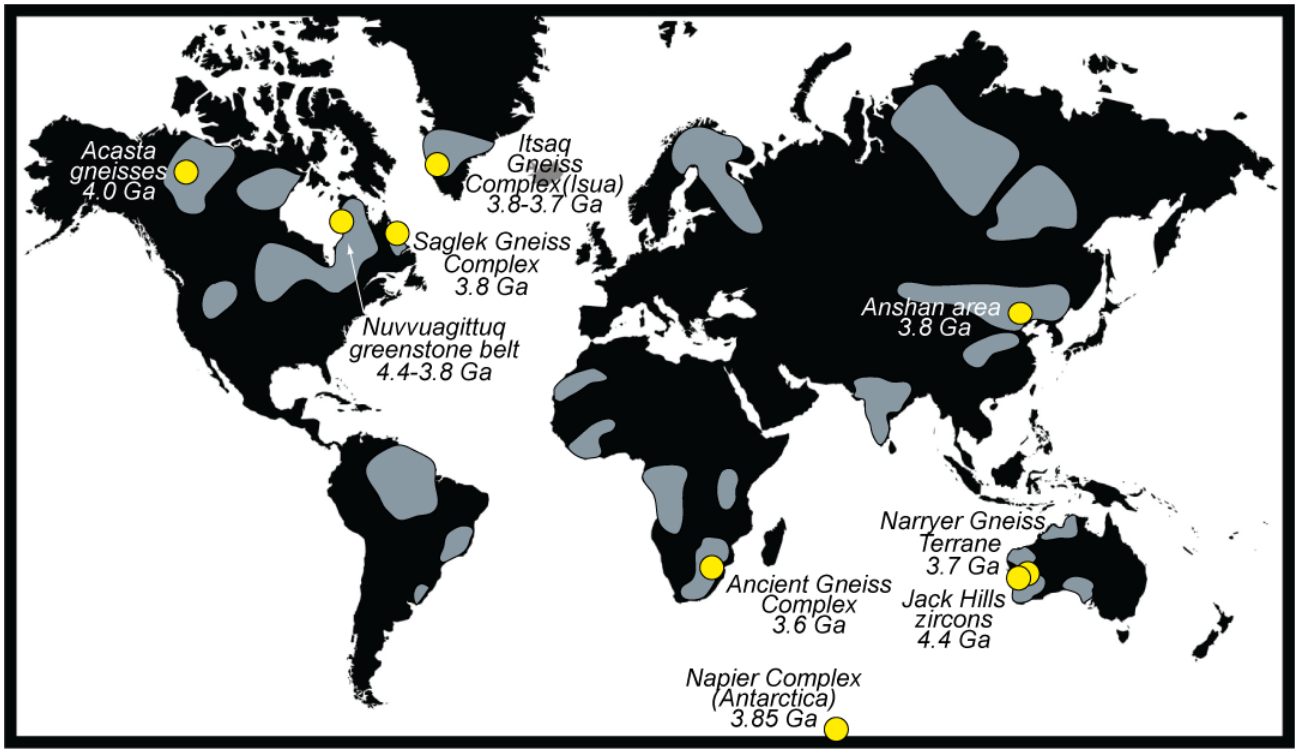


Figure 19.3.2: Archean Cratons (light grey) and locations of Earth's oldest rocks on a world map. (By Jonathan O'Neil, with permission, from: UOttawa Early Earth Blog)

Whenever it started, it is clear that by the end of the Archean, most of the cratons on Earth had formed, though the growth of continental cores like cratons has been shown to be episodic (McCulloch and Bennett, 1994), even in the Archean, meaning some sort of process must have periodically added felsic crust. The Wilson Cycle can help explain this pattern today and in the recent past. It is not as useful for a time before tectonics, nor at explaining that most cratons are dated to ~3 billion years old and older. One last item to consider: most of the cores of all modern cratons are Archean and subsequent rifting has mostly gone around them, not through them, presumably due to their strength. Were the processes that occurred before 2.5 billion years ago so fundamentally different, in that the only time in Earth history where cratons could form (known as 'cratonization'), was in the Archean? Some difference in how the Earth constructs its surface, from before the Archean to after, appears to be true. If true, it would be a significant exception to the idea of uniformitarianism.

Did I Get It? – Starting Tectonics and Making Continents Quiz

? Exercise 19.3.1

Which tectonic process makes continents today, and thus, it is important to know when it started to understand continental formation?

a. Subduction

b. Rifting

c. Seafloor spreading

d. Collision

Answer

a. Subduction

Exercise 19.3.2

Early "tectonics" focused on which kind of movement?

a. Circular (convective)

b. Equant (square)

c. Vertical (up/down)

d. Lateral (side to side)

Answer

c. Vertical (up/down)

Exercise 19.3.3

Which rock type, found only in recent rocks, can only form via the cold temperatures of subduction?

a. Ophiolite

b. TTGs

c. Kimberlites

d. Blueschist

Answer

d. Blueschist

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19.4: Conclusion

The geodynamics of the early Earth are controversial and difficult to ascertain. There are a few things that are certainly known:

1) Early on, Earth would have developed a magma ocean, cooling on its surface to form a very mafic crust.

2) Continents started to form early in Earth's history, by a process different than modern continental formation, indicated by somewhat strange early rocks.

3) The date when subduction began is highly debated, with the most commonly referenced time frame being around 3 billion years ago.

4) The start of modern tectonics is also debated, with most scientists agreeing that subduction might have been different in the Archean, including more slab break-offs.

5) This different 'tectonics' in the Archean includes TTGs formed via melting slabs and a lack of blueschists/UHPM rocks.

With this list of (relatively) known quantities, a picture of early Earth's tectonics (or lack thereof) is starting to take shape, though much remains to be resolved.

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CHAPTER OVERVIEW

20: (Case Study) Snowball Earth glaciations

20.1: Introduction
20.2: Evidence of glaciation
20.3: Evidence of low-latitude of glacial deposits
20.4: Cap carbonates
20.5: Return of the BIF
20.6: When were the Snowballs?
20.7: Starting and stopping a Snowball
20.8: Impact on animal evolution
20.9: Conclusion
20.10: Further reading

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20.1: Introduction

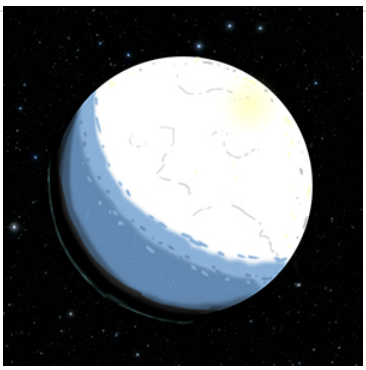


Figure 20.1.1: Our planet is thought to have been completely frozen over during the Neoproterozoic. From space, Earth would have looked like a big snowball. (Callan Bentley art)

The “Snowball Earth” glaciations were a series of ice ages during the Neoproterozoic era of geologic time, mainly confined to the Cryogenian period, but perhaps also into the Ediacaran period, too. These ice ages were thought to have been so profound that perhaps the entire surface of the planet froze over, all the way from the poles to the equator. In a 1992 paper, Caltech geophysicist Joe Kirschvink quipped that from a vantage point in outer space, the planet would have looked like a giant snowball. The evocative name stuck, and there has been an avalanche of scientific studies investigating the Snowball Earth glaciations ever since.

Depending on how you count, there were two episodes of Snowball Earth glaciation during the Neoproterozoic, or perhaps three.

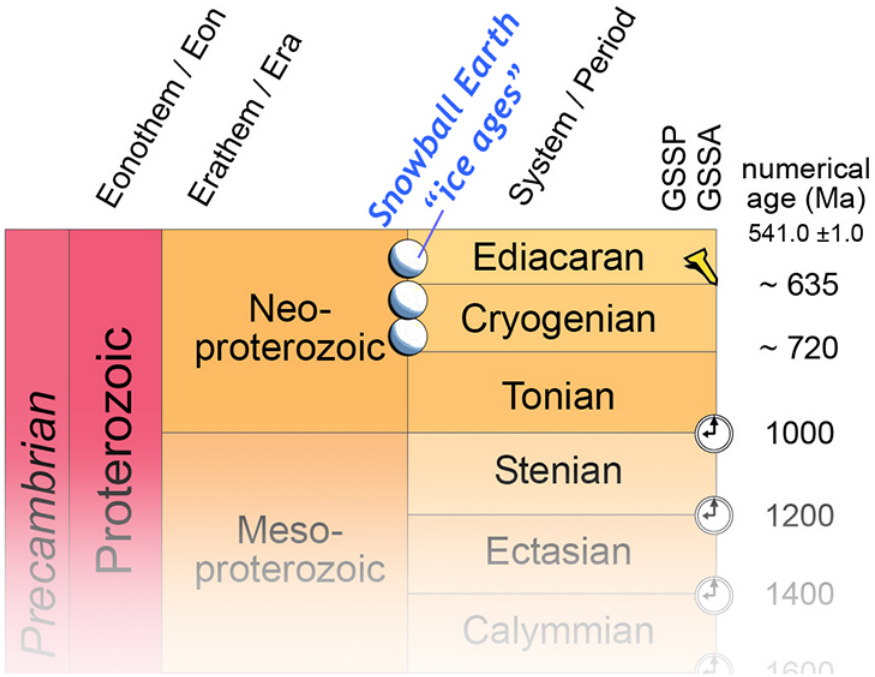


Figure 20.1.2: A close-up of the Neoproterozoic Era as depicted in the ICS timescale, showing the Cryogenian and Ediacaran periods.

They are called the **Sturtian** (about 720 to 660 Ma) and the **Marinoan** (about 645 to 640 Ma), with the **Gaskiers** (or **Ediacaran**, at about 580 Ma) being the possible third event in the sequence. There’s pretty good evidence that the Sturtian and Marinoan ice ages were true “Snowballs,” with glaciers flowing at sea level at the equator, but the Gaskiers event seems to have not been as extensive.

Did I Get It? - Quiz

Exercise 20.1.1

When were the "Snowball Earth" glaciations?

a. During the Paleoproterozoic era of geologic time

b. During the Neoproterozoic era of geologic time

c. During the Mesoproterozoic era of geologic time

d. During the Mesozoic era of geologic time

Answer

b. During the Neoproterozoic era of geologic time

Exercise 20.1.2

How many episodes of Snowball Earth glaciation happened?

a. 1, perhaps 2 (depending on what counts)

b. About 10, perhaps 12 (depending on what counts)

c. 3, perhaps 4 (depending on what counts)

d. 2, perhaps 3 (depending on what counts)

Answer

d. 2, perhaps 3 (depending on what counts)

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20.2: Evidence of glaciation

Why do geologists think there were ice ages during the Neoproterozoic? These cold times in Earth history are times when glaciers grew, flowed, and moved sediment around. Though the ice itself melts (and thus cannot be preserved though geologic time), its impacts on the sedimentary record are distinctive and durable. Ancient glaciers left behind three kinds of signatures: (1) striations, (2) tillites, and (3) dropstones. These signatures are found in Neoproterozoic strata in many places: for the Sturtian, there are 39 localities documented, on six continents. For the Marinoan glaciation, there are 48 places where glaciogenic signatures have been found.

Striations

Striations are scratches or grooves that result when a glacier takes a particle of rock (like a cobble or boulder) and drags it against the bedrock, gouging out a scratch mark that shows the direction the glacier was flowing. These are subtle features in cross-section, and they show almost no volume, so the best way to see them is on the ancient bedrock surface over which the glacier flowed. But this is rarely exposed, so this is the least common kind of evidence for Snowball Earth.



Figure 20.2.1: Striated glacial pavement (unconformity surface) below Neoproterozoic Smalfjord tillite, western Norway. (Photo by Galen Halverson; reproduced with permission.)

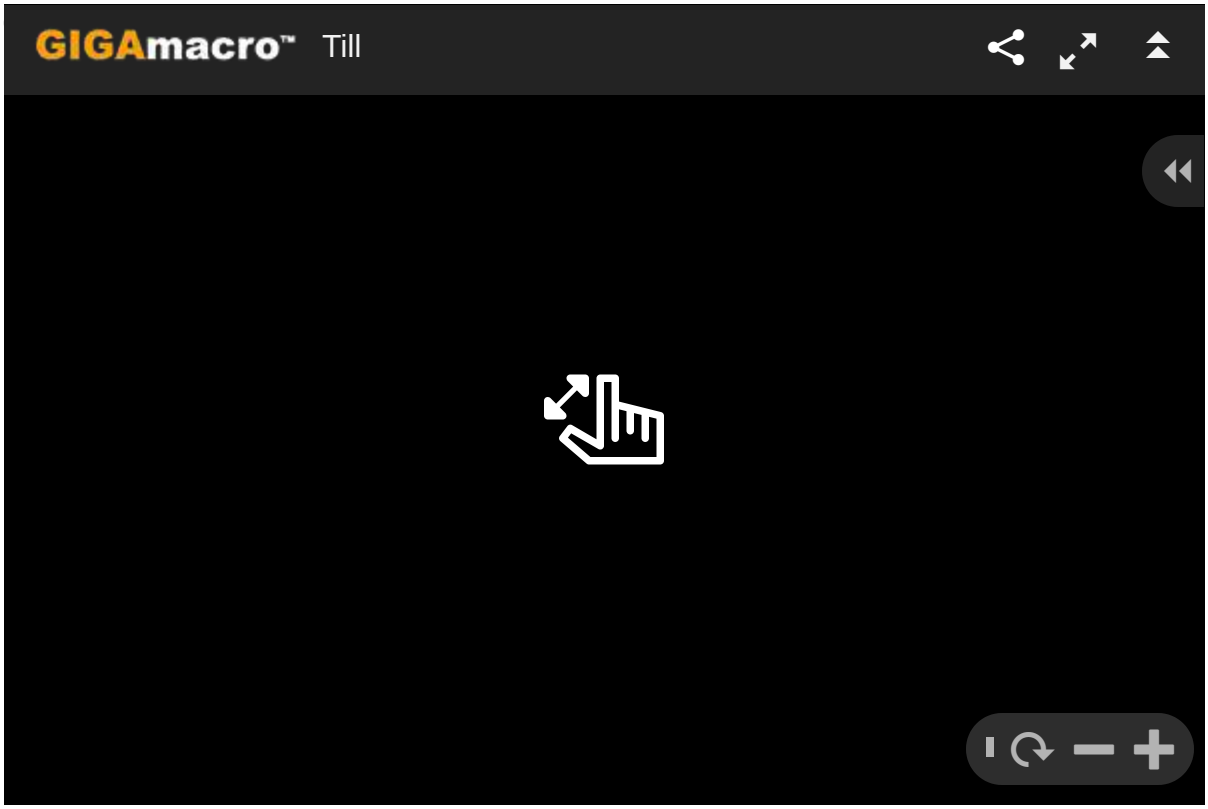


Figure 20.2.2: Diamicitite bearing faceted cobbles and boulders of granite, near Konnarock, Virginia. (Callan Bentley photo)

Glaciers produce deposits of exceptionally poorly sorted sediment: giant particles like boulders are dumped in the same spot as very fine particles like silt and clay. On the continents, this distinctive “boulder clay” is officially called “till.” Similar extremely poorly sorted deposits occur in the marine realm, where glaciers flow into the sea. To get a sense of this style of sorting, and a general sense of the diversity of rock types making up large clasts, as well as the shape of these clasts, we have compiled some representative imagery here.

Some of these examples are modern, and some are ancient. Regardless, they all share key characteristics worth noting.

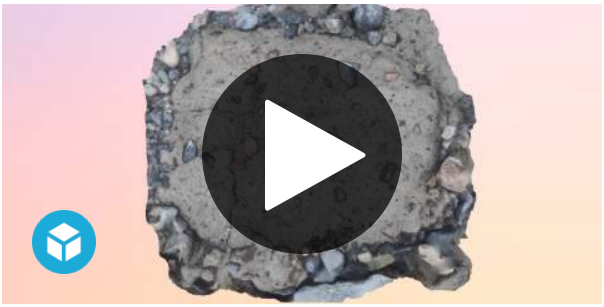
Check out this gigapixel panorama of Pleistocene-aged till from Alberta, Canada:



(Callan Bentley GigaPan)

You'll notice there that (1) it's really poorly sorted, (2) there's a lot of small stuff —the "matrix" — which "supports" the overall deposit and that means that (3) the biggest bits — the "outsized clasts" — are pretty much isolated from one another, making them look like they are "floating" in the soup of the surrounding clay and sand.

Here's a 3D model, showing a similar looking deposit from Ireland:



(Callan Bentley & Sadhbh Baxter 3D model)



Figure 20.2.3: Gaskiers tillite, Newfoundland. (Callan Bentley photo)

Again, this is Pleistocene rather than Neoproterozoic, but the darned-near-present is the key to the past. Note the large faceted clasts "floating" (*i.e.*, not touching each other) in all orientations within the finer-grained matrix. When till is lithified (turned to rock through compression and cementation), we call it **tillite**. The Snowball Earth glaciations are recorded by tillite deposits.

Another rock name frequently used in describing poorly sorted sedimentary deposits is **diamictite**. Diamictites are poorly sorted, but the name is purely descriptive, and doesn't necessarily imply a glacial origin for the deposit. For instance, a landslide could produce a very poorly sorted mix of sedimentary grains, a *diamicton*, which might then be lithified to make a diamictite. Both rock types contain large "out-sized" clasts (cobbles, boulders) is isolated from one another, surrounded by finer-grained, poorly sorted sediment. All tillites are diamictites, but not all diamictites are tillites.



Figure 20.2.4: A large (cobble-sized) clast pf porphyritic rhyolite within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.



Figure 20.2.5: A large (cobble-sized) faceted clast within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.



Figure 20.2.6: A large (pebble-sized) clast within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.



Figure 20.2.7: A large (cobble-sized) faceted clast within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.



Figure 20.2.8: A large (cobble-sized) clast of vesicular basalt within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.



Figure 20.2.9: A large faceted (boulder-sized) clast within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.



Figure 20.2.10: A large (cobble-sized) clast within diamictite of the Gaskiers Formation, western Avalon Peninsula, Newfoundland.

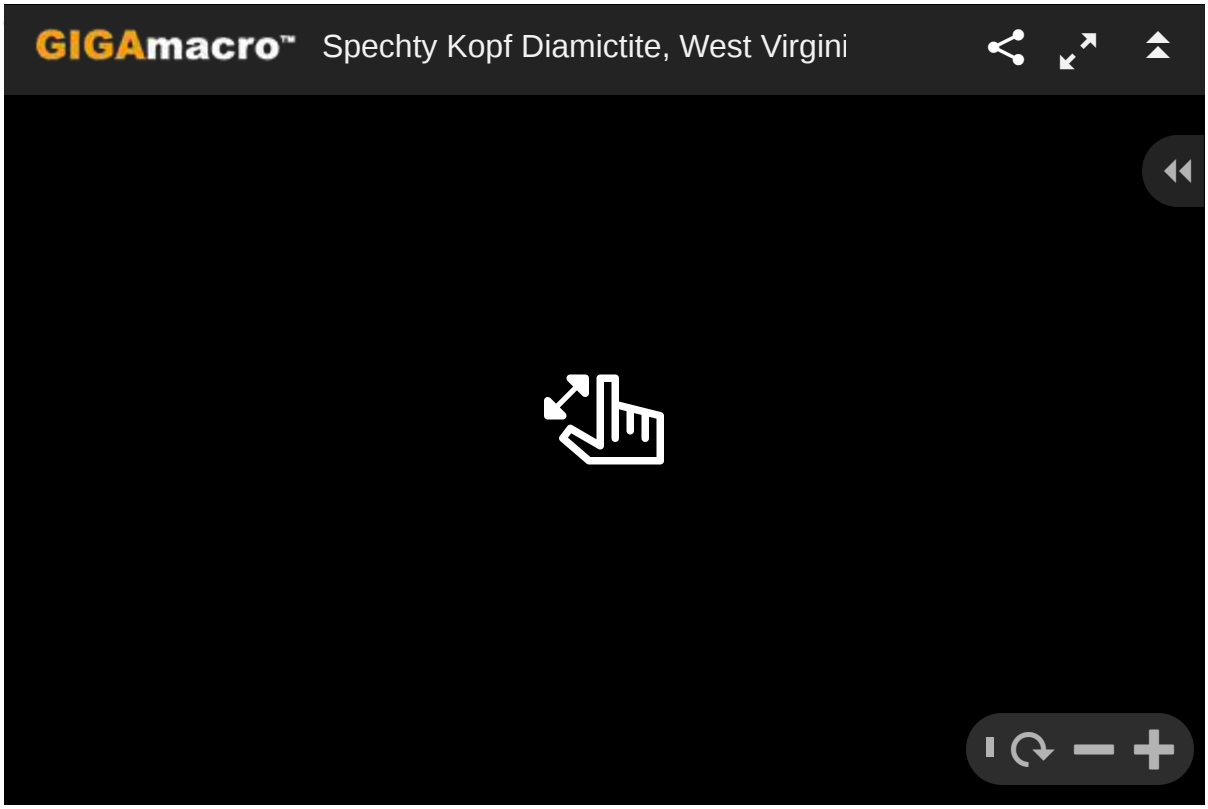


Figure 20.2.11: A faceted shape bearing intersecting scratches is strong evidence of a clast having been ground down through glacial action. (Callan Bentley photo)

One way to distinguish whether a diamictite had a glacial origin (that is, whether it is a tillite) is on the basis of the character of the larger clasts. Some of the boulders and cobbles in a true tillite will show faceted shapes, where they've have flat faces ground down by being dragged along over resistant surfaces. These facets will frequently show randomly-oriented scratches, too. These features together provide compelling evidence of the past grinding action of glacial ice.

Because scratched-up clasts are relatively uncommon, many geologists will be cautious and conservative when describing poorly sorted sedimentary deposits, referring to them by the broader and less specific "diamictite" even if they suspect the deposit is glacial in origin (glaciogenic; that is, a tillite or its marine equivalent).

Here are two hand samples of diamictite to examine. Which one is clearly glaciogenic? Which one is equivocal? Challenge yourself to explain the evidence you're keying into to make the distinction.



(Callan Bentley GIGAmacro)



(Robin Rohrback GIGAmacro)



Figure 20.2.12: Outsized clasts in foliated metadiamictite of the Fauquier Formation, Virginia Blue Ridge. Foliation is vertical in this photo. (Callan Bentley photo)

Of course, like any rock, diamictites are subject to metamorphism if mountain-building occurs in their region after they form. Metamorphism transforms diamictites into ***meta***diamictites, and tillites into ***meta***tillites. Metadiamictites show the original sedimentary poor sorting, but then ***also*** bear a foliated metamorphic overprint, giving them a palimpsest texture that indicates this more complicated history.

These are potentially-glacial deposits that were deposited in regions that later experienced mountain-building and recrystallization. If the regional metamorphism is intense, original primary sedimentary structures can be destroyed. If metamorphism is moderate, it can generate confusion, as metamorphic foliation wraps around outsized clasts in the same way that sedimentary laminations might in an unmetamorphosed dropstone-bearing deposit. Geologists working in metamorphic terranes have to be extra careful in making sedimentary interpretations of such rocks.

Compare the two formation mechanisms here:

Figure 20.2.13: Formation of a dropstone in layered sedimentary strata. Compare with the development of foliation that wraps around a metasedimentary porphyroclast. (Callan Bentley cartoon)

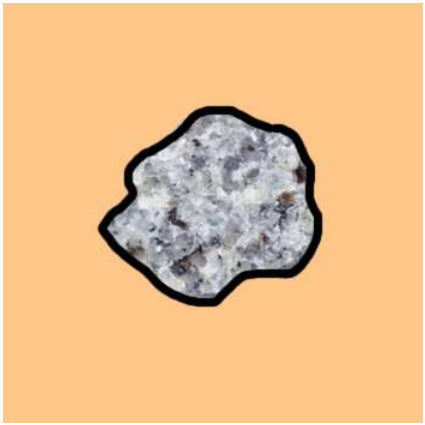


Figure 20.2.14: Formation of foliation wrapping around a porphyroclast in metasedimentary strata. The pattern resembles the relationship between a dropstone and laminated strata. (Callan Bentley cartoon)

Here is a gallery of metadiamictites to peruse:



Figure 20.2.15: Port Askaig tillite, Islay, Scotland.



Figure 20.2.16: Port Askaig tillite, Islay, Scotland.



Figure 20.2.17: Outsized clast in foliated metadiamictite of the Fauquier Formation, Virginia Blue Ridge.



Figure 20.2.18: Outsized clasts in foliated metadiamictite of the Fauquier Formation, Virginia Blue Ridge.

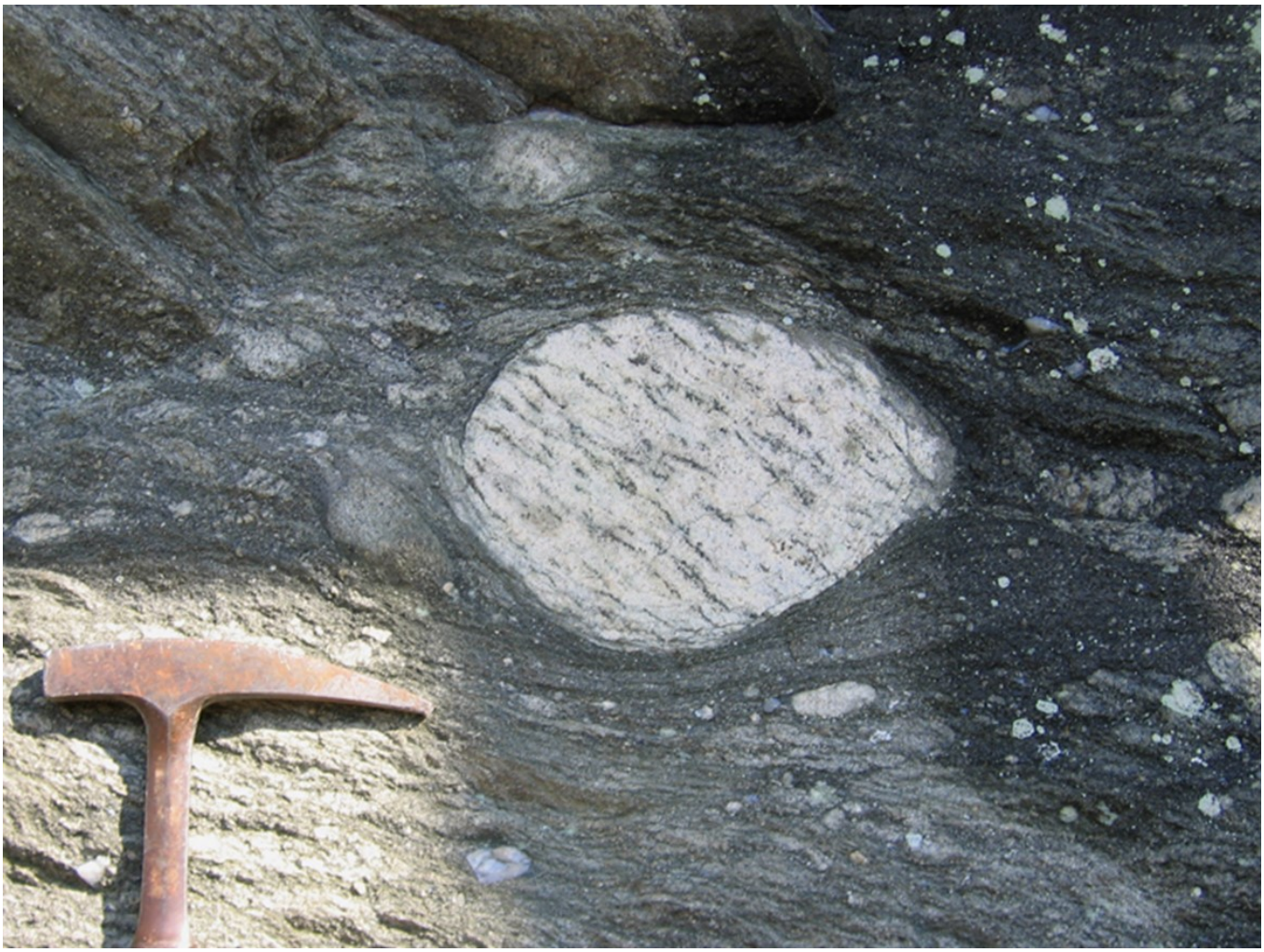


Figure 20.2.19: Outsized clast in foliated metadiamictite of the Fauquier Formation, Virginia Blue Ridge. Foliation and bedding are both horizontal in this photo.

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Figure 20.2.20: Outsized clasts in foliated metadiamictite of the Fauquier Formation, Virginia Blue Ridge.

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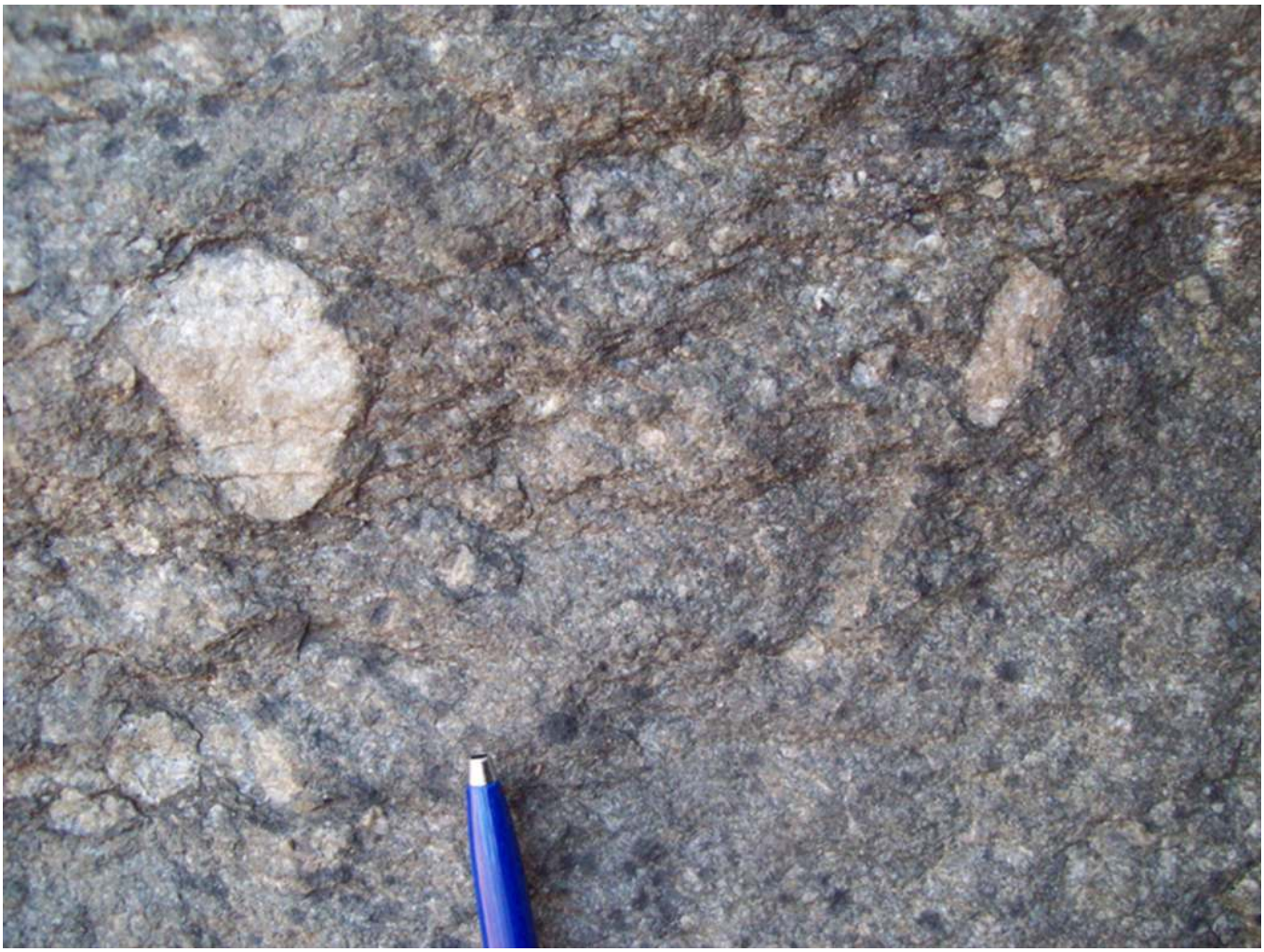


Figure 20.2.21: Outsized clasts in weakly foliated metadiamictite of the Mechum River Formation, Sharptop, Virginia Blue Ridge.

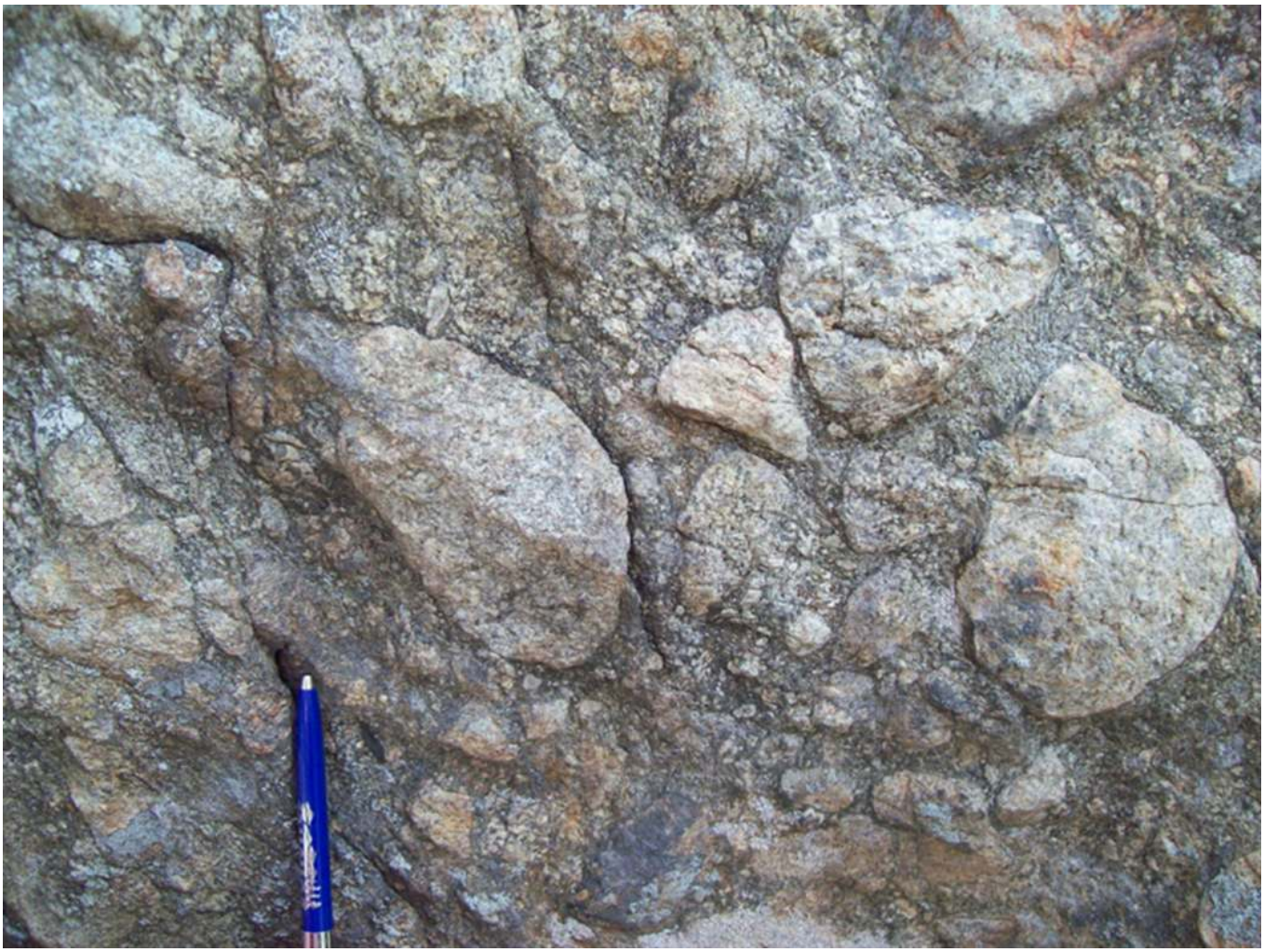


Figure 20.2.22: Outsized clasts in weakly foliated metadiamictite of the Mechum River Formation, Sharptop, Virginia Blue Ridge

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Figure 20.2.23: Detailed look at metamorphic foliation "wrapping" around an outsized clast in metadiamictite of the Mechum River Formation, Sharptop, Virginia Blue Ridge.

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Figure 20.2.24: Outsized clast in weakly foliated metadiamictite of the Mechum River Formation, Sharptop, Virginia Blue Ridge.

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Figure 20.2.25: Boulder-sized clast in weakly foliated metadiamictite of the Rockfish Conglomerate, Virginia Blue Ridge.



Figure 20.2.26: Boudinage of an internally-foliated outsized clast in foliated metadiamictite of the Mechum River Formation, near Sperryville, Virginia Blue Ridge. Note the vertical foliation in the clast vs. the horizontal foliation in the surrounding metadiamictite.



Figure 20.2.27: Deformation of a foliated outsized clast around another clast in metadiamictite of the Mechum River Formation, near Sperryville, Virginia Blue Ridge.

Did I Get It? - Quiz

? Exercise 20.2.1

Is this a non-glaciogenic diamictite or a tillite (with a glacial origin)?



a. It's a tillite, because this large clast shows scratches on at least one of its faceted surfaces.

b. It's a tillite, because this large clast is pink.

c. All we can say is that it is a diamictite, since the rock is poorly sorted.

d. All we can say is that it is a diamictite, since there is no distinctive glaciogenic evidence here.

Answer

a. It's a tillite, because this large clast shows scratches on at least one of its faceted surfaces.

Exercise 20.2.2

Summarize the steps in the making of a glaciogenic metadiamictite (a metatillite) in chronological order.

a.

1) Glaciation and erosion of pre-existing rocks

2) Metamorphism of the diamictite (tillite)

3) Lithification of the sediment into diamictite (tillite)

4) Deposition of poorly sorted sediment

b.

1) Glaciation and erosion of pre-existing rocks

2) Deposition of poorly sorted sediment

3) Lithification of the sediment into diamictite (tillite)

4) Metamorphism of the diamictite (tillite)

c.

1) Intrusion of granite

2) Metamorphism of the granite

3) Uplift and erosion of the meta-granite

4) Glaciation

Answer

b.

1) Glaciation and erosion of pre-existing rocks

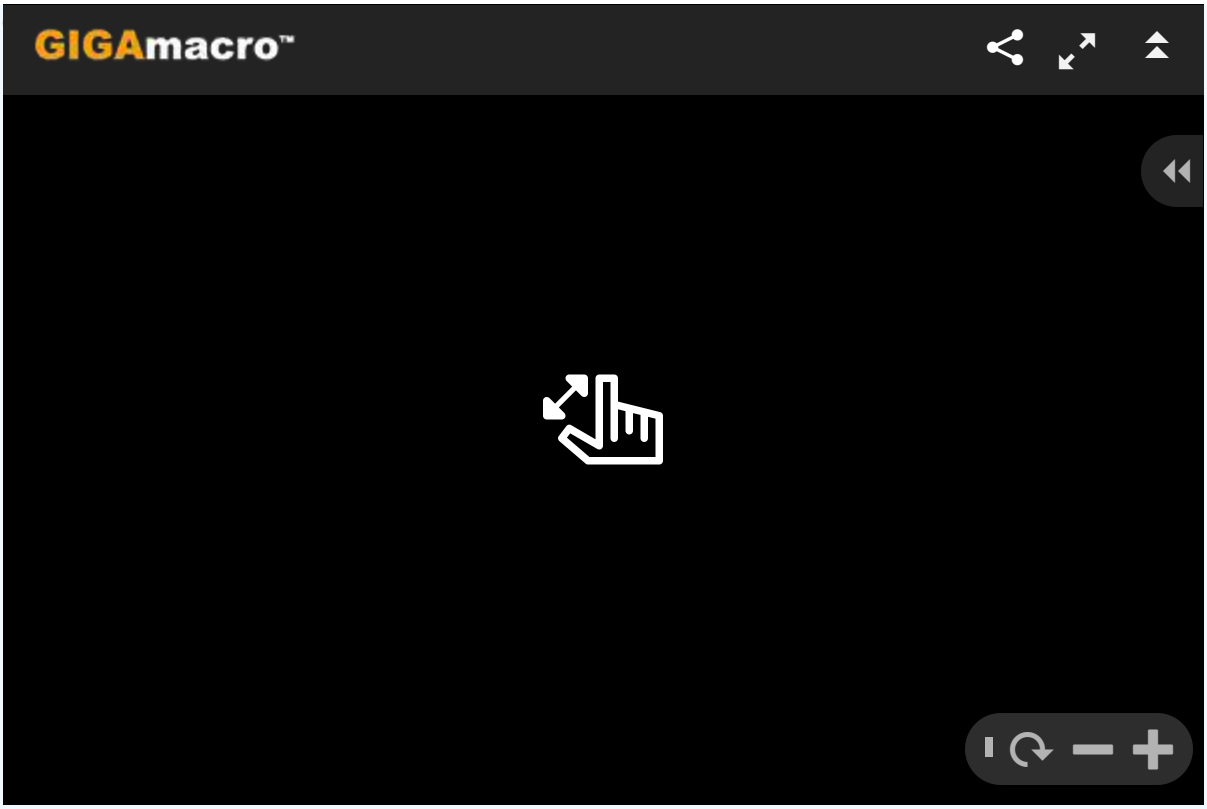
2) Deposition of poorly sorted sediment

3) Lithification of the sediment into diamictite (tillite)

4) Metamorphism of the diamictite (tillite

Exercise 20.2.3

Is this a tillite? Explain.



a. No, this is a limestone with mud-chip rip-up clasts.

b. No, it's better sorted than a tillite (the pebbles are touching, not "floating"), and there is no sign of scratching on the clasts.

c. Yes, this sample shows an extremely poorly sorted mixture of large "out-sized" cobbles that "float" in a matrix of much finer sediment. Several of the big clasts show scratches on their faceted surfaces.

d. Yes, it's dark colored and shows a mixture of rounded and semiangular clasts.

Answer

b. No, it's better sorted than a tillite (the pebbles are touching, not "floating"), and there is no sign of scratching on the clasts.

Dropstones

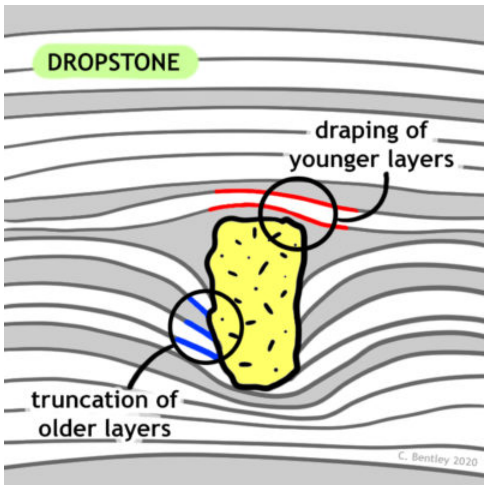


Figure 20.2.28: Cartoon cross-section of a dropstone, in cross-section, showing key relationships to the surrounding sedimentary layers. (Callan Bentley cartoon)

A third and final signature of glaciation in the Neoproterozoic era is the presence of dropstones in offshore marine sedimentary deposits. These are fine-grained, thin-bedded or laminated deepwater deposits of clay and silt, into which drop much larger particles, such as pebbles, cobbles, or boulders. The dropstones pierce into the soft pre-existing sedimentary layers, disrupting their continuity. We say that the pre-dropstone layers are **truncated**. Post-dropstone layers of fine sediment are draped atop them.

There's more than one way to make dropstones today. After all, you could paddle a canoe out to deep water with a few cobbles onboard, then toss them in with a splash. Or a tree could grow through many years, its roots wrapping around cobbles in the soil. If the tree were to be blown over in a storm, it might be washed out to sea, root-wrapped cobbles and all. When, at sea, the wood finally rots, the cobble would be freed to drop into deep water sediments.

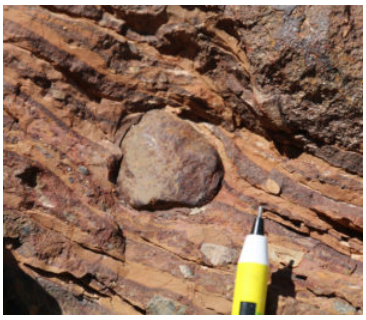


Figure 20.2.29: Dropstone in Kingston Peak diamictite, Sperry Wash, Death Valley region, California. (Callan Bentley photo)

But there were neither trees nor canoes during the Neoproterozoic. The only plausible candidate for transporting diverse clasts offshore where they could be dropped into thin sedimentary layers was ice. The idea is that when outlet glaciers from the Snowball Earth ice sheets reached the sea, where they calved off icebergs full of gravel, sand, cobbles, and silt. These icebergs floated out to deep water, melted, and let their load of sediment fall through the water below.

[Cartoon of larger-scale process here? Simplified version of this: https://www.nature.com/articles/s41467-019-13496-5/figures/2]

In the Neoproterozoic, ice was the only way to get dropstones, and so the **presence of Neoproterozoic dropstones = Neoproterozoic glacial ice.**



(Robin Rohrback GIGAmacro)

It's worth pointing out that there's no size implication with the term "dropstone." Dropstones can be huge, but they can also be small. Here's a very small example: a mere 40 microns across, imaged with a scanning electron microscope:



(Robin Rohrback SEM GIGAmacro)

Here is a gallery of dropstones to peruse:



Figure 20.2.30: An internally-laminated dropstone in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.



Figure 20.2.31: A BIG dropstone in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.



Figure 20.2.32: A layer rich in dropstones in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.



Figure 20.2.33: Two substantial dropstones in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.



Figure 20.2.34: Two small dropstones in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.



Figure 20.2.35: A dropstone measuring about 1m by 2m in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.



Figure 20.2.36: Large dropstone in Dwyka Formation of South Africa (Carboniferous/Permian age). (Photo by Christie Rowe, reproduced with permission.)



Figure 20.2.37: Two dropstones in Dwyka Formation of South Africa (Carboniferous/Permian age). (Photo by Christie Rowe, reproduced with permission.)

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Figure 20.2.38: A dropstone in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California. This one shows a piece of the underlying Beck Spring Formation, which has distinctive "giant ooids" (the sphere like shapes).



Figure 20.2.39



Figure 20.2.40: Dropstone in early Permian aged Pagoda Formation, Mt. Butters Massif, Shackleton Glacier, Antarctica. (Photo by Libby Ives; reproduced with permission.)



Figure 20.2.41: A dropstone in Kingston Peak Formation, exposed in Sperry Wash, near southern Death Valley, California.

Did I Get It? - Quiz

? Exercise 20.2.1



What feature is shown in this sample?

a. A dropstone in a diamictite

b. Striations on a glacial pavement

c. Cap carbonate

Answer

a. A dropstone in a diamictite

Exercise 20.2.2

What characterizes a tillite, as opposed to a diamictite?

a. Diamictites are lithified, but tillites are not.

b. Diamictites are poorly sorted sedimentary deposits that are lithified, but tillites are diamictites that are specifically glacial in origin.

c. Diamictites are well sorted, but tillites are poorly sorted.

d. Tillites are well sorted, but diamictites are poorly sorted.

Answer

b. Diamictites are poorly sorted sedimentary deposits that are lithified, but tillites are diamictites that are specifically glacial in origin.

Exercise 20.2.3

Relative to a tillite, we might look for striations _____.

a. ...on top of the tillite, on the surface separating it from overlying "cap carbonate" deposits.

b. ...below the tillite, on the unconformity surface separating it from the bedrock beneath.

c. ...within the tillite, between any dropstones.

Answer

b. ...below the tillite, on the unconformity surface separating it from the bedrock beneath.

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20.3: Evidence of low-latitude of glacial deposits

We have now established that there is plenty of evidence of glacial activity preserved in the rock record from the Neoproterozoic. But it is also true that there is plenty of glacial activity on the modern Earth, and no one would mistake our modern planet for a giant snowball. So the question becomes: why do we think that the Snowball Earth glaciations were ***global*** freeze-overs, not just ***local*** glaciations (as in the modern world)? The answer comes from examining paleomagnetic signatures within the sedimentary rocks.

Paleomagnetism

The study of ancient magnetic signatures within rocks is paleomagnetism. Geologists who pursue this important work are paleomagneticians, but they are a cheeky lot, and like to call themselves "paleomagicians." They collect very-precisely-oriented samples of either volcanic or sedimentary rocks, and then place those samples within a machine that can measure the orientation of their faint remnant magnetic field very precisely. This yields several kinds of magnetic information, but for our purposes we will be focused on determinations of paleo-inclination, which in turn can speak to the paleo-latitude at which the sample formed.

How does this work?

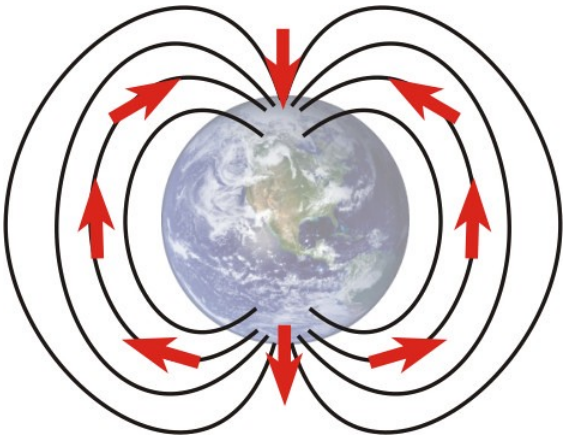


Figure 20.3.1: The shape of Earth's magnetic field is a torus. Lines of magnetic force are vertical at the poles, and horizontal at the equator, with intermediate values in between. (Callan Bentley cartoon)

The Earth's magnetic field is generated due to flow in its molten outer core, which is made of magnetically-conducting iron/nickel alloy. The magnetic force penetrates through the overlying mantle, crust, ocean, and atmosphere, and even extends out into space around the planet. The shape of that magnetic field is the relevant issue here. It is shaped like a very fat doughnut, with a dimple-like "hole" over the north magnetic pole and the south magnetic pole. Technically, this shape is called a "torus."

As expressed at Earth's surface, lines of magnetic force are vertical at the poles, and horizontal at the equator. In the diagram at right, consider the angular relationship between the "shaft" of the arrows and the surface of the Earth. At the equator, they are parallel, and at the poles you will note a perpendicular relationship. In between, the angles vary systematically. This angle (*ranging between 0° at the magnetic equator and 90° at the magnetic poles*) is called the angle of **inclination**. Because it varies by north-to-south position on the planet's surface, it is used as a record of ancient latitude (paleo-latitude).

Lava flows and sedimentary deposits form at Earth's surface as more or less horizontal sheets. Prior to consolidation (cooling and crystallization in lava flows, settling and lithification within clastic sedimentary strata), magnetically-susceptible grains such as magnetite are free to move, and they align themselves with the surrounding, permeating planetary magnetic field. Once lithified or crystallized, the grains have no more freedom of movement, and are locked in place as a durable record of the orientation of the Earth's magnetic field at the moment that layer formed.

Equatorial sea-level glaciers imply high albedo



Figure 20.3.2: Gently folded tidal rhythmites of the Elatina Formation of central Australia, strata that show both evidence of glaciation and paleo-latitudes as low as 4 degrees from the equator. (Photo by Galen Halverson; reproduced with permission.)

A series of papers in the mid-1980s to late 1990s established measurements of paleomagnetic inclination (and thus paleo-latitude) for Marinoan-aged sea-level marine strata in several locations, particularly central Australia. These strata formed about 635 Ma and include signatures of glacial influence (such as dropstones), but they also show tidal rhythms preserved as varves. The idea behind studying these special strata is that (1) we know they are glacial, but (2) they also show fine layers that we can rely on as representing "horizontal" at the time they formed. Collecting carefully-oriented samples from such rocks allows for the precise measurement of their subtle magnetic signatures. Luckily, these samples have also been undisturbed since the time that they formed: their magnetism hasn't been reset by a metamorphic event, for instance. The quality of the paleomagnetic measurements is robust, meaning that they are trustworthy and have passed several tests validating the results. And those results are: the sediments formed within 10° of the equator.

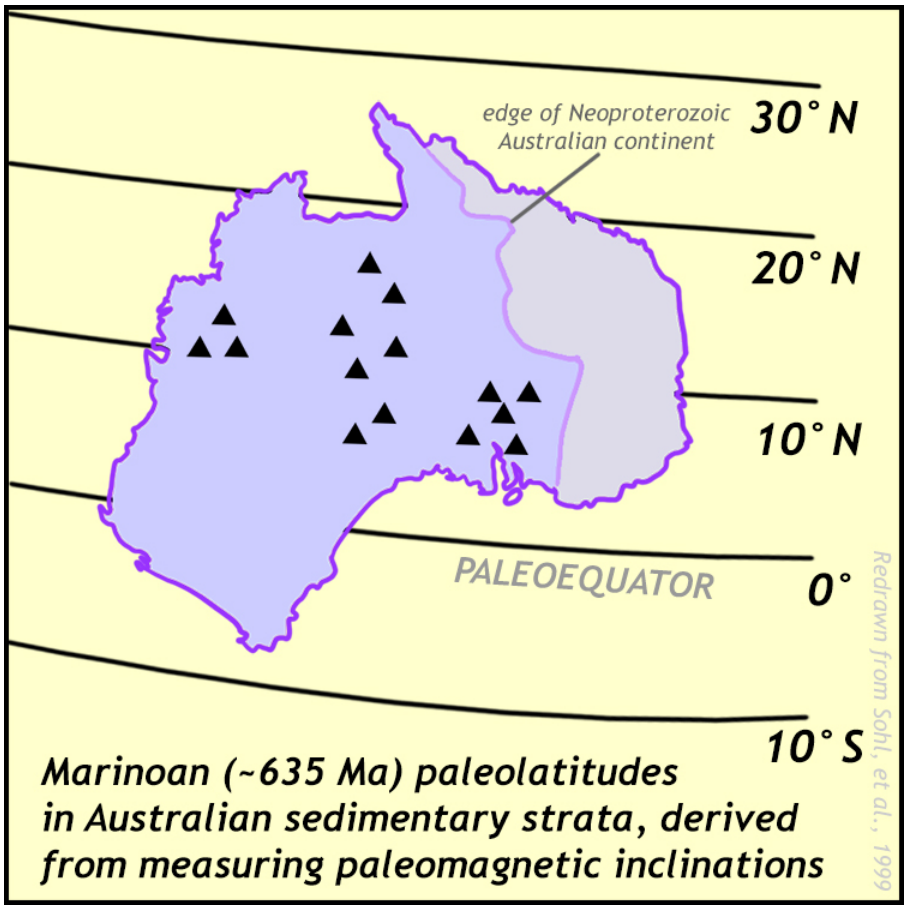


Figure 20.3.3: Marinoan-aged glaciogenic sedimentary rocks in central Australia show near-equatorial paleolatitudes. (Callan Bentley redrawing of a figure from Sohl, et al. (1999))

Now, this is a big deal, because the equator is usually pretty warm, and not a place where we would expect continental ice sheets to be flowing into ocean waters, shedding dropstone-laden icebergs.

Similar measurements of Sturtian (~710 Ma) sedimentary and volcanic rocks in ancestral North America show it too was positioned astride the equator during Neoproterozoic time.

This is important, because generally speaking the equator is a very warm part of the planet's surface: it's where the incoming rays of sunlight are most direct. As you move toward the poles from the equator, temperatures generally decrease. Therefore, if indeed there were glaciers flowing into the sea at such equatorial latitudes, it implies an exceptionally frigid climate for the planet as a whole.

Furthermore, because snow and ice are highly-reflective, such low-latitude ice would imply an overall very reflective planetary surface. In other words, an icy equator implies an icy planet, which would reflect away a lot of otherwise-warming incoming sunlight: it would have a high albedo. Among naturally occurring substances, snow and ice have the highest albedo (they are most reflective) and open ocean water has the lowest (it is most absorptive of incoming solar radiation).

Did I Get It? - Quiz

Exercise 20.3.1

Why are paleomagnetic records of low-latitude glaciation important?

a. They imply very low levels of oxygen in sea water during the time of the Snowball Earth.

b. They show us the ability of glaciers to erode the landscape into U-shaped valleys.

c. They show the sudden timing of the post-glacial melt-off, and ensuing deposition of "cap carbonate" strata.

d. If it was cold enough for glaciers at sea level at the equator, then it must have been cold enough for glaciers everywhere else on Earth.

Answer

d. If it was cold enough for glaciers at sea level at the equator, then it must have been cold enough for glaciers everywhere else on Earth.

Exercise 20.3.2

Which of the following has the highest **albedo?**

a. Snow

b. Ocean water

c. Rock

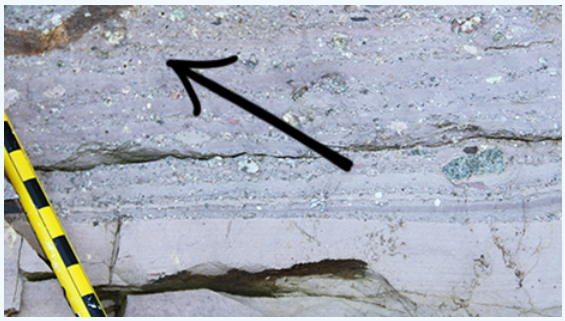
Answer

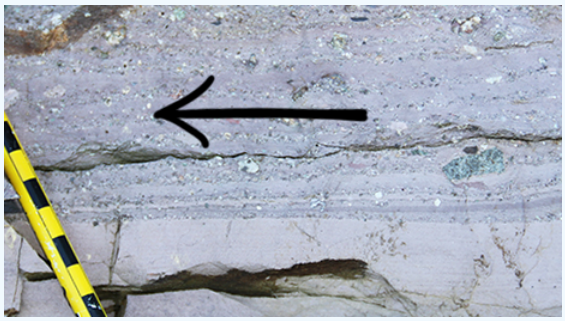
a. Snow

Exercise 20.3.3

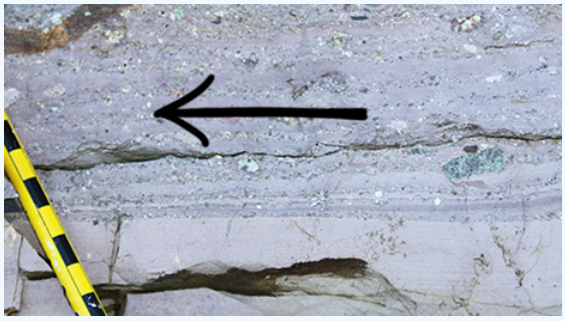
Consider these annotated photographs showing cross-sectional views of 3 hypothetical sedimentary layers. The arrows show the paleomagnetic inclination. Which of the strata formed **at the paleo-equator?**

a. 

b. 

c. 

Answer

c. 

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20.4: Cap carbonates

One surprising feature of Snowball Earth glaciogenic deposits is that many of these distinctive sedimentary rocks with their characteristic signatures of glaciation are comformably overlain by deposits of limestone and/or dolostone. Because the carbonate strata always appear on top of the glaciogenic strata, they are dubbed "**cap** carbonates."



Figure 20.4.1: Noonday dolomite cap carbonate comformably overlies glaciogenic Kingston Peak diamictite in Sperry Wash, near southern Death Valley. (Callan Bentley photo)

The most astounding thing that cap carbonates tell us is that temperature conditions abruptly switched from the chill of glacial conditions to balmy tropical warmth. Shallow water limestones form best today in places like the Bahamas, or offshore of Florida. The cap carbonates are conformable with respect to the glaciogenic sediments on which they are deposited. They are also much more extensive, covering vastly more area than the glaciogenic sediments alone, suggesting they were deposited during a major transgressive event. Several lines of evidence point to the rate of cap carbonate deposition being very rapid. The comformable relationship between pre-Snowball carbonates (warm) to during-Snowball glaciogenic sediments (cold) to post-Snowball cap carbonates (warm again) implies extreme temperature oscillations during the Neoproterozoic. These strata stretch our conception of the limits of climate change on Earth.

Overall, the cap carbonates show a transgression (as glacial ice melted and was added as water to the ocean) and a massive chemical precipitation of carbon-containing rock (which has major implications for climate controls). These extensive deposits indicate the ocean was in a state of carbonate oversaturation despite what must have been extraordinary levels of CO_2-induced ocean acidification in the wake of millions of years of volcanic outgassing with no carbon-removal processes operating on Earth's surface (see discussion below for details).

The cap carbonate transgressive sequences for Sturtian and Marinoan glaciations are distinctive and consistent in deposits of that age around the world.



Figure 20.4.2: Columnar “tubestone” stromatolites from Maieberg Formation, Namibia. (Photo by Jay Kaufman, reproduced with permission.)

Many cap carbonates show some primary sedimentary structures that suggest the post-Snowball hothouse Earth was a crazy place.

Not only was the temperature quite high, but apparently the cap carbonates were deposited very rapidly. This makes sense if we consider that the world’s surface consisted of a lot of pulverized rock interacting in shallow water with an atmosphere supercharged with carbon dioxide. As the CO_2 dissolved into the seawater to combine with calcium ions (and some magnesium ions too), huge quantities of carbonate were precipitated and blanketed the seafloor. As the glaciers melted, a lot of fresh water was added to the oceans, and sea level rose globally, flooding the recently-deglaciated land surface.

Overall, the cap carbonates include two major subunits: an initial dolostone, representing shallow conditions, and a later limestone, representing a deeper depositional setting. Within this transgressive sequence are several distinctive primary sedimentary structures:

1. tubestone stromatolites
2. giant wave ripples
3. aragonite crystal fans
4. barite seafloor cements



Figure 20.4.3: Columnar “tubestone” stromatolites from Sete Lagoas, Brazil. (Photo by Jay Kaufman, reproduced with permission.)

A distinctive primary sedimentary structure in the cap carbonates is the **tubestone stromatolite**. Stromatolites are fossilized microbial mats as we see in other carbonate rocks throughout geologic time. However, the cap carbonate stromatolites are often very thin, and grow in tight proximity to other another.

One interpretation is that these stromatolites have this distinctive form because they were growing in relatively shallow water (within the depths to which sunlight penetrates) and that the water was precipitating fine-grained carbonate rapidly, so that a form of natural selection takes over, and only those stromatolites that grow upward are those that don’t get buried survive to photosynthesize another day. The result is a dense collection of finger-like projections, with space in between, sometimes with short-lived microbial lamination bridging the gap over the fresh deposits of chemically-precipitated fine-grained carbonate. These finger-like stromatolites are sometimes organized into broad dome-like structures or mounds.

Above the tubestone stromatolites are accumulations of structures that resemble ripple marks, a signature bedform occurs-the **giant wave ripple**. These odd, large (1-6 m wavelength), symmetrical ripples grew upward without much sideways movement of the wave crest. They were initially thought to represent violent hurricane waves in the post-Snowball greenhouse climate, but this doesn’t explain why they grew straight upward. A preferred explanation is that they represent the record of a deeper wave base combined with the rapid cementation of the seafloor during cap carbonate deposition.

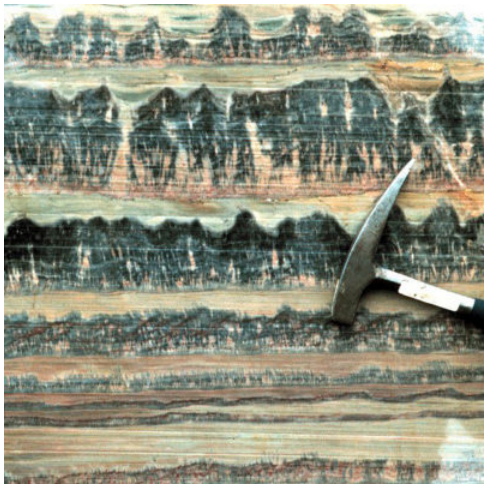


Figure 20.4.4: Multiple layers of aragonite crystal fans in cap carbonate at Sete Lagoas, Brazil. (Photo by Jay Kaufman, reproduced with permission.)

Another interesting feature that suggests very rapid deposition of the carbonate is layers full of **aragonite crystal fans**. (Aragonite is a polymorph of calcite.) These large blade-like crystals of $CaCO_3$ occur in layers within the overall cap carbonate sedimentary sequence. The long axis of the crystals is perpendicular to the ancient seafloor. Astonishingly, these crystals seem to imply such rapid crystallization of carbonate from the seawater that huge crystals several centimeters long grew upward into the supersaturated water column, and were buried as fast as they could form! It's very difficult to imagine conditions stable enough for 10 cm tall crystals poking up all over the seafloor, like a 'bed of nails,' and persisting for any length of time in violently oscillating ocean waves.

In Marinoan cap carbonates (but not Sturtian), **barite seafloor cements** appear exactly at the transition between the older, shallow-water dolostone deposits and the younger, deeper-water limestone deposits. In structure, these are similar to the aragonite crystal fans described above, but the mineral is different: it is a barium sulfate called barite, $BaSO_4$. The inference is that this distinctive layer formed when sulfate from surface meltwaters and barium (derived from feldspar weathering?) in deeper ocean waters mixed for the first time. If this is right, then the barite seafloor cements mark passage of the chemocline between these two reservoirs as transgression proceeded.

Did I Get It? - Quiz

? Exercise 20.4.1

At what rate were cap carbonates laid down?

a. Rapidly

b. Very gradually

Answer

a. Rapidly

? Exercise 20.4.2

Which of the following primary sedimentary structures are distinctive to cap carbonates?

a. Giant wave ripples, graded beds, and desiccation cracks (mud cracks).

b. Giant wave ripples, tubestone stromatolites, and seafloor crystal fans of aragonite.

c. Ball and pillow structure, tubestone stromatolites, and spinifex texture.

d. Aeolian cross-beds, coral reefs, and seafloor crystal fans of aragonite.

Answer

b. Giant wave ripples, tubestone stromatolites, and seafloor crystal fans of aragonite.

? Exercise 20.4.3

Cap carbonate sequences show an overall _____ sequence.

a. transgressive

b. conservative

c. regressive

Answer

a. transgressive

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20.5: Return of the BIF



Figure 20.5.1: A dropstone (of quartzite) in banded iron formation strata of the Rapitan Group, Northwestern Canada. (Photo by Galen Halverson; reproduced with permission.)

There is an additional chemical sedimentary rock that also appears in some Snowball Earth sedimentary sequences: banded iron formation (BIF). This is cited as evidence of radically altered oceanic chemistry during the Snowball glaciation, and therefore an indirect evidence of extensive sea ice cover.

As we saw in the Archean, the chemical precipitation of banded iron formation is a consequence of anoxic ocean waters. It is a broad generalization that BIFs are an “extinct rock” that “died out” at the Paleoproterozoic / Mesoproterozoic boundary with the build-up of copious free oxygen in the planet’s atmosphere and oceans. But that turns out to be not quite true.

During the Neoproterozoic Snowball Earth glaciations, BIFs make a reappearance in the geologic record. They were “resurrected” because the conditions that allow their formation occurred again, this time due to the Snowball.

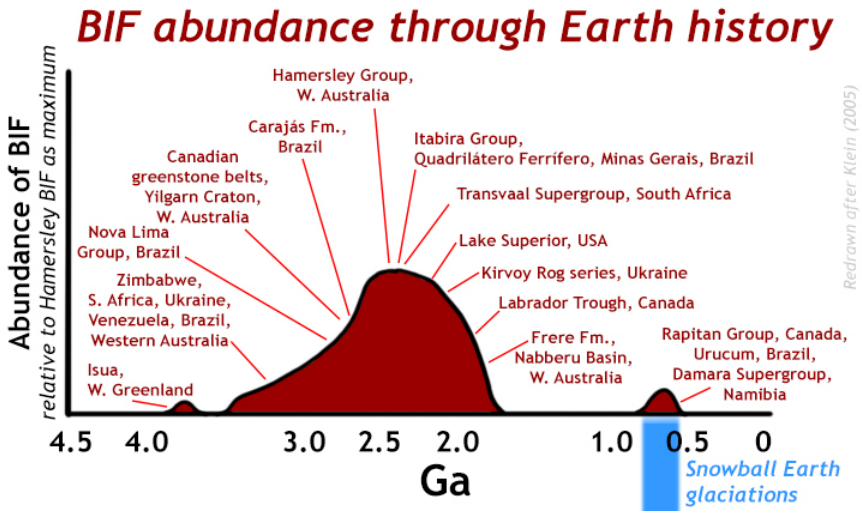


Figure 20.5.2: A plot showing the relative abundance of banded iron formation (BIF) through geologic time. Note the small “BIF blip” that coincides with the timing of the Neoproterozoic Snowball Earth glaciations. (Callan Bentley figure, redrawn from Klein (2005))

Because BIFs are the geologic signature of an anoxic ocean, their reappearance after a long absence is cited as strong evidence that sea ice over of Earth’s oceans was indeed global. That is, they were entirely or almost-entirely sealed off from interacting with the oxygenated atmosphere. Any dissolved oxygen in the oceans was consumed through chemical reactions, but was not replaced by diffusion from the atmosphere. This allowed deep sea black smokers to again permeate ocean water with dissolved iron.

Did I Get It? - Quiz

? Exercise 20.5.1

What rock type is evidence that significant portions of Earth's oceans were anoxic during the Snowball Earth glaciations?

a. cap carbonates

b. turbidites

c. banded iron formations

d. granites

e. evaporites

f. tidal rhythmites

Answer

c. banded iron formations

? Exercise 20.5.2

Why do anoxic ocean waters support the idea of a total planetary freezeover during the Snowball Earth?

a. It suggests that temperatures were too cold for oxygen to be stable as a gas, and instead it formed a liquid.

b. It suggests that the first animals (which were very cool) were sucking all the oxygen out of the oceans.

c. It suggests a lack of gas exchange between the atmosphere and the oceans: blocked by extensive sea ice cover.

Answer

c. It suggests a lack of gas exchange between the atmosphere and the oceans: blocked by extensive sea ice cover.

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20.6: When were the Snowballs?

We have already mentioned that when most geoscientists talk about the "Snowball Earth," they mean the glaciations that occured during the Cryogenian period of the Neoproterozoic era, specifically the Sturtian and the Marioan glaciations. The Sturtian occurred approximately 720 to 710 Ma, and the Marinoan around 660 to 635 Ma. Both the Sturtian and Marinoan glaciations show a global distribution of their sedimentary deposits. Their ages are not determined by fossils, but by relationships with rock units that are datable using isotopic methods (ash deposits, cross-cutting dikes, inclusions of datable rocks *etc*.).

Consider this hypothetical demonstration of the relevant information:

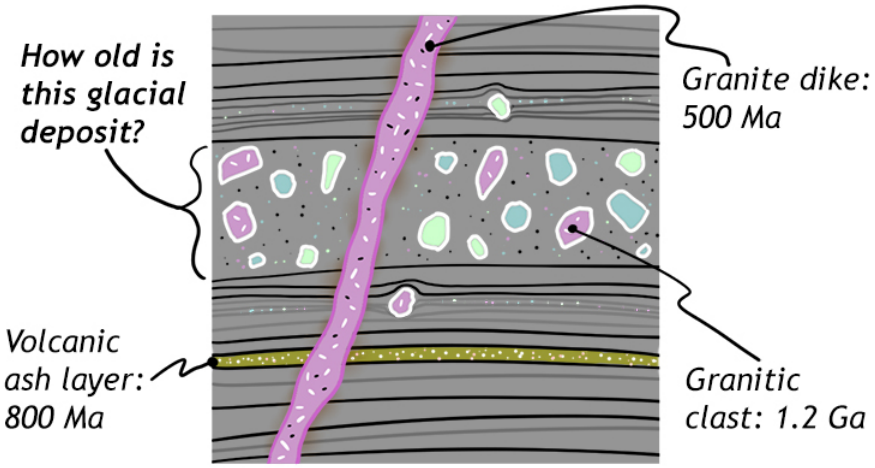


Figure 20.6.1: The age of this glacial deposit must be: (a) younger than the age of the clasts it contains, including the granite clast dated at 1.2 Ga, (b) younger than the volcanic ash layer stratigraphically below it (800 Ma), and (c) older than the granitic dike that cuts across it (which is 500 Ma). The inclusion date is redundant to the information provided by the ash in this case. We can constrain the age of the glacial deposit to be between 800 Ma and 500 Ma, based on the available information. (Callan Bentley cartoon)

These techniques are nothing special for Snowball Earth only, but merely the application of general relative dating principles to the question of the timing of Neoproterozoic glaciation.

Some scientists also consider the slightly younger Gaskiers glacation (635 to 542 Ma, during the Ediacaran period of the Neoproterozoic) as a "Snowball Earth" glaciation. However, the paleomagnetic signatures on Gaskiers deposits are relatively high-latitude, not tropical or equatorial. There are also a lot fewer sites that host Gaskiers-aged strata, though it is recorded from at least eight small paleocontinental landmasses. (This is in contrast to dozens apiece for the Sturtian and Marinoan.) Considering the Gaskiers is younger than the Marinoan and Sturtian, and therefore less likely to have been destroyed by subsequent rock cycle processes, this implies it was truly not as extensive in distribution. Furthermore the Gaskiers glaciation has no known associated banded iron formations, suggesting that ocean water continued to exchange oxygen with the atmosphere during glaciation, and therefore no global sheath of sea ice existed then. Finally, dating of volcanic ash deposits to constrain the age of the Gaskiers glaciation suggest that the entire ice age lasted for only about 340,000 years, which is inconsistent with the multi-million year expectation based on the "Snowball Cycle."



Figure 20.6.2: Gaskiers diamictite, Newfoundland. (Callan Bentley photo)



Figure 20.6.3: Gowganda tillite, Ontario. (Photo by Ron Schott)

If we go back in time, to the Paleoproterozoic era, we find that there was another episode of glaciation recorded, though whether it was also a true “Snowball” (in the sense of ice reaching all the way to the equator) is a difficult question to answer. It is sometimes called the Huronian glaciation or Makganyene glaciation. It dates to between 2.4 and 2.1 Ga, and is best preserved in North America in Ontario’s Gowganda Formation, a striking jet-black tillite that formed about 2.3 Ga.

(Being positioned in southern Ontario meant that the Gowganda’s glacial deposits were ideally positioned for plucking by Pleistocene ice age glaciers. This means that in some very *young* glacial deposits, we can find pieces of very *old* glacial deposits!)

Did I Get It? - Quiz

? Exercise 20.6.1

Which of the following "Snowball Earth" glaciations happened from 660 Ma to 635 Ma?

a. Marinoan

b. Sturtian

c. Ediacaran

d. Gaskiers

Answer

a. Marinoan

Exercise 20.6.2

What makes the Gaskiers glacation distinct from the Sturtian and Marinoan glaciations?

a. It is characterized by exceptionally mature deposits of quartz sand, well-rounded pebbles, and rock salt deposits.

b. It shows a more limited extent, has no banded iron formations, and is the youngest of the three.

c. It is the only "Snowball Earth" glaciation to occur in the Phanerozoic eon.

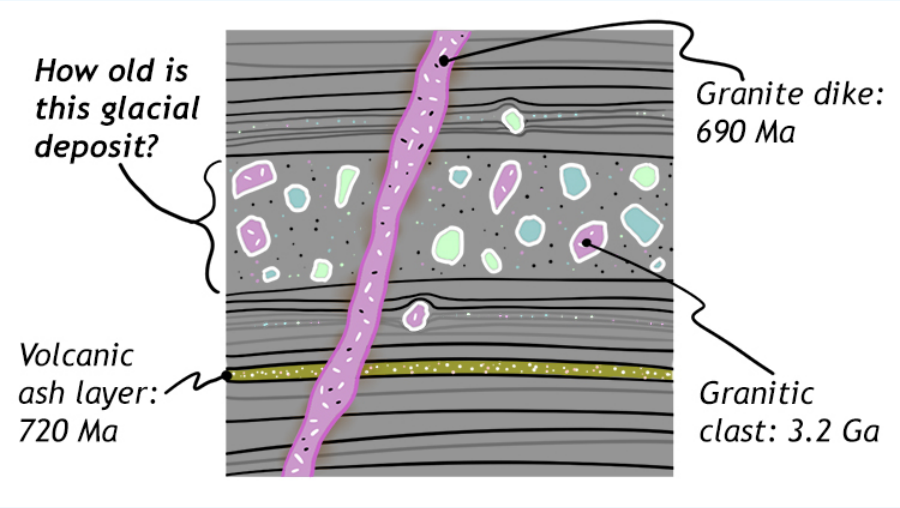
d. It shows a more extensive distribution of deposits, has the largest amount of associated banded iron formations, and is the oldest of the three.

Answer

b. It shows a more limited extent, has no banded iron formations, and is the youngest of the three.

Exercise 20.6.3

As precisely as possible, how old is the glacial deposit illustrated in this cross-section?



a. Between 3.2 Ga and 720 Ma (and thus likely Marinoan).

b. Between 720 Ma and 690 Ma (and thus likely Sturtian).

c. Between 690 Ma and 330 Ma (and thus likely Gaskiers).

d. Between 3.2 Ga and 690 Ma (and thus likely Sturtian).

Answer

b. Between 720 Ma and 690 Ma (and thus likely Sturtian).

Exercise 20.6.4

Are there any records of glaciation on Earth prior to the Neoproterozoic Snowball Earths?

a. No

b. Yes

Answer

b. Yes

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20.7: Starting and stopping a Snowball

The Snowball Earth glaciations have important implications for our understanding of the role of greenhouse gases in regulating Earth's climate. As you have learned, the planet's surface temperature depends on several key variables, including the amount of energy the Sun emits, the planet's position relative to the Sun, the ability of the planet to redistribute heat on its surface (mainly via ocean currents), and the ability of the planet's atmosphere to retain heat that would otherwise be radiated out into space.

It is in this last category that greenhouse gases matter. They allow visible light from the Sun to pass through them unimpeded, but block infrared radiation – the kind that the planet's rocks and oceans re-emit after basking in the sunlight for a while.

The most accepted model for the initiation of the Snowball Earth glaciations depends exquisitely on the level of the greenhouse gas carbon dioxide, CO_2, in Earth's atmosphere. The explanation goes like this: in the late Proterozoic, a supercontinent called Rodinia formed in Earth's equatorial region. Because of its direct rays of incoming sunlight, the tropics are today home to (1) warm temperatures and (2) a humid climate with lots of rain. Lots of water vapor can be stored in warm air. Assuming that these same warm, wet conditions applied to the tropical belt during the Neoproterozoic, there would have been enhanced chemical weathering of the rocks in that belt. The reactions of chemical weathering consume CO_2, such as the process by which feldspar (the most common mineral in Earth's crust) is converted to clay (a widespread constituent of sedimentary rock). This means that as chemical weathering proceeded, CO_2 levels in the atmosphere would have fallen. Typically, weathering-induced CO_2 drawdown is balanced by volcanic CO_2 emissions. But if the rate of tropical-weathering-induced CO_2 drawdown was greater than the volcanoes could compensate for, atmospheric CO_2 levels would have fallen. And because that means there would be less of a greenhouse effect, the temperature would have gotten colder.

The next stage of the story takes our attention away from the equator and focuses it instead on the polar regions, where incoming sunlight is much less direct. As a general rule, these are the coldest parts of our planet. As global CO_2 was drawn down, ice would have begun to grow at the poles, forming floating sea ice that grew and grew through time, reaching into progressively lower and lower latitudes.

Sea ice has a high albedo: about 92% of incoming sunlight is reflected away. Very little energy is absorbed from the incoming sunlight. In contrast, ocean water has an extremely low albedo: only about 8% of incoming sunlight is reflected away, and the rest is absorbed as heat energy.

As sea ice built up in the polar regions, its high albedo reflected away more and more of the energy the planet was receiving from the Sun. This constantly-increasing albedo was an amplifying feedback system that took the CO_2-induced cooling and made it even more pronounced. Eventually, when the ice reached about 33° of latitude, the albedo-induced cooling effect was unstoppable, and the remainder of the planet's surface, including the equator, froze over in as little time as a decade.

The Snowball Earth had begun. The average global temperature at this time is estimated to have been -50 °C.

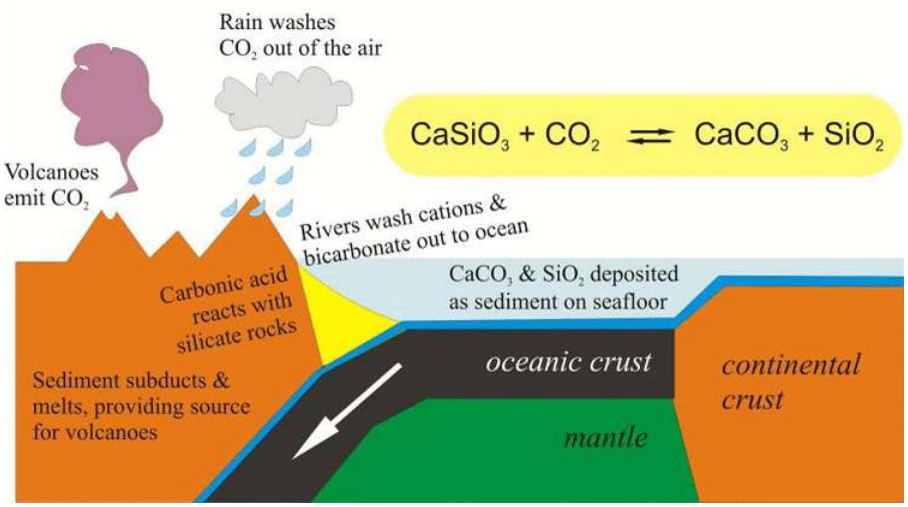


Figure 20.7.1: How the Neoproterozoic carbon cycle is thought to have changed to initiate a Snowball Earth glaciation (and sow the seeds for its eventual me b.gif">original by Paul Hoffman (2006)).

Sea ice sealed off gas exchange between the liquid ocean below and the atmosphere above, resulting in anoxic waters developing in much of the sea. It was at this time that banded iron formation was resurrected. Up above, during the unbelievable cold of the Snowball, much of the the carbon cycle would no longer be functional. Chemical weathering rates would be effectively nil. The planet was locked in an icy death grip, so reflective it could never warm up again… right?

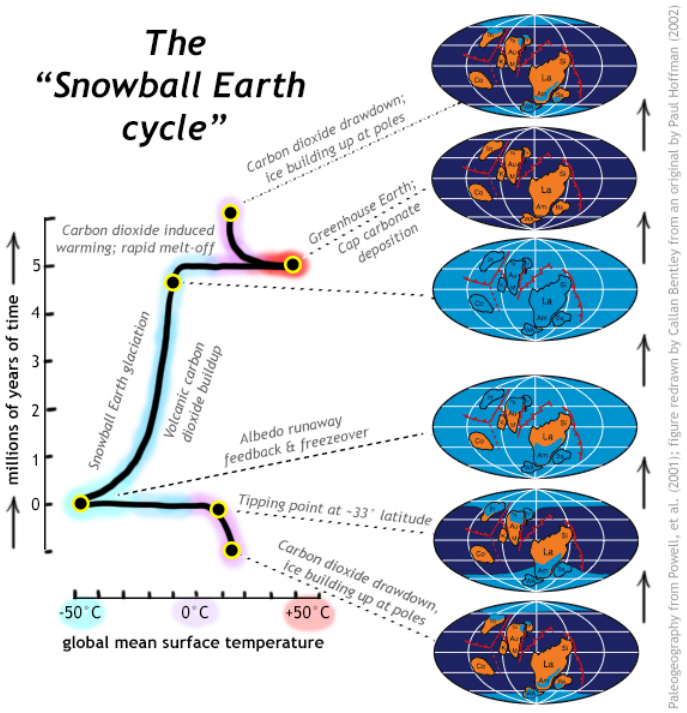


Figure 20.7.2: The theoretical cycle of events that turned on and off the two "hard" Snowball Earth glaciations. (Callan Bentley figure, redrawn from an original by Paul Hoffman (2002))

Wrong! (Obviously, since we're here to talk about it.) You will recall that plate tectonics is not driven by energy from the Sun. Instead, it operates from Earth's internal heat, driven in part by radioactive decay. So under the sea ice and under the glaciers, the plates continued their slow lateral movement. More to the point, subduction continued and seafloor spreading continued, and both of these processes resulted in volcanism. During the Snowball Earth, volcanoes continued to erupt, producing lava and ash but more importantly, emitting gases. These gases included water vapor (about 60% by volume) and carbon dioxide (about 15% by volume). The water vapor would have frozen, but the CO_2 would not.

As the volcanoes continued to erupt, CO_2 levels in the atmosphere began to build up again. With no weathering to extract the CO_2 from the air, the amount of this greenhouse gas began to increase unchecked. Eventually, after about 5 million years or so, enough CO_2 would have accrued that the resulting greenhouse heat retention would have overcome the albedo's cooling effects. A few patches of sea ice would have melted –probably at the equator — and the exposed ocean water would have efficiently absorbed incoming solar energy. This melted more ice, which exposed more ocean water, and once again an amplifying feedback loop took over, rapidly melting off the sea ice (and glaciers) in short order.

The Snowball Earth was over; the post-glacial hothouse had begun. With an atmosphere chock full of CO_2 and a low albedo allowing maximum absorption of incoming solar radiation, the average global temperature at this time is estimated to have been +50 °C.

The CO_2-charged atmosphere and pulverized glacial sediment (full of calcium ions) then equilibrated, reacting to produce a tremendous amount of carbonate sediment that rapidly precipitated in the oceans. As swollen, warm seas transgressed onto the continents, the cap carbonates were laid down, rapidly and perhaps violently. Gradually, CO_2 was drawn down again.

Overall, this story is thought to have repeated at least twice (Sturtian and Marinoan), and perhaps with lesser intensity a third time (Gaskiers). Because of its repetition, some geoscientists have suggested it as a kind of cycle: the Snowball cycle.

Exercise 20.7.1

What is the sequence of geological events/processes that is thought to have initiated the Snowball Earth glaciations?

a.

1) CO_2 drawdown through intense weathering of the exposed crust in the tropical weathering belt,

2) Global cooling causing sea ice build up in polar regions and growing into lower latitudes,

and finally,

3) Runaway albedo-driven amplifying feedback, tipping the planet's surface into a totally frozen state.

b.

1) Global cooling causing sea ice build up in polar regions and growing into lower latitudes,

2) Runaway albedo-driven amplifying feedback, tipping the planet's surface into a totally frozen state,

and finally,

3) CO_2 drawdown through intense weathering of the exposed crust in the tropical weathering belt.

c.

1) Runaway albedo-driven amplifying feedback, tipping the planet's surface into a totally frozen state,

2) Global cooling causing sea ice build up in polar regions and growing into lower latitudes,

and finally,

3) CO_2 drawdown through intense weathering of the exposed crust in the tropical weathering belt.

Answer

a.

1) CO_2 drawdown through intense weathering of the exposed crust in the tropical weathering belt,

2) Global cooling causing sea ice build up in polar regions and growing into lower latitudes,

and finally,

3) Runaway albedo-driven amplifying feedback, tipping the planet's surface into a totally frozen state.

Exercise 20.7.2

How do we think that Earth escaped from the Snowball Earth glaciations?

a. Meteorite impact

b. Being warmed by lots of volcanic lava

c. Volcanic ash deposits darkening snow cover and reversing the albedo-induced cooling effect

d. Volcanic CO_2-induced global warming

Answer

d. Volcanic CO_2-induced global warming

Problems with the traditional Snowball Earth model

The Snowball Earth is a controversial topic among some geologists. The evidence for widespread glaciation is incontrovertible, but whether the entire planet froze over entirely (a “hard” Snowball) is less agreed-upon. Perhaps it was cold, but not quite -50 °C. After all, somehow life survived through the Snowball, and it is hard to imagine how that would work. Perhaps there were “refugia,” protected places or warmer-than average places: adjacent to deep-sea hydrothermal vents, perhaps, or polynyas that are maintained by currents flowing from beneath. So there are ways of getting around that objection.

A harder problem to surmount is the way that glaciers flow. In the modern world, where we observe glacial ice, it flows downhill, from its zone of accumulation toward its zone of ablation. We have ample evidence that there was glacial flow, glacial erosion, and glacial deposition. But in order for that all to happen, a functional hydrological cycle is probably a requirement. Glaciers only flow in the modern world because they get fresh ice forming at their upstream end, due to precipitation of snow. But that precipitation wouldn’t be possible without evaporation of water from somewhere else. If the planet were truly frozen over during the Snowball Earth glaciations, there should have been dramatically reduced potential for evaporation, not only because of the low temperatures (meaning low amounts of energy to drive evaporation), but more importantly because of the impact of the sea ice acting as a physical barrier: sealing off potential exchange between the ocean and the atmosphere. If the Earth experienced a true Snowball, then how did the glaciers keep flowing?

For this reason, a substantial number of geoscientists prefer a milder “Slushball Earth” model as their interpretation for Neoproterozoic glaciation.

Did I Get It? - Quiz

? Exercise 20.7.1

A major argument against a "hard Snowball" model for Neoproterozoic glaciation is that...

a. ...no life survived these ice ages.

b. ...glaciers need a functional water cycle in order to continue to flow. The fact that we have plenty of evidence they did flow (and erode rock) during the Neoproterozoic suggests that there must have been open water somewhere to allow evaporation.

c. ...it just seems so hard to accept that the world could change that much.

Answer

b. ...glaciers need a functional water cycle in order to continue to flow. The fact that we have plenty of evidence they did flow (and erode rock) during the Neoproterozoic suggests that there must have been open water somewhere to allow evaporation.

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20.8: Impact on animal evolution

One of the most intriguing coincidences in the geological record is that “immediately” after the Snowball Earth glaciations, some sedimentary layers hold the very first evidence of multicellular animals preserved as fossil remains. These are the Ediacara, namesakes of the Ediacaran period.

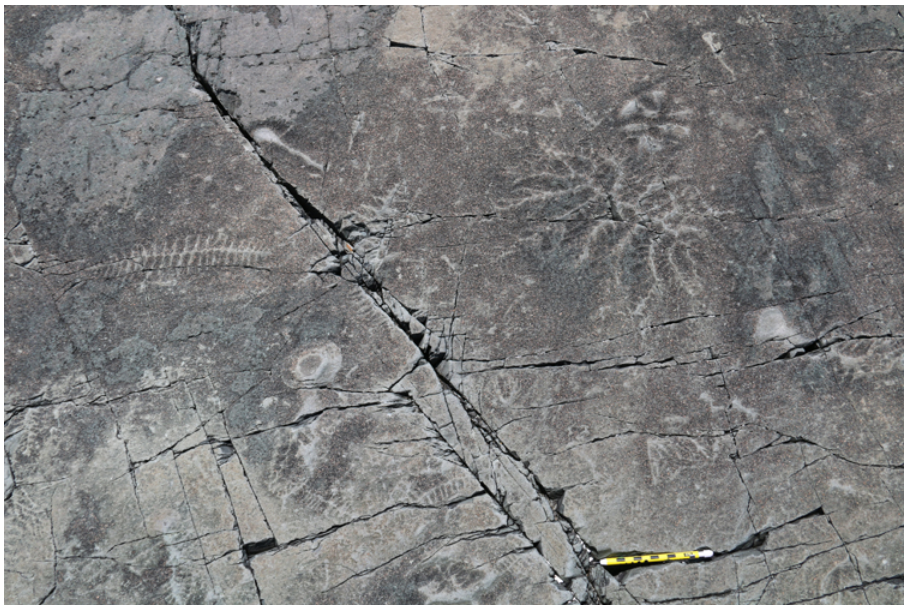


Figure 20.8.1: Diverse fossil Ediacara on a bedding plane at Mistaken Point, Newfoundland. Pencil for scale. (Callan Bentley photo)

Exactly *what* the Ediacara *were* isn’t entirely clear. Animals, probably. What we know is this:

- They were big enough to see with the naked eye, ranging between a couple of centimeters and a couple of meters in length.
- They had morphologically distinct body parts, suggesting that they had differentiated tissues, if not organs.
- They lacked hard parts such as bones, shells, horns or teeth. (The fossils are preserved as impressions of their soft bodies in shale and sandstone.)
- Many appeared to be sessile (rooted in one spot), though others were probably mobile, on the basis of track-like trace fossils associated with some of the body fossils. Some resemble jellyfish, and so may have floated in a similar fashion.
- Most showed bilateral symmetry, though there are others with apparent radial symmetry, and at least one has apparent trilateral (3-part!) symmetry.



Figure 20.8.2: A pair of fossil *Fractofusus*, distinctive Ediacaran organisms, on a bedding plane at Mistaken Point, Newfoundland. Pencil for scale. (Callan Bentley photo)

In Newfoundland, the end of the Gaskiers glaciation occurred at 579, and the oldest Ediacara fossils are about 5 million years younger. There are sparse examples for several million years, and then at Mistaken Point, a legendary Lagerstätte on the southeastern Avalon Peninsula, a 565 Ma ash bed smothered a huge number of Ediacara on a single bedding plane. A few meters above that, the same thing happened again. This unusual circumstance provides for exceptionally precise time constraints on the age of the organisms killed by the ash fall. Mistaken Point is the oldest known fossil assemblage of diverse, large, biologically complex fossilized animals. All told, Ediacara lived for 33 million years prior to the Cambrian "explosion." The fact that Ediacara evolved so soon (geologically speaking) after the end of the Snowball Earth suggests the great global catastrophe may somehow have nurtured the conditions for the evolution of these first animals.

But how? One obvious answer is by wiping out whatever came before, and "leveling the playing field" for up-and-coming animals with their new evolutionary innovations. Another straightforward variable was doubtless the increase in shallow sea habitat as glacial melting caused sea level rise, flooding the continents with shallow seas, perhaps an ideal habitat.

More esoteric explanations hinge on an increase in oxygen levels after the Snowball, and how those oxygen levels then provided sufficient fuel for oxygen-consuming animals to exists for the first time. Or perhaps because collagen is an important binding

compound that holds animal bodies together, and collagen requires plenty of oxygen to form, that may be the most important aspect of the rise in oxygen. Exactly why oxygen rose isn't entirely clear, though.

Obviously, being an animal comes with some big advantages, as large bodies with sensing tissues distributed throughout are able to sense gradients well (hot/cold, salty/fresh, toward a greater concentration of food molecules/away), and their large size (relative to microbes and single-celled protists) also potentially helps buffer against short-term hostile variations in environmental conditions. These are both innovations that natural selection could emphasize through time.

Did I Get It? - Quiz

Exercise 20.8.1

What is the next big milestone in Earth history after the melting of the last Snowball Earth glaciations?

a. The evolution of the first dinosaurs.

b. The evolution of the first arthropods.

c. The evolution of the first animals.

d. The evolution of the first plants.

e. The evolution of the first humans.

Answer

c. The evolution of the first animals.

Exercise 20.8.2

Which of the following were structures that Ediacaran organisms had?

a. Shells

b. Differentiated tissues

c. Teeth

d. Eyes

Answer

b. Differentiated tissues

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20.9: Conclusion

The evidence for widespread glaciation during the Neoproterozoic's Cryogenian and Ediacaran periods is indisputable. Several packages of glacial sediments (tillites, marine units bearing dropstones, *etc*.) have overlying "cap" carbonates, interpreted as warm-climate limestones laid down during post-glacial transgressions. One model to explain these glacial deposits is the "Snowball Earth," which suggests the glaciation was triggered by (1) CO_2 drawdown due to intense tropical weathering of the crust that then (2) led to a runaway ice-albedo feedback loop. The planet escaped the Snowball due to (3) volcanic outgassing building CO_2 levels up anew, and then in the aftermath of the glaciers melting, CO_2 dissolved into seawater, combined with Ca and Mg ions, and (4) laid down thick packages of carbonate rock. The Sturtian and Marinoan glaciations both show banded iron formations associated with their glacial deposits, suggesting widespread (if not global) sea ice cover triggering ocean anoxia. A third glaciation, the Gaskiers, is younger, shorter, and less widespread than the other two, but shortly after it ended, the first fossil animals (called Ediacara) show up in the geologic record.

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CHAPTER OVERVIEW

21: (Case Study) Endosymbiosis

A major innovation in living organisms came about through the cooperation of two early simple cells, working together for their common benefit. **Endosymbiosis** is a variety of mutual symbiosis in which one organism lives inside another. This process represents a merger of branches of the "tree of life," and is thus a form of horizontal gene transfer. The evolution of organelles (small, functional units within cells) resulting from endosymbiosis have led to major revolutions in the history of life, including the first protists and the eventual development of multicellular organisms such as plants, fungi, and animals.

21.1: Making eukaryotes from prokaryotes

21.2: A history of the idea

21.3: Mitochondria

21.4: Chloroplasts

21.5: Other organelles

21.6: Further reading

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21.1: Making eukaryotes from prokaryotes

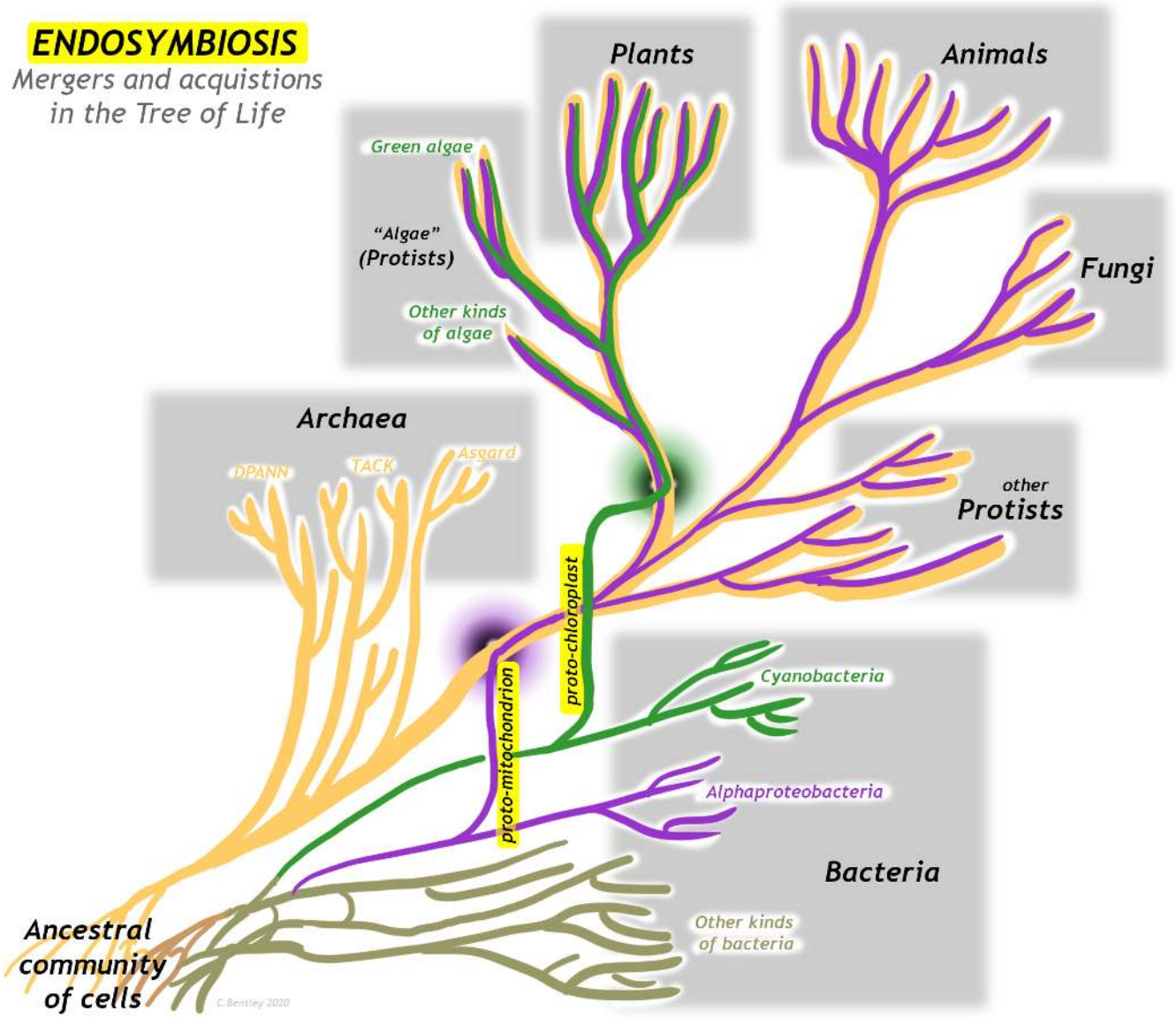


Figure 21.1.1: An interpreted "tree of life" showing proposed relationships between the two domains of prokaryotes (Bacteria and Archaea) and the Eukaryotic line. Note the two highlighted episodes of endosymbiosis.

Eukaryotic cells are the large, complicated cells that make up the bodies of animals, plants, fungi, and protists. With their well-defined nucleus, they are distinct from the smaller, simpler cells that comprise the bodies of single-celled microbes. Microbial cells are prokaryotes instead, and come in two varieties: bacteria and archaea. Bacteria and Archaea strongly resemble one another in size and form (including their lack of a nucleus), but are distinguished from one another on the basis of utterly distinct genetics and biochemistry.

Though the domain Eukarya contains a diversity of forms, the defining characteristic of a eukaryote has long been recognized as its membrane-bound nucleus, the site where its genetic material is stored (in the form of DNA) and transcribed (in the form of RNA). The *–karyotic* part of the names "eukaryote" and "prokaryote" translates as *kernel* or *nugget*, a reference to the distinct visibility of the nucleus in those cells that have one. However, the mitochondrion could be considered just as defining and ubiquitous a structure for eukaryotes. Only a few anaerobic eukaryotes lack mitochondria. The structure, function, and genetic signature of the mitochondrion gives every indication that it represents the "capture" of a bacterial cell by an archeal cell sometime around 1.45 Ga (Mesoproterozoic).

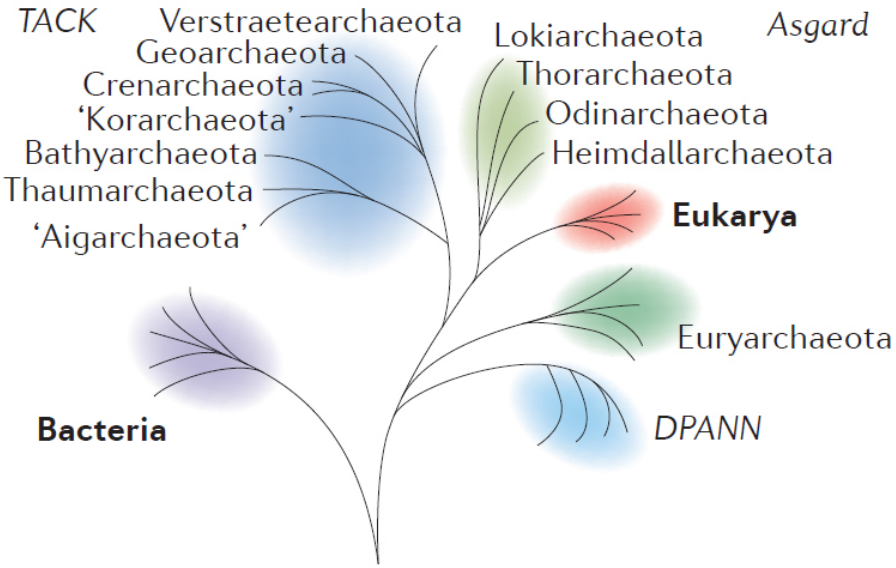


Figure 21.1.2: A conceptual "tree of life" showing the relationship between Bacteria (purple), Eukarya (red), and several groups of Archaea (blues and greens). The archeal clade called "Asgard" is most closely related to line that would become eukaryotic host cells. Not shown is the endosymbiotic merger between bacteria and ur-Asgardians that allowed the eukaryotic line to innovate and flourish. (Modified from Figure 1 of Eme, et al. (2017).)

There is great diversity among the Archaea, and several major groups within that domain had split off already by the time one sub-group had some members engage in endosymbiosis while others didn't. The group of Archaea most closely related to the eukaryotes are the superphylum Asgard. The Asgardians are more related to the "Eukaryotic" host cell line than they are to other groups classified as fellow archeons. Somewhere at or just after the branching point between Asgard and Eukarya was the moment when the first proto-mitochondrion was engulfed. The endosymbiosis between the Asgard-like archeon and its future mitochondrion became cemented into a permanent and irrevocable interdependency through geologic time. What was initially perhaps happenstance and casual eventually evolved into something permanent and obligatory, an innovation that was wildly successful in allowing diversification of life on Earth.

Endosymbiosis is often thought of as the larger organism "engulfing" the smaller, but it could just as legitimately be viewed as the "colonization" of the larger organism by the smaller one. Either way, each organism benefits from the relationship. The net result is a genetic chimera, a blending and a merger between two organisms that used to be distinct, but when combined make something new and different. All eukaryotes are chimeras, some several times over.

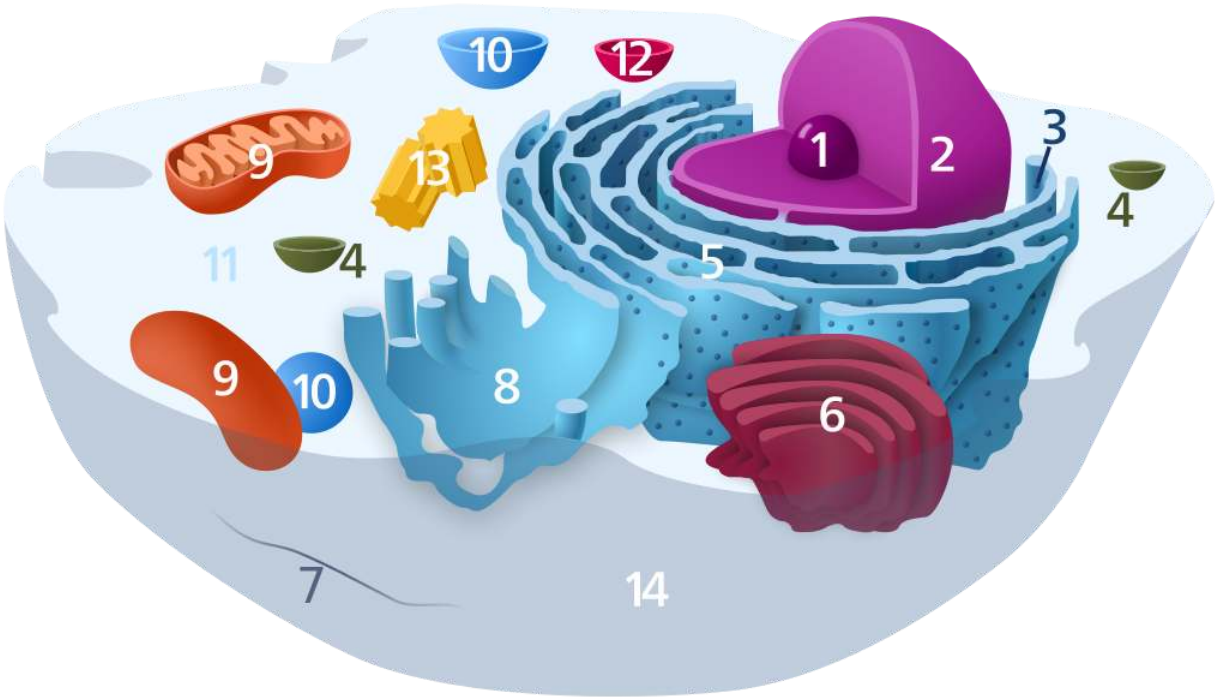


Figure 21.1.3: Components of a typical animal cell:
1) Nucleolus; 2) Nucleus; 3) Ribosome (dots as part of 5); 4) Vesicle; 5) Rough endoplasmic reticulum; 6) Golgi apparatus; 7) Cytoskeleton; 8) Smooth endoplasmic reticulum; 9) Mitochondrion; 10) Vacuole; 11) Cytosol (fluid that contains organelles; with which, comprises cytoplasm); 12) Lysosome; 13) Centrosome; 14) Cell membrane. (Public domain image via Wikimedia.)

What is true for mitochondria among all eukaryotes is also true for another organelle in some eukaryotes: plants and the various forms of "algae" *(which are protists)* contain chloroplasts, organelles that have the ability to conduct photosynthesis. In structure, function, and genes, the chloroplasts resemble free-living cyanobacteria, and are thus thought to represent a second episode of endosymbiosis unique to the photosynthetic eukaryotes.

In both cases, the smaller symbiont retains its own DNA, and reproduces independently of the rest of the host cell. Over geologic time, the genome of the internal symbiont will be pared down to eliminate redundancy, with the genome sometimes getting trimmed down to a third the size of the length of the DNA in their free-living relatives. Some of its genes may be copied or moved into the host genome. The two organisms may become locked into interdependency, where neither can survive without the other. At this point, it starts to make more sense to refer to the internal symbiont not as an "organism" in its own right, but as an organelle within the host cell.

There are two major organelles in eukaryotic organisms that have resulted from endosymbiosis, mitochondria (common to most eukaryotes) and chloroplasts (found in plants and algae only), as well as a few less obvious examples. After reviewing the scientific history of the idea, this case study will examine each organelle in turn.

Did I Get It? - Quiz

? Exercise 21.1.1

Which organelles in eukaryotes are thought to have started due to endosymbiosis of bacteria in archaea?

a. Mitochondria and vacuoles

b. Mitochondria and chloroplasts

c. Vacuoles and chloroplasts

d. The Golgi apparatus and mitochondria

e. Vacuoles and the Golgi apparatus

f. Chloroplasts and the Golgi apparatus

Answer

b. Mitochondria and chloroplasts

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21.2: A history of the idea

In the early 1880s, German botanist Andreas Schimper used the most cutting-edge microscopes available to study plant cells. He noted that chloroplasts grew and divided independently of the rest of the cell. This was a novel insight, and confounded expectations. Copying the experiments of his contemporary Louis Pasteur, Schimper demonstrated that a plant cell which has lost its chloroplasts cannot regenerate them. Like life itself, in the modern day, all chloroplasts descend from other chloroplasts. Schimper's 1883 conclusion was that chloroplasts were endosymbiotic cyanobacteria. Another botanist, this one from Russia, Konstantin Merezhkovsky (also spelled Mereschkowski) built on this idea, proposing in 1905 that algae and plant cells were chimeras between a protozoan (a protist) and a cyanobacterium nestled inside it. However, critical empirical tests were missing at the turn of the century: there was no DNA sequencing available then, nor sufficiently advanced microscopy to investigate Merezhovsky's ideas to the extent they deserved. Thus, the hypothesis of chloroplast endosymbiosis withered shortly after it was first suggested. The technology that would show it to be true had not yet been invented. An idea this radical needed testing to remain viable; Without it, the idea went dormant for several decades.

Shortly thereafter, American biologist Ivan Wallin developed an interest in mitochondria. In 1922, he noted that the little rod-shaped organelles were so similar to bacteria that the two were frequently mistaken for one another. Intriguingly, the same staining techniques that worked to highlight mitochondria also worked on bacteria. Wallin concluded that, "mitochondria are symbiotic bacteria in the cytoplasm of the cells of all higher organisms." Again, the idea was plainly stated but did not gain widespread acceptance due to the lack of confirmation from other more precise lines of evidence.

At the same time as Wallin's work in America, another botanist in Russia independently came up with the same idea expressed decades earlier by Merezhovsky. In 1924, Boris Kozo-Polyansky published a paper again documenting the similarities between chloroplasts and cyanobacteria, extrapolating the larger insight that, "new organisms occur not only by divergence of the lineages but also by their convergence and fusion." Again, the ground was not fertile for the reception of the provocative notion, and it once more faded into obscurity.



Figure 21.2.1: Lynn Margulis in 2005. (Photo by J. Pedreira via Wikimedia Commons.)

The idea of endosymbiosis would see a revival and ultimate validation in 1967 when Lynn Margulis (then Lynn Sagan) [*yes, that Sagan*] proposed that not only was the endosymbiosis notion true for chloroplasts, it also applied to mitochondria, and maybe flagella too. A rhetorical force of nature, when Margulis championed endosymbiosis, it stuck. Widespread scientific consensus wasn't only the result of Margulis's tireless advocacy, but also because of the advance in scientific technology since Merezhovsky's day. For one thing, electron microscopy now existed, and was capable of showing how similar the internal structure of chloroplasts was to the architecture of cyanobacteria. Moreover, it showed the double-membraned envelope of chloroplasts, a signature of having been enfolded from the exterior to the interior of the host eukaryote. Seeing these details simply wasn't possible prior to sufficiently detailed imaging technology.

More recently, the ability to extract and sequence DNA has shown the similarity between the genes of organelles like the mitochondrion and the chloroplast and free-living bacteria. The two structures shared many of the same genes, in the same way that familial cousins do. This was the detail that cinched the case for most scientists. With multiple lines of evidence converging on the same answer, it was time to accept that answer: **Eukaryotes are chimeras resulting from endosymbiosis.**

Because eukaryotes are capable of feats that could never be accomplished by a lone prokaryote, a philosophical corollary is that evolutionary innovation can be achieved by cooperation rather than solely through competition. The idea of endosymbiosis offers a different lens through which to view the evolution of life on Earth: it turns "survival of the fittest" on its head, and suggests instead "survival of the most well-integrated team."

Did I Get It? - Quiz

Exercise 21.2.1

Which scientist finally saw success in establishing the idea of an endosymbiotic origin for mitochondria and chloroplasts among the wider scientific community?

a. Ivan Wallin

b. Boris Kozo-Polyansky

c. Konstantin Merezhkovsky

d. Lynn Margulis

e. Andreas Schimper

Answer

d. Lynn Margulis

Exercise 21.2.2

What was the reason that the wider scientific community accepted mitochondria and chloroplasts as obligate endosymbionts in the late 1960s and 1970s?

a. The discovery of cyanobacteria, which were previously unknown

b. Advances in the technologies of microscopy and genetic sequencing

c. Documentation that mitochondria respond to the same laboratory staining techniques as bacteria

d. Records that chloroplasts reproduce independently of the rest of the cell

Answer

b. Advances in the technologies of microscopy and genetic sequencing

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21.3: Mitochondria

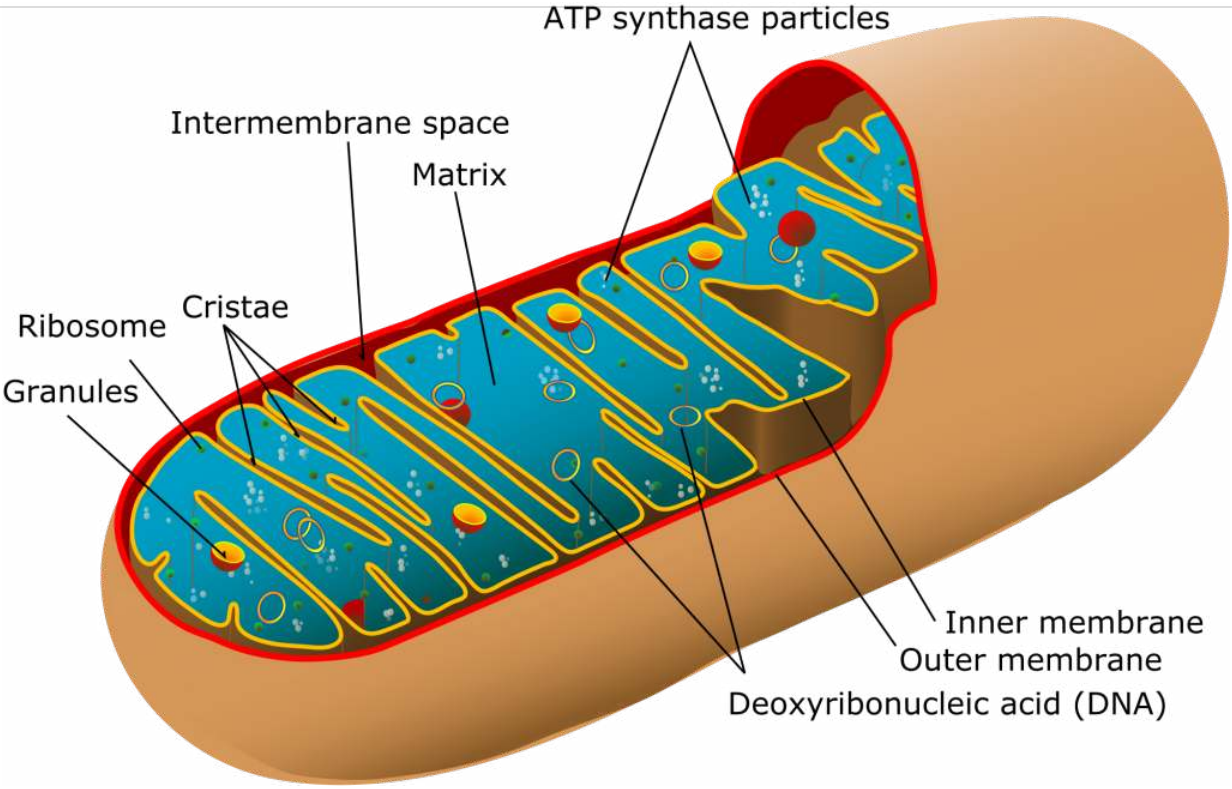


Figure 21.3.1: Structure of the mitochondrion. (Illustration by Mariana Ruiz Villarreal, via Wikimedia Commons.)

Mitochondria are organelles found in many eukaryotes, including animals, plants, fungi, and many varieties of protist. They are bounded by two membranes, the outer one smooth and pill-shaped, the inner one intensely folded into fin-like creases. This gives the inner membrane a very large surface area, crumpled into a small volume of space. On average, they are about 1 μm long. The number of mitochondria per cell can vary between zero (in some anaerobic protists) and upward of 2000. Functionally, mitochondria are critical because they produce the molecule adenosine triphosphate (ATP), a vital molecule used as a conveyance of chemical energy in the cell. To do this, they catalyze the reaction of free oxygen (O_2) with pyruvate *(a product of the digestion of glucose)*, and generate carbon dioxide (CO_2) and water (H_2O) as waste products in addition to the ATP. In most mitochondria in most eukaryotes, about 36 units of ATP are produced as outputs per unit of glucose ingested. In other words, mitochondria are where respiration happens: where energy is derived from the food we eat, “burning” it chemically by oxidizing it, and “charging up the batteries” of our bodies.

Mitochondria are probably descended from a common ancestor shared with modern free-living bacteria, specifically the group called alphaproteobacteria (sometimes written as α-proteobacteria). We think this for several reasons. First, mitochondria only come from other mitochondria. Like bacteria, they reproduce through binary fission. If they are removed from a cell, the cell cannot replace them on its own. Within the inner membrane, the mitochondrion contains DNA, RNA, and ribosomes: all the apparatus for storing genetic information and translating it into proteins. Second, comparison of the ribosomal RNA in mitochondria and other extant groups shows the mitochondria to be most similar to the alphaproteobacteria. The extant alphaproteobacterial genus *Rickettsia*, a tickborne pathogen, comes closest in relatedness to the proto-mitochondrion. Interestingly, the tetracycline antibiotic medications used to treat tickborne illnesses such as Rocky Mountain spotted fever (such as Doxycycline) can also negatively impact the functioning of mitochondria in animal cell test cultures.

Did I Get It? - Quiz

? Exercise 21.3.1

Mitochondria are valuable organelles in eukaryotes because _____.

a. they help fight off infections during times of cellular stress

b. they are the site of photosynthesis, the chemical process that builds glucose from smaller molecules using the Sun's energy

c. they make the energy-rich molecule ATP and share it with the rest of the surrounding cell

Answer

c. they make the energy-rich molecule ATP and share it with the rest of the surrounding cell

? Exercise 21.3.2

Which group of modern bacteria are most closely related to the proto-mitochondrion?

a. Alphaproteobacteria like *Rickettsia*.

b. Cyanobacteria like *Cylindrospermum*.

c. Betaproteobacteria like *Neisseria*.

Answer

a. Alphaproteobacteria like *Rickettsia*.

Recent analogues

The modern amoeba species *Pelomyxa* hosts endosymbiotic bacteria that serve the same function as mitochondria do in other eukaryotes. Interestingly, ribosomal RNA analysis of its endosymbionts reveals not one species but three! Moreover, two of them are bacterial, while one is a methanogenic archeon (Gutiérrez *et al*., 2017). This set of relationships is probably relatively recent, for it is not sufficiently entrenched as to be obligatory. The microbial symbionts can live freely on their own, and *Pelomyxa* has not yet evolved a system of passing them to its descendants. So the amoeba must capture fresh bacterial endosymbionts anew with each generation.

Strigomonas is another example. This trypanosomatid is related to the parasite that causes sleeping sickness in humans. It relies on an endosymbiotic betaproteobacterium, and cannot live without it, though the bacteria is capable of life outside *Strigomonas*. This is an interesting case where one relies on the other, but the feeling is not mutual!

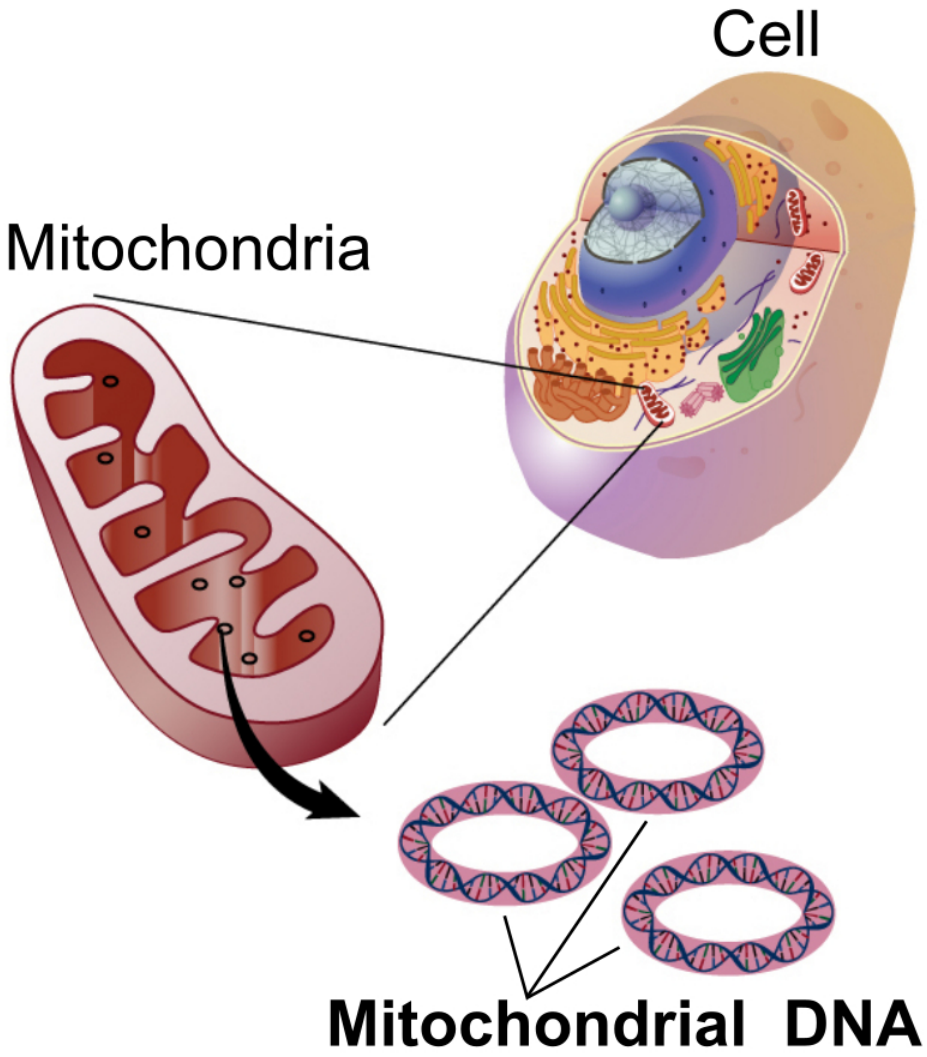


Figure 21.3.2: The context of the hoop-shaped DNA hosted by mitochondria. (Public domain; Illustration by National Human Genome Research Institute – National Institutes of Health.)

The DNA of mitochondria is not stored in chromosomes, the way that nuclear DNA is. Instead, it is found in circular "hoops" of DNA within the mitochondrion, similar to the shape in which DNA is organized in bacteria. In animal mitochondria, the length of a one of these hoops ranges from 11–28 thousand base pairs. (For comparison, the number of base pairs in the human genome is ~3.2 billion.)

In animals, mitochondrial DNA is passed from mother to offspring. The father contributes half of the nuclear genetic material to his children, but the mother supplies the other half of the nuclear DNA *plus* all of the mitochondrial DNA. You got your mitochondria from your mom, who got them from *her* mom (your maternal grandmother), who got them from *her* mom (your mom's maternal great-grandmother). Mitochondrial DNA is thus useful for tracing a line of matrilineal descent.

Studies of human genetics has shown a mitochondrial last common ancestor to all mitochondria in living humans was in a woman who lived in Africa between 120,000 and 156,000 years ago. This individual has been dubbed "mitochondrial Eve." [LINK TO HUMAN EVOLUTION CASE STUDY?]

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21.4: Chloroplasts

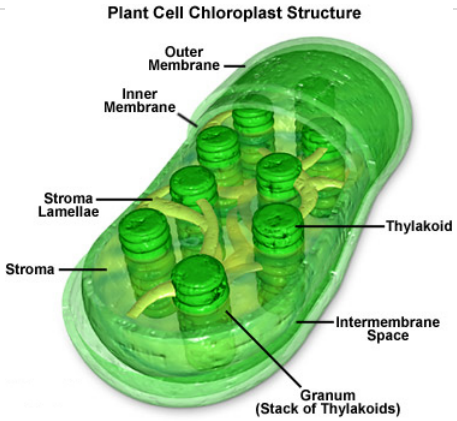


Figure 21.4.1: Structure of the chloroplast. (Image by MolecularExpressions at Florida State University; reproduced with permission.)

Chloroplasts are organelles within the cells of algae and plants that are the sites of the photosynthesis reaction. There can be between one and ~100 chloroplasts per plant cell. They are less than 10 μm in diameter and ~2 μm thick. Within each chloroplast are multiple stacks of poker-chip-shaped discs called thylakoids. These structures contain chlorophyll, the enzyme which utilizes incoming wavelengths of light in the blue and red portions of the visible spectrum to catalyze the reduction of atmospheric carbon dioxide (CO_2) with water (H_2O), generating glucose ($C_6H_{12}O_6$), along with free oxygen (O_2) as a waste product. *(Chlorophyll appears green because that is the portion of the spectrum* ***not*** *utilized for photosynthesis. Rather than being absorbed, it is* ***reflected****.)*

The anatomy of chloroplasts is shared by free-living cyanobacteria. Like cyanobacteria, chloroplasts have their own DNA. They reproduce independently of the rest of their host cell. Unlike cyanobacteria, they have two membranes separating them from their host cell. They make the inner of the two membranes, while the outer membrane is manufactured by the host cell.

We estimate that the endosymbiosis between an early eukaryote (already sporting mitochondria) and a cyanobacterium (the proto-chloroplast) to have occurred around 1.25 Ga. This date is supported both by genetic comparisons and the fossil record. It probably happened more than once, something that is thought to be more likely on the extreme population sizes and ubiquity of bacteria.

Did I Get It? - Quiz

? Exercise 21.4.1

What is the function of the chloroplast in the cells of algae and plants?

a. It is the site where the host cell stores (and copies) its genetic information.

b. It is the site of photosynthesis, the biochemical process where chlorophyll is used to catalyze a reaction between carbon dioxide and water, producing glucose using energy from sunlight.

c. It is the site of respiration, the the biochemical process where pyruvate and oxygen are utilized to make molecules of ATP, a molecule that serves as a major "currency" of energy within the cell.

Answer

b. It is the site of photosynthesis, the biochemical process where chlorophyll is used to catalyze a reaction between carbon dioxide and water, producing glucose using energy from sunlight.

? Exercise 21.4.2

Which group of microbes is thought to be most closely related to the proto-chloroplast species?

a. the "Asgard" superphylum of Archaea

b. Methanogenic Archaea

c. Alphaproteobacteria

d. Cyanobacteria

e. Betaproteobacteria

Answer

d. Cyanobacteria

Recent analogues



Figure 21.4.2: Two cyanobacterial endosymbionts within a single *Paulinella chromatophora* amoeba. (Photograph by Ferry J. Siemensma; reproduced here with permission.)

An illuminating analogue can be found in the amoeba *Paulinella*, which incorporated a cyanobacterium into its cells much more recently, sometime between 140 and 90 Ma. Based on genetic similarity, the internal symbiont was probably an ancestor to one of the extant marine cyanobacterial genera *Prochlorococcus* or *Synechococcus*. Though the symbiosis that led to proper chloroplasts is more obscured by the depths of geologic time, that the same process has been "done over" independently in other groups such as *Paulinella* suggests the validity of the idea of cyanobacterial endosymbiosis as a critical process in the evolution of photosynthetic eukaryotic life.

Another example is *Azolla*. Technically a fern, it looks nothing like most ferns, and instead floats on the surface of freshwater ponds all over the world. *Azolla* hosts the cyanobacterium *Anabaena* within its cells. The *Anabaena* has the ability to capture nitrogen (N_2) from the surroundings and "fix" it in a biologically useful form, ammonia (NH_3). This is beyond the abilities of a typical chloroplast, but being a plant the *Azolla* already has typical chloroplasts. The endosymbiosis of the *Anabaena* is an **additional** cyanobacterial acquisition, for the sake of the nitrogen it can sequester and share. This is a huge advantage to the *Azolla* host, which is extremely capable of growth as a result – it can double its biomass in two days. The relationship between *Azolla* and *Anabaena* is not haphazard or improvised: It's codified enough that the host *Azolla* cells pass the endosymbiont *Anabaena* cells directly to their offspring through cell division and reproduction, an example of "vertical" transmission.

Take two

There can also be multiple rounds of endosymbiosis. Imagine, for instance, a eukaryotic cell containing a double-membraned chloroplast. (This is the situation we have already described.) What happens when that cell gets engulfed by another, bigger cell? This is secondary endosymbiosis, a fresh iteration wherein the eater gets eaten. The chloroplast will now be quadruple membraned, with both the "original" two membranes, plus a third membrane from its first host and a fourth membrane from its newest, outermost host.

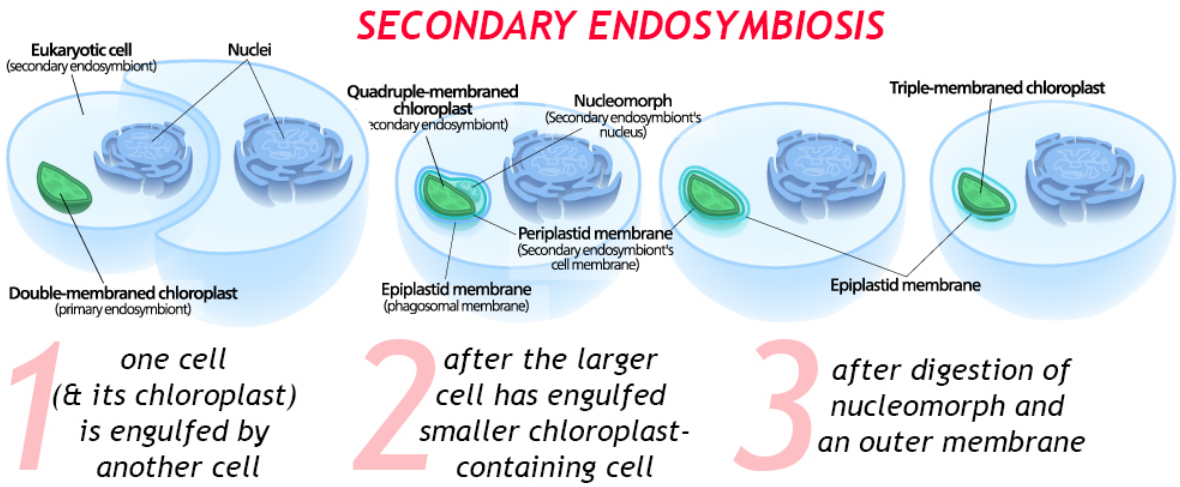


Figure 21.4.3: When a cyanobacterium is engulfed by a larger eukaryotic cell, you get a chloroplast. When that cell gets engulfed by another, it's a case of secondary endosymbiosis.

When this process is recent, there may be a residual nucleus-like body between the third and fourth membranes: this is the nucleus of the first host cell. In time, the second host cell may digest (or assimilate) that DNA, and perhaps one or more of the membranes surrounding the chloroplast.

There is even a well-documented case of ***tertiary*** endosymbiosis, in the dinoflagellates, where a cyanobacterium was engulfed by an ancestral eukaryotic host, making a unicellular alga, which was then swallowed by a protozoan, and then that was engulfed by a haptophyte alga! Geobiologist Andy Knoll refers to dinoflagellates as "biological Matryoshka dolls," for good reason. A good first guess about how many episodes of endosymbiosis a particular organelle has undergone can be arrived at by counting the number of membranes and subtracting 1. Just to drive the point home, it's worth noting that dinoflagellates are the so-called "zooxanthellae" that are endosymbionts in corals: Unless the coral undergoes "bleaching," evicting the dinoflagellates into the surrounding water, that's one more Matryoshka doll yet!

Did I Get It? - Quiz

? Exercise 21.4.1

A secondary endosymbiont will initially have how many layers of membrane surrounding it?

a. Four

b. Two

c. Six

d. One

e. Five

f. Three

Answer

a. Four

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21.5: Other organelles

Though mitochondria and chloroplasts are the most well known and important examples of microbial endosymbiosis leading to novel organelles, there are perhaps other examples too.

Consider the flagellum, a whip-like appendage that projects off the surface of many groups of microbes. Lynn Margulis suggested that the flagellum appears to be an organelle that represents a merger between a spirochete bacterium and a host archeal cell related to the genus *Thermoplasma*. This idea hasn't been as thoroughly validated as the endosymbiotic origins of mitochondria and chloroplasts, but it remains an intriguing idea worth further investigation.

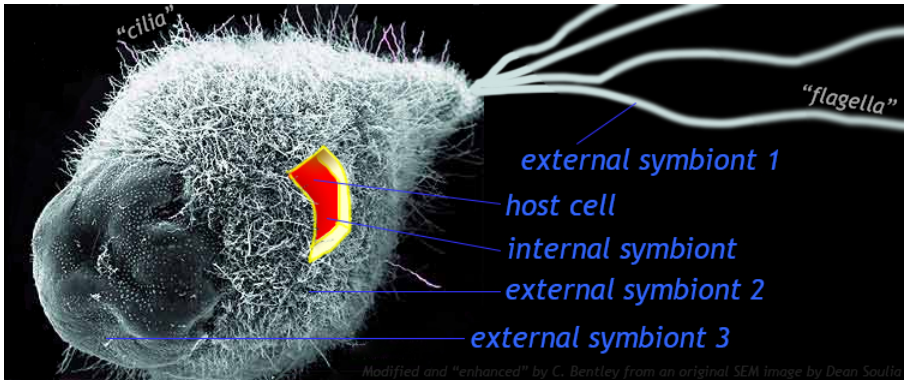


Figure 21.5.1: Mixotricha paradoxa is "one" organism with five genomes. (Modified from an SEM image by Dean Soulia.)

Mixotricha paradoxa is an interesting example that appears to corroborate the idea of a symbiotic origin for flagella. *Mixotricha* is a species of protozoan (a eukaryote) that lives inside the gut of certain Australian termites. Unusually for protozoans, it sports both cilia and flagella, which is why its name translates to "a paradoxical mixture of hairs." *Mixotricha* has five genomes: one for the main host cell, one for an endosymbiont that helps with the digestion of cellulose and lignin, and three on the surface (two spirochetes and one rod bacterium).

Some scientists have suggested that the nucleus itself, that iconic hallmark of the eukaryotic cell, arose through an even earlier episode of endosymbiosis. Perhaps that earliest, most fundamental merger occurred with a now-extinct group of microbes. Hartman and Federov (2002) identified hundreds of genes coding for proteins that are found in eukaryotic cells but have no equivalents in either Archaea or Bacteria. They interpret this as evidence of a third microbial line, the primordial host cell that would eventually lead to the eukaryotic line, dubbing it the Chronocyte (translation = *time cell*). Another possibility is that a large DNA virus may have been responsible; this is the viral eukaryogenesis hypothesis. Future research may shed more light on these intriguing possibilities.

An intriguing series of mergers and acquisitions permeate the history of eukaryotes on Earth, some more plain and others more obscure. When we consider ourselves as individuals, its humbling to realize we are in fact communities within communities, a whole that is greater than the sum of its parts.

Did I Get It? - Quiz

? Exercise 21.5.1

Mixotricha is a protist with what surprising aspect to its anatomy?

a. It helps termites digest the cellulose in wood.

b. It is really large.

c. It is shockingly handsome for a eukaryote.

d. It is a symbiotic relationship between five distinct organisms, each with their own genome.

Answer

d. It is a symbiotic relationship between five distinct organisms, each with their own genome.

Exercise 21.5.2

A traditional view of evolution is an endlessly-branching tree, with individuals competing fiercely against one another for limited resources. How does an endosymbiotic origin for key organelles bear on that idea?

a. It suggests that the only way to succeed is to outcompete every other organism.

b. It suggests another source of evolutionary novelty and success may be found through cooperation and community-building between formerly independent organisms.

c. It suggests that the traditional view is almost certainly (and exclusively) correct.

Answer

b. It suggests another source of evolutionary novelty and success may be found through cooperation and community-building between formerly independent organisms.

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21.6: Further reading

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CHAPTER OVERVIEW

22: (Case Study) Discovering plate tectonics

Plate tectonics is an overarching paradigm that explains a lot of independent observations about Earth surface dynamics. In this case study, we examine the **historical development of this important idea.** A separate chapter outlines a modern treatment of plate tectonics.

After sputtering suggestions of continental movement in the previous 150 years (most notably in 1915), new information about the seafloor between the continents was revealed in the aftermath of World War II's naval battles. This prompted a fresh look at crustal dynamics, and in the late 1960s and early 1970s, geoscientific consensus gelled around the idea that many different observations about our planet could be explained with a single model. The idea of "plate tectonics" put together old ideas about continental drift with new data showing seafloor spreading. The new theory was a revolutionary explanation that made the distribution of volcanoes, earthquakes, continents, and topography make sense in a way that no idea had achieved before. Fifty years of additional data and testing have confirmed plate tectonics as ***the*** key idea in modern geology.

22.1: Antecedents
22.2: Continental drift
22.3: Seafloor spreading
22.4: Hot spot tracks
22.5: Conclusion
22.6: Further reading

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22.1: Antecedents

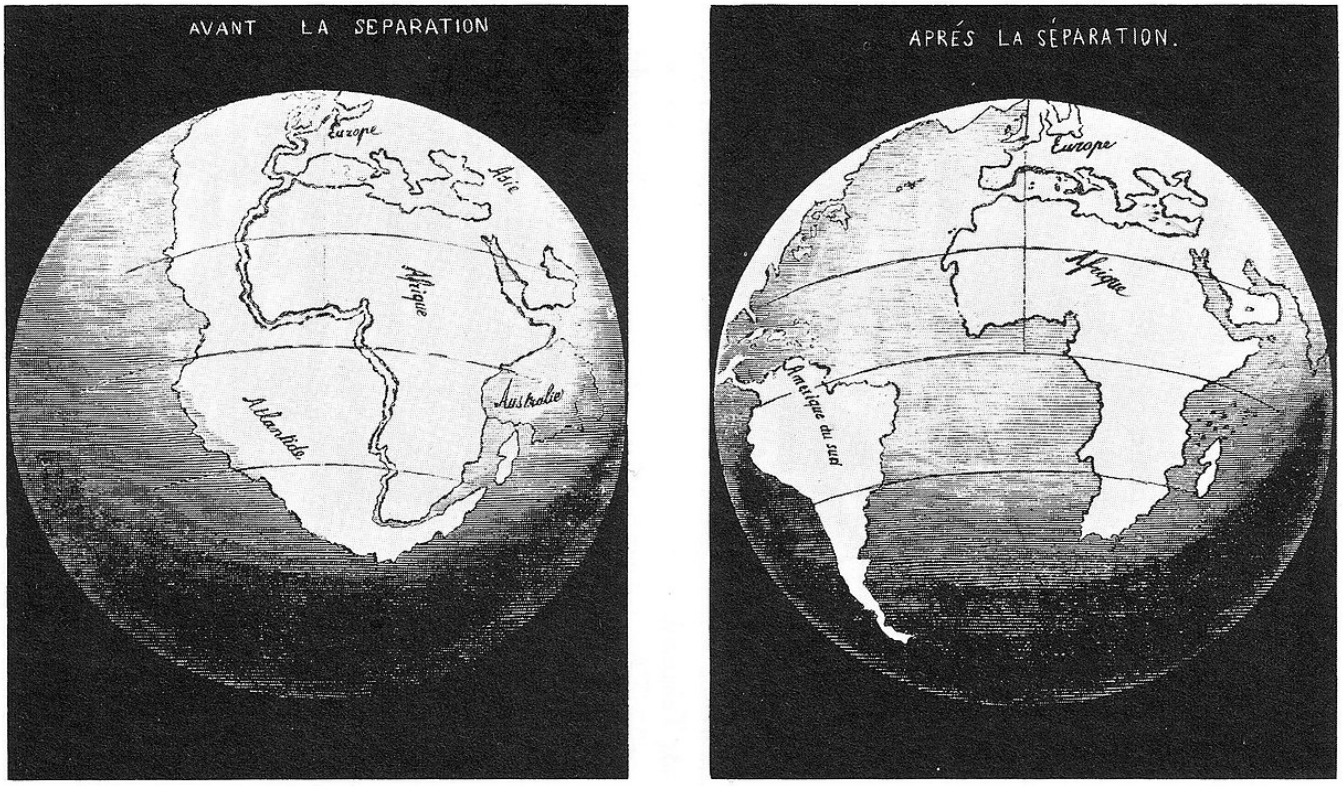


Figure 22.1.1: Antonio Snider-Pellegrini's hypothesis of continental drift in 1858, showing the first graphical depiction of the opening of the Atlantic.

The idea that continents can move is a rather audacious notion to emerge in the mind of an human who lives such a short life. Prior to the invention of satellite-based telemetry (GPS), there was not discernible way to directly measure such a thing. Yet various lines of evidence did suggest the notion to various thinkers in the European scientific tradition. A particular prompt came from the mapping of the world's landmasses as a consequence of the "Age of Exploration." *(That's a Eurocentric phrase if ever there was one! Sometimes this is called the "Age of Being Explored by Europeans," since that is how most of the world's population experienced it.)* Once the shapes of other continents were documented, they could be contemplated.

The earliest mention of what could later be called continental drift appears to have been Abraham Ortelius. In 1596, he mused on the shapes of the Atlantic coastlines of the Americas, Europe and Africa. Their shapes were so similar, he suggested that they may have been "torn away… by earthquakes and floods."

Almost 300 years later, in 1858, Antonio Snider-Pellegrini published his hypothesis of continental drift, using the shapes of the continents as his primary line of evidence, but also citing similar terrestrial fossils on widely separated continents. These lines of evidence would also be invoked by Alfred Wegener.

Did I Get It? - Quiz

? Exercise 22.1.1

Was Alfred Wegener the very first person to come up with the idea that continents could move?

a. No

b. Yes

Answer

a. No

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22.2: Continental drift

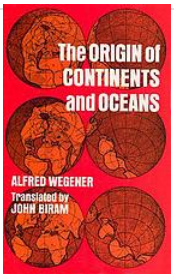


Figure 22.2.1: The 1966 English translation of Wegener's 1915 book.

German meteorologist Alfred Wegener compiled evidence for continental movement through time, first sharing these ideas in a 1912 talk and then later publishing a 1915 German book called *Die Entstehung der Kontinente und Ozeane ('The Origin of Continents and Oceans')*. In it, he postulated the existence of a supercontinent he called **Pangaea** which existed in the late Paleozoic Era. According to Wegener, since the early Mesozoic, Pangaea's various pieces have been drifting away from one another, with the Atlantic Ocean opening up in between. He is the person most associated with a full-throated articulation of the idea of **continental drift.**

Marshalling evidence for this audacious idea, Wegener compiled evidence of several varieties:

1. the shape of the continents
2. modern mountain belts
3. ancient mountain belts
4. sedimentary rocks that indicated paleo-latitude
5. glacial striations
6. the distribution of fossil organisms

Let us examine each in turn.

The shape of the continents

Like Snider-Pellegrini, Wegener first conceived of continental drift by noting the visual match between the **shapes of the continents**, particularly South America's eastern coast and Africa's western coast. The fit between these two is even more pronounced when the modern coastline is ignored, and instead the "puzzle pieces" are defined on the basis of the edge of the continental shelf. Though currently submerged by seawater, the edge of the shelf is where the continental crust truly ends.



Figure 22.2.2: Jigsaw puzzle pieces bear shape information as well as the design on their upper surface. (Charles Hamm photo, via Wikimedia.)

But while his predecessors stopped there, Wegener took the idea further. After all, jigsaw puzzles are solved not only only on the basis of the shape of their pieces, but the imagery on their surface too. So Wegener examined the geological evidence on the continents themselves, comparing these data across the oceans in an attempt to match them up.

Modern mountain belts

One compelling line of evidence for continental drift comes from **modern mountain belts**.

Once you substract the Atlantic Ocean, the Appalachian Mountains line up with other ranges, including the Caledonides in Scandinavia, the North-west Highlands of Scotland, and the Anti-Atlas Range in Morocco. The mountain belts running through Scotland and Newfoundland fill in some of the connective tissue, and suddenly this appears to be a single mountain range, running through the former heart of Pangaea, now ripped asunder into several pieces.

Here is a reconstruction of Pangaea presented as a paleogeographic map, so you can zoom in and examine the linear trend of the pre-Atlantic mountain belt for yourself:

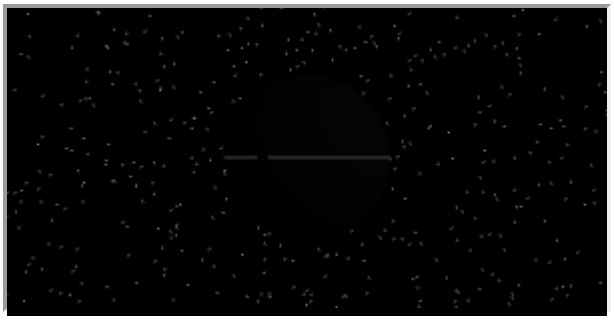


Figure 22.2.3: Paleogeography of the Permian (~250 Ma). (Maps by Christopher Scotese of the PALEOMAP project, with virtual globe by Ian Webster of Dinosaur Pictures.org; reproduced with permission.)



Figure 22.2.4: Patterns of Archean crust (cratons) and the Proterozoic-aged mountain belts that wrap around them are continuous across the South America / Africa break. (Callan Bentley cartoon.)

Mountain ranges aren't stable topographic entities. All that rock poking up into the air represents a lot of potential energy. Landslides and rockfalls tear mountains down over geologic time, reducing their topographic relief. But we can still tell where the mountains used to be on the basis of geological evidence including (1) belts of foliated metamorphic rocks, (2) granite intrusions, (3) zones of compressional deformation — folding and thrust faulting, all at the "roots" of **ancient mountain belts**, and (4) deposits of sedimentary rock adjacent to the mountain belt (and derived from it). Wegener used the pattern of Proterozoic-aged mountain belts that wrapped around more ancient blocks of continental crust as an additional line of evidence of the former connection of Africa and South America.

Sedimentary rocks that indicate paleolatitude

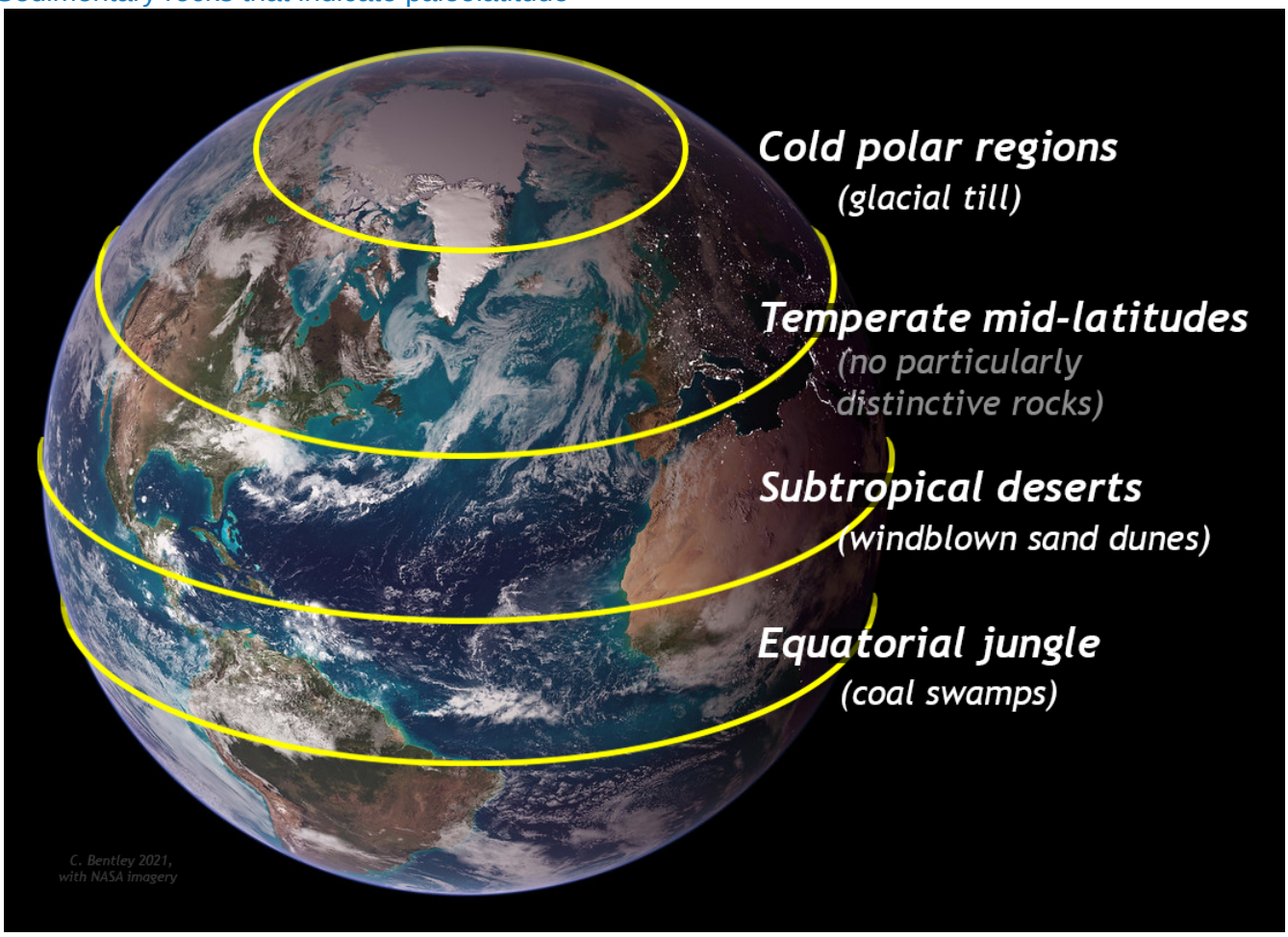


Figure 22.2.5: Certain sedimentary rocks may be indicators of paleo-latitude. (Callan Bentley graphic using NASA imagery.)

One of the big, obvious patterns about our planet today is that it's colder toward the poles and warmer toward the equator. Applying the principle of uniformity (a.k.a. "uniformitarianism") to the ancient past, Wegener reasoned that during the time of Pangaea, the same pattern applied. The idea here is that the positions of landmasses in the past might be indicated by **characteristic deposits of sediment** on their surfaces. In other words: certain sedimentary rocks form in cold settings only, while others are more likely in the tropics or mid-latitudes. Wegener deduced that Pangaea was a Permian-aged supercontinent, so he looked at the character of Permian-aged sedimentary rocks found on the various continents. South Africa showed diamictites of glacial origin, and therefore was in a polar position at Pangaean times. He also identified a belt of coal, which he interpreted as representing thick forest cover thriving at the Pangaean equator. Flanking the coal belt were two belts of windblown sand, stacked up in dunes that resembled those in the modern Sahara Desert. Wegener interpreted these as representative of the subtropical high-pressure zones where atmospheric circulation nurtures deserts today.

The key thing about all these observations is that they held up across continents, like a series of colored stripes extending across multiple jigsaw puzzle pieces.



Figure 22.2.6: Oorlogskloof Glacial Pavement, South Africa: Permian-aged glacial striations. Sunglasses for scale. (Photo by Evelyn Mervine; reproduced with permission.)

We mentioned sedimentary rocks that served as evidence of Permian-aged glaciation in the southern part of Pangaea. There is a tangent worth exploring here, too – or at least Alfred Wegener thought so! When glaciers flow over the landscape, they gouge into it. This can create spectacular U-shaped valleys, but on a smaller scale, the glacier carries a suite of boulders and cobbles and that get scraped over the landscape as the glacier flows. This leaves behind a series of parallel **striations**. These linear features show which way the glacier flowed. By comparing striation orientations from Permian-aged unconformities in different parts of southern Pangaea, Wegener was able to reconstruct a radial pattern, like the spokes on a bicycle wheel. He interpreted this as the pattern you would expect from a single ice sheet, like modern day Greenland, where a big domelike mass of ice flows downward and outward in all directions. *(The Greenland ice sheet was a place he was very familiar with.)* The center of this ancient ice sheet, Wegener reckoned, must have been the Permian South Pole.



Figure 22.2.7: Mesosaurus fossil from the Permian (~270 Ma) Paraná basin of Brazil. (Photo by Kevmin, via Wikimedia.)

Wegener had the insight that the distribution of certain time-specific **fossils** would also shed light on which landmasses were connected to which other landmasses at that specific time. He limited his accounting to terrestrial species, on the logic that marine species would be freer to move around through the waters of the ocean. Sharks wouldn't be a good test of the continental drift idea, Wegener thought, but something like a koala would.



Figure 22.2.8: Mesosaurus fossil in shale of the Whitehill Formation (Lower Permian) near Keetmanshoop, Namibia. (Photo by Olga Ernst & Hp. Baumeler, via Wikimedia.)

He highlighted a land plant and an aquatic (freshwater) reptile as two particularly useful examples. The reptile was *Mesosaurus*, known from two places on the planet: Brazil's Paraná basin, and the Aranos basin of southwest Africa. When the position of the South American and African continents was reconstructed to their pre-Atlantic position, this created a cluster of *Mesosaurus* fossils sites that overlapped both continents.

Wegener realized there were three plausible ways of explaining this distribution. Consider these multiple competing hypotheses:

1. The fossils represent two populations of the same species that got established in two continental locations thousands of kilometers apart, and somehow the original population surmounted the obstacle of the Atlantic Ocean to colonize the second location. Maybe the *Mesosaurus* scrabbled over there on a long skinny transatlantic land bridge, now eroded away? In this case, the continents don't need to drift.

2. The two groups of *Mesosaurus* fossils are an example of convergent evolution, independently derived on the two continents, and don't actually represent a single ancestral population. In this case, the continents don't need to drift.

3. There was one population of *Mesosaurus*, established within the Pangaean supercontinent, and when Pangaea broke up, the continent of Africa drifted away from the continent of South America, splitting the *Mesosaurus* collection in two. In this case, the simplest story for the fossil distribution ends up implying something huge about the way the Earth works: continents drift!

Wegener, of course, opted for the final hypothesis.

The plant was a seed fern [Update link seed fern case study], *Glossopteris*. Here is a GIGAmacro view of some *Glossopteris* leaves – Zoom in to see their details:



GIGAmacro by Robin Rohrback.

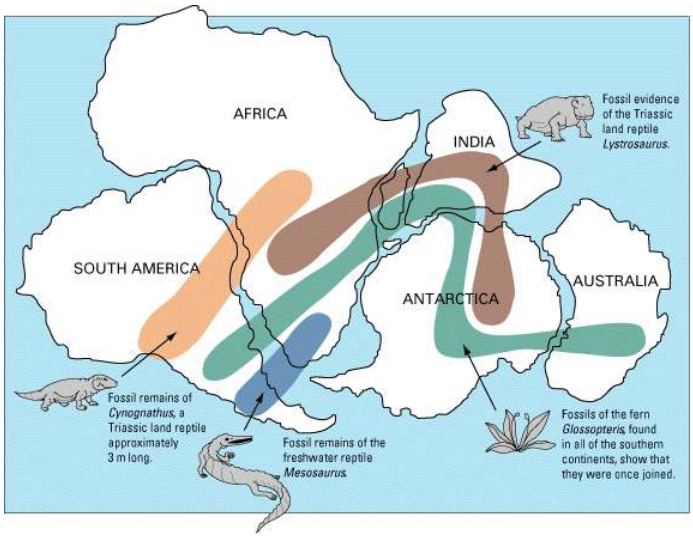


Figure 22.2.9: As noted by Snider-Pellegrini and Wegener, the locations of certain fossil plants and animals on present-day, widely separated continents would form definite patterns (shown by the bands of colors), if the continents are rejoined. (USGS map.)

Glossopteris has a late Paleozoic distribution across all the southern portion of Pangaea; what's sometimes called Gondwana. Fossils of *Glossopteris* have been found in South America, Africa, Madagascar, Antarctica, India, and Australia. How best to explain this distribution? The same multiple working hypotheses that applied to *Mesosaurus* apply to *Glossopteris*, and again, Wegener found the "one population of *Glossopteris*, ripped asunder into six pieces by continental drift" the most compelling explanation.

Many other Permian-aged species show this same pattern, suggesting the timing for Pangaea's existence.

Exercise 22.2.1

What was the name of the most recent supercontinent, the one Wegener focused on in his articulation of the evidence for continental drift?

a. Gorgonzola

b. Ur

c. Columbia

d. Oklahoma

e. Pangaea

f. Rodinia

Answer

e. Pangaea

Exercise 22.2.2

Which of the following was *NOT* on Alfred Wegener's list of evidence for continental drift?

a. sedimentary rocks that indicated paleo-latitude

b. the distribution of fossil organisms

c. glacial striations

d. the age of the seafloor

e. the shape of the continents

f. modern mountain belts

g. ancient mountain belts

Answer

d. the age of the seafloor

Exercise 22.2.3

If the Atlantic Ocean is conceptually removed, the Appalachian mountain belt continues in _____.

a. southern Africa and western South America

b. Antarctica, Australia, and southern India

c. Scotland, Scandinavia, and northwest Africa

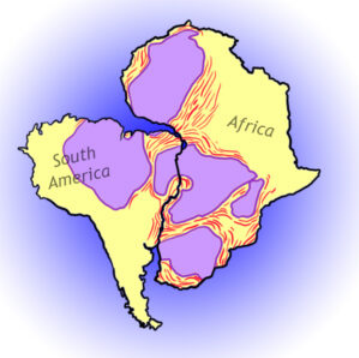
d. Siberia, Madagascar, and the Isthmus of Panama

Answer

c. Scotland, Scandinavia, and northwest Africa

Exercise 22.2.4

Which of Wegener's lines of evidence is shown in this map?



a. Glacial striations

b. Modern mountain belts

c. The distribution of fossil organisms

d. Ancient mountain belts

Answer

d. Ancient mountain belts

Exercise 22.2.5

Which of Wegener's categories of evidence for continental drift is shown in this GigaPan?



Tom Raimondo image

a. The distribution of fossil organisms

b. Permian-aged glacial striations

c. Sedimentary rocks that indicate Permian-aged paleo-latitude.

d. Proterozoic mountain belts

Answer

b. Permian-aged glacial striations

? Exercise 22.2.6

Where are Mesosaurus fossils are found today?



a. Africa only

b. Africa and South America only.

c. Africa, South America, Madagascar, India, Australia, and Antarctica

d. North America, Scotland, and Scandinavia.

Answer

b. Africa and South America only.

Landing with a thud

Though this evidence may seem compelling to us today, reception to Wegener's proposal was mixed. Some prominent geologists east of the Atlantic championed the idea (notably Arthur Holmes in the United Kingdom and Alexander du Toit in South Africa), but it was a hard sell to American geologists. Wegener himself brought this paradigm-shifting idea to geology from outside the field; he was a meteorologist by training. Though that shouldn't matter, it did. Scientists are people, and there is frequently a tendency to discount a radical idea offered by someone with less formal training that yourself. They waved off his compilation of evidence as mere coincidence, or sought to explain it by other means (the fossil distributions could be explained by since-vanished land bridges, for instance).



Figure 22.2.10: Alfred Wegener. (Public domain.)

Wegener also made some calculations about the rate of continental movement, and got speeds that were two orders of magnitude faster than reality. Though geologists of his day had no clue what the real rate was, they did know that Wegener's rate of 2.5 m/year seemed outrageously zippy. (For contrast, modern estimates of Atlantic Ocean divergence are ~2.5 cm/year.)

The fatal flaw though was that Wegener only documented evidence that the continents had moved. He did not take the idea further into a model for why they moved. Wegener did not suggest ***how*** they might have moved, only that they ***had***. Did the continents plow through the ocean basins like icebreakers crossing a frozen lake? If so, what would compel them to do so? The lack of a conceivable driving mechanism for the horizontal movement of continents led most American geologists to reject the notion outright.

It is worth noting that Arthur Holmes did suggest mantle convection as a driver, as early as 1919. He even suggested a "bulge" where warm mantle material rose beneath the seafloor, presaging insights that would come in the 1950s and 1960s. But even this failed to overcome skepticism in the scientific community.

Finally, there was as yet no way to reconstruct the path a continent had taken through time, so a vital test was lacking. Technology would eventually catch up (paleomagnetism; see below), but too late for Wegener to be vindicated while he was alive. The idea was shelved, and when Wegener was killed on a meteorological expedition in Greenland in 1930, continental drift lost its most effusive champion.

Did I Get It? - Quiz

? Exercise 22.2.1

One issue with continental drift as Wegener proposed it, was that he had the continents moving _____.

a. in the wrong direction

b. far too quickly

c. far too slowly

d. vertically, not horizontally

Answer

b. far too quickly

? Exercise 22.2.2

The principal flaw in Wegener's presentation of continental drift is that he failed to accompany his robust list of many different kinds of evidence with _____.

a. a prediction of what would come in the future, if the continents kept drifting as they were currently moving

b. a well-articulated discussion of ocean currents induced by the presence of such a large landmass as Pangaea

c. any proposal of a mechanism which could plausibly move such large slabs of rock over substantial horizonal distances

Answer

c. any proposal of a mechanism which could plausibly move such large slabs of rock over substantial horizonal distances

Other supercontinents

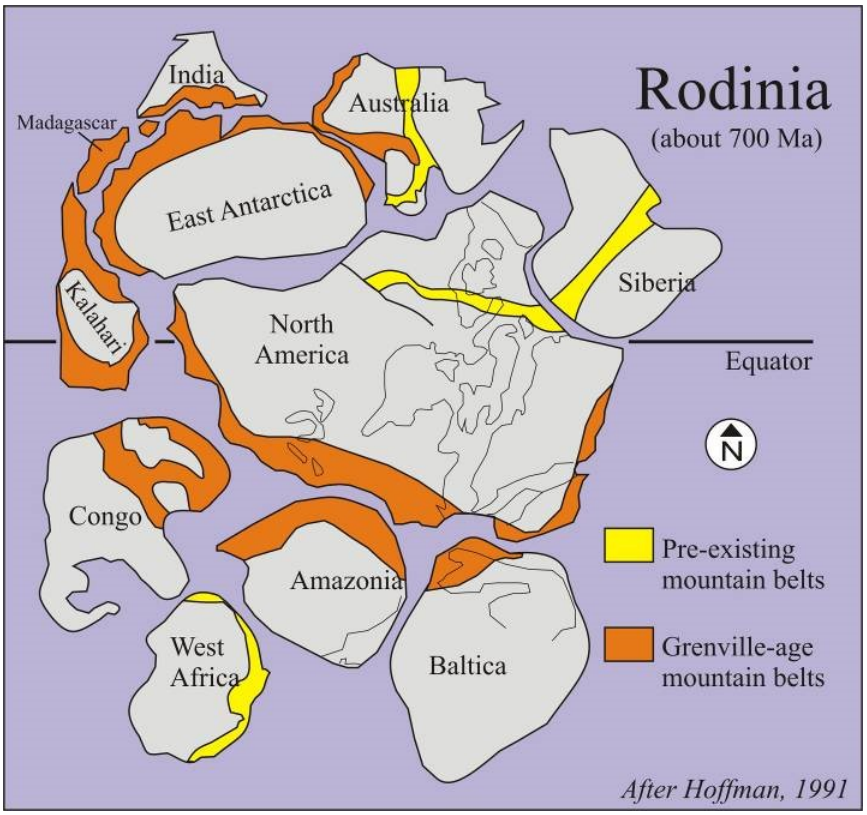


Figure 22.2.11: The Proterozoic supercontinent Rodinia at the time of its breakup, around 700 Ma. (Callan Bentley graphic after an original by Paul Hoffman.)

A brief note: We now know that Pangaea was not the ***only*** supercontinent, merely the ***most recent***. Over the course of geologic history, continents have joined up into larger landmasses, and broken apart again repeatedly. We have a pretty clear view of Pangaea's size and shape because that view is only sullied by its breakup. Earlier supercontinents are harder to make out, since each successive tectonic episode tends to overprint evidence of its predecessors. The various processes of "the rock cycle" (weathering and erosion, metamorphism, melting) tend to destroy older rocks, or modify them in ways that obscure the information that they previously carried within them. Nonetheless, we've been able to make out previous supercontinents, the more recent ones clearer than than the more ancient ones. There was Rodinia (~1 Ga), Nuna (~1.8 Ga), Ur (~2.6 Ga), and several others.

Did I Get It? - Quiz

Exercise 22.2.1

Was Pangaea the only supercontinent to ever exist on this planet?

a. Yes: it represents a pristine primordial state from which the world has slowly descended into chaos and confusion.

b. No, there were several previous to Pangaea, but they are not as well documented.

Answer

b. No, there were several previous to Pangaea, but they are not as well documented.

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22.3: Seafloor spreading

Wegener's catalog of evidence for the past motion of the continents was revitalized when a new context was offered in the aftermath of World War II. The new context shifted the focus from the continents (~30% of the planet's surface) to the rocks that floor the ocean basins (~70% of Earth's surface). The new idea was that the seafloor was an ephemeral entity, forming at oceanic ridge systems, and destroyed elsewhere through the process of subduction.

The shape of the ocean basins

World War II was fought on many fronts. The relevant aspect of the conflict for the current discussion is submarine warfare. The United States Navy fought the German U-boats in the Atlantic Ocean, attempting to prevent tragedies like the sinking of the *Lusitania*. To aid in the war effort, sonar was utilized to map the seafloor to a greater degree than had previously been accomplished. When the data was declassified after the war ended, it revealed a fascinating aspect of the planet that had previously been hidden: the seafloor wasn't flat!

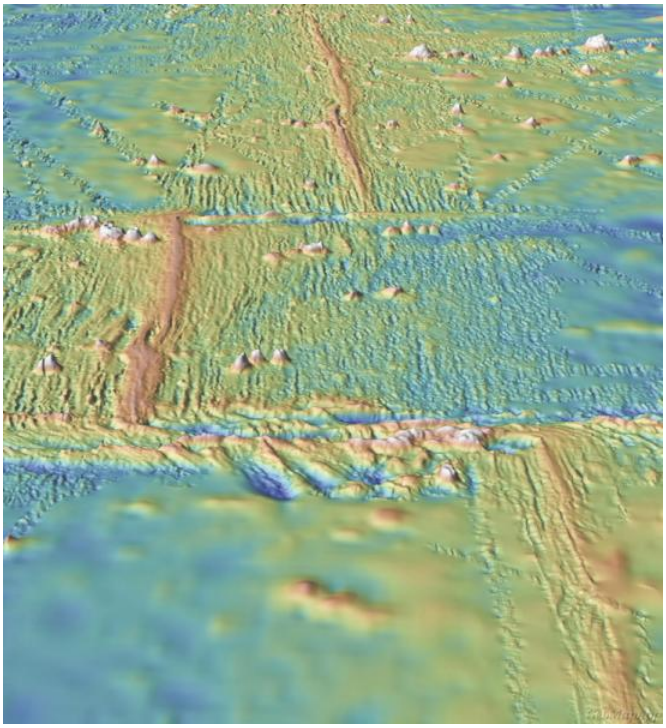


Figure 22.3.1: The oceanic ridge system comes in many segments, offset along perpendicular fracture sets. (Visualization via GeoMapApp.)

Instead, the longest range of mountains on the planet, the **oceanic ridge system**, wrapped around the planet much like the seams wrap around a baseball. Even the stitching on the baseball is relevant here, as it evokes the fracture zones that emanate from the ridge, perpendicular to the ridge's trend. And right there are the crest of this odd ridge system was a valley, a rift valley. It looked like a scar.

The oceanic ridge system is 70,000 km (43,000 miles) long. In both the Atlantic and Indian Ocean basins, the ridges were in the middle of the ocean, equidistant from the continents on either side. This led many geoscientists to adopt the term "**mid**-ocean ridge" to describe them, but it's worth noting that the East Pacific Rise is not in the middle of the Pacific Ocean. So oceanic ridge is the more inclusive term.

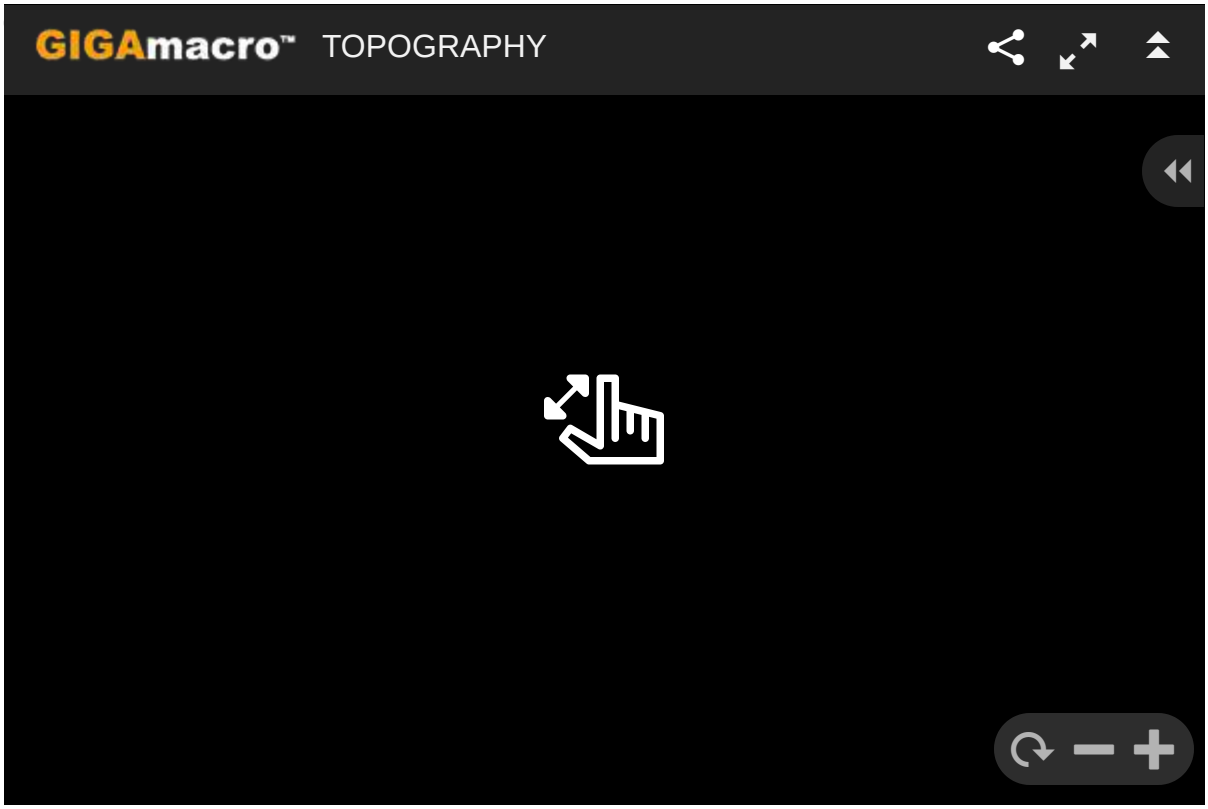


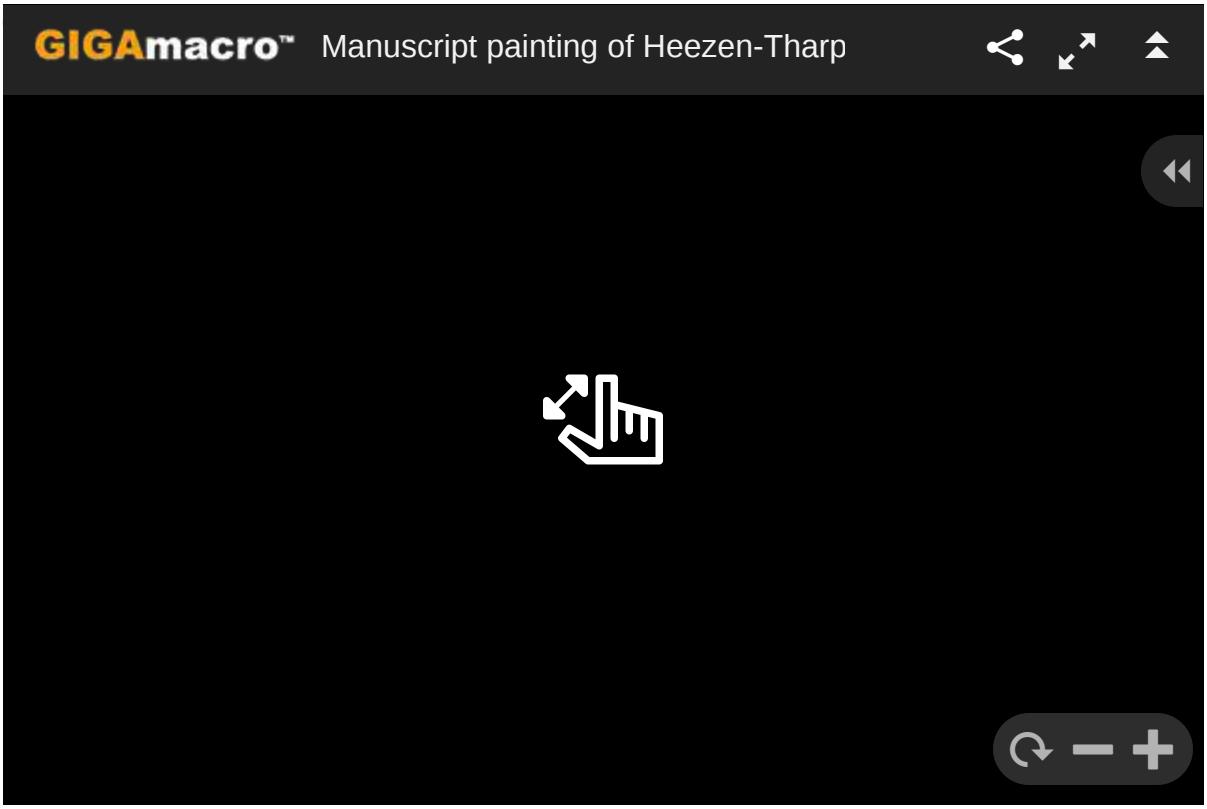
Figure 22.3.2: “Discovering Plate Boundaries” topography/bathymetry map (Dale Sawyer)

There were also very deep spots in the ocean – long furrows that were quickly dubbed “**trenches**.” Curiously, these trenches paralleled chains of active volcanoes, and were in areas known to have lots of earthquakes.



Figure 22.3.3: Marie Tharp and Bruce Heezen, with their map.

At Columbia Lamont-Doherty Earth Observatory, Bruce Heezen and Marie Tharp collaborated on detailed mapping of the seafloor, systematically constraining the variation in bathymetry with a series of profiles. They are celebrated today not only for their meticulous measurements, but because of a landmark visualization of these findings; they commissioned artist Heinrich Berann to paint a rendition of their **seafloor map** that popped in a way few scientific visualizations have ever matched. The finalized Heezen & Tharp map (with labels, *etc*.) was published in 1978.



Painting by Heinrich Berann, via the Library of Congress

How to explain this exciting new information about the shape of the largest region on our world?

An essay in geopoetry

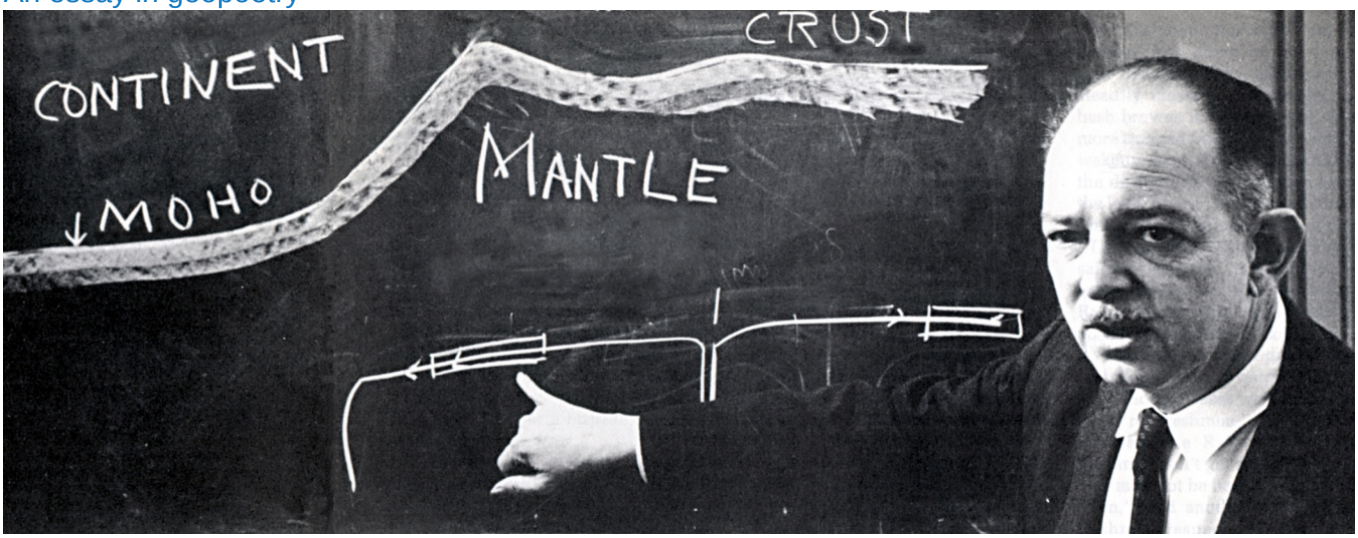


Figure 22.3.4: Harry Hess, the Naval captain and 'geopoet' who suggested seafloor spreading.

Harry Hess was a a geophysicist on the faculty at Princeton, and served as a captain in the United States Navy during World War II. He had experience with seafloor mapping using sonar on his own ship, and was deeply impressed with the Heezen & Tharpe ridge-top rift data when Heezen presented it at a departmental seminar in 1957. Hess's paper outlined the case to be made that new seafloor was generated at the ocean ridges: that active spreading at those sites triggered magmatism that sealed the crack shut with new seafloor. But Hess went further: he didn't think the planet was getting larger over time through ever-expanding ocean basins, but thought instead that a steady state ocean coverage could be maintained if some seafloor was recycled into the "jaw crusher" (his phrase) of the newly discovered oceanic trenches. Alpine geologist André Amstutz called these **subduction zones**, and though he

wrote in French, the name was soon adopted in geoEnglish, too. Powering it all was convection in the underlying mantle: warm (but solid) rock rising beneath the oceanic ridges, and cold rock sinking beneath the subduction zones. Hess was conscious of how revolutionary these ideas were, and he modestly covered up the audaciousness of seafloor spreading by hedging his bet with self-deprecating humor. He waved his arms, and called his paper “an essay in geopoetry.”

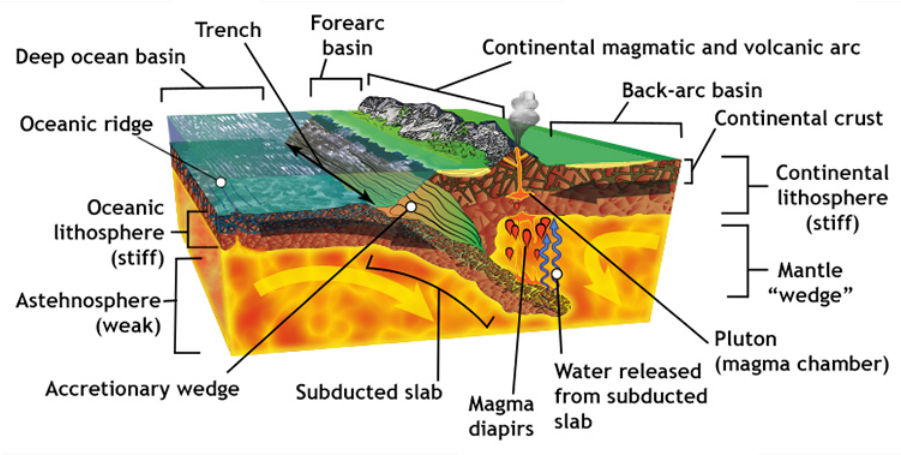


Figure 22.3.5: Key features in a subduction zone. (Modified from an original by KDS4444.)

Hess need not have been so coy; his ideas have been validated ever since. The seafloor spreads, adding new oceanic crustal material (basalt and gabbro) at the oceanic ridge system into the gap between diverging continents. Old oceanic crust is destroyed through subduction at deep sea trenches.

Did I Get It? - Quiz

? Exercise 22.3.1

Which geoscientist is recognized as first articulating the concept of seafloor spreading?

a. Matthew Vine

b. Alexander du Toit

c. Marie Tharp

d. Alfred Wegener

e. Harry Hess

Answer

e. Harry Hess

? Exercise 22.3.2

What is found halfway across the Atlantic Ocean, mid-way between eastern North America and northwestern Africa?

a. A deep sea trench

b. A continental transform fault

c. An oceanic ridge

d. Glacial striations

Answer

c. An oceanic ridge

Exercise 22.3.3

Which team of geophysicists made a map that presented the newfound structure of the seafloor in visually astonishing detail?

a. Alfred Wegener and Antonio Snider-Pellegrini

b. Marie Tharp and the Deftones

c. Bruce Heezen and Harry Hess

d. Harry Hess and Alfred Wegener

e. Marie Tharp and Bruce Heezen

Answer

e. Marie Tharp and Bruce Heezen

Exercise 22.3.4

What is a "subduction zone?"

a. A place where new oceanic lithosphere is generated through the process of seafloor spreading. It's marked by a ridge and a central rift valley at its crest.

b. A place where old oceanic lithosphere dives down into the mantle. It's marked by a deep sea trench and a parallel volcanic arc.

c. A place where short segments of oceanic lithosphere on adjacent plates slide past one another. The crust on one side is measurably older than on the other side.

Answer

b. A place where old oceanic lithosphere dives down into the mantle. It's marked by a deep sea trench and a parallel volcanic arc.

Exercise 22.3.5

What conflict led to a new push for detailed mapping of the seafloor?

a. The Viet Nam War

b. The American Civil War

c. World War I

d. The conquests of Ghenghis Khan

e. World War II

f. The War of 1812

Answer

e. World War II

Age of the oceanic crust

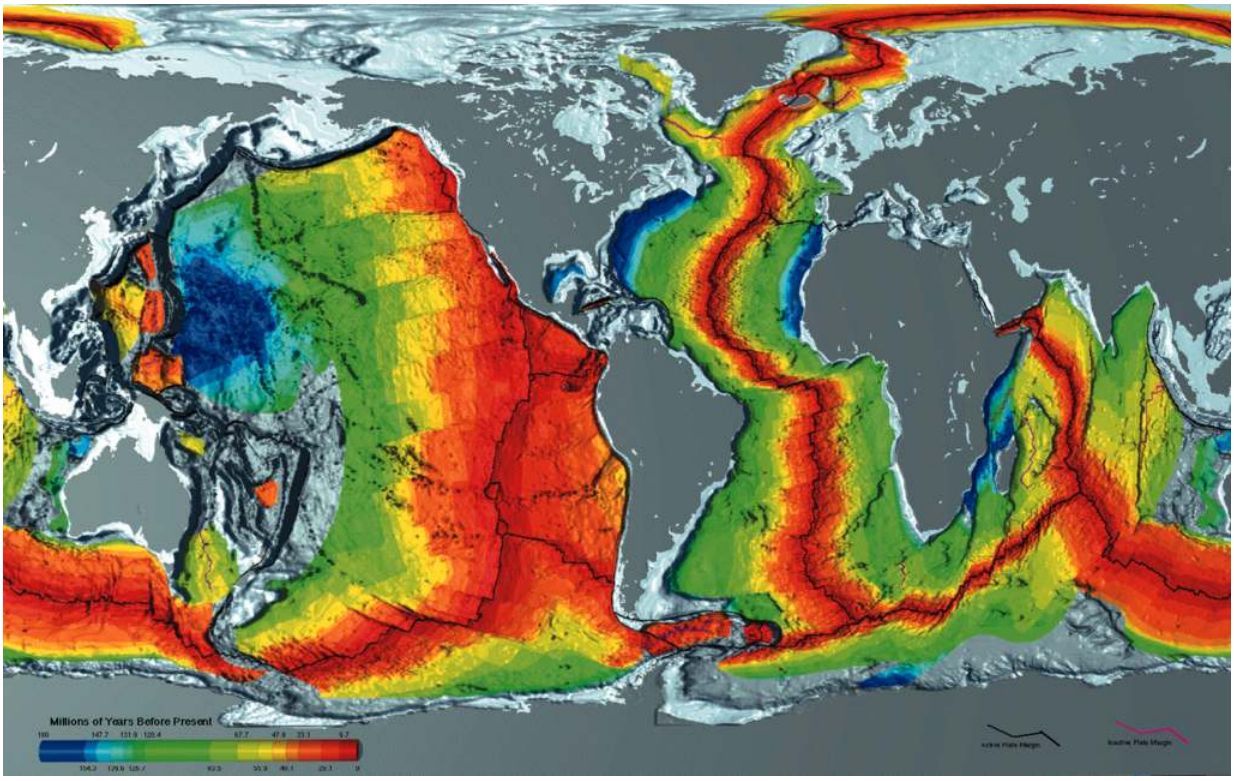


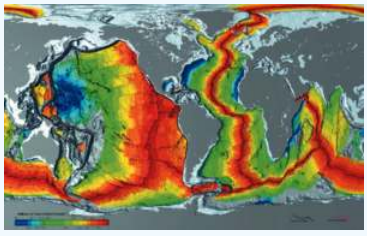
Figure 22.3.6: Age of the oceanic crust on Earth. (NOAA map.)

Oceanographic expeditions such as the Deep Sea Drilling Program sampled the seafloor in many places, gathering cores of seafloor sediment down to the top of the oceanic crust. The age of the oldest sediment could be deduced through study of the microfossils it contained, and that would constrain the age of the crust beneath to be older. Scientists then used correlation with key features of the seafloor (ridges, fracture zones, etc.) to extend points where the age of the oceanic crust was known to cover wider regions. The overall pattern was that the youngest crust was located atop the oceanic ridges, and it got older perpendicular to the ridge, in both directions. In the Atlantic Ocean, for instance, the pattern offered a beautiful symmetry, with crust of 0 Ma age at the Mid-Atlantic Ridge, and Jurassic (~180 Ma) oceanic crust immediately adjacent to the North American and African continental shelves.

Did I Get It? - Quiz

? Exercise 22.3.1

Based on the age of the the seafloor, which of the following continents did Africa separate from *first*?



a. South America

b. Greenland

c. Arabia

d. North America

Answer

d. North America

Exercise 22.3.2

Almost all the oceanic crust on this planet is less than _____ old.

a. 5 million years

b. 50 million years

c. 200 million years

d. 6000 years

Answer

c. 200 million years

Exercise 22.3.3

The oldest seafloor is _____ to the oceanic ridge of that same plate.

a. adjacent

b. beneath

c. closest

d. furthest

Answer

d. furthest

Paleomagnetism

Like many planets, Earth generates its own magnetic field. This magnetic field manifests from flow of a magnetically-conductive fluid: the molten iron/nickel alloy of the outer core. What happens in the core doesn't stay in the core: the magnetic field penetrates the mantle, crust, oceans, and atmosphere, and extends far out into space. This protects our planet from the atmosphere-eroding effects of the solar wind.

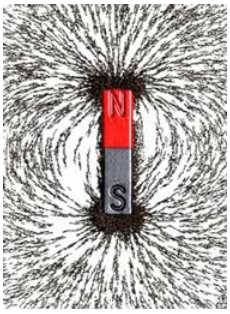


Figure 22.3.7: Bar magnets have two poles, like the Earth's magnetic geodynamo.

Closer to the surface, where rocks form that can later be studied by historical geologists, the magnetic field also makes an impression. Certain minerals are magnetically sensitive. There are several, including ilmenite and hematite, but for our purposes here, let's focus on **magnetite**. Crystals of magnetite have freedom of movement within a molten lava flow or within a body of sediment settling through a water column. Given the freedom to move, they will rotate into alignment with Earth's prevailing magnetic field. When the rest of the surrounding lava is crystallized, or when the surrounding sediment is lithified, the grains of magnetite are no longer free to move. At this point, they are "fossil" evidence of Earth's magnetic field, and if they are moved to

new locations, or new orientations, they take their "stamp" of the old magnetic field with them. We call the study of old magnetic orientations **paleomagnetism**.

Geophysicists who study paleomagnetism (who cheekily call themselves "paleomagicians") collect very precisely oriented samples of magnetite-bearing igneous or sedimentary rocks and bring them back to a lab where a well-shielded magnetometer measures the remnant magnetism.

Inclination

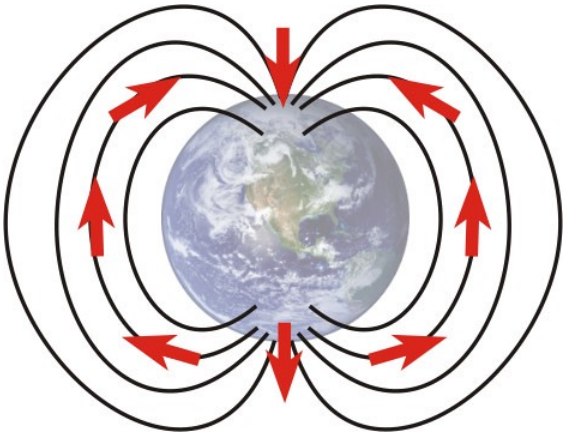


Figure 22.3.8: The shape of Earth's magnetic field is a torus. Lines of magnetic force are vertical at the poles, and horizontal at the equator, with intermediate values in between. (Callan Bentley illustration.)

Latitude on Earth today (position on Earth's surface between the Equator and the poles) corresponds strongly with the **inclination** of Earth's magnetic field. The lines of magnetic force emanate from the South magnetic pole, wrap around the planet, and dive back into the planet at the North magnetic pole. The lines are vertical at these polar locations (*i.e.*, perpendicular to Earth's surface), but horizontal at the magnetic Equator (*i.e.*, parallel to Earth's surface).

To interpret paleolatitude, we apply this modern observation to the past through the lens of uniformitarianism. If we find a sedimentary rock layer with magnetite crystals that show a bedding-perpendicular orientation, we would infer a polar latitude of origin for the original sedimentary deposit. (We apply the principle of original horizontality to infer the bed was laid down more or less horizontally, so a bedding-perpendicular paleomagnetic inclination means "vertical" when it formed.) Conversely, if we find a sedimentary rock layer bearing magnetite crystals that have their magnetic orientation parallel to the bed, then that would imply the sediment was deposited near the paleomagnetic Equator. Intermediate latitudes have values that are intermediate between horizontal and vertical, and the variation occurs in a systematic way.

Declination

Now consider another behavior of a free-spinning magnet in a magnetic field: a compass needle. Like a crystal of magnetite, the compass needle is magnetically susceptible, and will align itself with the planet's magnetic field, pointing toward magnetic north. Fossil magnets pull this same trick. We call the variation between a given direction and north by the name **declination**. If you want to track how a given landmass has moved through geologic time, paleomagnetic declination comes in handy. In this case, you want a sequence of sedimentary strata and/or lava flows, spanning the relevant portion of geologic time. If you are successfully in

extracting paleomagnetic declination information from them, and you can successfully constrain the ages of the layers, you will get a series of time-stamped declination directions. For instance:

500 Ma	"The north magnetic pole was at 190° to the modern north pole, at a latitude distance of 40°."
350 Ma	"The north magnetic pole was at 230° to the modern north pole, at a latitude distance of 60°."
200 Ma	"The north magnetic pole was at 300° to the modern north pole, at a latitude distance of 80°."
50 Ma	"The north magnetic pole was at 350° to the modern north pole, at a latitude distance of 100°."

In this hypothetical example, four dates/orientations spanning half a billion years constrain the movement of that block of crust increasingly southward (crossing the equator sometime between 200 Ma and 50 Ma), and rotating clockwise all the while.

This "testimony" from the strata of a given location is called an **apparent polar wander** path. It is basically how one particular spot on a given continent might view the north magnetic pole through time.

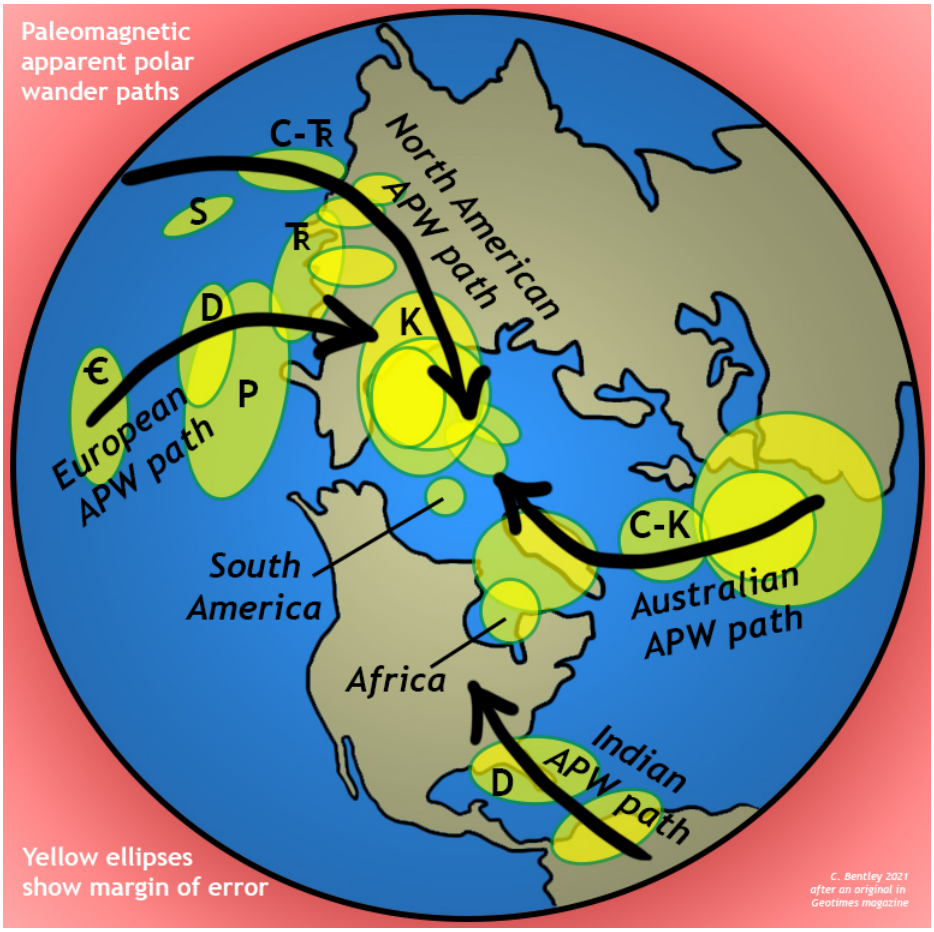


Figure 22.3.9: Apparent polar wander (APW) paths for several continents. Orange ellipses show uncertainties. (Callan Bentley graphic, after an original from *Geotimes*.)

But different continents tell different stories – they have apparent polar wander paths that do not make any sense if the continents are eternally fixed in place. If continents were not permitted to drift, then the implication from a half dozen different continents' apparent polar wander paths is that in the past there were numerous magnetic north poles, arrayed all over the place, and over geologic time they have gotten closer to the modern north pole, and merged into one. This makes no sense, in terms of what we

know about how magnets work. But, if the continents are free to drift, and the seafloor to spread in between them, or subduct in front of them, suddenly you only need one planetary magnetic field to make sense of the continents' plurality of parochial perspectives.

In other words, paleomagnetic inclination gives us a way to quantify the past motion of a given plate (continent) relative to a more or less stationary magnetic north pole.

Polarity reversals

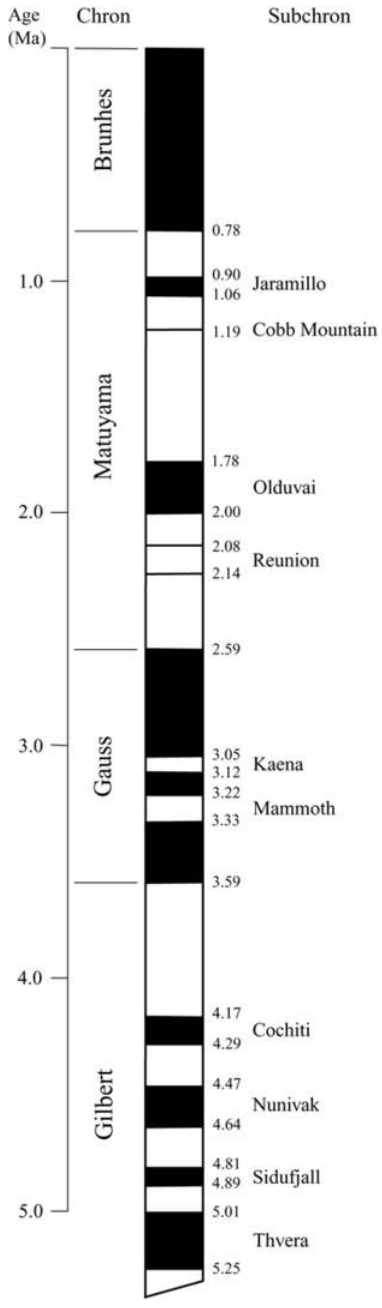


Figure 22.3.10: The last 5 million years of magnetic reversals. (Public domain via Wikimedia.)

The final aspect of paleomagnetism that bears consideration has to do with reversals of the magnetic field. We've previously discussed (a) inclination and (b) declination as being related to angles between a line of magnetic force and (a) Earth's surface and (b) the direction to the north pole. But that line is not just a line; it is an arrow. In other words, all lines of magnetic orientation have a direction to them – a direction in which they point. A rock forming at the modern day south pole would have a vertical magnetic inclination, but so would one forming at the north pole. How can we tell them apart? Earth's magnetic flow exits the planet at the

south magnetic pole and re-enters at the north magnetic pole. So the modern South Pole rock would have an arrow pointing “up” (away from Earth’s surface), while the modern North Pole rock would have an arrow pointing “down” (toward Earth’s interior).

By paying attention to which way these arrows point, we arrive at an interesting insight: they frequently switch back and forth! The reason ***why*** these reversals happen is not entirely clear, but it is plain that they do — and we can use that to our advantage. Not only does it open up a whole new avenue for dating rocks (reading their “magnetostratigraphy” like a bar code) but it also provides a tool for confirming that seafloor spreading is, in fact, real.

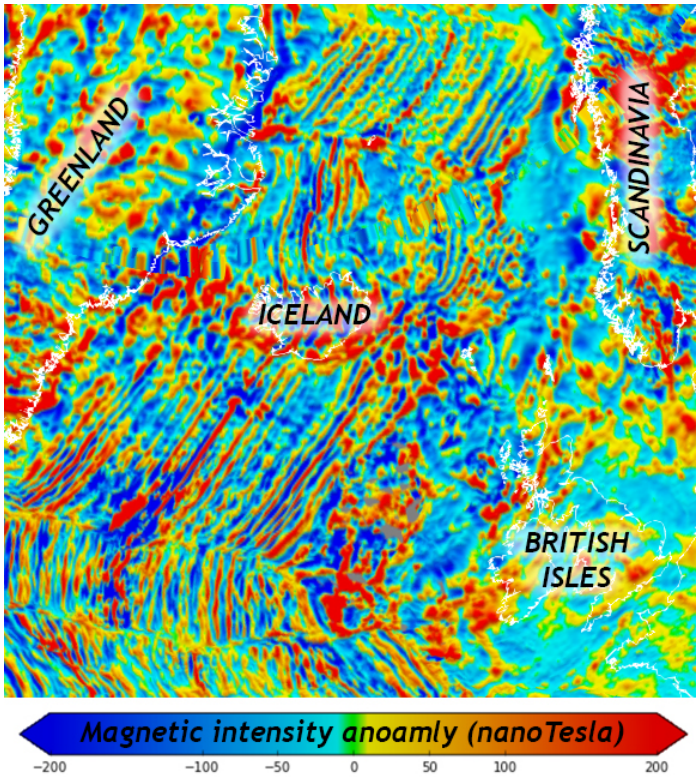


Figure 22.3.11: Magnetic intensity anomaly map of the North Atlantic. Note the “mirror imaging” of the anomaly pattern in the seafloor on either side of the Mid-Atlantic Ridge. Explore more here. (Modified from a presentation of the EMAG2 model by Maus, et al. (2009).)

When new oceanic crust is generated, magma wells up from the mantle below, where it is produced through decompression partial melting. Within that magma are magnetically-susceptible minerals (like magnetite). The magma injects into the crack between the two diverging plates, and solidifies. At that moment, the magnetic signature of that moment in Earth history is locked into the plate. Then tensional stresses build up anew, the crust cracks again, and more magma squirts in to seal it shut, again, taking on the magnetic signature of the Earth. If a reversal happened in between dike-injection events, that will be preserved as “stripes” of oppositely-oriented magnetism on the seafloor — something that we can readily measure using a submersible magnetometer towed behind a ship.

It turns out that the pattern of magnetic reversals in Earth’s oceanic crust is recorded twice – once in each direction leading away from the oceanic ridge system. This is a prediction of the Vine-Matthews-Morley hypothesis (named for three geophysicists who suggested it):

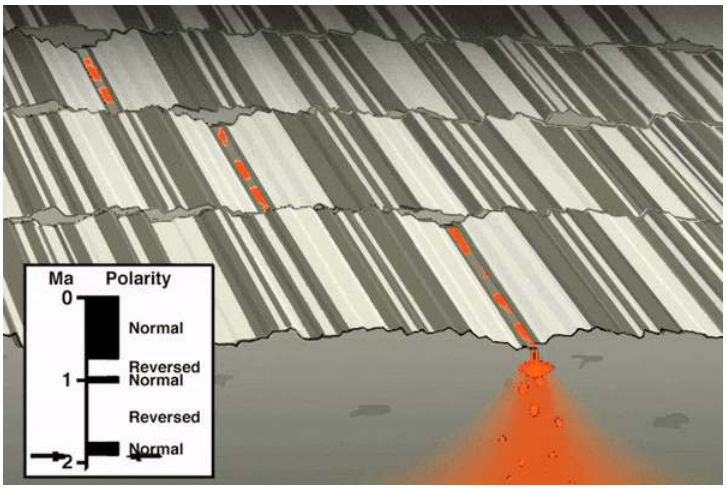


Figure 22.3.12: Spreading along several mid-ocean ridges, showing magnetic striping symmetry. (Animation by Tanya Atwater, UCSB.)

Paleomagnetic stripes in the seafloor are like a doubly-redundant tape recording of the reversals in Earth's magnetic field. The "normal" polarity of the Bruhnes chron that we are currently in can be found astride the ridges all over the world, but head a short distance away (in either direction) and you'll soon encounter the "reversed" signature of the older Matuyama chron. The two records are like mirror images of one another.

These patterns of magnetic "stripes" parallel to the oceanic ridge system and faithfully recording the changes in the planet's magnetic polarity were perhaps the most compelling line of evidence yet in support of seafloor spreading, and thus plate tectonics.

Did I Get It? - Quiz

? Exercise 22.3.1

What is the origin of Earth's magnetic field?

a. The Moon

b. A million mantle magnets

c. Convective flow in molten iron of the outer core

Answer

c. Convective flow in molten iron of the outer core

? Exercise 22.3.2

The study of ancient magnetic fields preserved in rocks is _____.

a. Historical magnetology

b. Fossilferrum

c. Dipole deduction

d. Paleomagnetism

Answer

d. Paleomagnetism

Exercise 22.3.3

Inclination is _____.

a. The angle at which a magnetic field line intersects Earth's surface.

b. The orientation of the planet's magnetic field (i.e., north magnetic pole at south geographic pole, or vice versa).

c. The horizontal direction toward the magnetic North Pole.

Answer

a. The angle at which a magnetic field line intersects Earth's surface.

Exercise 22.3.4

What is declination?

a. The horizontal direction toward the magnetic North Pole.

b. The orientation of the planet's magnetic field (north magnetic pole at south geographic pole, or vice versa

c. The angle at which a magnetic field line intersects Earth's surface.

Answer

a. The horizontal direction toward the magnetic North Pole.

Exercise 22.3.5

What does "apparent polar wander" describe?

a. The north magnetic pole dances around over time, and every continent gets its own north magnetic pole.

b. From the perspective of fossil magnets preserved within stratified rocks on a given landmass, over time the relative position of the north magnetic pole changes, sometimes closer, sometimes further away, sometimes this way, sometimes that way. "Apparently," the pole is wandering, from the perspective of a stationary continent. Landmasses aren't actually stationary, though so really the continent is doing the wandering.

Answer

b. From the perspective of fossil magnets preserved within stratified rocks on a given landmass, over time the relative position of the north magnetic pole changes, sometimes closer, sometimes further away, sometimes this way, sometimes that way. "Apparently," the pole is wandering, from the perspective of a stationary continent. Landmasses aren't actually stationary, though so really the continent is doing the wandering.

Exercise 22.3.6

In the geologically recent past, what have all continents' apparent polar wander paths come closer to?

a. The South Geographic Pole

b. The Mid-Atlantic Ridge

c. Middle Earth

d. The North Magnetic Pole

Answer

d. The North Magnetic Pole

Exercise 22.3.7

What are the "stripes" of normal and reversed paleomagnetic orientation in the oceanic crust parallel to?

a. Deep sea trenches

b. The San Andreas Fault

c. The oceanic ridge system

d. Mountain belts in the continental crust

Answer

c. The oceanic ridge system

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22.4: Hot spot tracks

Remember that one of the errors Alfred Wegener made was to estimate too-fast rates of speeds for his drifting continents. Now that the idea of seafloor spreading had been established, a test could be made. Numerical dating of igneous rocks had become validated as a technique during the interregnum between Wegener's book and Hess's paper, and attention soon turned to island archipelagos such as Hawaii and the Galapagos.

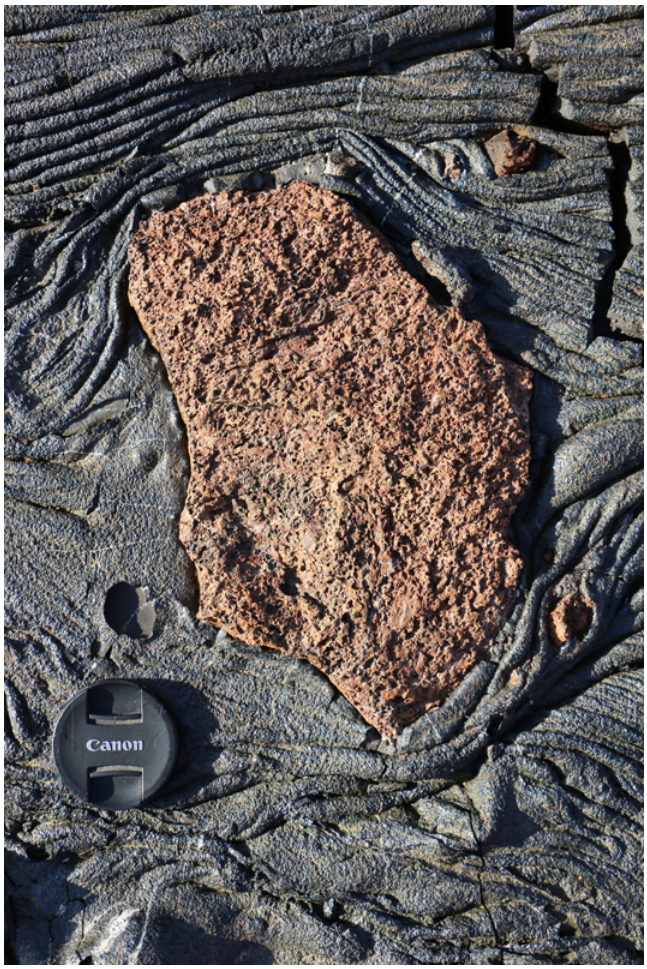


Figure 22.4.1: Xenolith of orange scoria in pahoehoe basalt, Santiago Island, Galapagos. (Callan Bentley photo.)

Unlike volcanic island arcs, which feature a trench-parallel line of volcanoes all erupting at various times, on and off, over and over, Hawaii and the Galapagos are different. In each archipelago, the volcanoes show an age progression. At one end of the chain is an active volcano or two, erupting lava and ash in the present day. But as you work your way toward the other end of the chain, the volcanoes are inactive and more weathered. Sure enough, radiometric dating confirms our suspicion: they get older in one direction! In Hawaii, the volcanoes get older toward the northwest. In the Galapagos, the volcanoes get older toward the east. One explanation that has been offered to explain this age progression is that there is a source of magma unrelated to plate tectonics, piercing upward periodically with injections of magma as the plate of oceanic lithosphere moves steadily overhead. This simple notion has a simple name: it's a **hot spot**.

Hot spot volcanism implies that if a plate, such as the Pacific Plate, is moving at a steady rate in a given direction over a stationary hot spot, periodically magma from the hot spot will burst through, making volcanoes on the surface of the plate, which will then trundle them along, away from the hot spot. The volcano will go dormant, begin to weather and erode, and will subside. To the northwest of the main Hawaiian islands are a series of seamounts: not nearly as much landmass pokes up above sea level, but they still represent a massive bump on the face of the Pacific Plate. Advocates of the hot spot interpretation for Hawaii would interpret these seamounts as places that used to be on top of the hot spot, but have long since been dragged away.

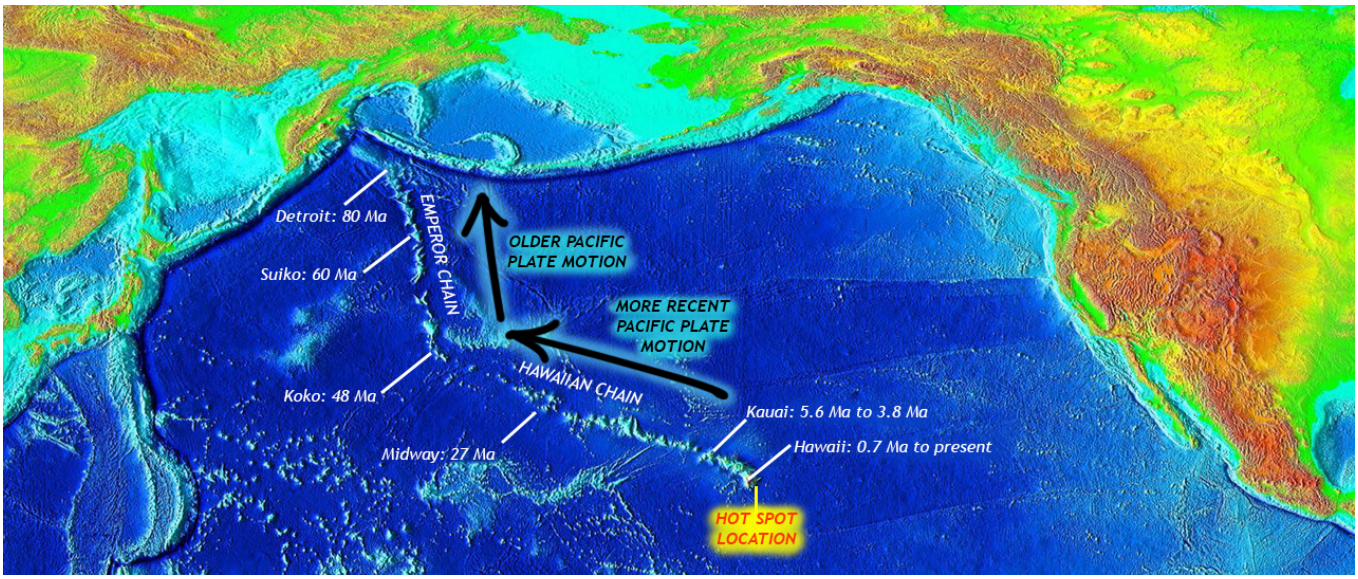


Figure 22.4.2: The Hawaii-Emperor seamount chain. (Callan Bentley modification of USGS map.)

A kink in the seamount chain can be found northwest of the Midway Atoll, where the Emperor Seamount chain trends off toward the north. This reflects a change in direction of the Pacific Plate that took place around 48 Ma.

The Pacific is a fast-moving plate. In some places, it's moving about 10 cm/year relative to a fixed frame of reference. Plates can dawdle along at a slower pace, too; For instance, the Atlantic is widening at only 2.5 cm/year. Even a fast moving plate is still too slow for the human eye to perceive. We frequently liken the pace of plate movement to the rate of fingernail growth. You can't see it happen, but you know it happens. I suppose nail clippers are the subduction zone in that analogy…

Did I get it? - Quiz

? Exercise 22.4.1

Prior to the inception of GPS, how did hot spots help constrain plate motion?

a. When continents collide, the crust is thickened and gets hot, and this helps encourage the growth of metamorphic minerals -- minerals that can be numerically dated to figure out when the orogeny happened.

b. As a plate moved over a hot spot, every now and then a volcano (with numerically-datable minerals in its lavas) would erupt and solidify, and then be carried away.

c. The hot spots generated a magnetic field which was imprinted into the oceanic lithosphere.

Answer

b. As a plate moved over a hot spot, every now and then a volcano (with numerically-datable minerals in its lavas) would erupt and solidify, and then be carried away.

? Exercise 22.4.2

Which of the following is thought to be a hot spot volcanic track?

a. the Aleutian Islands of Alaska

b. Cuba, Hispaniola, and Puerto Rico

c. Hawaii and the Emperor Seamounts

d. the Great Rift Valley of East Africa

Answer

c. Hawaii and the Emperor Seamounts

Exercise 22.4.3

How fast do plates move?

a. 2-20 m per year

b. 2-20 km per year

c. 2-20 cm per year

d. 2-20 mm per year

Answer

c. 2-20 cm per year

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22.5: Conclusion

Scientific understanding proceeds erratically. Alfred Wegener's attention to the continents was valuable, but wasn't capable of achieving intellectual liftoff on its own. Our understanding of Earth surface dynamics needed a shot in the arm – an infusion of fresh data. When geophysical surveys of the ocean basins were revealed in the aftermath of a great global conflict, seafloor spreading was born. The new idea about the ocean basins was mixed with continental drift to yield the theory of plate tectonics.

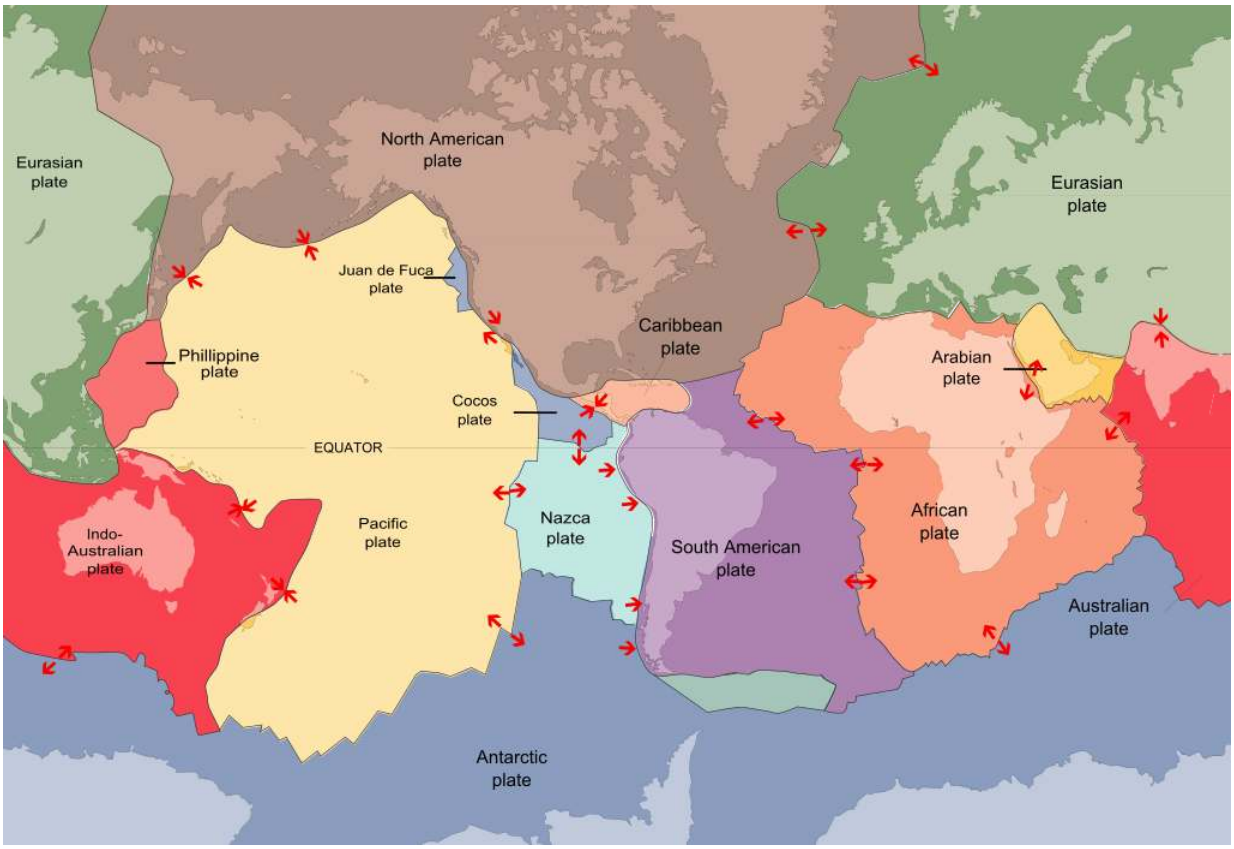


Figure 22.5.1: Plates and plate boundaries with a general sense of motion identified.

Plate tectonics says that the outermost rocky surface of our planet is broken into a series of slabs called "plates." These plates may be made of continental lithosphere, oceanic lithosphere, or (usually) both. The plates move around, and the action that occurs at their boundaries (convergent, divergent, and transform) are vital in producing a suite of major geological phenomena: mountain belts, volcanic arcs, deep sea trenches, oceanic ridges, rift valleys, laterally offset geologic units and belts of earthquake activity.

Now you have a sense of the scientific history of this important idea. But plate tectonics isn't stuck in the 1960s. The past half a century has repeatedly tested and refined our understanding of the behavior of Earth's lithosphere, leading to great insights in Historical Geology. If you haven't already read it, it's now time to examine the discussion of modern plate tectonics.

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22.6: Further reading

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CHAPTER OVERVIEW

23: (Case Study) The Grenville Orogeny and the rifting of Rodinia

23.1: The Grenville Orogeny and the rifting of Rodinia

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23.1: The Grenville Orogeny and the rifting of Rodinia



Did I Get It? - Quiz

Exercise 23.1.1

Granite gneisses (1.1 Ga) and granites (1.0 Ga) are evidence for what tectonic event?

a. The rifting of Rodinia

b. The Taconian Orogeny

c. The rifting of Pangaea

d. The Grenville Orogeny

Answer

d. The Grenville Orogeny

Exercise 23.1.2

The rifting of Rodinia opened which ocean basin?

a. the Pacific Ocean

b. the Iapetus Ocean

c. the Tethys Ocean

d. the Atlantic Ocean

Answer

b. the Iapetus Ocean

Exercise 23.1.3

What kinds of rocks record the break-up of Rodinia?

a. Granite and charnockite

b. Limestone and garnet schist

c. Basalt and immature clastic sediment

d. Granite and granite-gneiss

Answer

c. Basalt and immature clastic sediment

? Exercise 23.1.4

What is the nature of the contact between the Blue Ridge basement complex and the Swift Run and/or Catoctin Formation?

a. Unconformity

b. Conformable sedimentation

c. Thrust fault

Answer

a. Unconformity

? Exercise 23.1.5

What primary volcanic structure can be found in the Catoctin Formation?

a. Cooling columns

b. Pahoehoe texture

c. Stratovolcanoes

d. Fossil tree trunks

e. Volcanic bombs

Answer

a. Cooling columns

? Exercise 23.1.6

What is distinctive about the charnockites of the Blue Ridge basement complex, compared to other granitoids?

a. They have gold in them.

b. They have pyroxene in them.

c. They have garnet in them.

d. They have silicate minerals in them.

Answer

b. They have pyroxene in them.

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CHAPTER OVERVIEW

24: (Case Study) Tippecanoe Sequence - A sequence stratigraphic view

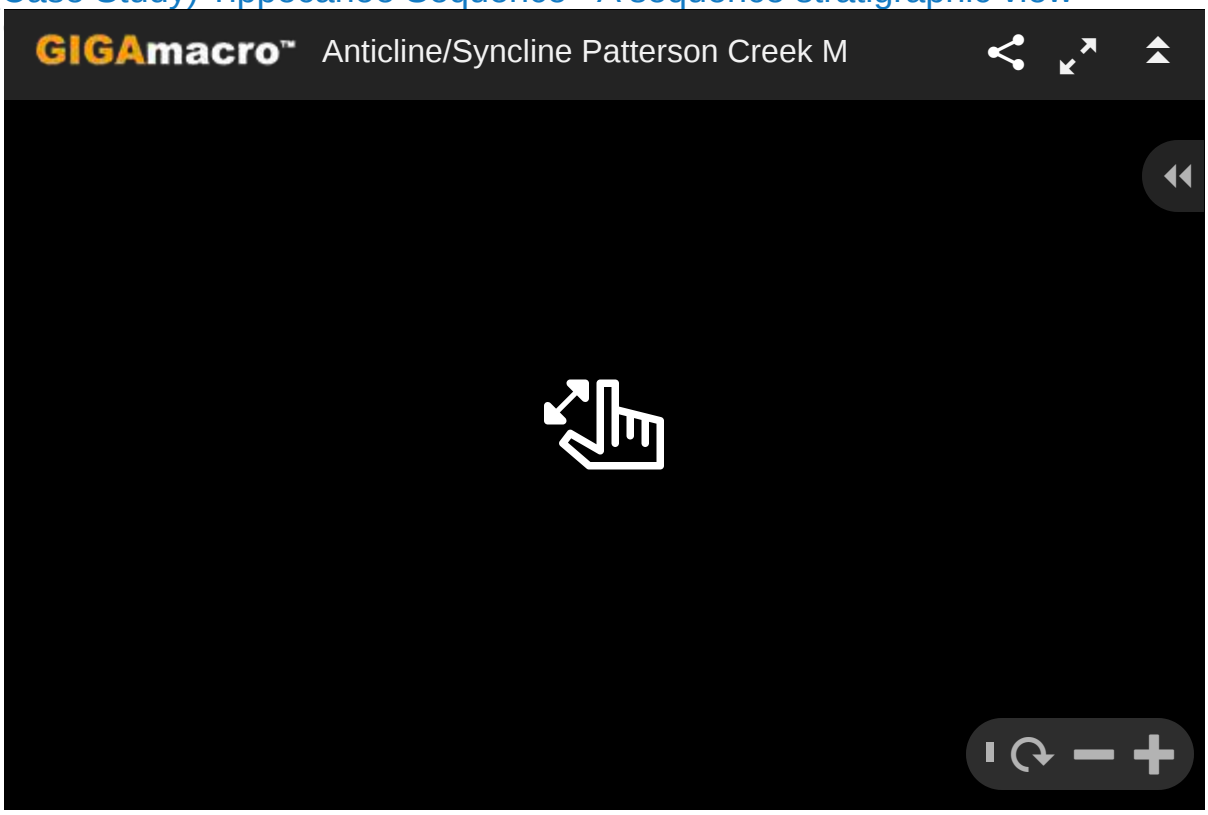


Figure 24.1: Outcrop of the Silurian Tonoloway Formation, located along Corridor H (US48) in West Virginia. This formation is a part of the Tippencanoe Sequence.

Key Concepts

After reading this chapter, students should be able to:

- Identify and describe the three North American cratonic sequences;
- Describe the stages of the Wilson Cycle.
- Distinguish between sea level transgression and regression.
- Describe the various Systems Tracts that comprise a sequence.

Note

This study is best understood after spending time studying Sequence Stratigraphy basics. For a primer, please read: Stratigraphic Tools: Basic Sequence Stratigraphy.

24.1: Sequences of strata in ancient Eastern North America
24.2: The Tippecanoe Sequence - Transgressing Seas, Taconian Mountains
24.3: Final Thoughts
24.4: References

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24.1: Sequences of strata in ancient Eastern North America

The Appalachian region of the United States east coast provides an excellent record nearly the entire Paleozoic Era in its rock record. A drive along Corridor H (US 48) through eastern West Virginia will start in the Cambrian carbonate strata and end in Carboniferous coals and sands. During this time, the North American craton would see the opening and closing of the proto-Atlantic Ocean (Iapetus Ocean), plate movement from southern subtropical latitudes to tropical latitudes, and the coming and going of epicontinental epeiric seas. The rock record in this region is diverse and important. The ebb and flow of sea level and facies have left a very detailed record of changes over time.

Sedimentary Sequences of the Paleozoic

Four large sequences of sea level transgression took place around the Laurentian craton from the late Proterozoic Era through the end of the Paleozoic Era. The oldest, the Sauk Sequence, is dominated by carbonates. The youngest two, the Kaskaskia and Absaroka, record the closing of the Iapetus Ocean Basin. In between these and the Sauk exists the Tippecanoe Sequence. Based upon sediment thickness and volume alone, it is likely that this transgression was the deepest of all of them in terms of water depth. It also records the evolution of the ancient Taconian Mountains.

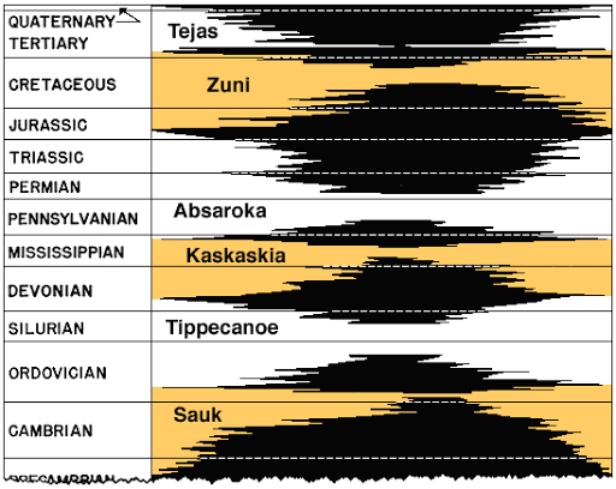


Figure 24.1.1: Sedimentary “Sloss” Sequences of North America. In the Mid-Atlantic region, the Sauk, Tippecanoe, and Kaskaskia are well exposed. White and yellow areas represent major marine transgressions onto the craton while black areas represent major unconformities (sequence boundaries). The Sauk is bounded on the bottom by the “Great Unconformity”, the boundary between the Sauk and Tippecanoe is referred to as the “Knox Unconformity”, and the boundary between the Tippecanoe and Kaskaskia is the “Wallbridge Unconformity”. All six of Sloss’ Sequences are 2nd order supercycles. (Sloss, 1964)

In 1963, a geologist named Lawrence Sloss from Northwestern University would become famous because of a paper published as a part of a symposium held by the Kansas Geological Survey and collectively published in KGS Bulletin 169. Sloss identified six sequences of sedimentation that occurred along the margins and interior of the of ancient North America, known as Laurentia, from the late Neoproterozoic through the early Cenozoic Eras. These sequences were given names. The oldest was the Sauk, followed by the Tippecanoe, Kaskaskia, Absaroka, Zuni, and Tejas. The names of these have their various origins, but geological names are always linked to a physical location. The Sauk Sequence get its names from rocks in the vicinity of Sauk County, Wisconsin. The Tippecanoe Sequence is named for rocks in the vicinity of the Tippecanoe River in the U.S. State of Indiana. The Absaroka Sequence is named for rocks described in vicinity of the Absaroka Range in Wyoming. The name Tippecanoe really refers to the sea that existed in the Iapetan Basin that these layers of strata represent. This sequence of rock layers records the history of this sea at that location. As the Tippecanoe Sea surrounded all of Laurentia, there are many places in North America where you can see a progression of this sequence in the stratigraphic record, including, for example, in the Grand Canyon.

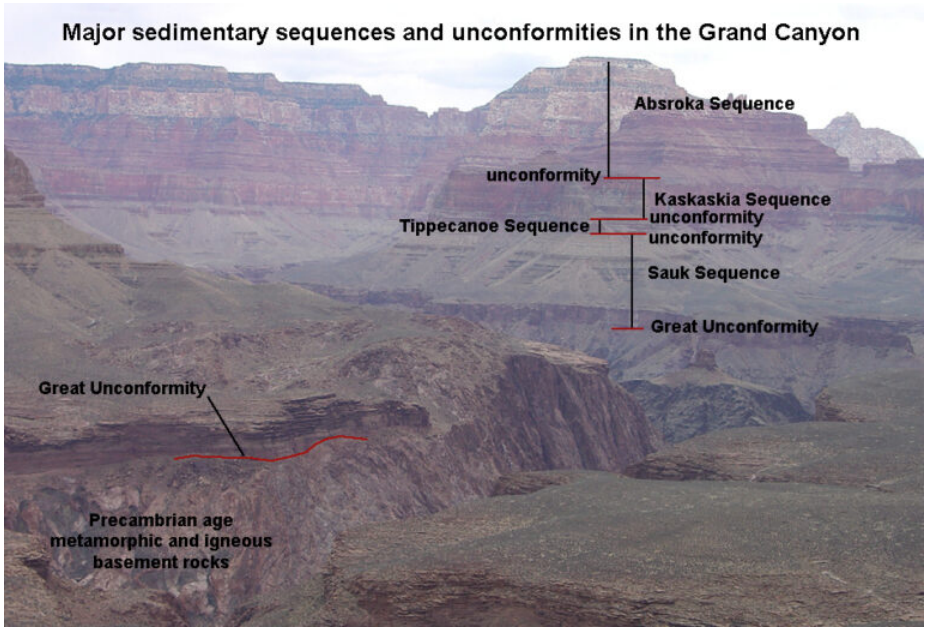


Figure 24.1.2: Cratonic Sequences, or Sloss Sequences, as they exist in the Colorado Plateau and Grand Canyon region in the American southwest.

Sloss (1964) identified these sequences through the development of curves based upon data on land subsidence (lowering), cratonic interior uplift, and sedimentation. The lower plot in the figure below, is the basis for the famous sea level curve eventually developed by other researchers working for Exxon in the 1970s. The curve explores global (eustatic) sea level changes as they relate to base level. Each Supersequence, Sauk, Tippecanoe, etc., is similar. The sequence begins with high eustatic sea levels above the base level, progresses through a time of much lower sea levels, and then ends with a major transgression of the sea to end the sequence. Each sequence begins with a sequence boundary, a regression and progradation of the coastline, and then ends with a major transgression. In the middle of all of these sequences are major orogenic (mountain-building) events, as depicted above.

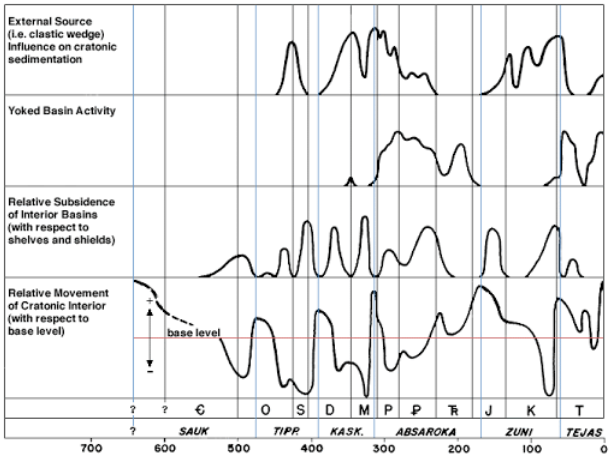
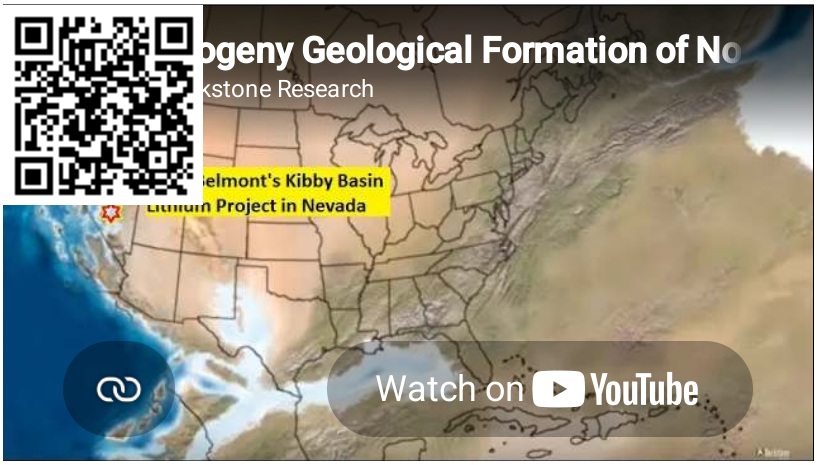


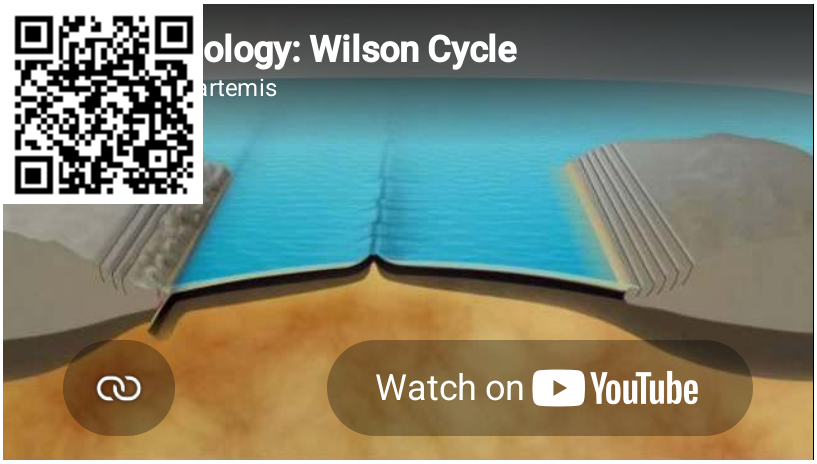
Figure 24.1.3: Four plots depicting relative tectonic activity across ancient North America. From the top down, Sedimentation fluctuations, Yoked basin activity, Relative land subsidence, and Relative movement of the cratonic interior. Tectonic, sedimentation, and global (eustatic) sea level activity would contribute to the deposition of genetically-related strata being deposited over the course of six periods of time as sequences of sediment packages separated by erosional sequence boundaries. (From Sloss (1964).)

Watch this great video that puts together the Sloss Supersequences with the tectonic and environmental changes taking place in North America. Pay attention to the time periods in the Paleogeographic images. The Sloss Supersequence diagram is much clearer if you watch it in full screen however, you can match the diagram in the video with the diagram above for better clarity.



The Wilson Cycle

Four of these six sequences (Sauk through Absaroka), all were deposited during one very long 1st order megasequence. This megasequence marks the opening and closing of the Iapetus (ancient Atlantic Ocean) as a whole. It is also known as a "Wilson Cycle". The Wilson Cycle, named after the Canadian geophysicist J. Tuzo Wilson, is a model that explains how an ocean basin opens and closes, beginning with a divergent boundary and ending with collisional subduction between two continents. In his seminal work (1966), Wilson built upon earlier observations of fossil similarities existing on either side of the Atlantic by proposing that there also existed a "proto-Atlantic Ocean" between the continents during the Paleozoic Era. This ocean closed by stages, bringing dissimilar rocks together as the supercontinent Pangaea formed. This simple model can then be applied to other contexts, such as the closing of the Tethys Sea as India moved northward toward Asia, etc. Watch this simple schematic video of the Wilson Cycle below.



Stages of the Wilson Cycle

There are nine stages to a Wilson Cycle, as described in the diagrams below. Stage A begins with a stable continental craton and progresses to Stage B, where continental rifting begins. This rifting progresses through stages C through E with passive margins on both sides of the new ocean basin.

Closing of the basin begins with Stage F and then runs through Stage I, where the two continents come together to form a new supercontinent. In the meantime, Stages E through H record the formation of orogenic events and the accretion of terranes. This cycle will be described in more detail in another chapter. The Tippecanoe Sequence takes place in this model between stages E through G.

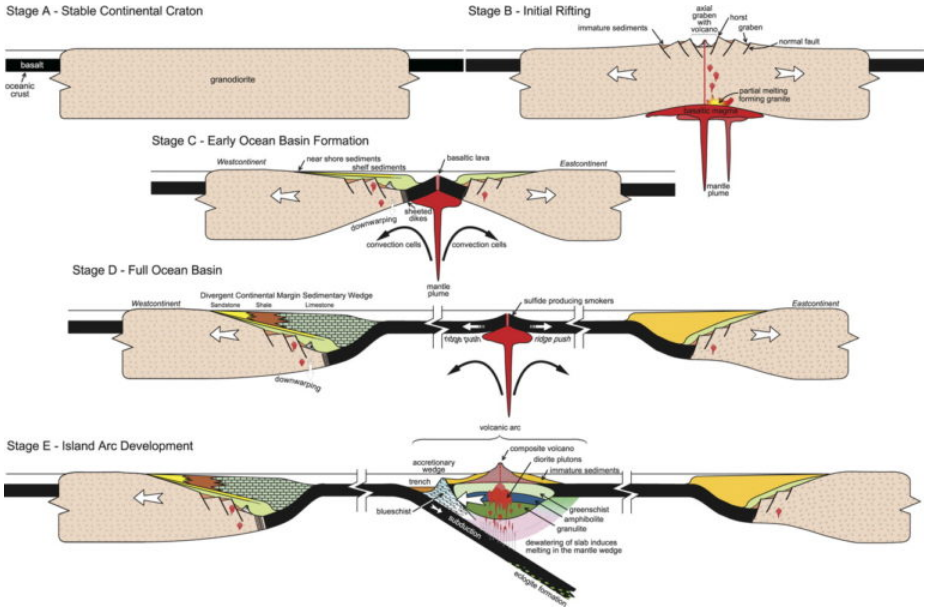


Figure 24.1.4: Stages A through E of a Wilson Cycle (Whitmeyer et al., 2007)

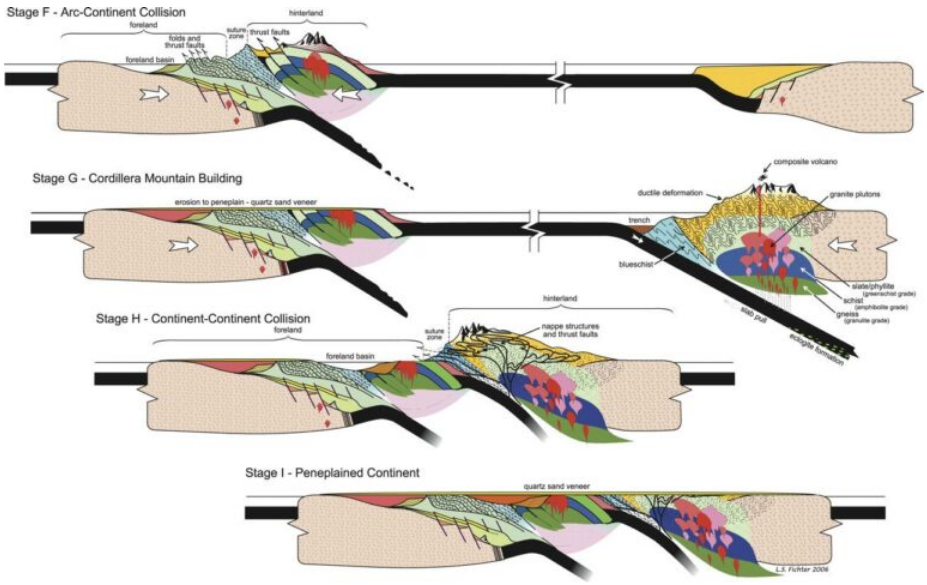


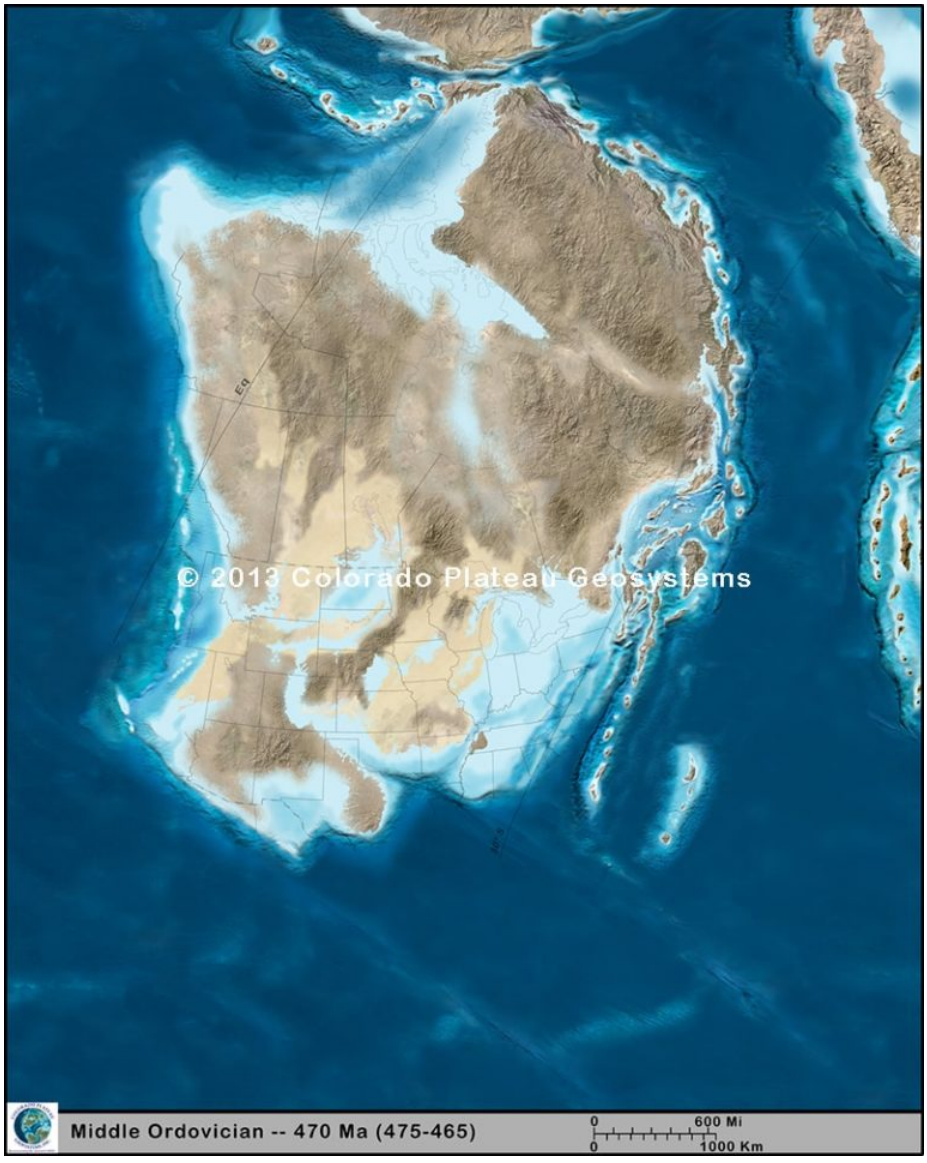
Figure 24.1.5: Stages F through I of a Wilson Cycle (Whitmeyer et al., 2007).

The Tippecanoe Sequence begins with missing rock, an disconformity. Called the Knox Unconformity, it serves as the sequence boundary across the craton. It can be seen in the strata of the Grand Canyon as well as it is seen in the strata of western Virginia. It is an erosional surface resulting from a rapid drop in sea level. Throughout the Tippecanoe, sea level is transgressing onto the craton. As average sea level rose, there were fits and starts of smaller sequences, higher order sequences, within the Tippecanoe. These are due to a variety of factors. The story of the Tippecanoe includes Bahamian-like offshore reefs and tidal expanses, extensive muddy submarine landslides, sandy beaches, and a return to carbonates. It includes volcanic ash from Taconian eruptions and a variety of important body fossils.

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24.2: The Tippecanoe Sequence - Transgressing Seas, Taconian Mountains

The story of the Tippecanoe is the story of the beginning of the end of the Iapetus Ocean. During the late Proterozoic, a supercontinent referred to as Rodinia would break up. The breakup of this supercontinent would create a rift basin into which seawater would flow, eventually widening into a very large ocean. The event left its mark across the Laurentian craton, in the form of a basal sequence boundary. Known as the Great Unconformity, this erosional boundary would mark the start of deposition of sediments along the margin of the Sauk Sea that existed from the late Precambrian through the early Ordovician. The end of the transgression of the Sauk Sea (Highstand Systems Tract) marked a second important sequence boundary (Falling Stage Systems Tract). Known as the Knox Unconformity, it marks the base of the sequence of rocks that record the Tippecanoe Sea's interactions with North America from the middle Ordovician Period through the middle Devonian. (See stratigraphic column below). Before it was all over, it be interrupted by the Taconian Orogeny, a major orogenic event. Laurentia at this time sat mostly south of the paleoequator with the modern eastern United States along its southern flank. Eventually, the Tippecanoe Sea reaches its penultimate highstand in the middle Devonian and, as the Falling Stage Systems Tract commences the formation of the Wallbridge Unconformity, deposition of Tippecanoe sediments comes to an end.



Eventually, the Taconian Mountains formed off of the coast an arc of Japan-like volcanic islands. Evidence for this is contained within the lithostratigraphic record as widespread and in many cases very thick volcanic ash deposits. While the entire sequence is

transgressive across the craton, tectonic change and sedimentation changes taking place due to the process of orogeny along the southern portion of the craton would lead to more localized changes. Relative sea level, sedimentation, and tectonics would lead to lower orders of sea level sequences (3rd and higher). Embedded within the Tippecanoe as a whole then are successively smaller sequences of shorter duration that record events taking place throughout this time period.

Middle Ordovician Tippecanoe Sea. Note the position of the Paleoequator. The Taconian Island arc is still off of the southern coast of Laurentia (modern eastern North America) and the Tippecanoe Sea lies to its north (forearc basin). Sediments deposited in the forearc basin include the early Tippecanoe Carbonates into the Martinsburg Formation.

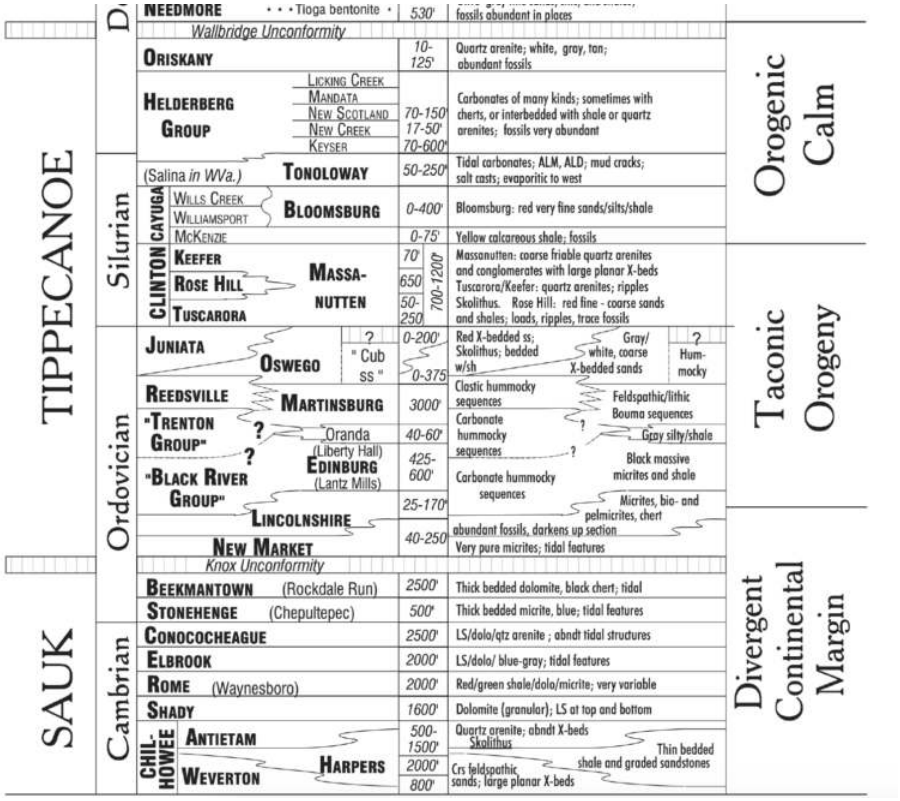


Figure 24.2.1: Generalized stratigraphic column for the central and northern Shenandoah Valley of Virginia and eastern West Virginia,
encompassing the lithologic units visited on Day 2 of the field trip. After Fichter and Diecchio (1993).

Tippecanoe Sequence 1

We can divide the Tippecanoe into two smaller sequences. The first records deposition directly following the Knox Unconformity and ended with the onset of coarse sandstones. This 3rd order sequence includes the carbonate bank sediments of the lower Tippecanoe, the New Market Limestone through the early Martinsburg Formation. It also includes the very distal upper Martinsburg Formation to the base of the Oswego Formation, where an unconformity existed across some of the area. The Oswego Formation has places in it that are rich in river cobbles, a good indication that this represents sediment that was being deposited very near land. Below is an accommodation space plot for this 3rd order sequence:

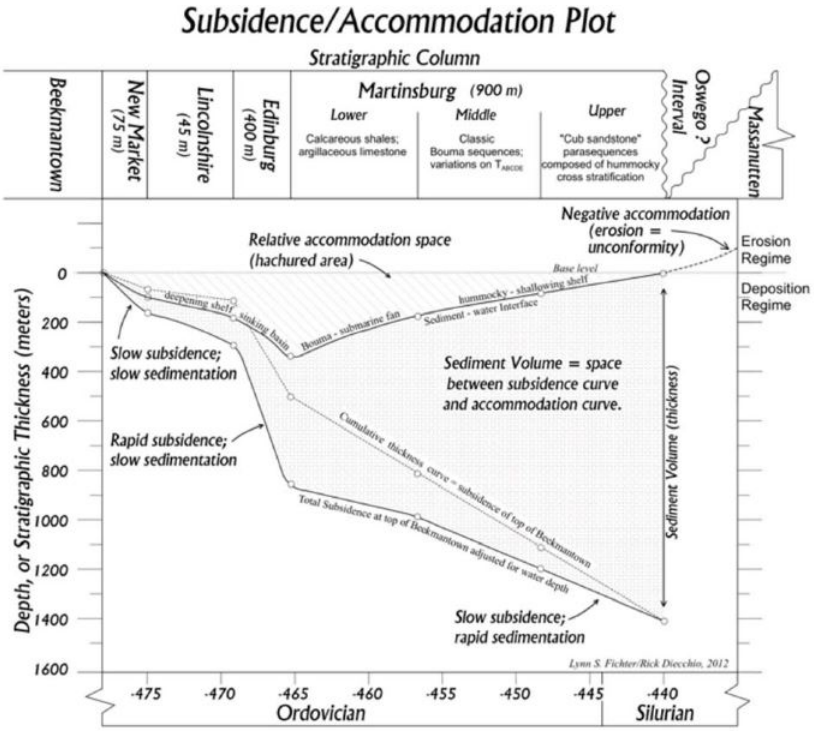


Figure 24.2.2: Subsidence and Accommodation Plot for the first half of the Tippecanoe Sequence in the Mid-Atlantic of North America (Whitmeyer et al., 2015).

This plot might be a bit confusing at first. But, you are looking at the cross section of the Taconian forearc basin over time during this portion of the Ordovician Period (left to right). As time passed the shelf basin was deepened and accommodation space increased (hachured area). Carbonate deposition was still the norm, as it was during the latest portion (uppermost) of the Sauk Sequence.



Figure 24.2.3: The contact between the lower Ordovician New Market Formation and Lincolnshire Formation. These carbonate deposits are typical of rocks deposited at a part of the “Great American Carbonate Bank”. They also represent shallower-water conditions. This is because of their carbonate lithology, fossil content, and structure.

The sequence began with the lowstand systems tract (LST), as recorded by the New Market Limestone. The sequence boundary already existed (Knox Unconformity) and the bulk of sediment being deposited was being formed in place by organisms secreting carbonate for their shells, tests, and some perhaps through evaporation. The seas at this time were conducive to calcite, rather than aragonite, being the primary shell-forming mineral. Abundant seafloor spreading is keeping magnesium levels stable and a warm global climate favored widespread biochemical carbonate production.

This sequence began to deepen as the onset of the Transgressive Systems Tract (TST) approached, marking the progression of deposition into the Edinburg Formation. This deepening, or subsidence, was the result of the downward flexure of the forearc basin. The transgression of relative sea level was due to the approaching Taconian island arc. Subsidence continued until about 468 million years ago, when deposition of the lower Martinsburg Formation began. Terrestrial sediments took over, eventually drowning out carbonate production.



Figure 24.2.4: Martinsburg formation, Page County, VA. The Martinsburg is composed of cyclical packages of turbidite sequences. These “fining-upward” sequences are interpreted as representing offshore mixed carbonate-siliciclastic conditions below wave-base. The dominant lithologies are siltstones and shales. Carbonate-rich materials are richest at the base of the formation, where the transition from the Edinburg Formation below is gradational in many places (Stickley Run Member, Martinsburg Fm.).



Figure 24.2.5: Graptolite fossils in the Martinsburg Formation, Page County, VA. Graptolites are fossilized here as carbon films. They break into small pieces very easily. Preservation in such detail is indicative of rapid burial, which would be typical in turbidite deposits.

During the Highstand Systems Tract (HST), submarine turbidite (submarine landslide) deposits became the norm. Accommodation space began to diminish as the Taconian highlands approached. During HST, sediment supply from land typically causes a prograding shoreline situation. This was likely the case for the Martinsburg Formation as its deposition progressed. Beginning as a very distal carbonate/mud in the lower Martinsburg, it eventually prograded into more proximal thick-bedded deposits as terrestrial sediment supply continued to increase and relative sea level dropped. By the end of Martinsburg Formation deposition, over 900m of sediment had been deposited and an erosional boundary formed at the base of the Oswego Formation.

At this point, the Falling Stage Systems Tract (FST) was completed and an entire sequence stratigraphic cycle is complete.

During this time, the craton as a whole experienced the results of an overall rise in global (eustatic) sea level. But, the paleocoastline (of today's mid-Atlantic) saw a period of transgression (New Market Lm to Edinburg Lm) followed by a period of regression (Lower to Upper Martinsburg Fm). The end of this first 3rd order sequence of the Tippecanoe brought to an end the Ordovician Period and the full Taconian Orogeny was underway.

Tippecanoe Sequence 2

The latter half of the Tippecanoe is marked by another 3rd order sequence.

Beginning with the landward Massanutten Formation, replete with its *Skolithos* sp. and *Arthrophycus* sp. burrows marking it as sandy shoreline facies, this entire sequence ends beneath the Wallbridge Unconformity. The Taconian mountains reached their peak elevation during the late Ordovician and very early Silurian and were eroding into the nearby passive margin basin.



Figure 24.2.6: Massanutten Sandstone, Woodstock Tower, Massanutten Mountain, VA. The image below is of a bedding plane surface on a nearly vertical slab of Massanutten Sandstone. Replete with what are very likely Arthrophycus sp. trace fossils, the Massanutten Fm. represents shallow water conditions typical of the base of a sequence.



Figure 24.2.7: Hand sample of the Massanutten Sandstone. Note the coarse-grained nature of the specimen. Unlike the base of the other 3rd order sequence of the Tippecanoe, the Massanutten Sandstone is not a carbonate.

This passive margin (FST to HST) brings with it an eventual break from siliclastic sedimentation, marking the erosion of the Taconian mountains now to a peneplain. With the eventual onset of the Acadian Orogeny coming at the end of the Tippecanoe, there is time for a carbonate bank to form once again in this still very tropical sea, south of the Paleoequator. The transition from the coarse sandy Massanutten Sandstone to the fine-grained intertidal Bloomsburg and Tonoloway Formations and into the Helderberg Limestones that represent this new carbonate bank is very typical of TST conditions. Sea level is rising, even if the rocks represented here are, overall, more proximal to shore. Deepening is occurring.

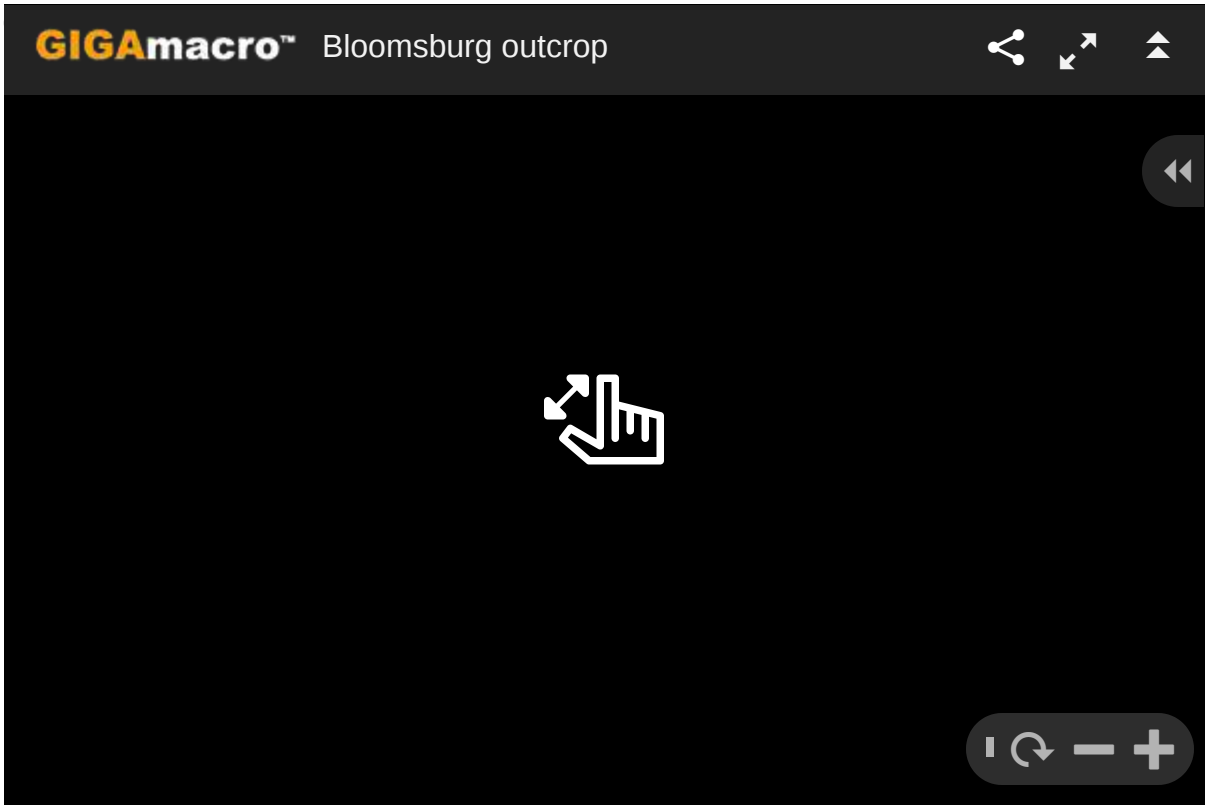


Figure 24.2.8: Bloomsburg Formation, Fort Valley, VA. This iron-rich sandstone was mined as a source of iron during the 19th century in the Shenandoah Valley of Virginia. Finer-grained than the Massanutten Sandstone, it represents the early onset of TST.



Figure 24.2.9: Tonoloway Formation, hand samples from Corridor H in West Virginia. The left image shows evaporite structures in the form of salt casts. There is a short-term cyclicity evident in the sample also, as the environment alternated from carbonate production to sabkha. The image on the right is of dessication cracks, a cross-cutting feature indicative of very dry conditions.

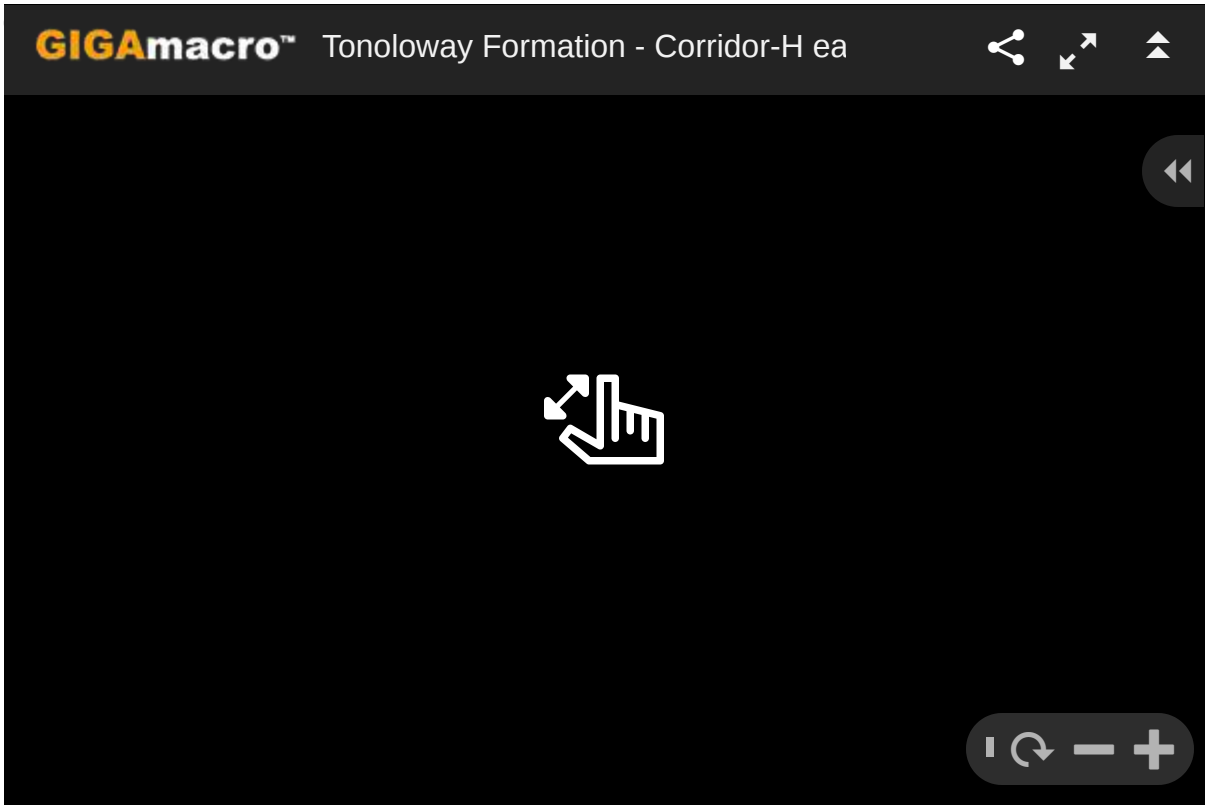


Figure 24.2.10: The Tonoloway Formation, Corridor H, West Virginia. This outcrop of the Tonoloway formation well represents the higher-order cycles that exist just within this single formation. Carbonate production occurring in marine conditions is alternated with periods of intense evaporation.

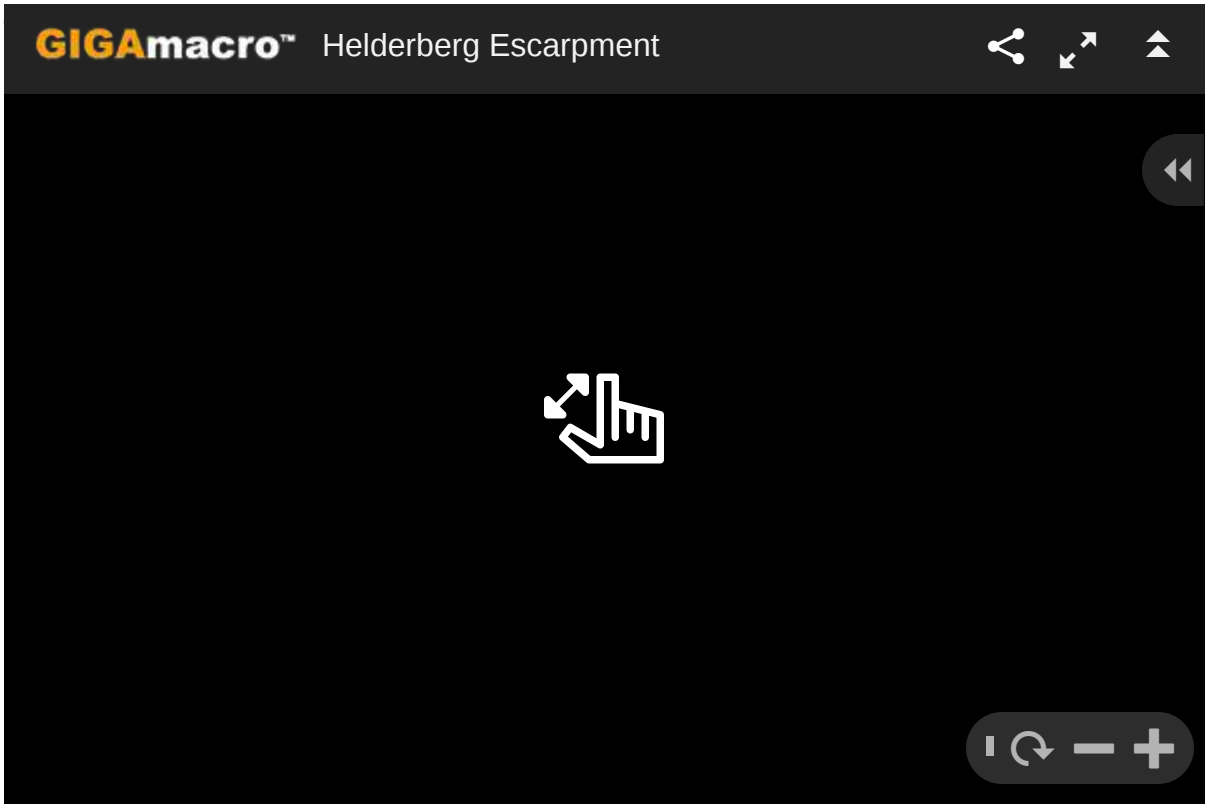


Figure 24.2.11: Helderberg Limestone escarpment, New York. These massive limestones represent the deepest waters of the HST.

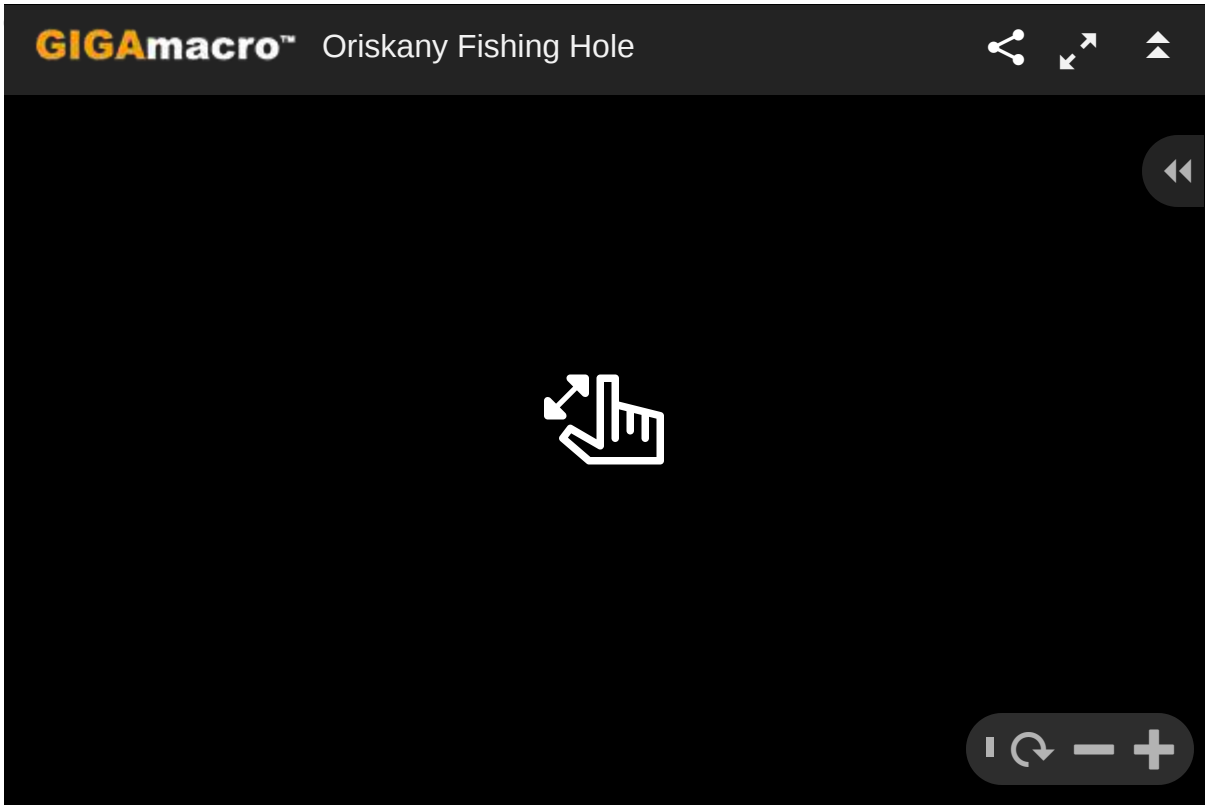


Figure 24.2.12: Late Silurian (Taconian Orogeny) Laurentia.

The sequence boundary is best placed at the top of the Oswego Formation. LST existed at the time of the deposition of the Massanutten (Tuscarora) Sandstone. The coarse sands and trace fossils of the Massanutten are indicative of shallow water. The Transgressive Systems Tract, TST, began when the first carbonates appear in the stratigraphy, interbedded with the very iron-rich red beds of the Bloomsburg Formation. The seas were beginning to rise and accommodation space was increasing. The transgression continued with the tidal Tonoloway Formation. This very interesting formation, at once full of ostracode fossils, salt casts, and mud cracks, is interpreted as representing a sabkha-like environment. It was a time of relative aridity with intense evaporation. While the salt casts of the Tonoloway Formation were being generated in what is now the Virginias, major salt deposits currently residing under Lake Erie were also forming.

As this new carbonate bank developed, it was deepened as sediment supply from the land diminished and accommodation space increased. This led to the formation of the very fossiliferous Helderberg Group limestones of the late Silurian and early Devonian. Having arrived now at HST, the waters of basin were as deep as they were going to get. As the sequence matured, a new moutain-building event began its onset, the Acadian Orogeny. The first clean quartz arenites show up in later Helderberg deposits, marking the coming of another terrestrial highland, called “Avalon”. These arenites mark the beginning of the prograding deposition so indicative of HST periods. The Acadian Orogeny got its start with the deposition of the Oriskany Sandstone, a fossiliferous and *Skolithos* sp.-rich sandy shore facies.

This sequence again ends with the formation of a sequence boundary as FST once again begins due to the sedimentation and then tectonic changes that lead to the formation of the new sequence boundary, the Wallbridge Unconformity. After a time, these Acadian mountains eroded to a plain. Following some amount of time and the Wallbridge sequence boundary, the Kaskaskia supersequence progressed.



4th Order and Higher Cycles

Within these two 3rd order sequence that have been described above, there are certainly 4th and perhaps 5th order parasequence sets and parasquences to be found. Identifying such short-term packages of sediments would add much to the story of the Tippecanoe. Any in-depth analysis of that level of detail goes beyond the scope of this case study, however.

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24.3: Final Thoughts

As illustrated in the 2nd order Tippecanoe Sequence and the two 3rd order cycles contained within it, sequence stratigraphy can provide a model for how to interpret the changes we see in facies. These changes, resulting from the complex interplay between tectonics, eustacy, sedimentation, and relative sea level make up the story of the strata. The application of these models to the this stratigraphic story brings a richness to the history of a basin that is less apparent than it would be otherwise. The tools of sequence stratigraphy, on other words, are necessary for hearing what the rocks are telling us about their story!

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CHAPTER OVERVIEW

25: (Case Study) The Taconian Orogeny

25.1: Introduction

25.2: Tectonic context

25.3: The roots of the mountains

25.4: The sedimentary signature

25.5: Conclusion

25.6: *What's in a name?

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25.1: Introduction

In the late Ordovician period of geologic time, subduction pulled a volcanic island arc and ancestral North America (Laurentia) together. The result was an extensive mountain belt that produced a large amount of eroded sediments. Both the resulting sedimentary rocks and the metamorphic and igneous rocks that formed at the roots of the mountain belt are used by geoscientists today to examine this episode of mountain-building, known as the Taconian ("Taconic"*) Orogeny.

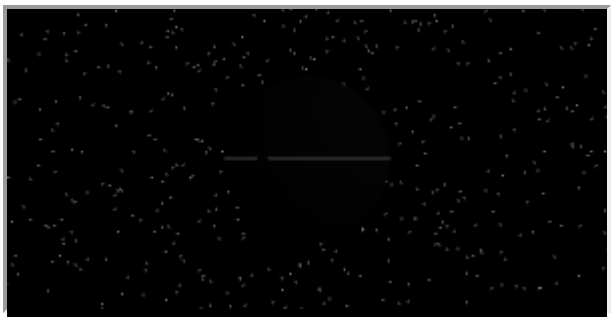


Figure 25.1.1: Earth as it appeared in the late Ordovician period of geologic time. Find ancestral North America ("Laurentia") and note the mountain belt along its southern edge. This is the Taconian Mountain Belt. This dynamic visualization was created by Ian Webster using tectonic and paleogeographic maps by C.R. Scotese's PALEOMAP Project, and is embedded here with permission. Grab it and spin it around! Zoom in and out! Explore!

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25.2: Tectonic context



Figure 25.2.1: The Port Deposit Tonalite of Cecil County, Maryland, is a classic Taconian rock. It has a magmatic crystallization age of 515 Ma (U/Pb in zircon) and a metamorphic age of 490-480 Ma (Rb/Sr in biotite). It formed offshore, in a magma chamber beneath one of the volcanoes of the Taconian volcanic island arc, and was metamorphosed when that arc collided with ancestral North America during the Taconian Orogeny.

The cause of the Taconian Orogeny was a collision between two tectonic plates: the ancestral North American plate's continental leading edge, and another plate of oceanic affinity, now deceased. The oceanic plate was one of the plates that floored the Iapetus Ocean, and as it moved toward the ancestral North American plate, the oceanic lithosphere that was part of the North American plate subducted, down and under the overriding plate of oceanic lithosphere. This resulted in a volcanic island arc, out in the middle of the Iapetus Ocean.

Part of the context of the orogeny is therefore on the ancestral North American continent, and part is out in the volcanic island arc. Rocks that formed in that island arc journeyed toward ancestral North America and accreted to the continent during the orogeny. Isotopic ages reflect this two part history: an initial crystallization from magma in the arc, and a later metamorphic age from the orogeny. The Port Deposit Tonalite, a metamorphosed granitoid, provides a nice example. Prior to the Taconian Orogeny, it was not yet metamorphosed: just a granitoid, under a volcano, moving along at a few cm per year, getting closer and closer to the Laurentian continental slope.

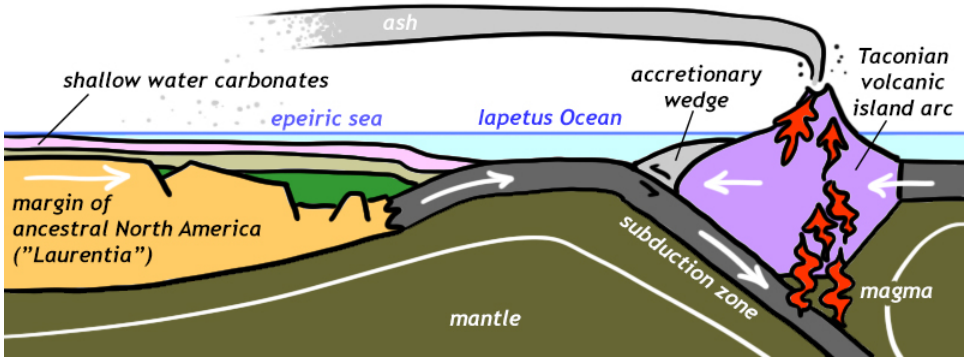


Figure 25.2.2: The tectonic situation that would lead to the Taconian Orogeny: subduction of the oceanic margin of the ancestral North American plate resulted in a volcanic island arc that drew ever closer, building up an accretionary wedge of Iapetan ocean floor and deepwater sediments.



Figure 25.2.3: Ooids from the pre-Taconian Cambrian-aged Conococheague Formation, Shenandoah County, Virginia.

Prior to the orogeny, the edge of ancestral North America was a passive margin setting: it was the edge of the continent, but not the edge of the plate. Through the Cambrian and well into the Ordovician, there was no tectonic activity anywhere nearby, and there had not been for a very long time. Submerged under an epeiric sea, it was the site of limestone and dolostone deposition in a Bahama-like carbonate bank setting. Primary sedimentary structures such as ooids and stromatolites testify to shallow water depths. The proportion of clastic detritus such as clay and silt was quite low. There are abundant fossiliferous limestones from this time full of brachiopods, bryzoans and other common Paleozoic filter feeders, indicating clean water: a lack of excess runoff and sedimentation.

But not for long...

Did I Get It? - Quiz

? Exercise 25.2.1

What was the cause of the Taconian Orogeny?

a. The accretion of a volcanic island arc to the edge of ancestral North America

b. The collision between India and Eurasia

c. Hotspot volcanism

Answer

a. The accretion of a volcanic island arc to the edge of ancestral North America

? Exercise 25.2.2

What was happening in eastern North America immediately prior to the Taconian Orogeny?

a. Passive margin sedimentation of carbonate strata

b. Erosion of lots of exposed basalt

c. Rifting related volcanism and sedimentation

Answer

a. Passive margin sedimentation of carbonate strata

Exercise 25.2.3

What was happening in the Taconian volcanic arc immediately prior to the Taconian Orogeny?

a. Rifting related volcanism and sedimentation

b. Passive margin sedimentation of carbonate strata

c. Subduction, arc volcanism, and build-up of an accretionary wedge

Answer

c. Subduction, arc volcanism, and build-up of an accretionary wedge

---

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25.3: The roots of the mountains

Since the Taconian mountain range itself is long gone, we can consider the orogeny from two different perspectives: (1) that of the eroded roots of the mountains, and (2) that of the sedimentary basins “next door,” which received sloughed-off sediment eroded from the mountains.

Let’s first examine the roots of the mountains, which can be found in the Piedmont geologic province of Virginia, Maryland, Washington, D.C., Pennsylvania, New Jersey, and New York, as well as assorted provinces in New England [TKTKTKTK]. The rocks in question are the crushed, cooked remains of the Iapetus Ocean and the Taconian volcanic island arc.



Figure 25.3.1: Relict graded bed in Mather Gorge Formation metagraywacke, near Potomac, Maryland.

The rocks of the Piedmont have been metamorphosed to various degrees, from greenschist facies all the way up to partial melting. Their protoliths range from basalt and gabbro (oceanic crust) to mudstone, graywacke, and limestone (oceanic sediments), as well as the volcanic rocks of the volcanic island arc (both intrusive and extrusive, both mafic and felsic). In some cases, the subsequent metamorphic recrystallization had a light enough touch that primary structures are still preserved, both volcanic and sedimentary. Graded beds in meta-graywacke of the Mather Gorge Formation are a nice example of a primary sedimentary structure that speaks specifically of oceanic processes. These graded beds formed from deep submarine deposition of clastic sediment by turbidity currents in the Iapetus Ocean.

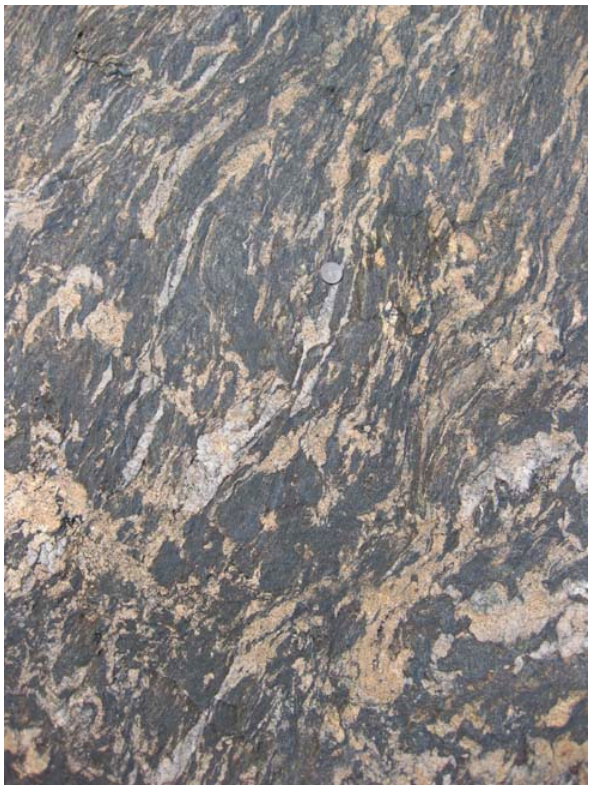


Figure 25.3.2: Migmatite exposed in Chesapeake & Ohio Canal National Historical Park, near Potomac, Maryland.

We can estimate the timing of the Taconian Orogeny by looking at metamorphic ages for these rocks (K/Ar, Ar/Ar, and Rb/Sr methods), as well as crystallization ages for the migmatites produced by partial melting (U/Pb). In both cases, the answer returned is ~460 Ma, a late Ordovician age. The Piedmont is also home to many plutons of felsic igneous rock like the Occoquan Granite, the Georgetown Intrusive Suite, and the Kensington Tonalite, and these all also yield isotopic ages in the 474 to 450 Ma range.

Explore this gigapixel panorama of a sample of migmatite from Orange County, Virginia, and search for pockets of speckled granite. These “leucosomes” represent the formerly-molten portion of this rock, which is otherwise a schist. Similar partial melting is occurring today beneath active modern mountain belts such as the Himalayas.


GIGAmacro™ Migmatite



Figure 25.3.3: Folded metamorphosed turbidites: alternating schist & metagraywacke layers (former shale & graywacke) were folded by Taconian mountain-building. Outcrop in Chesapeake & Ohio Canal National Historical Park, near Potomac, Maryland.

Deformation was another major signature of mountain-building in the Piedmont region. Primary structures were distorted by folds and disrupted by faults as the Taconian volcanic island arc docked with ancestral North America, compressing the Iapetan sediments caught in between.

In New England, Iapetus seafloor rocks including both oceanic lithosphere and overlying deepwater sedimentary deposits moved upward onto continental rocks, and westward for close to 5o km of distance. A major thrust fault allowed this relative motion. Today the trace of this fault is called "Cameron's line," after the geologist who first described it. Isolated klippen of the overthrust rocks remain in the eponymous region of the Taconic Mountains, but the trace of the fault also runs through New England and even through downtown New York City.

Did I Get It? - Quiz

? Exercise 25.3.1

Isotopic dating of metamorphic minerals and igneous plutons indicate that the Taconian Orogeny happened when?

a. ~200 Ma

b. ~1.2 Ga

c. ~460 Ma

d. ~840 Ma

e. ~500 Ma

f. ~330 Ma

Answer

c. ~460 Ma

Exercise \(\PageIndex{2}\)

Pre-Taconian rocks that were heated up sufficiently to partially melt made _____.

a. limestone strata, including the microbial mat fossils called stromatolites

b. mafic lava flows called basalts, featuring cooling columns

c. granitic magma, which is preserved in small pockets in the rocks called migmatites

Answer

c. granitic magma, which is preserved in small pockets in the rocks called migmatites

Exercise 25.3.3

A signature of compressional forces that operated in the Taconian mountain belt is: _____.

a. graded bedding

b. immature sedimentary deposits

c. granite

d. folding

Answer

d. folding

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25.4: The sedimentary signature

When mountains are raised, they erode. The erosion of the mountains produces clastic sediment, and a lot of it. While this does not accumulate on-site (*i.e.*, atop the mountain belt), neighboring sedimentary basins may be sufficiently low-lying that they can receive and preserve this sediment through geologic time. Long before geologists understood the thermal or tectonic origins of metamorphic rocks and granites, orogenies were known from their clastic sedimentary signature. Gravel, sand, and mud don't just magically spring into existence, after all – they require a source. A large amount of clastic sediment showing up in a stratigraphic sequence implies that there must have been a lot of nearby mountainous rock to be eroded.

The resulting clastic sediment (my students like to call it "mountain dandruff") comes in two essential varieties: a deep marine turbidite package that Alpine geologists call "flysch," and a terrestrial red-bed package dubbed "molasse." Though these European terms are a bit out of vogue in modern America, they very nicely summarize the sedimentary signature of the Taconian Orogeny. We find both Taconian flysch and Taconian molasse in the stratigraphic sequence of sedimentary rocks in the Valley & Ridge province.

Taconian Flysch

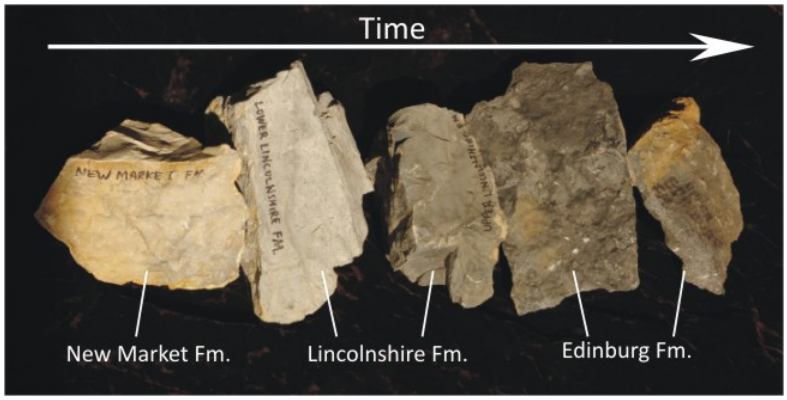


Figure 25.4.1: Over the course of the late Ordovician, light-colored shallow limestones gave way to increasingly dark deep-water limestones and shales.

The pre-orogeny limestones get dirtier and dirtier as the end of the Ordovician approaches. Their increased clay and silt content is seen as the first indication of the clastic onslaught to come, like a whiff of smoke before a forest fire. Over time, going up through the stratigraphic sequence, these passive margin carbonates give way to limy shales and then clastic shales with no calcite, and finally to graywacke turbidites interbedded with shale. The interpretation for this "dirtying upward" pattern is the increasing proximity and prominence of the Taconian mountain range, shedding more and more sediment the larger it grew. This flysch in the marine sedimentary record of Taconian mountain-building (and erosion).

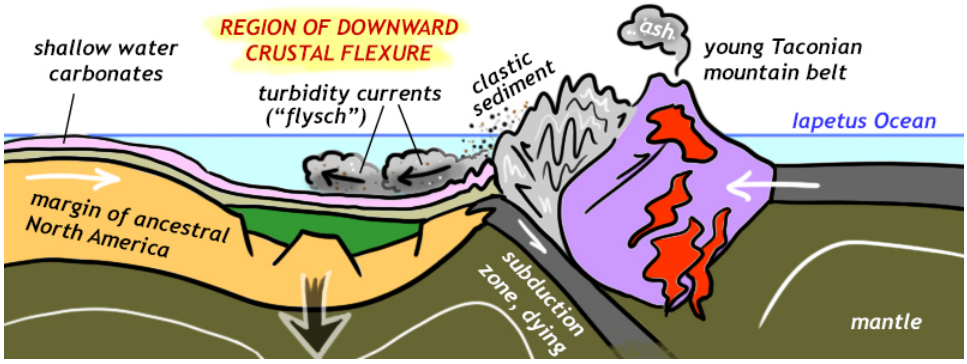


Figure 25.4.2: The deepening of the sedimentary basin adjacent to the young Taconian mountain belt was accomplished as the edge of ancestral North American continent flexed downward. Turbidity currents flowed into this deepened basin, depositing shale and graywacke: the Taconian "flysch."

The record of these turbidity currents is a series of graded beds in graywacke, separated by layers of shale. These Bouma sequences are distinctive deep sea sedimentary sequences that speak about submarine avalanche after submarine avalanche, delivering huge

quantities of sand and mud into the oceanic deep:



Here is a hand sample of rock showing a Bouma sequence:



The transition from pre-Taconian shallow-water carbonates to during-the-Taconian deepwater turbidites suggests that the water got deeper. There may have been a role for crustal flexure here: where the tectonic loading of the Taconian arc and its accretionary wedge onto the edge of ancestral North America caused the crust to sag downward under this extra weight, deepening the sedimentary basin next door.

In the Mid-Atlantic region of the Valley & Ridge province, the major geologic unit showing full-on flysch is the Martinsburg Formation. Fossils in the Martinsburg Formation allow us to constrain the timing of the mountain building from both biostratigraphic and paleoecological points of view. As the Ordovician limestone platform sediments get dirtier and more clay rich, shallow water filter feeders are replaced with species that are better suited for muddier and deeper conditions. Here are two examples showing deeper water faunas, one showing graptolites and one showing brachiopods, crinoids, and a nautiloid; both shown as gigapixel panoramas:





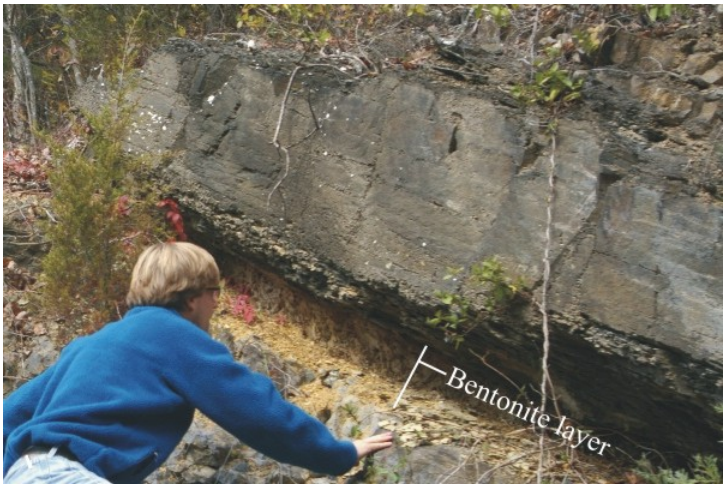


Figure 25.4.3: Late Ordovician bentonite layer between limestone layers in the Valley & Ridge province of northern Virginia.

Layers of ash are preserved too, presumably sourced to the approaching volcanic island arc. These ash layers weather today to a yellowish, crumbly clay material called bentonite, but they include zircons that can be dated, and that helps constrain the age of the sedimentary strata above and below the bentonites. Two widespread bentonite beds, named the Deicke bentonite (457 Ma) and the Millbrig bentonite (454 Ma), are found in a vast swath of Appalachia and the Midwest. They can be correlated all the way from southern Minnesota and Texas to Alabama and Georgia to upstate New York.

Taconian Molasse

Once the flysch basin filled up, rivers draining the Taconian mountain belt stretched out across the flysch, reaching westward toward the Tippecanoe epeiric sea. As they flowed, they transported sediment. The sediment built up in river channels and flood plain deposits. In the Mid-Atlantic region, these occur mainly in the Juniata Formation.

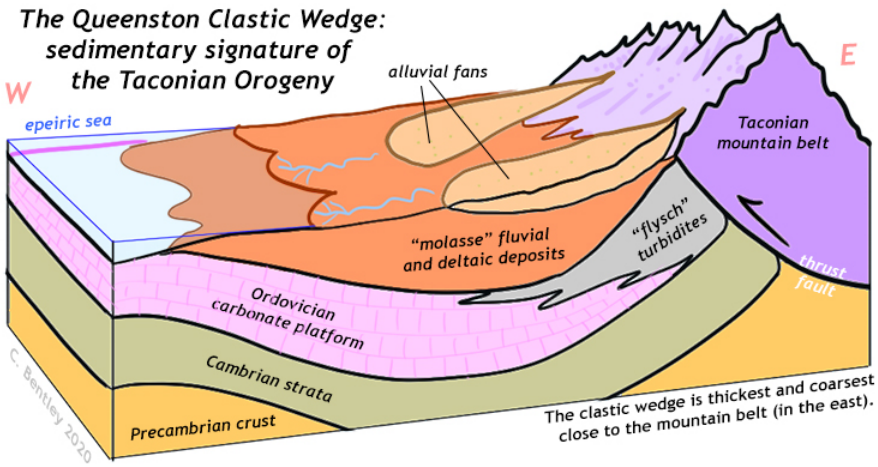


Figure 25.4.4: The Queenston Clastic Wedge was deposited west of the Taconian mountain belt. The molasse is thickest and coarsest close to the mountain belt to the east, and thins and fines to the west.

Here's a Google Maps Street View of one such exposure:

Note the sandstone-filled channel edge poking up from the grass at the right side of the screen, like half a smiley face. There are half a dozen red sandstone/shale layers to its left. A bit further left still, you can see just red shale (no sandstone). This is a small snapshot of the relationship between a river and its floodplain. The river is the channel sandstone with the smiley-face shape, and the red shale represents its floodplain. The transitional zone with the many small sandstone/shale couplets are interpreted as crevasse splay deposits, places where the river overflowed its banks in a flood, and spilled over its own natural levee.

The Juniata Formation is part of a more massive arcuate deposit of terrestrial deposits, called the Queenston Clastic Wedge. Some geologists dub it the “Queenston Delta,” though that’s probably not literally accurate. It was probably more like an alluvial plain fed by many rivers draining the Taconian mountain belt. In map view, it has a big fan-like shape, but in cross section, the name “wedge” makes more sense: it’s thickest (and coarsest) in the east, and then thins systematically to the west, pinching out to a feather edge in Michigan.

The Queenston Clastic Wedge is reckoned to be about half the sediment that was shed off the Taconian Mountains (with the other half having gone off east of the mountain belt, into the Iapetus). If this is correct, an estimate of the volume of the mountains can be made: 600,000 cubic km of rock. Since we know the width of the metamorphic belt (the “roots” of the mountains, as outlined in the previous section), this allows up to convert our volume estimate into an interpretation of height. As with the estimates from metamorphic pressures, this calculation suggests Taconian peaks on the order of 4000 m high.

After the Taconian Mountains had been worn down, in the Silurian, conditions returned to passive margin sedimentation, and a new layer of carbonate was laid down in the Silurian and into the Devonian. This was a temporary reprieve from active margin conditions, which would resume with the Acadian Orogeny in the middle to late Devonian.

Did I Get It? - Quiz

? Exercise 25.4.1

Which package of sediments is a general term for deposits of deep marine clastic sediment such as black shale and graywacke from turbidity currents?

a. evaporite

b. flysch

c. molasse

d. migmatite

e. carbonate

Answer

b. flysch

? Exercise 25.4.2

With their fluvial sandstones and redbed shales, the strata of the Queenston clastic wedge are _____.

a. Acadian molasse

b. Acadian flysch

c. Taconian molasse

d. Taconian flysch

Answer

c. Taconian molasse

? Exercise 25.4.3

Crustal flexure of the ancestral North American margin resulted in which of these?

a. Uplift of the sedimentary basin immediately adjacent to the Taconian mountain belt.

b. Deepening of the sedimentary basin immediately adjacent to the Taconian mountain belt.

Answer

b. Deepening of the sedimentary basin immediately adjacent to the Taconian mountain belt.

For a detailed look at the sediments shed off the Taconian mountain belt, see the Massanutten Synclinorium VFE.

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25.5: Conclusion

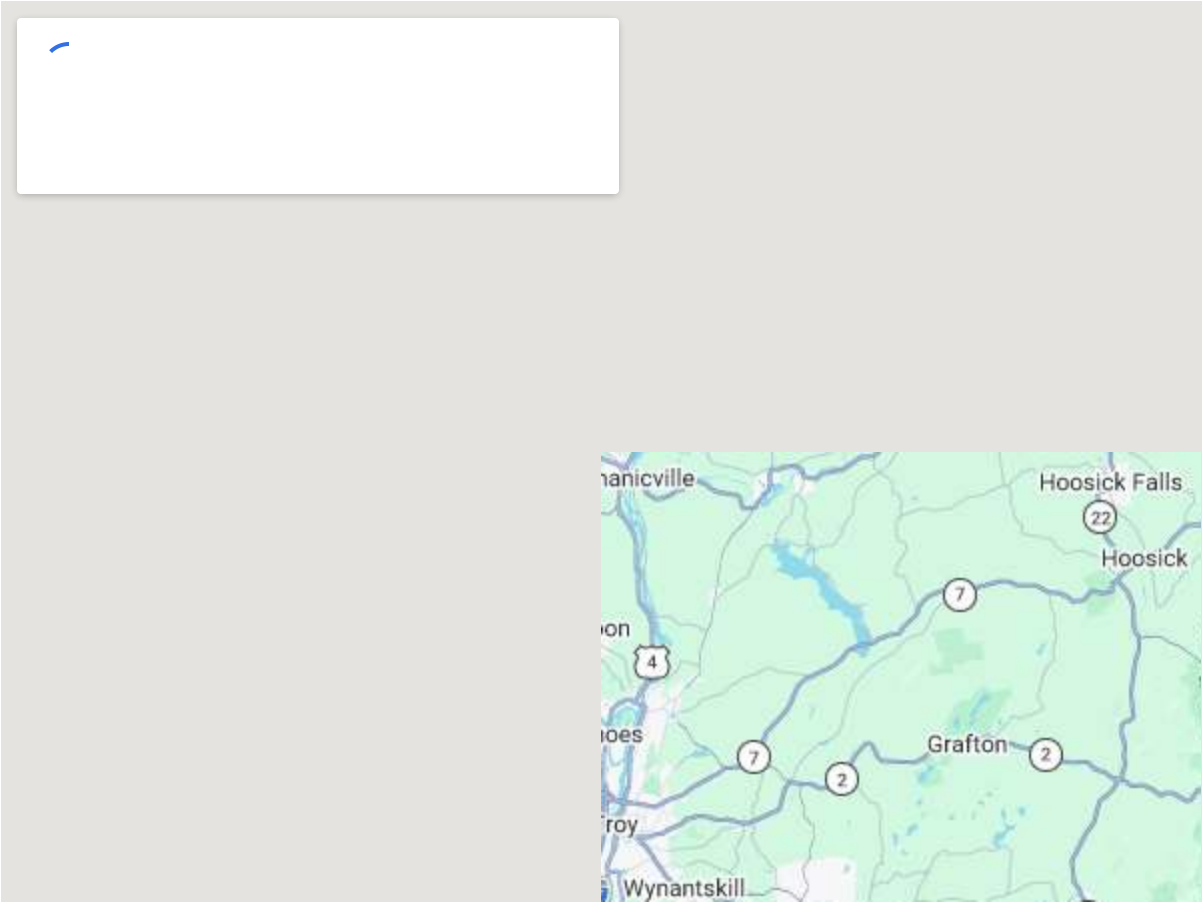
Multiple lines of evidence point to the accretion of a volcanic island arc with (what is today) eastern ancestral North America during the late Ordovician period of geologic time (about 460 Ma in the Mid-Atlantic region). This tectonic collision, called the Taconian Orogeny, resulted in an extensive metamorphic belt marking the roots of a mountain chain thousands of km long. As they were worn down by erosion, these ancient mountains shed copious quantities of sediment, which accumulated in neighboring low-lying basins (both marine and terrestrial).

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25.6: *What's in a name?

The Taconian Orogeny is also called “the Taconic Orogeny” by enough geologists that it’s probably worth exploring the different names here. The authors of this text believe that “Tacon**ian**” is the better term, and that “Tacon**ic**” is misleading. Let us briefly explain why...

The Taconic Mountains are a small, modern mountain range in upstate New York are located east of Albany, on the border with Massachusetts, close to the southwestern corner of Vermont:



Foundational work on understanding Ordovician mountain-building was first completed in these (modern) mountains, and thus the local landmarks provided the name for the orogenic episode. However — and this is the key point — the entire mountain chain from the Ordovician was **not** limited to the area of the modern Taconic Mountains. Instead, the ancient mountains extended from Newfoundland in eastern Canada all the way down to Alabama.

Not only were the ancient Taconian Mountains much longer as a range than the modern Taconic Mountains, they were taller, too. The highest peak of the Taconic Range today is only about 600 meters tall. In contrast, estimates from metamorphic minerals formed during the Taconian Orogeny suggest that the peaks of the Ordovician-aged Taconian Range must have been much taller. Peak metamorphic pressures of 1.5 GPa imply something on the order of 20 km of overlying crustal material. The Taconian Mountains, in other words, would have been an Alpine-scale range. In the modern Alps, the highest peak is more than 4000 meters tall.

Those ancient mountains are gone now, eroded away over geologic time. We can observe their eroded roots, and we can observe the sediment that resulted from that erosion, but the mountains themselves as topographic features are long gone. That ancient range deserves its own name, and that name should be distinct from the name that’s applied to the modern range. If the modern range is the Tacon***ic*** Mountains, then the Ordovician mountains need a different name: Tacon***ian***.

Let’s summarize with a quick comparison and contrast: The **Taconian** Mountains were an ancient landscape feature, thousands of kilometers long, with peaks that likely once exceeded 4000 m elevation, and are now completely eroded away. The **Taconic** mountains are a modern landscape feature, about 20 kilometers long, with maximum elevation of merely 400 m, and the mountains are not yet completely eroded away.

The Taconic Mountains are where the Taconian Orogeny was first described, but we should not confuse the piddly modern mountain range with its mighty Ordovician-aged predecessor. Vastly different in age, height, and extent, they deserve different names.

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CHAPTER OVERVIEW

26: (Case Study) California’s Coast Ranges - a Mesozoic accretionary wedge complex

26.1: Modern physiography suggests an ancient subduction zone

26.2: Rocks and structures associated with the subduction of oceanic crust

26.3: The hard rocks - mantle and crust

26.4: The sedimentary journey

26.5: Deformation of these rocks into geological structures

26.6: The end of subduction

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26.1: Modern physiography suggests an ancient subduction zone

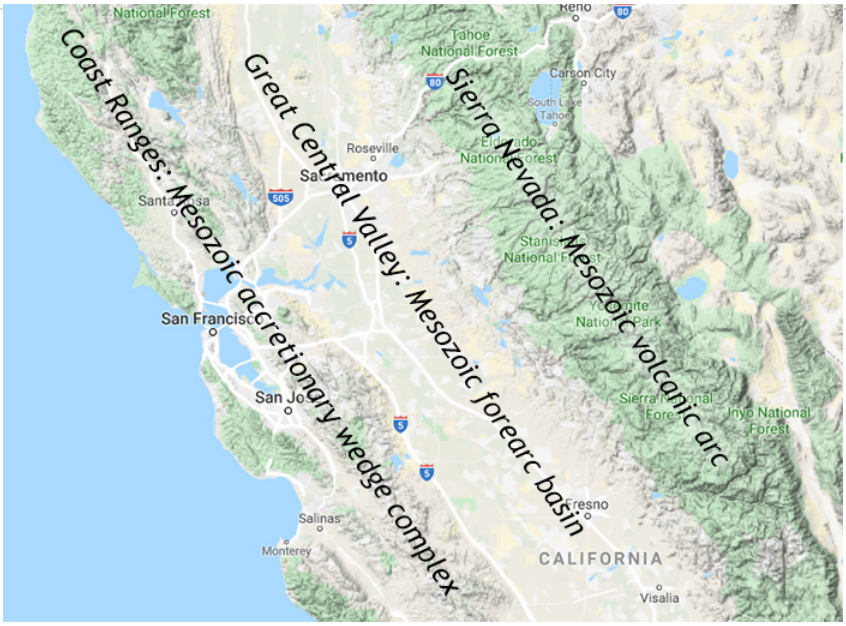


Figure 26.1.1: North-central California shows a three-part physiography that corresponds to the Mesozoic origins of this part of North America as a subduction zone at a convergent plate boundary.

At the latitude of San Francisco, California has three main physiographic provinces. From the Pacific coast on the west to more inland positions in the east, these are: the Coast Ranges, the Great Central Valley, and the Sierra Nevada. In spite of their subsequent geologic histories, these are all remnants of California's former history as a convergent margin.

Today, the region is famous as a transform boundary, where the Pacific Plate and the North American Plate slide laterally past one another along the San Andreas Fault.

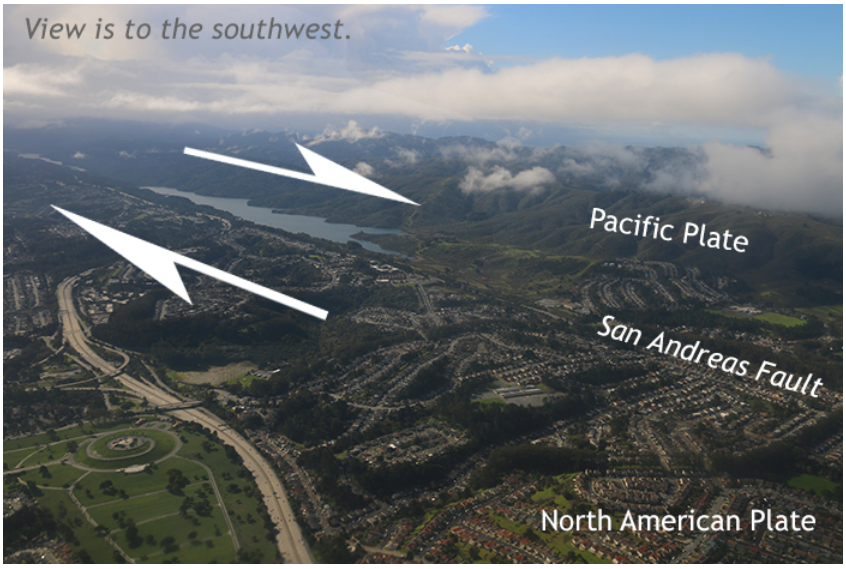


Figure 26.1.2: The modern plate boundary in the San Francisco region is a right-lateral transform plate boundary, seen here south of San Francisco, in an oblique view along the trace of the San Andreas Fault.

But the *modern* plate boundary in Califronia is not the same as the *ancient* plate boundary. During the Mesozoic, there was a third plate offshore: the Farallon Plate, which was in between the Pacific and the North American Plates. The relative motion between the Farallon and North America was quite different. Back then, coastal California was a convergent plate boundary, and the Farallon was subducting beneath western North America.

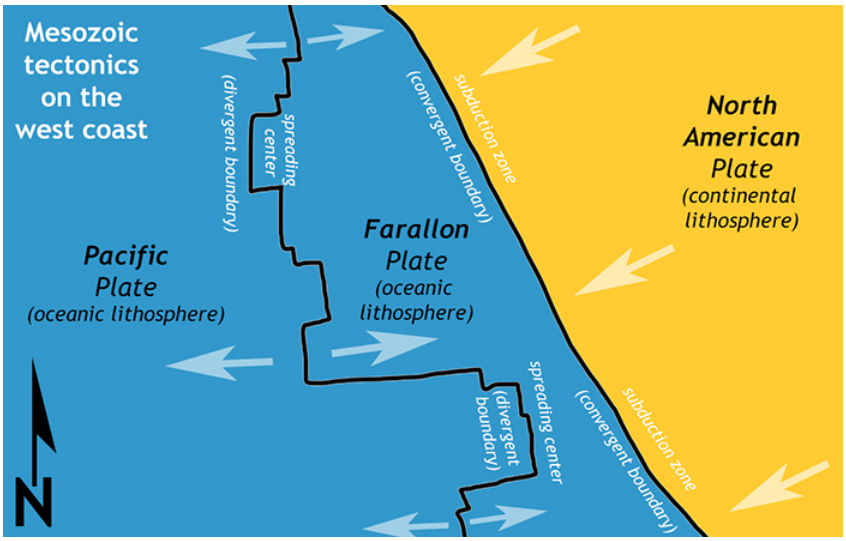


Figure 26.1.3: The tectonic setting of the west coast of North America in the Mesozoic was quite different than today, as shown in this interpretive map view. There was another plate present then, of oceanic lithosphere, between the Pacific Plate (also oceanic) and the North American Plate (continental lithosphere).

The boundary between the Pacific Plate and the Farallon Plate was an oceanic ridge, the site of seafloor spreading. In other words, it was a divergent plate boundary between two plates of oceanic lithosphere. The newly-minted Farallon traveled a short distance eastward, and then subducted beneath continental lithosphere of the North American Plate, producing key features of a convergent plate boundary: the accretionary wedge (focus of this case study) but also partial melting of the overlying mantle wedge. This melting was triggered by the release of water from the subducted plate. The resulting magma rose to produce a continental magmatic (and volcanic) arc. Weathering and erosion of the arc's mountains produced sediment that was transported downhill, and accumulated in both forearc and back-arc basins.

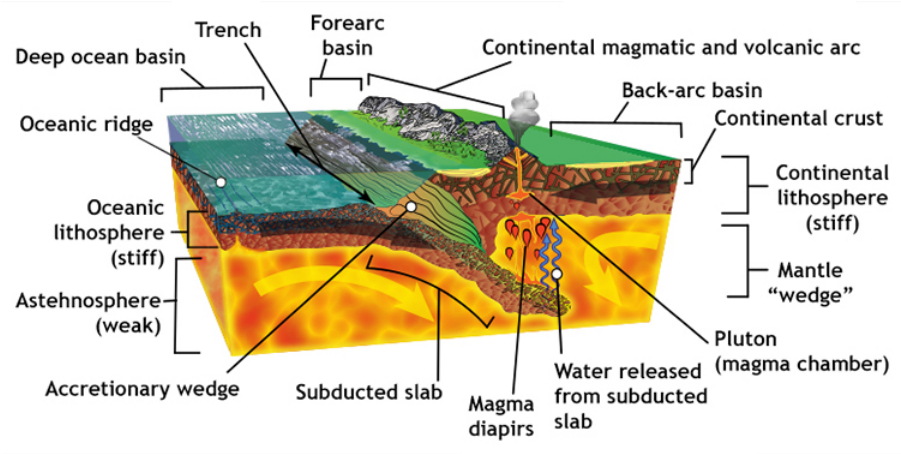


Figure 26.1.4: A diagram showing key features in a Mesozoic-California-style subduction zone. (Modified from an original by KDS4444.)

The rocks of the Coast Ranges preserve a record of their past. Let's take a look at both rock types and structures. To make the story somewhat simpler, geologists divide the rocks of the region up into packages of rocks that have seen relatively similar stories of formation, deformation, metamorphism, and exhumation. We call these packages terranes. On the peninsula of San Francisco itself, east of the modern plate boundary (*i.e., the San Andreas Fault*), there are three distinct terranes, separated by two map-scale bodies of sheared-out rocks called mélange:

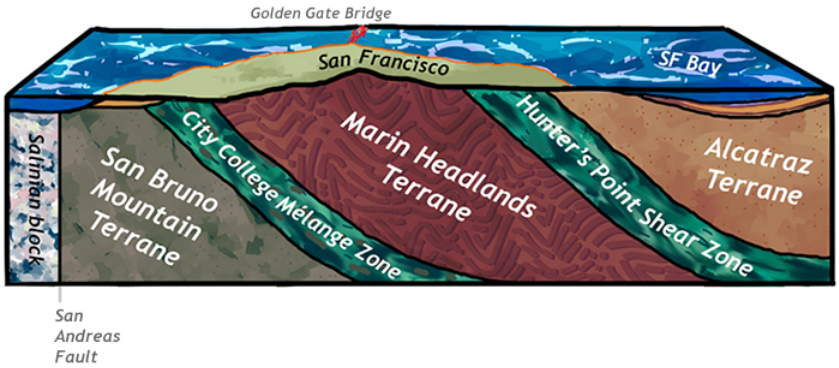


Figure 26.1.5: A west-east cross-section of the rocks beneath San Francisco, California: east of the San Andreas Fault are a series of east-dipping terranes, each showing a distinct rock suites and metamorphic and deformational history. Zones of sheared-out rock (called mélange) separate the terranes.

Note the difference in orientation of the terrane contacts (each of them a manifestation of one instance of the plate boundary between North America and the Farallon Plate) and the modern plate boundary. In the Mesozoic, the plate boundary was a subduction zone dipping to the east under North America, and that shape is preserved as powerful evidence of the fundamental geometry of that boundary. The modern transform plate boundary, in contrast, is vertical.

Further to the north and east, there are more terranes still. Each terrane is a sliver-shaped package of rocks that was subducted (some rather shallowly, some quite deeply) and then transferred from the subducting Farallon Plate to the overriding North American plate.

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26.2: Rocks and structures associated with the subduction of oceanic crust

The original seafloor rock types that were fed into the subduction zone included:

- Mantle peridotite
- Oceanic crustal rocks: Basalt and gabbro
- Deep sea sediments: chert + shale
- Turbidites: shale + graywacke
- Continental margin sediments: Sandstone + conglomerate

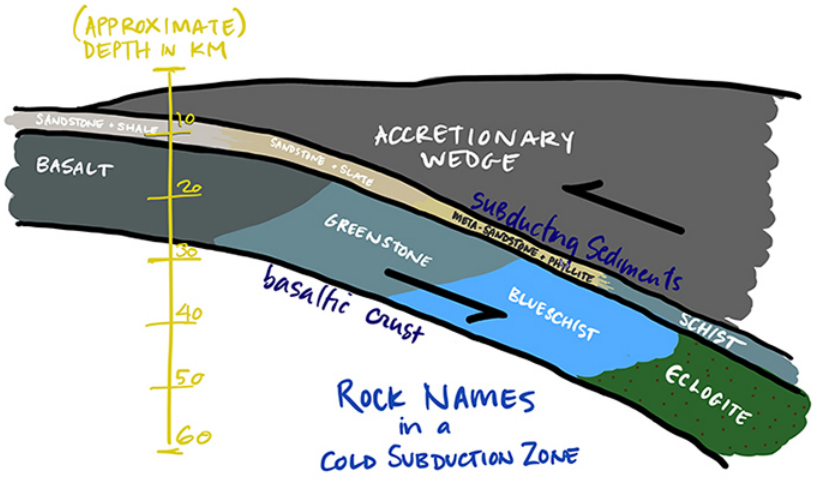


Figure 26.2.1: A general diagram showing the position of various meta-sedimentary and meta-volcanic rocks within a classic "cold" subduction zone relative to the accretionary wedge that forms on the leading edge of the overriding plate.

Once subducted, these rocks metamorphosed to various degrees, making new rocks:

- Serpentinite
- Greenstone (green*schist* where foliated)
- Blueschist
- Eclogite
- Metachert
- Phyllite
- Schist
- Metaconglomerate

Besides metamorphosing, being subducted wasn't easy on these rocks. They accrued several kinds of deformation as a result of the tectonic stresses they experienced:

- Chevron folds
- Thrust faults
- Mélange

Let's examine the metamorphic transformations in turn: first the oceanic lithosphere itself, and then the sedimentary cover. We'll look at the protolith, and then at the metamorphic rock (or rocks) it can become. Finally, we'll examine the deformational structures that record evidence of the stresses of subduction.

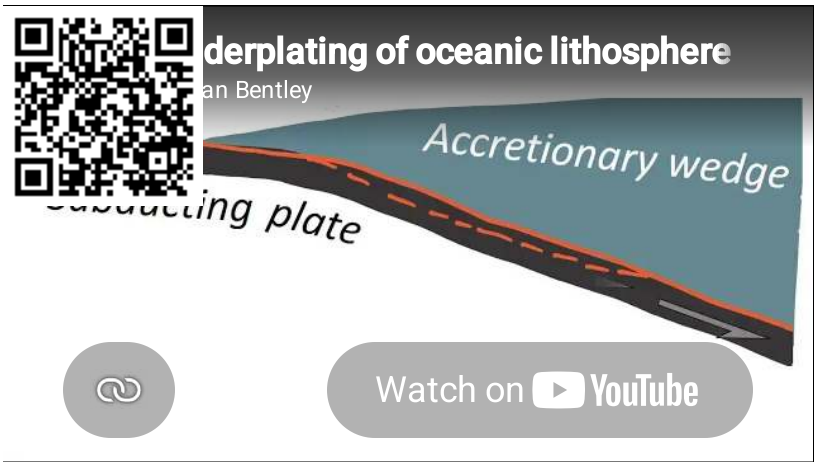
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26.3: The hard rocks - mantle and crust

MANTLE: Peridotite to serpentinite

The upper mantle is made of a rock called peridotite. Peridotite is dominated by the mineral olivine, though some peridotites also have plenty of pyroxene and plagioclase in them. Peridotite formed long ago when the early Earth cooled off and solidified in its uppermost reaches. Thought it is solid, the mantle still flows slowly. Warm mantle peridotite rises due to its low density, while cold peridotite sinks due to its higher density. This relative motion sets up convection related "cells" in the mantle that are thought to be one of the important driving forces that power the surficial movement of lithospheric plates, including the manifestation of convergent motion that we call "subduction."

Not all of the lithospheric mantle necessarily gets recycled during subduction. The process of "underplating" can cause slivers of the subducting plate (including lithospheric mantle) to be ripped off and added to the underside of the overriding plate. This 'subduction accretion' builds up the accretionary wedge from below, and is the major source of mantle rock introduced into the subduction complex.



When the peridotite is broken off and mixed into the accretionary wedge, it can come into contact with water, and metamorphose. This creates serpentinite, a metamorphic rock dominated by the mineral serpentine. Serpentine is a hydrated magnesium silicate. Serpentinite is the California 'state rock,' though because some of the minerals in serpentinite (such as chrysotile) have an asbestiform habit, it makes lawyers itchy, and there are periodic proposals to replace its official 'state rock' legal status.



Figure 26.3.1: Pillow lavas are a signature of underwater eruption. Here, relatively lightly-metamorphosed (greenstone) meta-basalts are exposed on Black Sands Beach in Marin Headlands.

OCEANIC CRUST: Basalt & its three metamorphic descendants

When the hot mantle is decompressed, it is capable of partial melting, which derives a mafic magma form the ultramafic source rock. Estimates are that mafic magmas derive from somewhere between 1% and 25% partial melting of source peridotite.

The resulting magma could cool at depth to generate a gabbro, or could be erupted on the seafloor to make basalt. When lava erupts at the bottom of the ocean, the high heat capacity of the surrounding seawater wicks heat away efficiently. This chills the edge of the lava instantly, forming a solid crust resembling a loaf of bread.

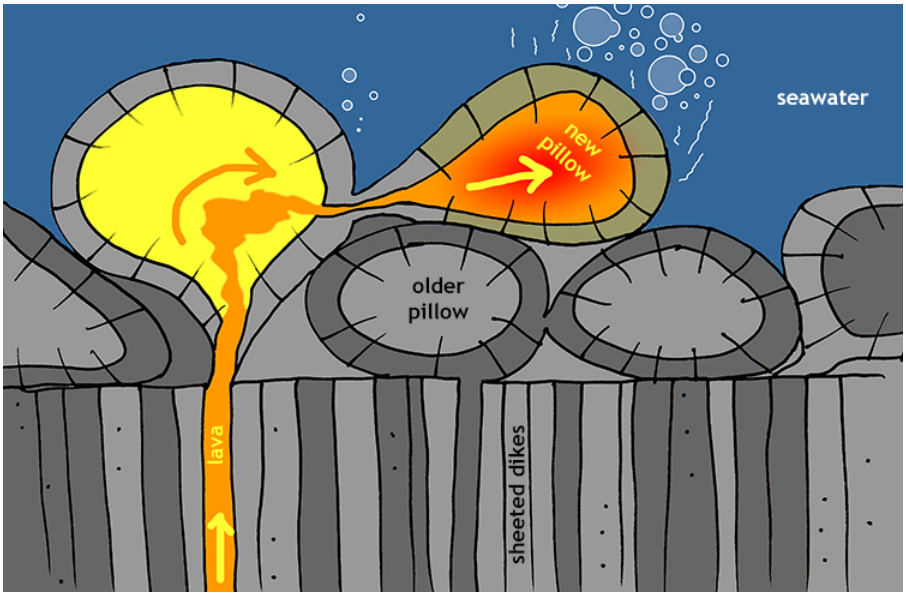
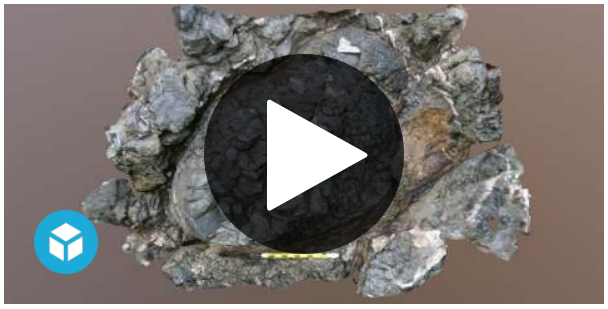


Figure 26.3.2: The formation of pillow structures occurs when basaltic lava erupts underwater.

The interior of the lava “loaf” is still molten, and if it is under sufficiently high fluid pressure, it will crack open the crust and squirt another lobe of fluid lava out into the water, repeating the process. Played out time and time again over geological time, this results in a pile of “pillow” shaped features. These “pillow lavas” are a therefore a primary volcanic structure; They are the signatures of seafloor eruptions of new oceanic crust.



When basalts are subducted, they serve as particularly good recorders of the pressure and temperature they encounter. If they are shallowly subducted, they will make greenstone. If they are subducted a bit deeper still (with high pressure), they will transform into blueschist, and if they are subducted really deep (with the highest pressure), eclogite will form.

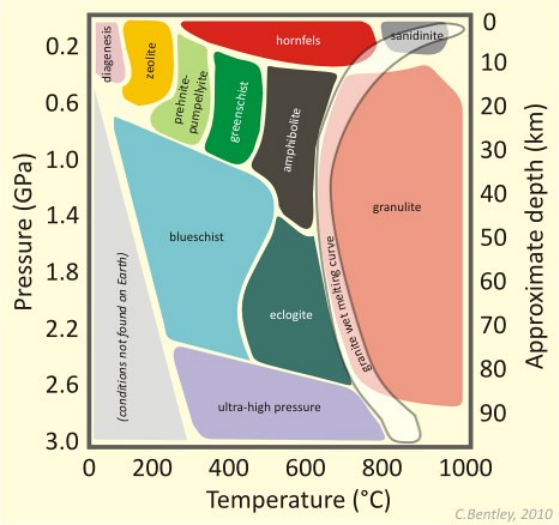


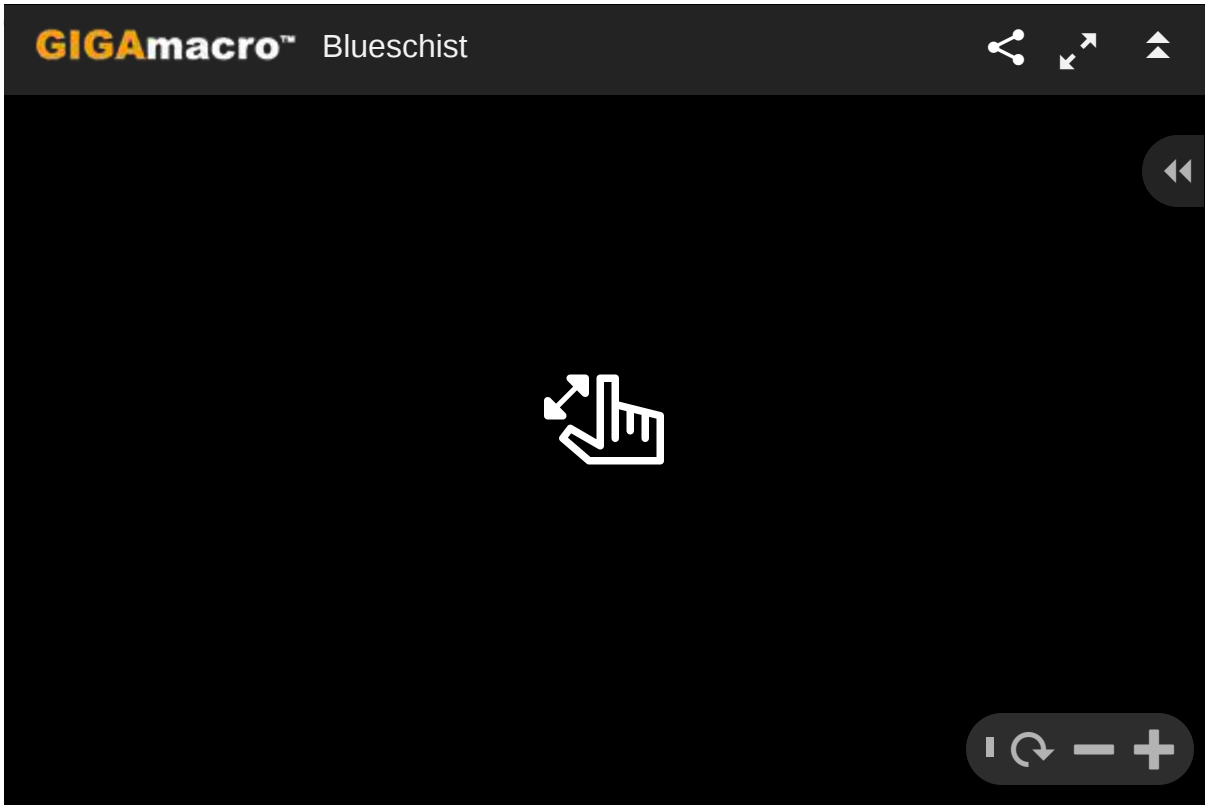
Figure 26.3.3: A plot showing the various metamorphic facies that result from different combinations of temperature and pressure. Note that blueschist and eclogite result from relatively low-temperature but relatively high-pressure metamorphism: conditions we would expect in a subduction zone.



Figure 26.3.4: Greenstone-facies meta-basalt, showing the preserved forms of lava pillows, Jenner, California.

Greenstone is the lowest grade of subduction related metamorphism. Very few of the basalts seen today in the Franciscan complex are still basalt. Many/most have been metamorphosed to make greenstone. When the minerals that make up basalt are subjected to depths around 20 to ~35 km in a subduction zone, the pressure and temperature are sufficient to make the plagioclase and pyroxene react and generate the minerals chlorite and actinolite and sometimes (if water is present), epidote. The jet black basalt is now turned a pale green in color, and this is quite a common sight in the California Coast Ranges. If tectonic shear stresses are low, large-scale primary features such as lava pillows may be preserved, despite the metamorphic recrystallization that has reorganized the rock on smaller scales.

Blueschist is the rock that results when basaltic protoliths are driven deeper into the subduction zone, to perhaps ~35 to 50 km depth. The rate of subduction is greater than the rate of thermal equilibration; as a result the subducted basalt soon finds itself at conditions where the pressure is quite high, but the temperature is still relatively cold. These are the conditions at which the minerals lawsonite and glaucophane form. Metabasalts in the Angel Island Terrane have seen these conditions, and where basalt was present, it has been recrystallized to make blueschist.



It's important to note that blueschist cannot form from "just any rock." It must have basalt (*or greenstone, which is of course made of the same blend of elements*) as its protolith. If a sandstone is subducted to the same depths and temperatures, it cannot make a blueschist, as it doesn't have the right starting ingredients. Such a sandstone is said to have been metamorphosed "at blueschist facies," but it is neither blue nor a schist. We can still find crystals of lawsonite in it if we examine it in thin section (under a petrographic microscope), but to the naked eye, it frequently looks just like a normal sedimentary sandstone!

Eclogite represents the highest set of temperatures and pressures; the protolith basalt's atomic ingredients reorganize to form distinctive "Christmas tree" colored minerals: the rich green of omphacite (a kind of pyroxene), and the cranberry-red of garnet (the variety called pyrope). Eclogites are astonishingly dense, which is perhaps not surprising considering they form to be stable at pressures corresponding to depths of at least ~50 km!

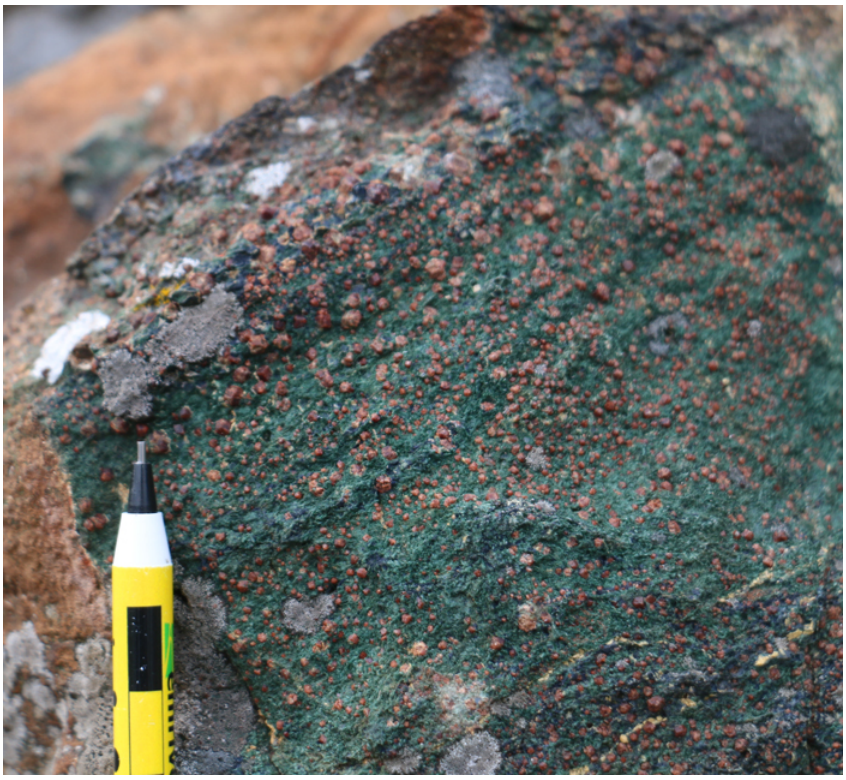


Figure 26.3.5: Eclogite consists of green omphacite (a variety of pyroxene) and prominent red pyrope (a variety of garnet). This outcrop appears on Ring Mountain, near Tiburon, California.

Eclogite is very dense. Depending on the location sampled, the density of eclogite ranges from 3.45 to 3.75 g/cm^3. Recall that the mantle's peridotite clocks in at around 3.3 g/cm^3, so this means that once eclogite forms, it can act as a dead weight, pulling on the rest of the subducted plate. This downward pull is thought to be a primary driver of "slab pull," a force partially responsible for the motions of lithospheric plates.

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26.4: The sedimentary journey

On top of the oceanic crust is where sediments can be deposited.

ABYSSAL DEPTHS: Chert & shale

In the offshore realm, far out in the Pacific *Ocean* (but atop the Farallon *Plate*), deep sea sediments of chert and clay shale accumulated. The source of the material for the chert was the dead bodies ("tests") of radiolaria, which rained down through kilometers of sea water as "marine snow." Unlike calcite, which tends to dissolve in the deep sea below the carbonate compensation depth, the silica that makes up the radiolarian skeletal material tends to be stable.



Figure 26.4.1: Chert of the Marin Headlands Terrane crops out on Marshall's Beach, south of the Golden Gate Bridge in San Francisco.

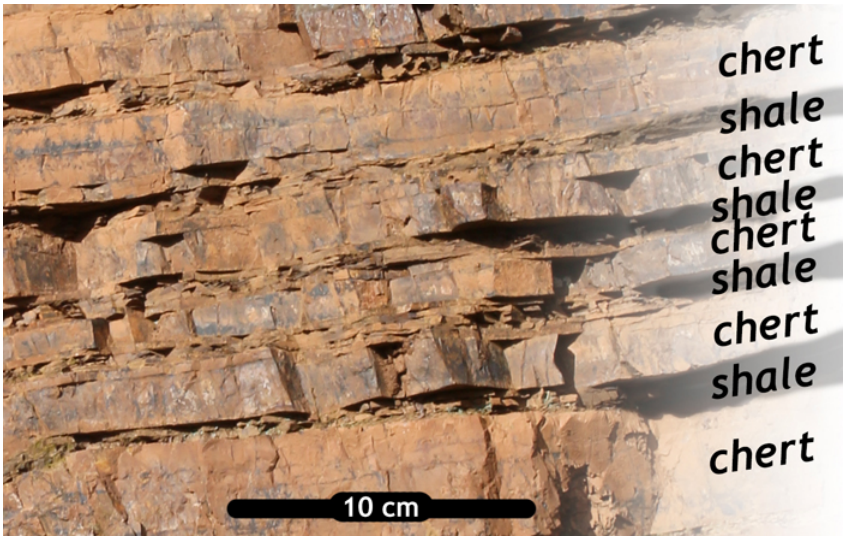


Figure 26.4.2: Chert and shale, rhythmically interbedded, as exposed at Marin Headlands, California.

We don't fully understand the alternation between chert and shale that characterizes the Marin Headlands Terrane, but it may be due to annual or seasonal cycles, or Milankovitch cycles in Earth's orbit. Some workers have suggested it's a diagenetic separation of the two sediment types: that is to say, the radiolaria skeletal remains and the clay are deposited at the same time, and only later separated once buried, by diagenetic processes.

Because chert is made of silica, it is both chemically stable and rather hard, so it stands up well to weathering and erosion. As a result, chert can be found underlying some of the highest landforms in the Coast Ranges.

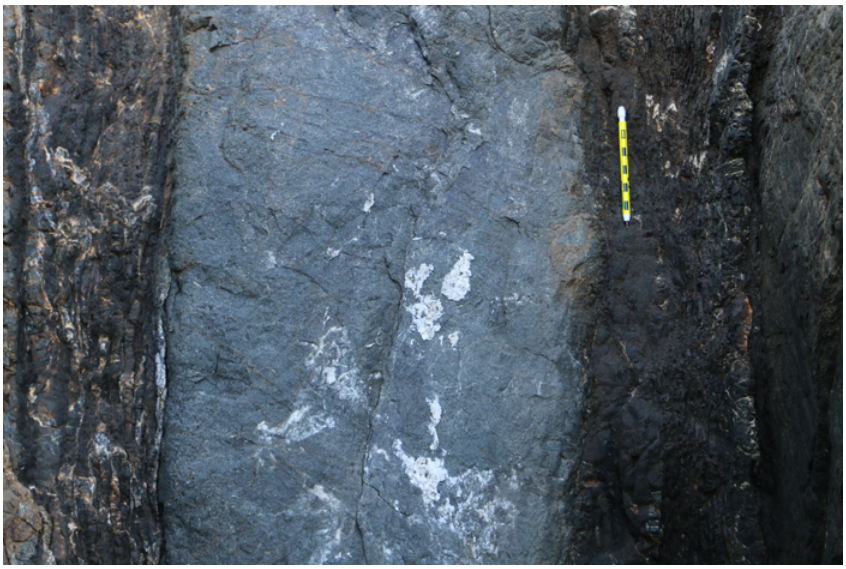


Figure 26.4.3: Graded bed in turbidite, Usal Beach, Mendocino County, California. Bedding is vertical, and fines (gets younger) to the right.

TURBIDITES: sand & mud via underwater avalanches

Closer to shore, the weathering and erosion of the continent brings a lot of sediment into the ocean. Along the continental slope, turbidity currents deliver slurries of sand and mud in suspension into the deep water. Though they come from surface waters, they are dense due to all their entrained sediment, and they tend to hug the bottom as they advance. Occasionally, they will travel along submarine channels, and spread out into complexes of abyssal fans. As these turbulent currents slow down, they drop their sedimentary load, with larger (heavier) particles dropping out first, and finer (lighter-weight) particles staying suspended in the water column for longer, eventually dropping out last. The resulting graded beds are signatures of this mode of deposition. We call the overall packages of graywacke sandstone and black shale deposited by turbidity currents “turbidites.”



Figure 26.4.4: This conglomerate bears a diverse assemblage of clasts, including blueschist and sandstone. Overall, the clasts do not appear to record much deformation. In other words, this appears to be a primary sedimentary conglomerate. Kayak Beach, Angel Island State Park, California.

NEARSHORE FACIES: River-borne sand & gravel

Closer to the coast, and in shallower waters of the continental shelf, rivers debouching into the sea drop deltaic deposits. For the most part, this is sand and mud, but in many places pebbles and cobbles (gravel) make it all the way to the shore too. These sedimentary deposits result in rocks such as shale (former mud), sandstone (former sand), and conglomerate (former gravel). Sometimes plant fragments are preserved in these rocks as little scraps of coal.

The ideal rock sequence for the rocks making up the subducted seafloor is therefore something like this:

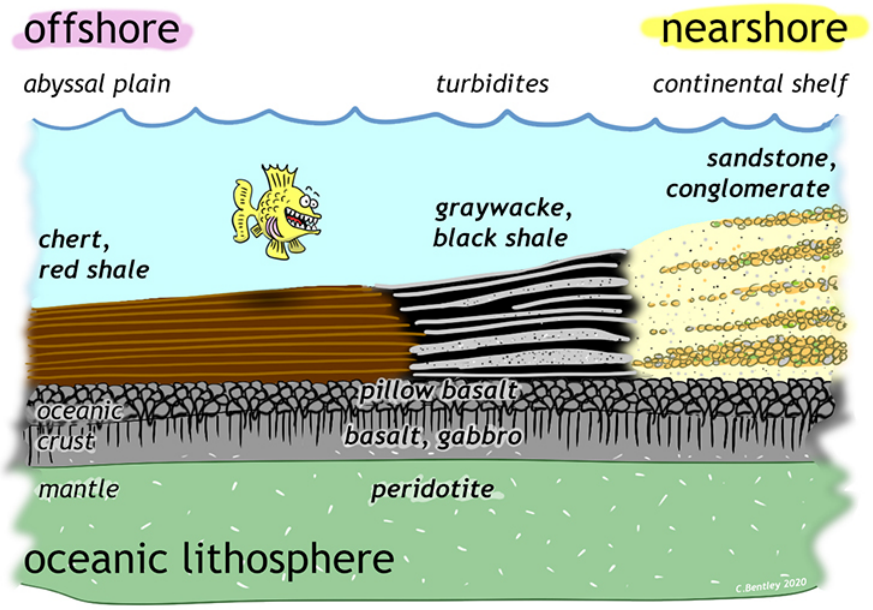


Figure 26.4.5: A cartoon summary of the original relationship between various rock units of the oceanic sequence, prior to being fed into the subduction zone (with accompanying metamorphism and deformation).

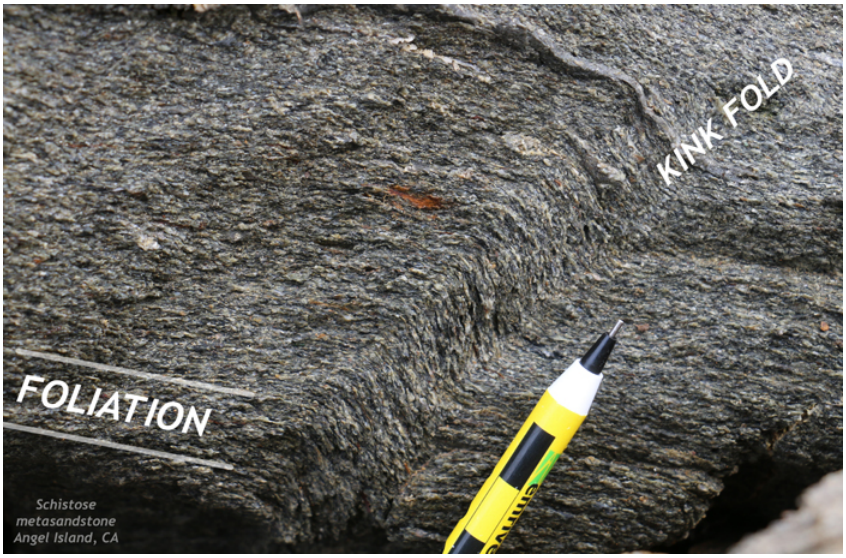


Figure 26.4.6: An example of a schisty texture, developed in a blueschist-grade metasandstone on Angel Island, California. A small kink fold emphasizes the trend of the shistocity (foliation).

METASEDIMENTS: schist and metaconglomerate

When they metamorphose, these sedimentary rocks transform into phyllite, schist or metaconglomerate. Mudrocks like shale can recrystallize to make shiny, foliated phyllite, which will turn to schist with higher grades of metamorphism. Sandstones turn into "metasandstone," which looks much like sandstone does (but contains key metamorphic minerals when examined in thin section). Some metasandstones can also develop a schistocity (scaly foliation, as seen in the example here).

Conglomerates made of a variety of clasts can demonstrate an interesting response to shearing stresses: the weaker rock types will smear out into ribbon-like shapes, while the stronger rock types will resist deformation as rigid chunks or else break into pieces. These “stretched pebble conglomerates” record *some* deformation, but not so much as to qualify the metaconglomerate as mélange.

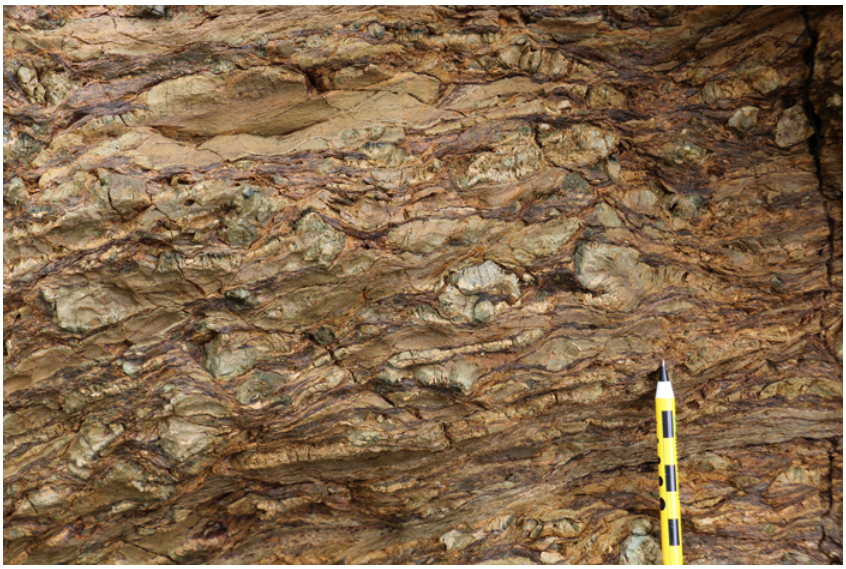


Figure 26.4.7: Coastal outcrop on Angel Island, showing a brown and green conglomerate, with pebbles of various sizes showing different degrees of shearing. Some are apparently intact and retain something close to their original shape, while others have been broken by (vertical) brittle fractures. Still others have been smeared out making a moderately strong (horizontal) foliation. These are continental margin deposits that have been fed into the subduction zone.

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26.5: Deformation of these rocks into geological structures

Convergent plate boundaries are stressful places: the rocks and sediments fed into the subduction zone are subject to compression and shearing, and the frequently deform in response to these stresses. Examples of deformation typical of subduction zone are well recorded in the rocks of California's Coast Ranges: folding, faulting, and extensive zones of shearing that result in "mixed-up" rocks called mélange. Let's examine each in turn.



Figure 26.5.1: Chevron-folded cherts of the Marin Headlands Terrane, exposed at Kirby Cove.

CRUMPLED LAYERS: Folding

Folding is particularly obvious in the regularly-layered cherts of the Marin Headlands Terrane, though it can also be spotted where bedding can be detected in the sandstone and shale layers in other adjacent terranes. Mechanically differentiated rock layers, such as chert/shale interbeds, are particularly susceptible to folding when compressed parallel to their layering: they buckle and crumple. Because the shale is weak and the chert is relatively stiff, layers of chert slide relative to one another on a "geological banana peel" of shale. This flexural slip allows much more folding than a homogeneous body of chert would be able to accommodate.

Folding occurs on many scales in California's Coast Ranges. As the example below shows, millimeter-thin layers can be folded, but so too can entire terranes on the scale of kilometers. For instance, the Angel Island Terrane is near the axis of a regional synform, with more recently accreted terranes on either side of it as well as wrapping around below it.

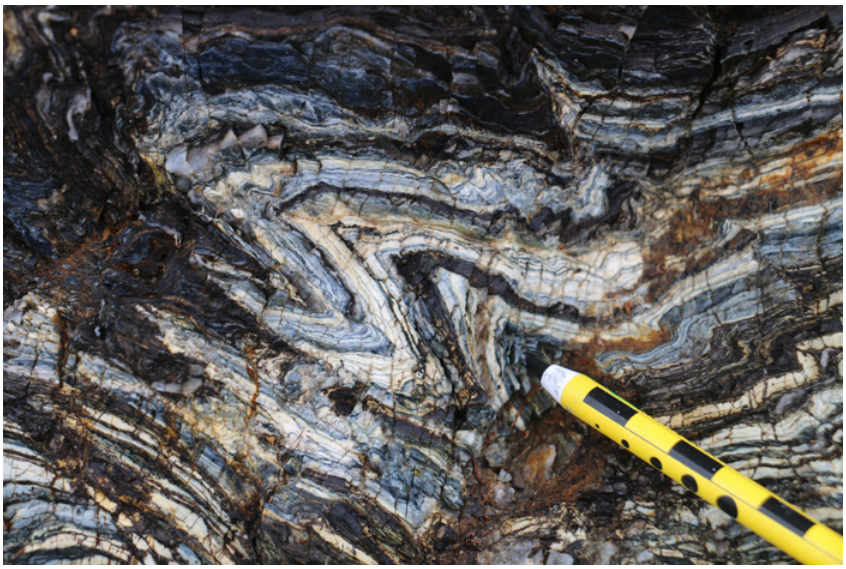


Figure 26.5.2: Folds in deep sea vent deposits, Kayak Beach, Angel Island, California.

SLIP ALONG FRACTURES: Faulting

Another form of deformation that is common in the accretionary wedge is faulting. Through faulting, new slivers of the oceanic section are slathered onto the bottom of the accretionary wedge complex, building it up from below.

At Marin Headlands, north of the Golden Gate Bridge, there are no fewer than 17 thrust faults that have placed deeper rocks of the oceanic crust on top of near-shore and off-shore sediments, repeating the overall oceanic section again and again.

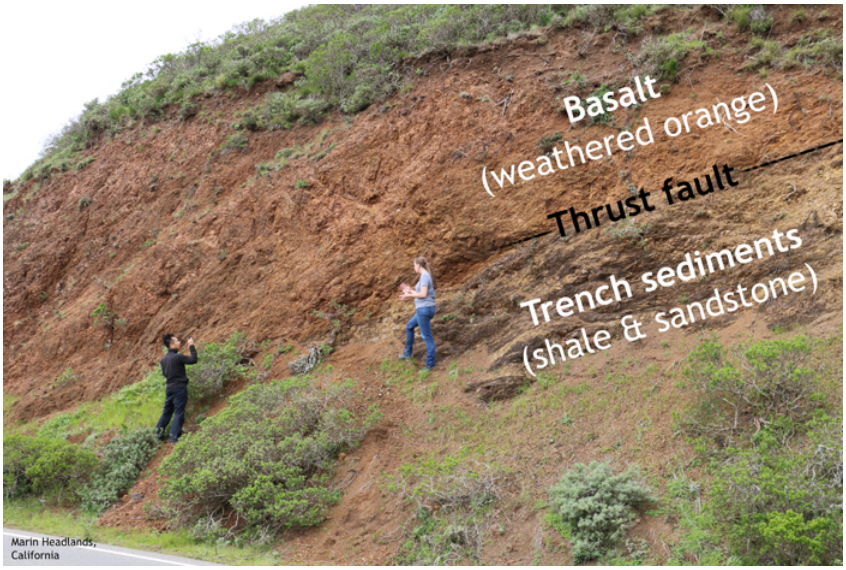


Figure 26.5.3: At Marin Headlands, basalt (rock of the oceanic crust) has been thrust atop trench sediments (sandstone and shale, sediments that must originally have been deposited on top of the crust). Sixteen such thrust faults have been mapped out at Marin Headlands, each repeating the oceanic "section" within the accretionary wedge.

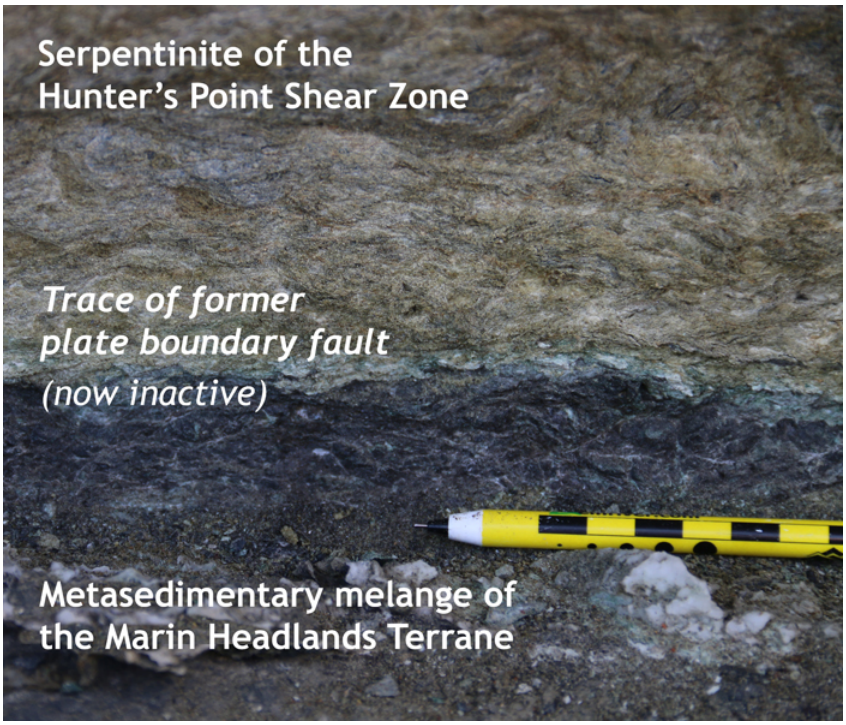


Figure 26.5.4: At Marshall's Beach in San Francisco, just south of the Golden Gate Bridge, visitors can observe the trace of the former plate boundary fault. This pencil-thin fault separated rocks already accreted to the North American Plate (in this case, former mantle that has been metamorphosed to serpentinite) from those that were being fed beneath, into the subduction zone (in this case, seafloor sediments of the Marin Headlands Terrane).

Sometimes these faults are discrete and crisp, where two vastly different rock units have been brought into contact along the slip surface. An example of this can be seen at right, where former mantle rock (serpentinite of the Hunter's Point Shear Zone) has been emplaced atop seafloor sediments (shale and sandstone mainly) of the Marin Headlands Terrane. During the Mesozoic, this fault was for a while the plate boundary between North America and the Farallon Plate. As subduction proceeded, new faults formed within the subducting plate, and the relative motion between the two plates transferred from a shallower fault to a deeper one. Newly subducted sediments were shoved underneath the mantle sliver, as the Farallon moved downward relative to North America. Later, another new fault formed, becoming the new plate boundary, and this fault was abandoned, left inert in the geologic record, with the footwall rock below it now also accreted to the North American Plate.



Figure 26.5.5: Geologist John Wakabayashi of California State University, Fresno, examines the fault contact between metasedimentary rocks of the Marin Headlands Terrane (footwall, to the left/west) and serpentinite of the Hunter's Point Shear Zone (hanging wall, to the right/east). This fault places mantle rocks on top of seafloor sediments. Sometime in the Mesozoic, this was the boundary between North America and the subducting Farallon Plate.



Figure 26.5.6: "Block in matrix" texture is characteristic of many Coast Ranges mélange outcrops. Marshall's Beach, San Francisco, California.

THE MIXED-UP ROCK: tectonic shearing to make mélange

Other rocks may be sheared out, deforming in a mass. Serpentinites are particularly weak rocks under shear stress, and break into thousands of tiny faults, giving the overall rock a “scaly fabric,” which looks at first like a cleavage. However, many of the little flakes shows slickenlines on their surfaces, indicating they are little faults that have seen some slip. A large body of serpentinite can thus be transformed to a broad shear zone. If other rock types get mixed into the sheared-out serpentinite, a serpentinite mélange will result. Mélange is a term for tectonically-mixed-up rock, where varying blocks or “lozenges” of varying rock types get entrained in a sheared matrix. We call this texture “block in matrix.” Most commonly the blocks are of coherent blocks of the matrix rock, but in other cases, all sorts of exotic blocks can be introduced. At Ring Mountain in Tiburon, for instance, a serpentinite matrix hosts blocks of blueschist, eclogite, amphibolite, peridotite, and meta-chert.



Figure 26.5.7: Serpentinite mélange showing block-in-matrix texture. Perles Beach, Angel Island, California.



Figure 26.5.8: A mix of protoliths in mélange exposed at Pacifica State Beach, California. Mélange can have a metavolcanic or a metasedimentary matrix. Here, it's both!

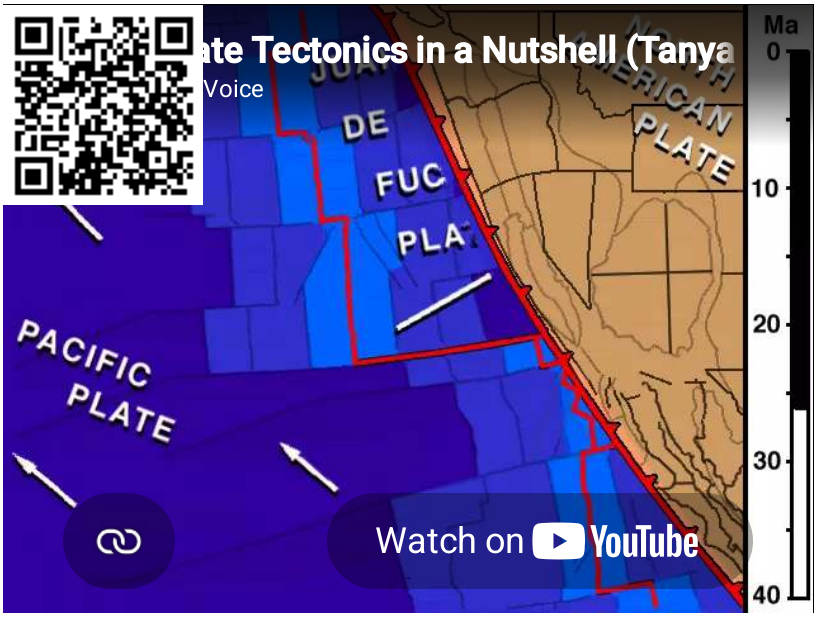
This pulpy mess is extraordinarily incompetent, and is the source of frequent landslides in coastal California. Shockingly, the south tower of the Golden Gate Bridge is anchored in serpentinite mélange of the Hunter's Point Shear Zone.

Shale acts in the same slippery fashion as serpentinite, as the flakes of clay it contains slip relative to one another, creating a scaly fabric. Shale-based mélange is just as common as serpentinite-based mélange. In San Francisco itself, the Hunter's Point Shear Zone is a wide swath of serpentinite mélange, which the City College Fault Zone has more of a shale basis to the scaly matrix. Near Pacifica, a town further south along the coast, the mélange has a distinctive mix of metavolcanic and metasedimentary source rocks, resulting striking mixtures of pale green and dark gray.

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26.6: The end of subduction

Subduction of the Farallon Plate continued until the around 25 million years ago. By the point, the faster rate of subduction had caught up to the slower rate of seafloor spreading, which consumed the last bit of young, fresh Farallon crust, and then the oceanic ridge itself. This brought the North American Plate into contact with the Pacific Plate for the first time. This "moment" was the birth of the San Andreas Fault, the modern transform plate boundary between North America and the Pacific Plate. The shape of the Farallon Plate was such that it was narrowest in the middle, and wider both north and south of that. So the first spot the subduction zone gobbled up the spreading center was in the middle of the Farallon Plate. This broke it into two. Over time, the Farallon's remnants (the Juan de Fuca Plate in the north, and the Cocos Plate in the south) continued to subduct, but they get smaller and smaller with each passing millennium. Between them, the San Andreas Fault continues to get longer and longer. Here's an animation showing how this has played out:



That covers what happened horizontally after the end of subduction, but there were also *vertical* motions of the crust. With the cessation of the drag of the subducting slab, the accretionary wedge was no longer pulled downward. It was allowed to rebound upward, and eventually rose to poke up above sea level, making the California Coast Ranges. Similarly, the forearc basin uplifted to become the flat-as-a-pancake Great Central Valley, and erosion of the volcanoes of the continental volcanic arc revealed their deeper plumbing system: the plutons of granite and granodiorite that we now see in the Sierra Nevada range.

Anderson, Don (2007). "The Eclogite Engine," Chapter 5 in *New Theory of the Earth* (book), Cambridge University Press.

Atwater, Tanya (1970). "Implications of plate tectonics for the Cenozoic tectonic evolution of western North America," *Bull. Geol. Soc. Amer.*, v. 81, p. 3513-3536. DOI: 10.1130/0016-7606(1970)81[3513:IOPTFT]2.0.CO;2

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Warhaftig, Clyde (1984). *A Streetcar to Subduction, and other plate tectonic trips by public transportation in San Francisco.* American Geophysical Union: book.

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CHAPTER OVERVIEW

27: (Case Study) The mixed-up quartzites of Cape Agulhas

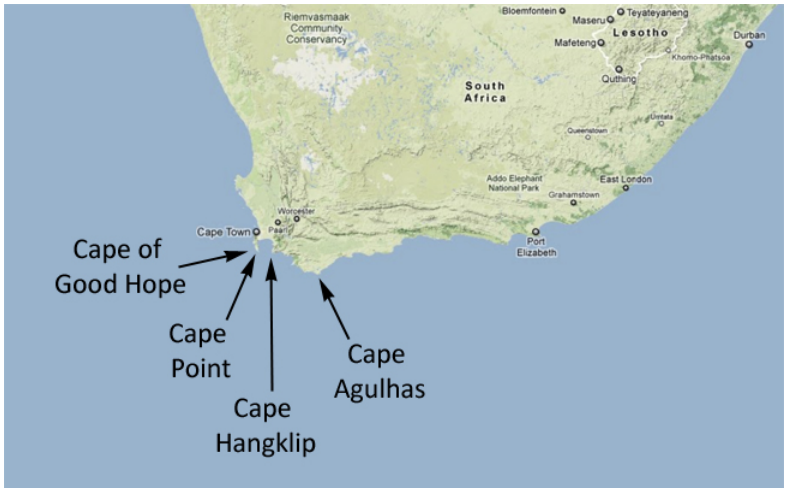
27.1: The mixed-up quartzites of Cape Agulhas

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27.1: The mixed-up quartzites of Cape Agulhas

This case study has you going off on a grand adventure!

Here, we examine what you will see if you visit the southernmost point of the African continent. There are various protuberances of land in southern South Africa, each of which is a “cape” with its own proper name. The Cape of Good Hope is probably most famous by virtue of its awesome, mythological-sounding name and its proximity to the urban center of Cape Town, but there are others, too. For instance, Cape Point is only a kilometer or two from the Cape of Good Hope. On the other side of False Bay, there is Cape Hangklip, and beyond that a ways lies the object of this case study, Cape Agulhas:



There is a plaque there showing that this is Africa’s southernmost point, and pointing east to the Indian Ocean, and west to the Atlantic Ocean:





The word “Agulhas” means “needles,” a Portuguese reference to the fact that when it was named in the year 1500, the cape was about where magnetic declination was 0°; in other words, the longitude at which magnetic north and true north were the same. (Since the magnetic pole wanders, however, this name is no longer as accurate as it was half a millennium ago.)

It is a pretty enough place, as you can see in the images below, but not nearly so striking as Cape Point or the Cape of Good Hope. Cape Agulhas is the southernmost protrusion of a mildly curving coastline, rocky and wavestruck, but it is not a crisp peninsula or anything like that...





The rocks here are quartzites, though their surface is covered with an assortment of lichens above water and seaweed below, and other encrusting organisms.

Imagine that you visit. You can wade out to one of the small outcrops of quartzite just offshore:



And then you notice something...



What pattern is apparent in this outcrop? (Ignore the orange lichens growing on it; look deeper into the structure of the quartzite.) Hint: it's a primary sedimentary structure with geopetal value...

This is **cross-bedding**, and it's **up-side-down!**

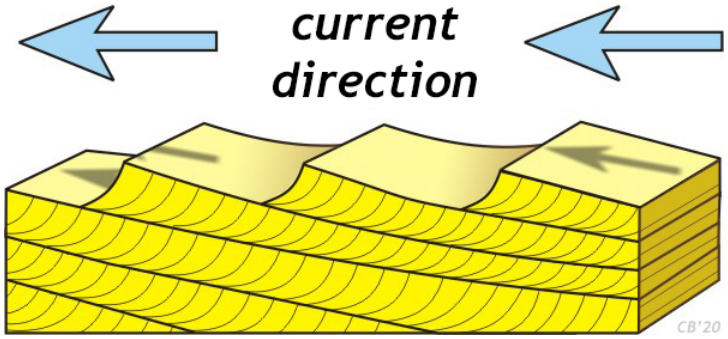
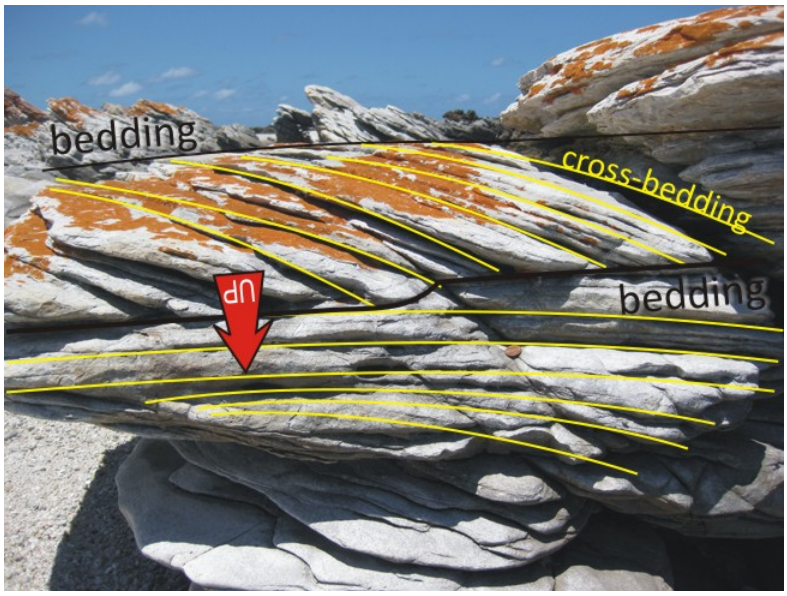


Figure 27.1.1: Cross-bedding and ripples are part of the same depositional setting, forming in a flowing directional current.

Cross-bedding is formed when layers of sediment are deposited on the leeward side of a ripple or dune that is moving grain by grain in a uni-directional current.

Notice how the cross-beds are at an angle to the main beds, and that the cross-beds are concave-up. They are **tangential** to the main bed at the bottom (meaning they curve gently to be parallel with the main bed) and they are **truncated** by the main bed at the top (meaning that they stop at an abrupt angle with the bottom of the overlying bed). In other words, cross-beds look different right-side-up than they do up-side-down. They can thus serve as good **geopetal indicators**: little geologic clues in stratified rocks about "which way was up" when the beds were deposited.

Looking at this outcrop through the eyes of a geologist, then, you see this:



A few dozen meters away, you find a right-side-up set of cross-beds, inclined steeply to be sure, but not inverted. You can tell this because the cross-beds are concave up, and truncated on their upper surface, and tangential on their lower surface:





This observation piques your interest, and you spend a pleasant half hour in the sunshine, searching around for cross-bedding in different orientations. You're not bored for long; soon you find a bunch of examples...

Your geological training allowed you to make the observation that sent you off on this quest. If you hadn't been educated to recognize cross-bedding when you saw it, you likely would have looked at the Cape Agulhas outcrops as "just a bunch of rock."

So now that the observation has been made, and then validated with additional exploration, you are probably beginning to wonder why. What hypotheses pop into your mind as you glide from outcrop to outcrop, seeing dozens of examples of inverted cross-bedding.

Scroll through these additional examples you encounter:

Example 27.1.1



https://geo.libretexts.org/@go/page/23092

Example 27.1.2



Example 27.1.3



Example 27.1.4



So there are a variety of orientations for the cross-bedding that you observe. How can you explain this?

One solution that unfolds in your mind is that the beds in which the cross-beds are included are themselves in multiple orientations, and in many places overturned. One way to accomplish this is by folding the beds up rather intensely. Like this (sketch drawn on my iPad's "Whiteboard" app):

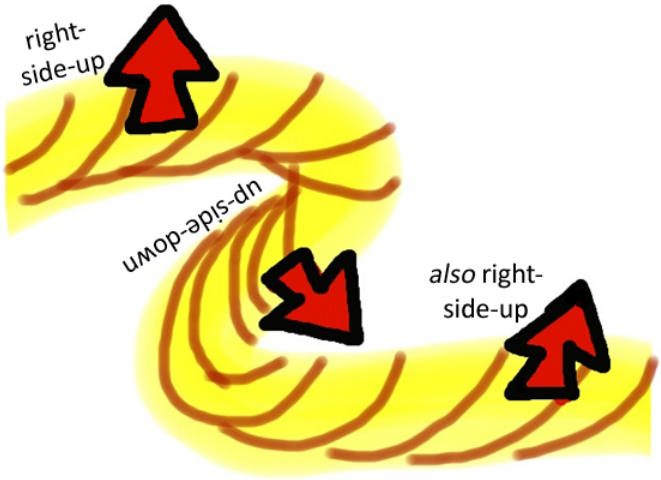




Figure 27.1.2: Small-scale fold in quartzite beds at Cape Agulhas.

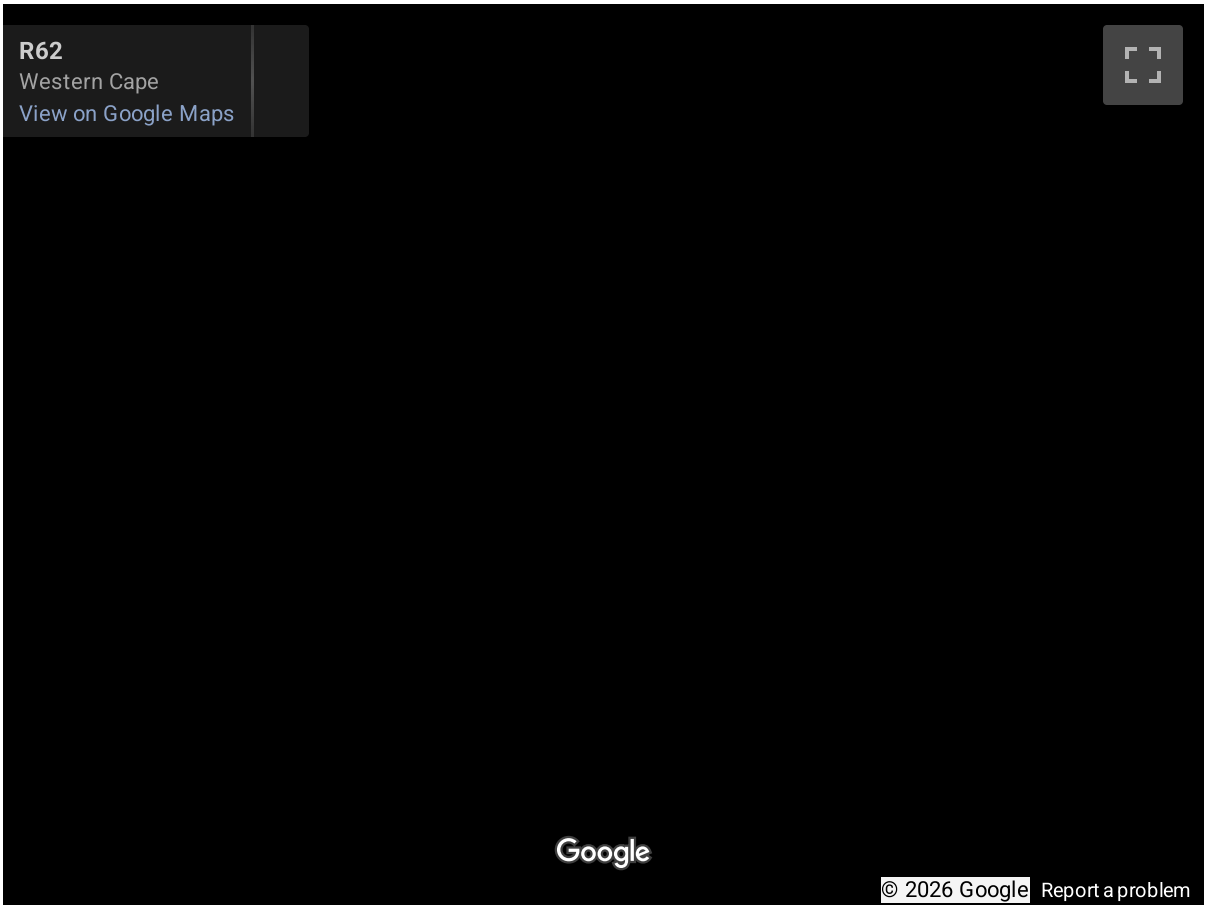
Sure enough, once you spend a little time looking around, testing your “folded” hypothesis, you find some confirming evidence; a few tight small scale folds. Applying Pumpelly’s Rule (the notion that small-scale structures mimic and reflect larger-scale regional structures), you deduce that you may be in a zone of folded sedimentary rocks: a ‘fold and thrust belt’ associated with an ancient episode of mountain building.

Indeed, these folds (and the flipped-around beds they deform) are part of the Cape Fold Belt, an east-west trending mountain belt that formed along the southern coast of South Africa when South America mashed into it during the amalgamation of the southern supercontinent Gondwana. These quartzite strata are the same geologic unit that holds up the mesa-like form of Table Mountain immediately south of Cape Town, but here they are much more deformed.



Figure 27.1.3: Folds exposed in a mountainside above Hermanus, South Africa. These strata have been deformed as part of the Cape Fold Belt.

Even driving the roads in this part of the world, there is strong evidence that these strata have experienced extreme tectonic compression. Explore this Google Maps “Streetview” scene below:



In this case, noticing the up-side-down cross-beds (a small-scale observation) led to a regional-scale insight: These rocks have been messed up by a tectonic collision!

Well done.

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CHAPTER OVERVIEW

28: (Case Study) Palimpsest outcrops

Many outcrops or samples of rock are simple and straightforward: they bear characteristic compositions, textures, or features that evoke a straightforward geologic interpretation.

But some are more complicated than that. Some outcrops or rocks show evidence of more than one “story” of formation. The successive chapters of its history are superimposed on its features, like a warrior accruing scars through a long and glorious career.

28.1: Palimpsest outcrops

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28.1: Palimpsest outcrops

Introduction: What does "palimpsest" mean?

The concept of palimpsest can serve us well in making sense of such rocks. “Palimpsest” means “something reused or altered but still bearing visible traces of its earlier form.” The origin of the word is a little different: originally it meant “scraped again and ready to use,” which refers to ancient writing on sheets of parchment. These ancient “sheets of paper” (actually animal skins, specially treated) were fairly precious and not to be discarded. When the writing on them was considered expendable, the parchment was recycled in another round of use. Ancient scribes would use a sharp blade to scrape the dried ink off the stiff sheets of parchment, and make them more or less usable again. This was before “delete” buttons and even erasers. The thing that captures the essence of palimpsest is the fact that a faint ghostly version of the initial script could still be discerned “behind” the brand-new bold lettering. You can seeing through one aspect of the manuscript to peer into a deeper level of its more ancient history.

An important example is this one:



Figure 28.1.1: A photograph of a two-page spread from the famous “Archimedes Palimpsest.” (Source: public domain image on Wikimedia by Matthew Kon.)

That photos shows a palimpsest which includes an older notebook of the Greek natural philosopher Archimedes that was later re-purposed by Christian monks as a prayer book. Archimedes’s work runs horizontally in this view, and the later religious overprint is in the lines of text that run vertically. The beauty and magic of a palimpsest such as this is that both texts can be extracted from the single page.

Bear this in mind as you contemplate this outcrop:

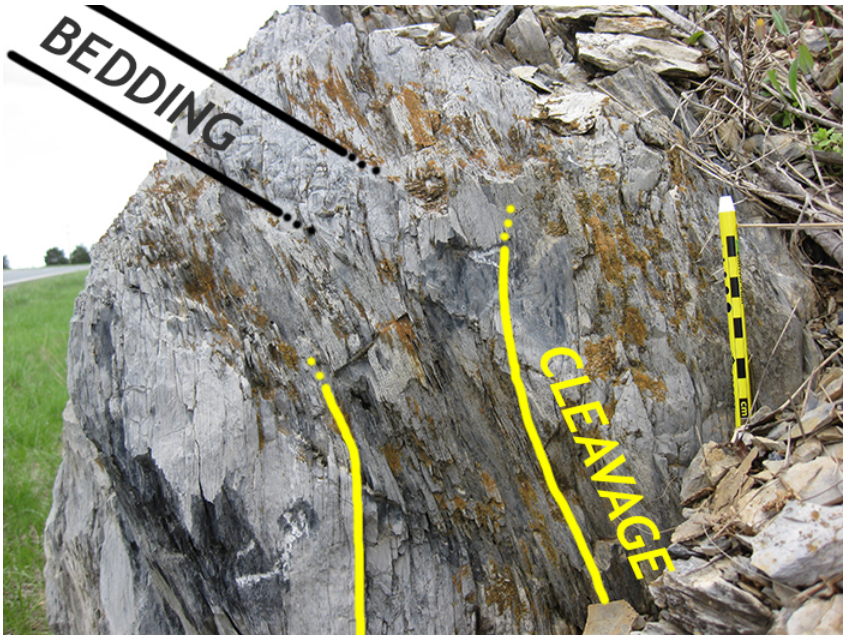


Figure 28.1.2: Both primary bedding and secondary cleavage can be discerned in this outcrop of Edinburg Formation in western Virginia. The bedding is Ordovician is age, but the cleavage was superimposed later, during late Paleozoic mountain-building.

This is an outcrop that shows both a "primary" feature (sedimentary bedding) and a later, "secondary" feature (the tectonic cleavage that overprints the bedding). It's one rock body that tells a story with two parts. *It's a palimpsest rock!* As with the Archimedes Palimpsest, we can see through one aspect of the rock (the cleavage) to peer into a deeper level of its story (the more ancient sedimentary bedding).

Many, many rock units have **more than one story** to tell. They've been through a lot, and they have the scars to prove it. Let's look at a few examples of these "palimpsest rocks."

Example 28.1.1: Keels, Newfoundland

Start by examining this rock outcrop, located near the little coastal village of Keels on the Bonavista Peninsula of eastern Newfoundland.

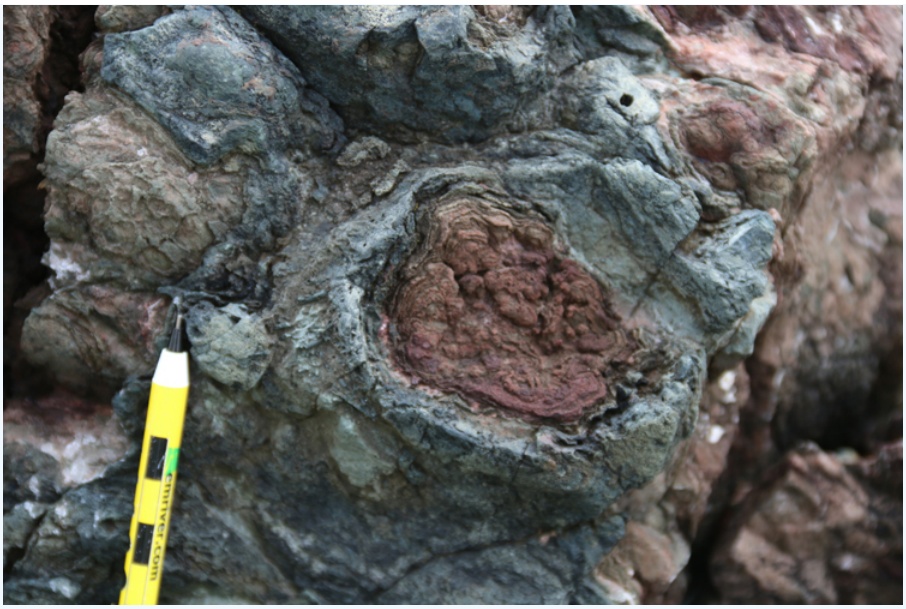


What you see here is a cross-sectional view of small stromatolites that are developed in sedimentary layers. The different layers are marked by the color change from red to green.

Here is a view looking down on the bedding plane, so you can see the stromatolites popping out in 3D:



Not that in some places, the green layer has broken open, revealing the maroon layers beneath with their concentric rings. Here is a close-up:



Stromatolites are ancient microbial mats that domed upward toward the sunlight as they grew. They were laid down in sticky layers of slime by photosynthetic microbes, probably cyanobacteria. But that's not all we see here. The tectonic cleavage is also a prominent feature of these rocks. Here, you can see bedding marked out by the different colored layers going left-to-right across the outcrop, while cleavage is more or less vertical, and etched out by seaside weathering:



This cleavage is caused by tectonic squeezing, compression from the left and the right sides of the outcrop. This causes pressure solution to develop in the rock, as minerals like calcite dissolve in the direction of maximum pressure, and reprecipitate in directions of lower pressure. The rock is reorganized under the tectonic compression, shortening from left to right and elongating top to bottom.



Figure 28.1.3: An animation showing the development of slaty cleavage (tectonic foliation) in an original stromatolite-bearing rock.

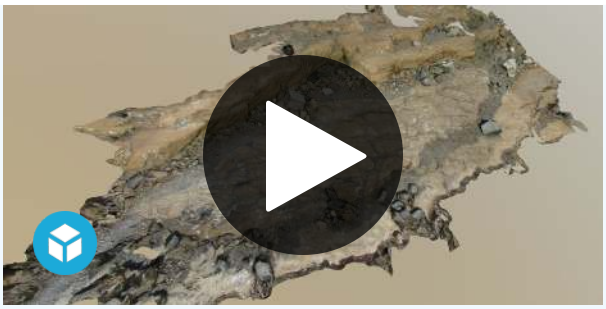
The combination of the stromatolites' shapes *and* the tectonic overprint (garnished with a final surface dressing by the various processes of weathering) gives us the final rock that we see today: the cumulative result of *all* these different processes playing out over time. This rock tells a story first of deposition of sediment in a shallow body of water (stromatolites need light to grow, so it cannot be too deep), followed by lithification and tectonic compression (resulting in the foliation), followed by uplift, exposure, and weathering.

Here's a sample of similar rock, also from Newfoundland, showing a Cambrian-aged *Olenellus* trilobite with a slaty cleavage overprint:



Example 28.1.2: Helena Formation at Grinnell Glacier, Glacier National Park, Montana

Now consider this outcrop in Glacier National Park, adjacent to the Grinnell Glacier:



We see more stromatolites here – bigger ones. *(Note that the same pencil from the Keels photos again serves as a sense of scale!)* The fossils appear in honey-colored rock rather than maroon or green. Again, the outcrop surface is more or less parallel to the bedding plane, so the stromatolites appear as "bull's eye" concentric rings. There is no tectonic cleavage on this outcrop, so there is no need to invoke an episode of mountain building on the basis of what we see here.

However, there is a clue that something else has happened in the history of this outcrop: the shallow grooves that go across the outcrop from left to right. Notice how they are all parallel to each other. These are glacial striations. This outcrop was underneath a flowing alpine glacier as recently as a hundred years ago. As the glacier flowed over the rock, bits of sediment entrained in the ice gouged into the rock, producing the striations.

These striations are not a "deep" part of the rock, but merely a carving of the surface of the outcrop. They are subtle and ephemeral: another ten thousand years of erosion might grind them away to an even surface. But while they last, they tell us about a recent cold period that managed to gouge this rock before melting back again.

Glaciated stromatolites: another palimpsest rock!

Example 28.1.3: Epidotized volcanic breccia of the Catoctin Formation

In Shenandoah National Park, astride Virginia's Blue Ridge, there are many outcrops of a strange looking green rock with spots. It looks like this:

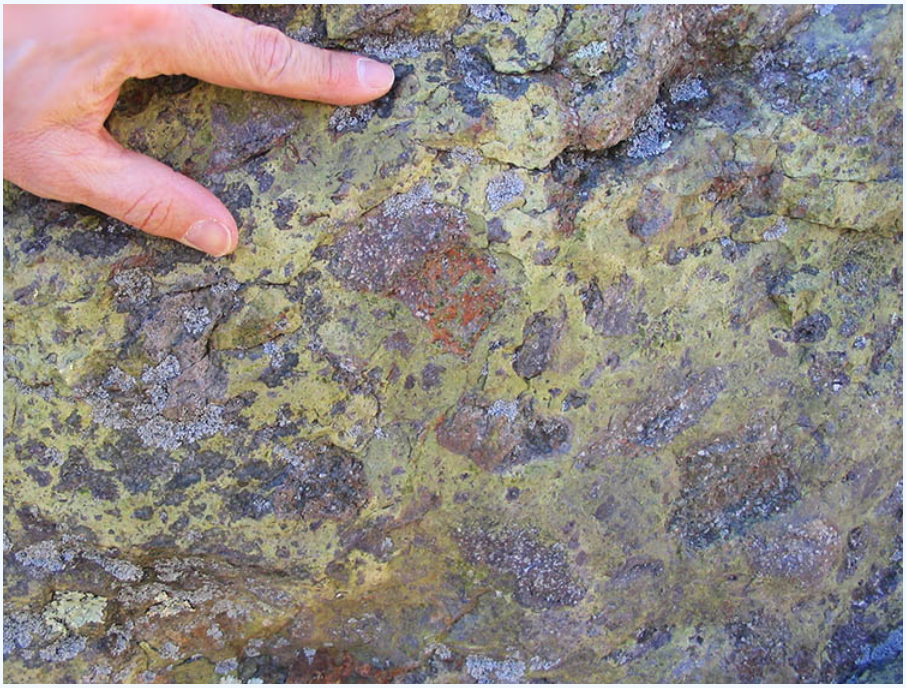


Figure 28.1.4: Ancient lahar deposits of the Catoctin Formation, as seen at the Franklin Cliffs Overlook on Skyline Drive. The purple chunks with white dots are clasts of meta-basalt with infilled amygdules. The green material surrounding them is epidotized volcanic ash.

This rock must be metamorphic, because it's chock full of the pistachio-green colored metamorphic mineral epidote, the metamorphic product of reactions between plagioclase and pyroxene in the presence of water. But this epidote seems to serve as a kind of "matrix" surrounding discrete chunks of purple rock with white dots. These purple chunks are meta-basalt. Their vesicles have been filled in with white mineral deposits, making them amygdules. The overall texture, therefore, looks a lot like a conglomerate. What gives? Is this a sedimentary rock, based on this poorly sorted texture? An igneous rock, because of the basalt and its ancient gas bubbles? Or a metamorphic rock, as the epidote would imply?

...*Yes*.

This outcrop is part of the Catoctin Formation, a Neoproterozoic aged volcanic sequence that formed upon the rifting and breakup of the supercontinent Rodinia. Most of the Catoctin consists of basaltic lava flows, but there are pyroclastic deposits (tuffs) too, and even some sedimentary layers (stream conglomerates, lake muds) in between the lava flows, indicating the passage of a relatively long time between eruptions. There are also layers of volcanic mudflow deposits – what we would today call "lahars." These slurries consisted of wet volcanic ash mixed with rock fragments. Most of the rock fragments were locally derived, and hence derived from the earlier part of the same volcanic sequence. Lahars are essentially sedimentary deposits of volcanic origin. We call the resulting deposits "volcanic breccias." Whether the amygdules formed before or after the basaltic clasts were caught up in the mudflow is unknown.

However, the formation of the epidote (a typical greenschist-facies metamorphic mineral) is not something we would expect to form in a volcanic setting. It is almost certainly due to the (much later) Alleghanian Orogeny, an episode of mountain-building caused by the collision of ancestral Africa (Gondwana) with ancestral North America (Laurentia) during the late Paleozoic era. At that time, the rocks of the Mid-Atlantic were subject to regional metamorphism and deformation. The ashy matrix of the volcanic breccia was more susceptible to metamorphic change than were the solid clasts contained in the matrix. This explains the difference in how pervasive recrystallization was in the one part of the rock compared to the other.

This Catoctin meta-volcanic breccia is thus a sedimentary deposit of volcanic origin, that was then metamorphosed.

Example 28.1.4: MacDuff, Scotland

Finally, consider this outcrop:



Figure 28.1.5: An outcrop in Scotland's Buchan block, near MacDuff. (Photo by Tim Johnson.)

There's a lot to look at here! Let's name some of these features, so that we can use them to tell the composite story of this outcrop:

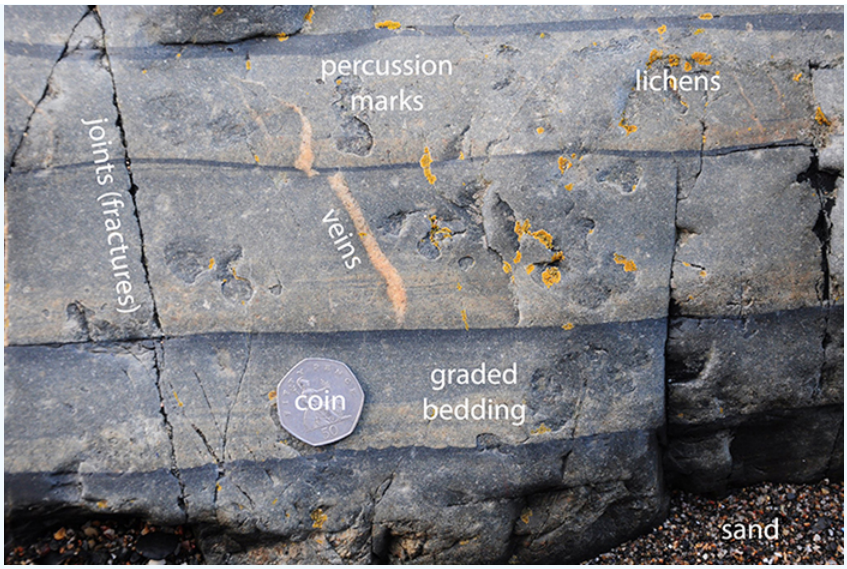


Figure 28.1.6: An annotated copy of the photograph, highlighting various features that together make up the modern outcrop.

Joints are fractures in the rock. They are different from faults, because there hasn't been any movement of the rocks on either side of a joint. **Veins** are healed fractures: the rock breaks, then mineral deposits seal the crack shut again. **Percussion marks** are "dings" that occur when a hefty particle of rock, such as a cobble, smashes against another rock. A cone-shaped package of fractures propagates a short distance into the impacted rock. It's very similar to a piece of gravel hitting a car windshield and producing a spider-web-shaped pattern of cracks. **Sand** is draped up against the base of the outcrop, but is a much younger deposit than the sedimentary layers seen in the outcrop itself. These show **graded bedding**, a primary sedimentary structure that forms when a mixed-size deposit of sediment settles out from the water column: biggest particles drop out first, and finest-grain particles settle out last. **Lichens** are lifeforms (a symbiotic mix of an alga and a fungus) that colonize rock surfaces. The **coin** serves as a sense of scale, so we know how big all the rest of this is.

It's quite possible that we only care about the graded bedding, and the rest of these details are just distractions.



Figure 28.1.7: The same photo, digitally "cleaned up" to focus on the graded bedding.

With modern digital manipulation, it's easy to remove some of the "distracting" elements from the image, to focus in on the characteristic that we find most relevant to our study – the graded bedding, say. But while Photoshop might make it easier to focus on the graded bedding, real rock outcrops look more like the first photo: with scuffs and scars that are evidence of part of the rock's full story. If you are lucky enough to do geologic field work, you'll have to figure out what is worth paying attention to. This will be determined by the sorts of questions you're asking: are you interested in the original depositional setting of the

> sediment that later lithified to make this rock? Then the graded bedding is relevant. Are you an economic geologist interested in whether these rocks might have gold or other precious metals in them? Then likely the vein will be the characteristic you're most interested in. If you're interested in the stresses associated with uplift, the joints might be the most relevant variable. If you're a lichenologist, none of that matters, and the rock is only a substrate for the lichens to colonize. You'll soon have to practice the skill of seeing past the irrelevant information and focusing on what you most care about, but part of your practice should be to try and see the whole rock, the whole outcrop, as a written record of a much longer, much more complicated story.

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CHAPTER OVERVIEW

29: (Case Study) Extremadura - interpreting a geological cross-section

29.1: Extremadura - interpreting a geological cross-section

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29.1: Extremadura - interpreting a geological cross-section

This case study presents a single image and challenges you to interpret it. This diagram shows a geological cross-section through rocks of a park in Extremadura, Spain.

Cross-sections give us a “sideways” view into the structure of a region, showing the variation of rock units with depth as well as distance. They are essential tools for interpreting geological histories.

So examine this diagram. How many geological events can you postulate having happened, in order to produce the sequence of rocks and structures seen here?

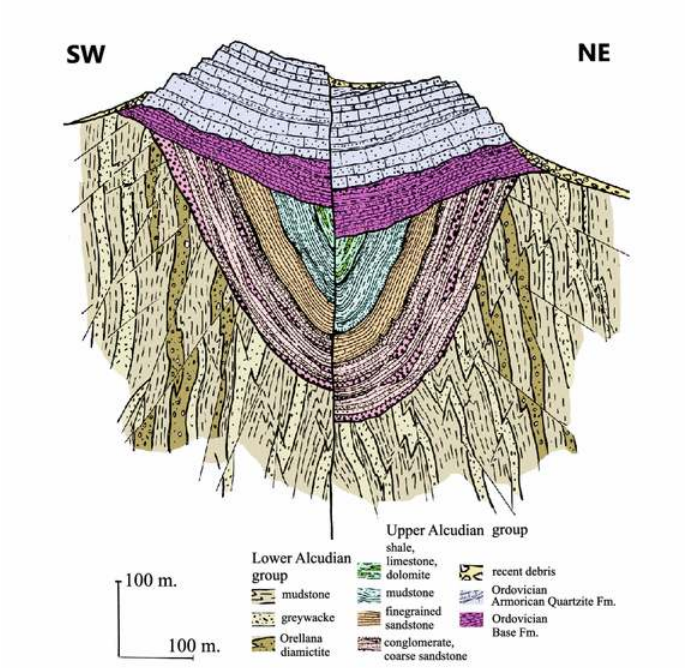
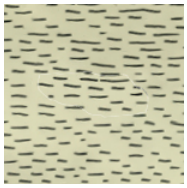
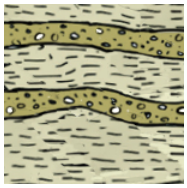
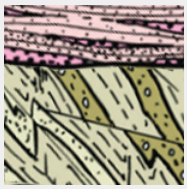
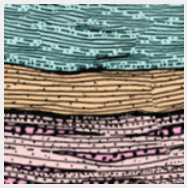
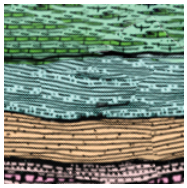


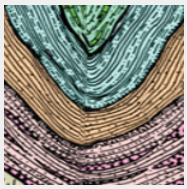
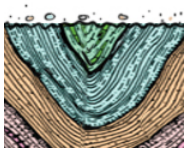
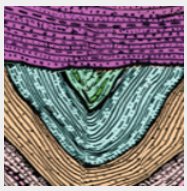
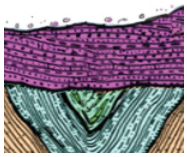
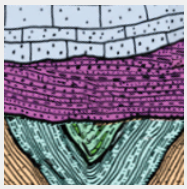
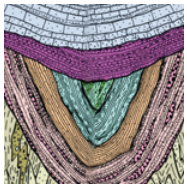
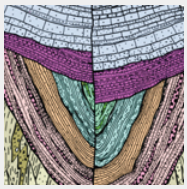
Figure 29.1.1: Cross-section through the rocks at Villuercas-Ibores-Jara UNESCO Global Geopark in Extremadura, Spain. (Modified from a figure by J.J. Álvaro, S. Jensen, and T. Palacios.)

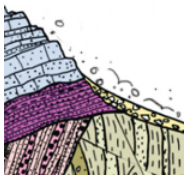
Here is one possible answer, starting with the oldest units (at the bottom of the diagram):

Table 29.1.1

1		*Once upon a time*, there was a sedimentary basin where most of the time, mud was deposited. According to the figure's explanation, this mud would go on to be preserved as the "**mudstone**" of the Lower Alcudian Group. The exact nature of this basin isn't clear from the data available to us, but depositing mud requires calm water. Fast moving water keeps small particles moving rather than laying them down, so probably this was a deep setting, perhaps a deep marine basin. That mud had to come from somewhere, so the presence of the mud deposit implies the weathering and erosion of rock, somewhere else.
2		Periodically, turbidity currents would bring in layers of dirty sand. Turbidity currents are gravity flows of sediment and water, and they tend to hug the bottom of the ocean due to their density. They can travel laterally for substantial distances before slowing and depositing their sediment, often in graded beds. This would later lithify to **greywacke**. *(Usually this word is spelled gr**a**ywacke in America, and spelled with an 'e' in Europe. Since this is a European example, we'll retain the source nomenclature.)*
3		At some point, either a glacier dumped a bunch of debris in, or there were some submarine landslides that resulted in an exceptionally poorly sorted deposit of **diamictite**. *(We can't tell from the data available here whether the clasts in that diamictite are faceted and striated or not; that would allow us to distinguish whether a glacier was responsible for the diamictite.)* In unfaulted sections, there are two layers of diamictite, separated by mudstone; so whatever process made the diamictite, it must have happened twice.
4		Then there was an time of **folding and thrust faulting**, probably accompanying an episode of mountain-building. Roughly speaking, and with the diamictite acting as our marker unit, the strata were warped into what appears to be a regional antiform in structure. The strata would not have been as steeply inclined then as they are now, so let's call this a broad antiform. *(I use the*

		term "antiform" instead of "anticline" here because we don't have information about the "paleo-up" direction of these Lower Alcudian strata. Strictly speaking, anticlines preserve superposition age sequences – youngest is "up" — but antiform is a more agnostic term; it implies that the strat could young in either direction.)
5		This tectonic squeezing resulted in **uplift**, subjecting our now-near-vertical strata to a time of **erosion**. The mudstone, greywacke, and diamictite layers were truncated against the ancient Earth's surface. *(This is the first step in making an unconformity.)*
6		Time went by and subsidence occurred, allowing sea level to creep back up and over the site. Higher-energy fluvial or shoreline (delta) deposits were laid down. Based on the lens-shaped deposits of **conglomerate** within the **coarse sandstone**, it looks like this was either terrestrial or transitional. The base of the unit makes an **angular unconformity** with the layers described in steps 1-5.
7		On top of that is a layer of **fine-grained sandstone**, perhaps indicating that the next step was sand being deposited on a beach or in a nearshore marine setting.
8		On top of that is a layer of **mudstone**, implying that deposition shifted back to a more offshore setting, and…
9		Topping that is a layer of **limestone, dolostone, and shale** – the most offshore facies yet. We can now apply our understanding of stratigraphy in relating the deposition of these different types of sediment. In total (steps 6, 7, 8, and 9: the Upper Alcudian Group), this package represents a classic fining-upward **transgressive sequence**.

10		Then another episode of tectonic compression happened, **folding** all these layers into a big syncline.
11		Once again, this **uplifted** the edges of some of the layers to the surface, where they were **eroded**. *(We're working our way into a second unconformity.)*
12		Again, subsidence and/or sea level rise (during the Ordovician, according to the figure's explanation) brought deposition back to the site, resulting in the **"Base" Formation**. An **angular unconformity** exists now between the base of the Base *(see what I did there?)* and the underlying steeply-dipping strata of the Upper Alcudian Group.
13		Based on truncation of some of the upper layers of the Base Formation against the overlying unit, I'm guessing there was another round of **erosion**. This was not as extreme an episode of erosion as the two we've seen previously, apparently lacking any significant tilting or major uplift.
14		Deposition resumed anew, with layers of quartz sand, also during the Ordovician. Later this will be lithified and apparently metamorphosed to make the **Amorican Quartzite** formation. This makes an **disconformity** with respect to the underlying Base layers.
15		Then there was a little bit of **folding** (pretty subtle compared to its predecessors), again affecting all layers, but most noticeable in the non-horizontality of the Ordovician package. Again, it's taken on the form of a syncline.
16		Then **faulting** down-dropped the northeast side relative to the southwest.

17		Then (or plausibly accompanying step 16) **uplift** brought it all above the surface, where **erosion** went to work again, breaking down the mountainous outcrop and laying down temporary (?) deposits of Recent debris on the flanks of the mountain.

Seventeen steps are needed in order to generate the succession and position of the rock units of this part of Spain. Were you able to key in to all of them?

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CHAPTER OVERVIEW

30: (Case Study) Diversity in the marine fossil record

30.1: Diversity vs. abundance

30.2: Counting diversity in shells and bones

30.3: Marine cohorts - the evolutionary faunas

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30.1: Diversity vs. abundance



Figure 30.1.1: Look at the diversity and abundance of Ordovician fossils in this limestone. How would you count everything in this rock sample? (Photo by K. Layou)

The fossil record is full of different characters that provide rich details about the story of Earth's story–multiplying bacteria, scuttling trilobites, darting fish, stomping *T. rex*, flowering plants. Certain taxa follow one another through time and the vertical sequence of strata in a principle defined by William Smith as faunal succession. Major appearances and disappearances of vertebrate groups were used in part to create the first versions of the geologic time scale. But a deeper consideration of the fossil record can yield many fascinating questions about the history of life.

Counting taxa

First, there is the consideration of diversity–typically this is defined as the number of taxa present within some defined timeframe, geographic area, or stratigraphic layers. Let's break that down a bit. Recall a taxon is one level of the Linnean taxonomic hierarchy–domain, kingdom, phylum, class, order, family, genus, or species. Paleobiologists can ask questions at a variety of levels–when did the major phyla originate? Why do some genera have more species than others? All of this is dependent on the traits that place an organism within the hierarchy. At higher taxonomic levels, key traits are relatively easy to identify. For example, all arthropods have a segmented body, or all echinoderms exhibit 5-fold (pentameral) symmetry. But as you dig down deeper into the lower taxonomic levels, defining criteria become less clear, particularly at the species level.



Figure 30.1.2: Sexual dimorphism in spiders. Larger female on left, smaller male on right. (Creative Commons Attribution 2.5 Generic license.)

The species concept is a long-discussed, little-resolved aspect of biodiversity. There are numerous ways taxonomists, biologists, paleontologists, and evolutionary scientists have defined a species–we will only highlight a few of the most common here. The *biological species concept*, which most people are familiar with, is based on reproductively isolated populations of individuals–for example, fish in mountain lakes at high elevations. However, this definition is highly biased toward sexually reproducing eukaryotes. There are many eukaryotic organisms, and entire domains of prokaryotes that reproduce asexually. This definition is also problematic for the fossil record, as we cannot directly observe which fossil organisms were able to successfully lreproduce with each other. Paleobiologists, therefore, commonly apply the *morphological species concept*, using fossilizable characteristics of shape and form to define species. Some specimens in museums are often described as the "type" specimen of species because that fossil was used to formally describe the species. Can you think of any issues with this method of defining a species? There are definitely a few–consider the issue of sexual dimorphism (difference in males and females of a species, say in size, presence of antlers), or individual age (size of juveniles versus adults), or even ecophenotypic variation (variation in shell thickness in calm vs. rough currents). There is also the tendency of the taxonomist at work to be a "lumper" or a "splitter" in terms of recognizing

nuances of specimens. Some organize specimens into fewer, more inclusive groups (lumpers), while others default to defining a greater number of more exclusive groups (splitters). The *evolutionary species concept* focuses on a clade, or single lineages of a population and its descendants that are distinct from others. Again, this is challenging to directly apply to the fossil record, and will still be dependent on fossil morphological analysis and rRNA molecular clock data.

Counting individuals



Figure 30.1.3: Turritella sp. snails, Yorktown Formation, Virginia. Full snails are about 2.5 cm long from apex to aperture. (Photo by K. Layou)

Another consideration of understanding life in the fossil record is fossil abundance. While thinking about diversity provides a list of all the species present in a sampling unit, paleobiologists may also want to consider how many individuals of each species exist as well. There are many ways to collect abundance data and the method used is typically dependent on the hypothesis being tested. In some cases, relative abundance counts are appropriate. This is often done using the terms rare, common, and abundant, but these terms should be given some guideline amounts. For example, rare might be 1-3 specimens; common 4-10 specimens; abundant, more than ten specimens. The number of specimens using this method will vary from study to study though. In many cases, quantitative analyses of ecological paleocommunities require absolute specimen counts. Once again, depending on the fossils, the study needs to define how counts are made. For whole individuals, like a snail, it is pretty easy to say one shell equals one individual. But for bivalved shells like clams, or complex vertebrate skeletons, it may be impossible to say if two shell or bone fragments are from one or more individuals. Additionally, many arthropods like trilobites molt their exoskeletons as they grow, so one individual could leave behind many exoskeletons! Therefore, attention to details like the number of right vs. left valves or individual skulls may be helpful.

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30.2: Counting diversity in shells and bones

Sepkoski's Phanerozoic diversity curve

Fossil diversity throughout the Archean and Proterozoic is challenging to track because of the lack of hard parts for most of the organisms alive during these times. Once paleontologists recognized the shift to skeletonized animals during the Cambrian at the beginning of the Phanerozoic, many big questions bubbled to the top of paleobiological research. Has diversity constantly been increasing through time? If periods of mass extinction (intervals when many genera or even families die off simultaneously) exist, are there similar periods of diversification, or origination, when many species or higher taxa appear at once? How long is the average species around on Earth? To tackle some of these questions, Jack Sepkoski, a paleontologist at the University of Chicago, developed what is now one of the most famous graphs in the world of paleobiology, his Phanerozoic diversity curve.

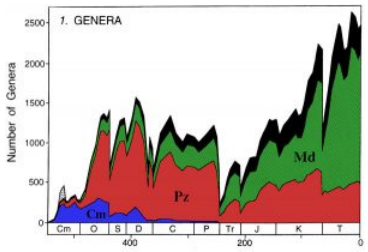


Figure 30.2.1: Sepkoski's (2002) Phanerozoic marine diversity curve. Total height of graph is total number of genera. Blue shading represents the Cambrian fauna, red shading is the Palozoic fauna, green shading is the Modern fauna, and black represents microfossils.

This graph shows the number of marine *genera* through the Phanerozoic. It is important to note this graph does have a subtle taphonomic bias to it, as it focuses on marine animals after the development of a shelly exoskeleton. When looking at the curve, several major features can be identified: 1) the Cambrian Explosion, a time of significant diversity increase through the Cambrian; 2) the Paleozoic Diversification and Plateau, a time of extensive radiation of invertebrate groups followed by relatively stable diversity from the Ordovician through Permian; and 3) the Cenozoic Rise, a time of exponential diversity growth from the Triassic to today. The five major mass extinctions are also quite noticeable as rapid, sharp declines in diversity at the end of the Ordovician, Devonian, Permian, Triassic, and Cretaceous.

After Sepkoski's work was published, a new question arose—is the diversity curve a true representation of biological diversification through time, or is the pattern influenced by taphonomy and the quality of fossil preservation? For example, some researchers suggested that the Cenozoic Rise is simply due to the "Pull of the Recent", meaning that there are more well-preserved, Cenozoic-aged rocks and sediments on Earth simply because there hasn't been enough time for them to go through the geologic wringer of the rock cycle yet. In the late 1990s, a large group of paleobiologists got together to begin developing the Paleobiology Database.

Checking the data: the Paleobiology Database

The Paleobiology Database (PBDB) is an online clearinghouse of occurrence data of fossils around the world for most of the Phanerozoic. Researchers scour published scientific literature for reports of fossil occurrences at described localities. The PBDB is an amazing resource because it includes taxonomic lists from a variety of spatial scales (sample within a bed, outcrop) but also typically pairs detailed sedimentary, stratigraphic, and taphonomic descriptions with the fossil occurrences. Want to know more about Jurassic bivalves in limestones within 200 miles of Paris? The PBDB has you covered. Since 1999, the PBDB has continued to grow as researchers have added data on marine microfossils, invertebrates, and vertebrates, terrestrial reptiles, birds, mammals, and plants, and more.

After a few years of PBDB data accumulation and analyses, a consensus group of paleontologists examined the data to see if Sepkoski's Phanerozoic diversity patterns were replicable. Even after correcting for variations in sample preservation, and chunking the geologic time scale into standardized 10 million year sections, these researchers were able to show the increase in taxonomic diversity during the Cenozoic is indeed a true biological signal in the fossil record.

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30.3: Marine cohorts - the evolutionary faunas

When Sepkoski was gathering his data, he recognized another fascinating set of patterns in his Phanerozoic diversity curve. Doing some basic analyses, he was able to identify groups of marine taxa with hard skeletal material that tend to originate at the same time, co-exist for a few geologic periods, and then go extinct at relatively the same time. He called these groups evolutionary faunas. The Phanerozoic diversity curve is divided into three of these evolutionary faunas: the Cambrian fauna, the Paleozoic fauna, and the Modern fauna. It is important to recognize that these groups of organisms are not mutually exclusive, and indeed, throughout much of the Phanerozoic, they coexist. The idea is that, typically, one of the groups is most dominant in ecosystems at particular points of time. Therefore, if you have a general sense of which fossil groups "hang out together" you will be able to make quick deductions about the age of the rocks containing the specimens.

The Cambrian fauna



Figure 30.3.1: Asaphiscus wheeleri, Cambrian trilobite from Utah. (Public domain image.)

The Cambrian fauna is responsible for the incredible increase in diversity seen in the fossil record, known as the Cambrian explosion. Some major taxa included in the Cambrian fauna are trilobites, inarticulate brachiopods, and early representatives of echinoderms. Trilobites are a superphylum (*Trilobitomorpha*) of arthropods that are a hallmark fossil of not only the Cambrian, but also much of the Paleozoic. Many of the taxa of the Cambrian fauna, particularly the trilobites, were hit hard by the end-Ordovician mass extinction, and trilobites are one of many taxa that go extinct during the end-Permian mass extinction.

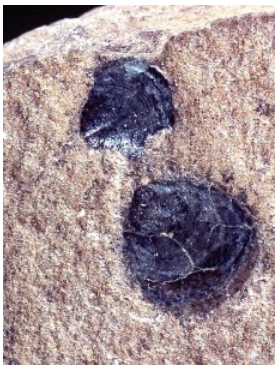


Figure 30.3.2: Inarticulate brachiopod *Lingula* sp. from the Cambrian Conasauga Formation of Alabama. (Photo courtesy of Encyclopedia of Alabama)

Inarticulate brachiopods are rather distinctive and typically fossilize with flattened, dark-colored shell with distinctive curved lines on the shell surface. Some genera, like *Lingula sp.* possessed a long, fleshy stalk to burrow into shallow marine sediments. While still alive today, inarticulate brachiopods are much less common than they were in the early Paleozoic.

The Paleozoic fauna



Figure 30.3.3: Diorama of a Late Ordovician seafloor with many representatives of the Paleozoic fauna. (Creative Commons Attribution 2.0 Generic license)

The Paleozoic fauna includes many of the taxa that are responsible for creating and maintaining the Paleozoic plateau of diversity from the Ordovician to the end of the Permian. These groups developed during the Great Ordovician Biodiversification Event (GOBE) as noted from the steep incline on the graph at the onset of the Ordovician. The GOBE is a summation of many increases in diversity among planktonic, benthic, and reef communities during the Ordovician. The Paleozoic benthic seas were dominated by a wide array of relatively abundant invertebrates, including trilobites, articulate brachiopods, bryozoans, reef building tabulate and rugose corals, crinoids, and blastoids. Interestingly, after initial bursts of species diversity of benthic organisms, there are also increases in the use of ecospace through the Paleozoic. Many of the taxa that evolved during the GOBE were filter feeding organisms, and despite diversification and abundance of phytoplankton and other prokaryotes and unicellular eukaryotes at the base of the food web, there was an increase in competition for resources. Therefore, as the Paleozoic continues, there is an increase in the number of organisms burrowing deeper under or growing taller (*e.g.*, stalked crinoids and blastoids) above the seafloor. These adaptations also aid in maintaining diversity as sediments eroding into epicontinental seas from tectonic activity increase turbidity. Fish also became dominant during this time, including sharks and other groups of armored fish. These armored fish were likely diversifying in part due to evolutionary selective pressures from other fish and large arthropods of the Paleozoic oceans, such as eurypterids.

The Modern fauna



Figure 30.3.4: “Duria Antiquior,” a famous watercolor by the geologist Henry de la Beche depicting life in ancient Dorset based on fossils found by Mary Anning. (Public domain image.)

The Modern fauna follows the ecological reset that occurs after the end-Permian mass extinction. Because so many taxa from the Cambrian and Paleozoic faunas were exterminated during the end-Permian event, many ecological niches were open and available for evolutionary innovations. A number of planktonic clades that are found in modern oceans evolved at this time, including silica-based diatoms and radiolarians, and calcite-based coccolithophores and foraminifera (note forams first evolved in the Cambrian). Modern (scleractinian) corals arose as the primary reef building organisms, along with echinoids (sea urchins and sand dollars) as the main benthic community echinoderms. Most importantly, however, was the rise of molluscs—bivalves, snails, and cephalopods diversified significantly and dominated benthic communities. Bivalves include a wide array of common shells you mind find on the beach or at the seafood market, such as clams, scallops, mussels, and oysters. Bivalves and snails also eventually worked their way into terrestrial ecosystems. Shelled (nautiloids, ammonoids, belemnoids) and unshelled (*e.g.*, octopus, squid) swimming cephalopods competed for food with marine vertebrate groups of the Modern fauna in the pelagic realm for fish and other swimming invertebrates. These vertebrate groups included a wide diversity of fish, swimming reptiles (*e.g.*, icthyosaurs, plesiosaurs, sea turtles), and swimming mammals (*e.g.*, whales and their ancestors, dolphins, sea lions).

To learn more about important fossil groups through time, see the Digital Atlas of Ancient Life from Paleontological Research Institute.

Further reading:

Sepkoski Jr., J.J. 2002. A Compendium of Fossil Marine Animal Genera. *Bulletins of American Paleontology* (363), 16 Sep 2002: 1-560. [Online]

Servais, T., Harper, D.A.T., Munnecke, A., Owen, A.W. and Sheehan, P.M. 2009. Understanding the Great Ordovician Biodiversification Event (GOBE): Influences of paleogeography, paleoclimate, or paleoecology? GSA Today 19(4): 4-10. doi: 10.1130/GSATG37A.1

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CHAPTER OVERVIEW

31: (Case Study) Reptile radiation

Learning Objectives

After reading this chapter, students should be able to:

- Define what the term ‘reptile’ really means
- Discuss the early evolution of amniotes
- Understand the path from synapsids to modern mammals
- Interpret early sauropsids and their skull shapes
- Differentiate between important diapsids, like archosaurs, squamata, pterosaurs, and sauropterygia

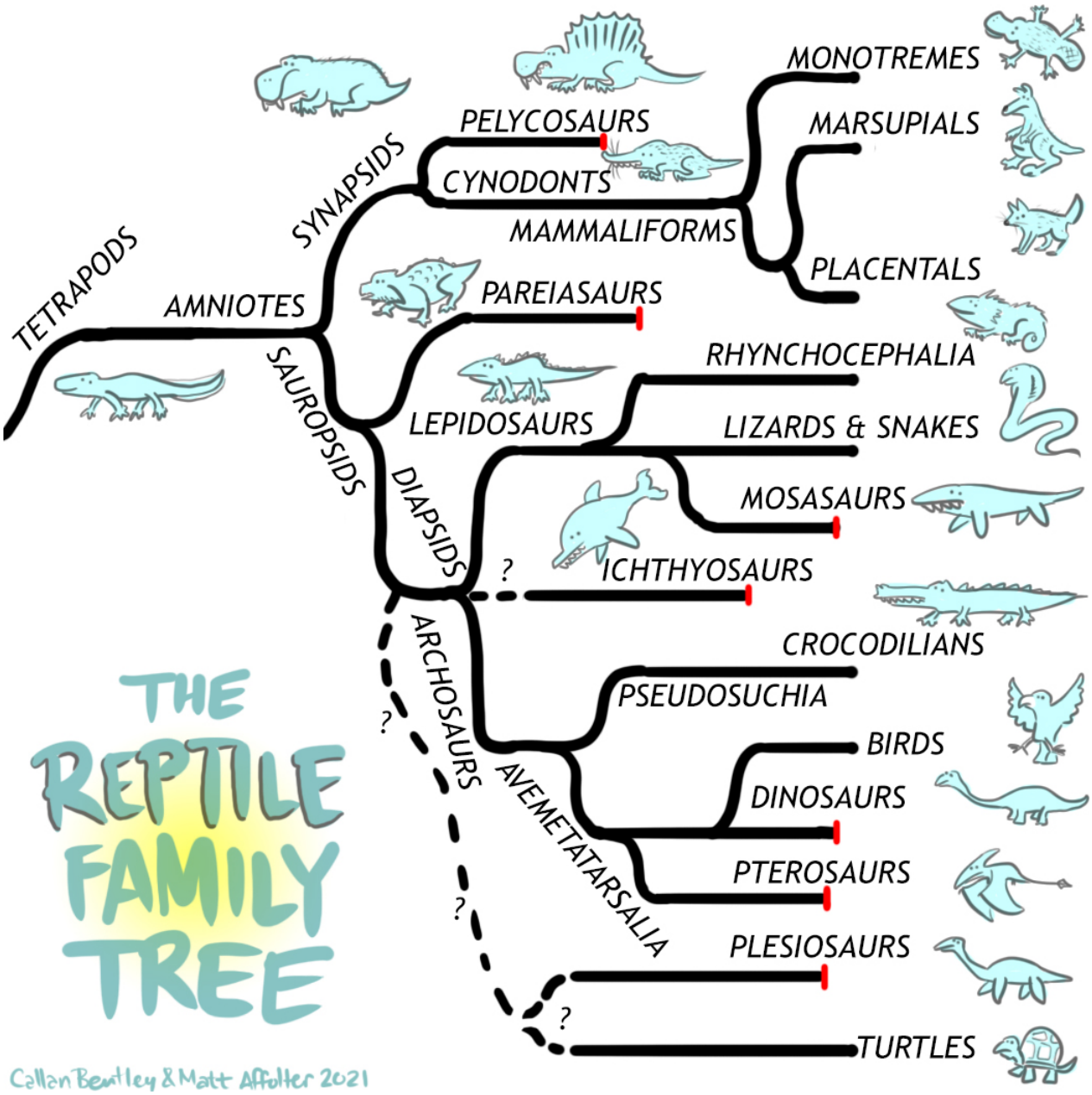


Figure 31.1: Cartoon of the reptile family tree, showing major amniote branches, mammal outgroups, and important and famous reptile lineages. Red lines at end of branches represents extinct groups. (Made by Callan Bentley and Matt Affolter.)

31.1: Reptile Radiation - Tetrapods Branch Out

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31.1: Reptile Radiation - Tetrapods Branch Out

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Introduction

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Figure 31.1.1: The famous Berlin Specimen of Archaeopteryx. (Picture by H. Raab (User: Vesta) via Wikimedia.)

The term ‘reptile’ seems straightforward, but when discussing evolutionary history, the term becomes much more complex. Reptile, as a biologic term, is actually paraphyletic. This means that all reptiles are not in the same simple family tree, or, some members of the group are excluded. For example, consider birds. The term bird includes the first bird (such as *Archaeopteryx* or a related animal) and all of the descendants, including all living birds. However, birds evolved from dinosaurs, which are reptiles. Does that mean birds are reptiles? This is one reason why the word reptile is problematic. For some researchers, birds are reptiles. For others, they are not.



Figure 31.1.2: Skeleton of Eryops, an early large carnivorous amphibian, from the Smithsonian’s National Natural History Museum. (Image by Daderot via Wikimedia.)

Another issue is where does the ‘reptile’ branch begin in the tree of life? To understand this, first one has to consider how reptiles evolved from non-reptile ancestors. The ancestors of the animals that eventually became reptiles were amphibians. During the mid-Paleozoic, the first fish to walk on land eventually turned into a common group of animals still alive today: amphibians. Amphibians were the first tetrapods (four-limbed animals) and they took advantage of the warm and tropical climate in the Devonian and Carboniferous and proliferated into many terrestrial niches with large and strange forms, like *Eyrops*.

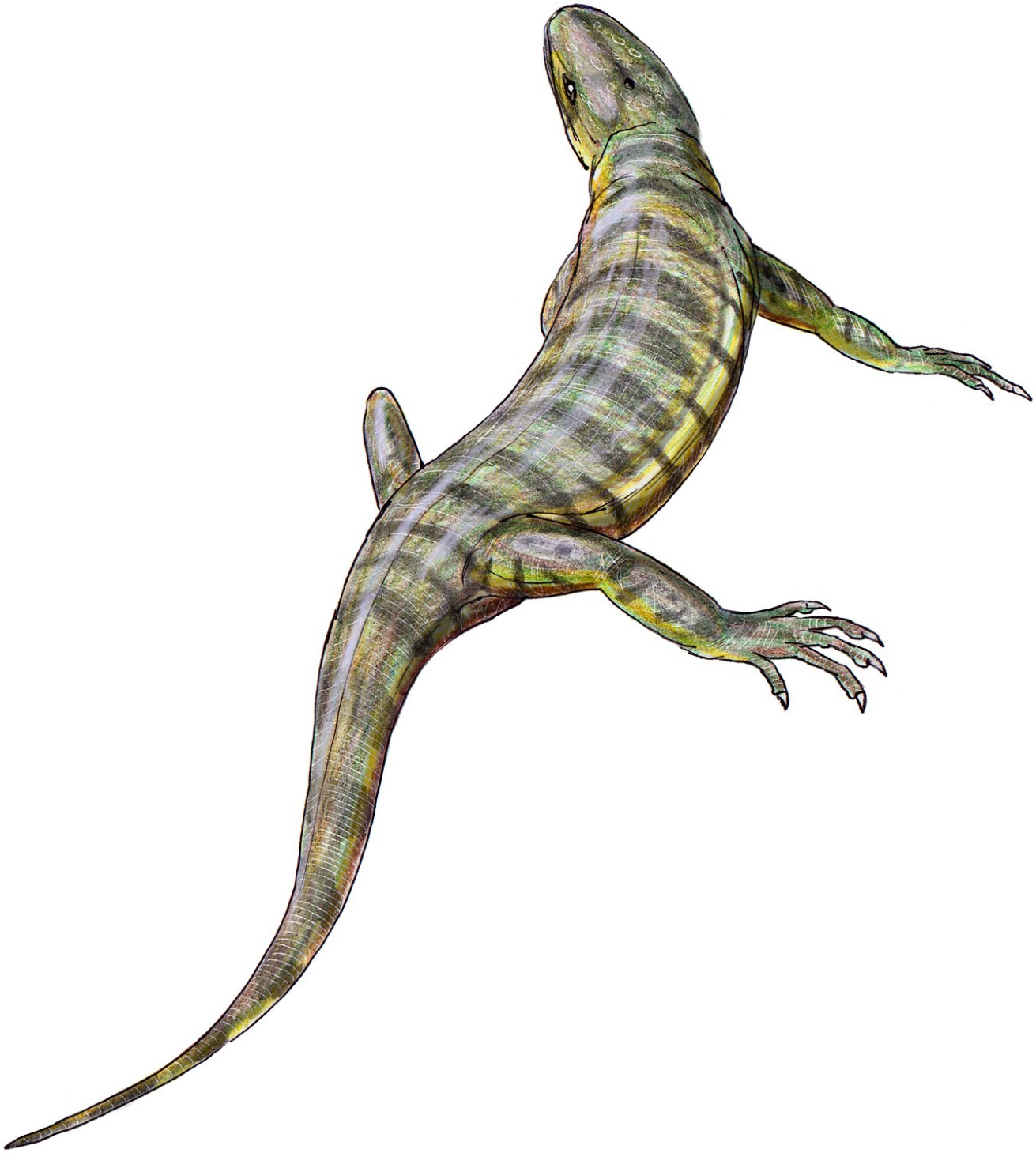


Figure 31.1.3: Casineria reconstruction (by ДиБгд via Wikimedia.)

Towards the end of the Paleozoic, as Pangea assembled and the climate dried out, amphibians were at a disadvantage. A new evolutionary leap occurred that allowed some animals to lay waterproof eggs on land, something their amphibian relatives were unable to do. This new group is the **amniotes**. Exactly which animal qualifies as the first amniote is a subject of debate, but *Casineria* or something similar is a good candidate. If this animal scurried in front of you right now, it probably would look like a modern reptile. However, it is actually not that closely related to modern lizards and it technically would not even be considered a true reptile.

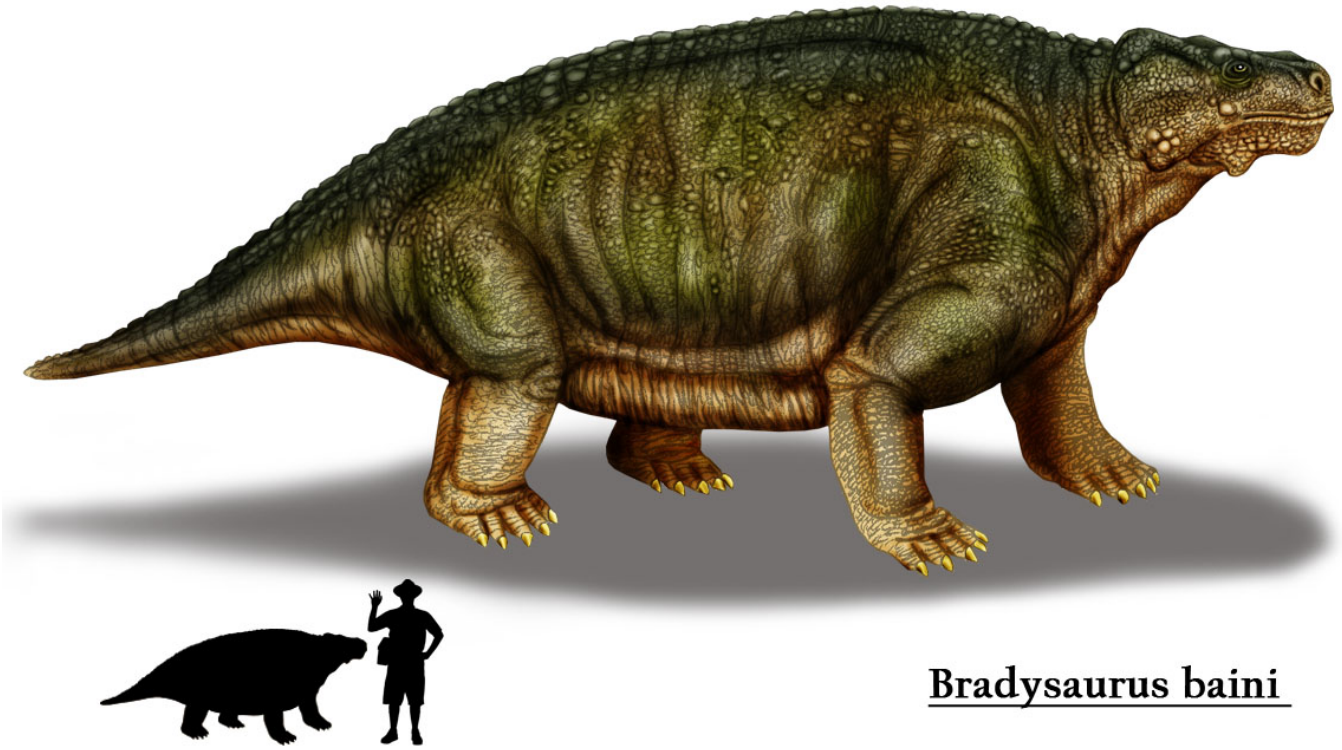


Figure 31.1.4: Bradysaurus, a parareptile (pareiasaur), with a size comparison to an adult human. (Image by Karkemish via Wikimedia.)

Very early in the amniote lineage, a crucial divergence occurred that has a profound effect on our modern world. One branch became the synapsids, the group that would eventually become mammals. The other branch, called the sauropsids, contain the non-mammal tetrapods: modern reptiles, birds, dinosaurs, and a large array of extinct lineages. Sauropsid, therefore, is potentially the best synonym for the word reptile. However, some scientists exclude parareptiles, an early extinct clade that was common at the end of the Paleozoic, from the sauropsid lineage.

Synapsids and mammals



Figure 31.1.5: A typical synapsid skull, this of Dimetrodon. Note the hole toward the base of the skull, here seen as the furthest left hole above the jaw line. (Photo from the University of Michigan Natural History Museum by Daderot via Wikimedia.)

The synapsids, the group of tetrapods which would eventually lead to mammals and humans, started out with a very reptilian-like body plan. Synapsid means fused arch, referring to the hole near the base of the skull. This hole in the synapsid skull is correlated to the temple in your skull. The first group of synapsids to diversify and specialize were the pelycosaurs. The pelycosaurs were the dominant and largest group of tetrapods in the Permian, and were formerly known as the mammal-like reptiles. However, since they are not true mammals or reptiles, this name has fallen out of favor.



Figure 31.1.6: Dimetrodon reconstruction. The posture is controversial, but here shown upright, based on possible Dimetrodon trackways. (Reconstruction by Max Bellomio via Wikimedia.)

The most famous member of this group is *Dimetrodon*. Its name means 'two-measure tooth' because it had different teeth for different uses. This was one of the earliest animals found in the fossil record that have differentiated teeth for different purposes, which is a hallmark of mammals. It is more closely related to humans than dinosaurs. Though it is often portrayed with dinosaurs in kids' books, it lived 40 million years before them. It lived in the Permian, and the first dinosaurs were Triassic. From track sites, it has been hypothesized that *Dimetrodon* could do a 'high walk' similar to what crocodiles can do, which is more efficient.

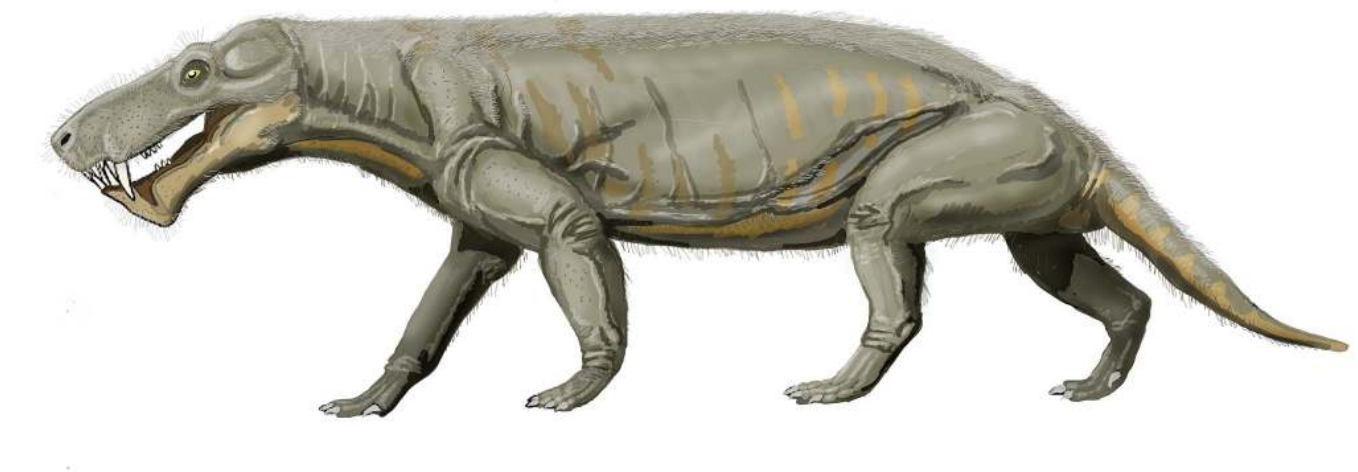


Figure 31.1.7: Gorgonops, a large Permian predatory therapsid. (Reconstruction by Dmitry Bogdanov via Wikimedia.)

One of the branches of the pelycosaurs evolved into a new form called therapsids. These replaced pelycosaurs as the dominant land animals toward the end of the Permian. The therapsids had several advantages over their less derived pelycosaur relatives, features more akin to their modern mammal descendants. They had a fully-upright stance, with their limbs always beneath the body. There is also circumstantial evidence that they were endothermic (commonly known as warm-blooded) and had at least some hair.

Scientists have found whisker pits in the skull, which is good evidence of hair. Examples of therapsids include the ferocious *Gorgonops*, one of the first saber-toothed animals.



Figure 31.1.8: Skull of Cynognathus. (Photo by Ghedoghedo via Wikimedia.)

Most of the therapsids went extinct at the end-Permian mass extinction, and even more went extinct at the end-Triassic mass extinction. One subgroup that survived both were the cynodonts. Meaning 'dog tooth,' cynodonts possessed even more mammalian traits.

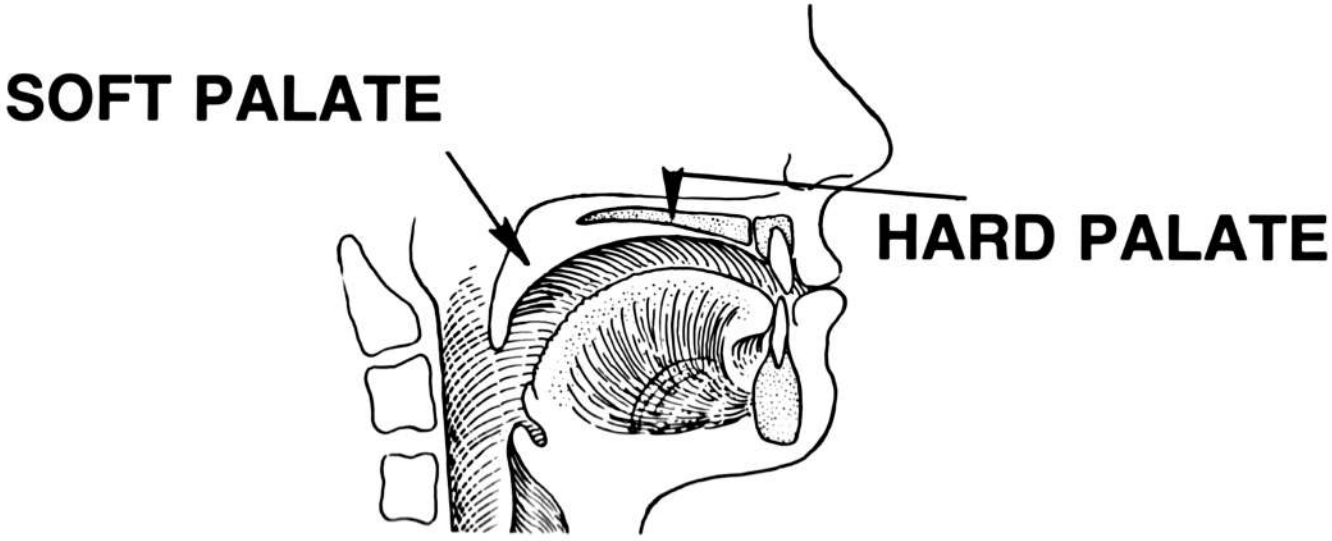


Figure 31.1.9: This cross-section of a human shows the hard palate and soft palate that all cynodonts and their relatives (including mammals and humans) have. The hard palate is partially made of bone, and the soft palate is not. The hard palate can be found in the fossil record because it can be fossilized more easily because it is made of bone. (Image by Pearson Scott Foresman via Wikimedia.)

An important but inconspicuous adaptation they evolved was a bony palate on the roof of their mouths. This has been attributed to a higher metabolism which matches its endothermy (warm-blooded), either by allowing breathing while chewing and/or allowing more complex chewing to help break down food.



Figure 31.1.10: Life restoration of Morganucodon (by FunkMonk via Wikimedia.)

By the mid Triassic, some cynodonts had evolved into true mammals. Technically, the first mammal is the last common ancestor of all living mammals, even if other animals (now extinct, called mammaliforms) had similar traits. Since many characteristic features of mammals (*e.g.* milk production (lactation), fur, neocortex of the brain) do not fossilize well, scientists commonly focus on jaw bones to identify and classify mammals.. Multiple bones found in the jaw of early synapsids migrated to form the bones of the inner ear. Since these ear bones are the smallest bones in mammals, they are also hard to find. However, the jaw is larger and more durable, so when the mandible fused into a single bone, it is a clear indication of a mammal or close ancestor. A common choice for the earliest animal to be considered a mammaliaform is *Morganucodon.*



Figure 31.1.11: The Short Beaked Echidna, one of four living echidna species and one of five living monotremes. (Photo by JJ Harrison in Mount Field National Park, Tasmania via Wikimedia.)

The earliest mammals are most likely related to monotremes, whose living examples are limited to the platypus and echidna. Monotremes are mammals, yet they lay eggs and do not have nipples, and instead lactate by 'sweating' milk from their chest. They also have a single orifice used for urination, defecation, and reproduction, which is what the name 'monotreme' refers to. These more primitive features also support the evidence of monotremes being the first mammals. The details of the evolution of the mammal branches of the family tree, like marsupials (Metatheria) and placentals (Eutheria), is complicated. Comparing the genomes of the living groups put the divergence much further back than the fossil record does. Scientists can do this with molecular clock studies. By looking at the amount of mutations in related species, experts can estimate how long it has been since they evolutionarily diverged. Adding to the complexity is the fact that the earliest fossils of mammals are controversial as to their taxonomic placement. For example, the earliest proposed placental mammal (Juramaia) is Jurassic (Zou et al., 2011), and yet the earliest known monotreme (Teinolophus) is Cretaceous (Rich et al., 2001), even though placental mammals are more derived and should appear after monotremes.

Early Reptiles & Synapsids: Did I Get It?

? Exercise 31.1.1

Why can birds be considered reptiles?

a. They are as old as reptiles

b. Birds and reptiles both evolved from amphibians

c. They evolved from reptiles

d. The have scales on their legs

Answer

c. They evolved from reptiles

Exercise 31.1.2

Synapsids eventually evolved into which group of modern animals?

a. Birds

b. Reptiles

c. Turtles

d. Mammals

Answer

d. Mammals

Exercise 31.1.3

Which group of modern mammals are most primitive, and most like early members of the linage?

a. Placentals

b. Monotremes

c. Marsupials

d. Canids

Answer

b. Monotremes

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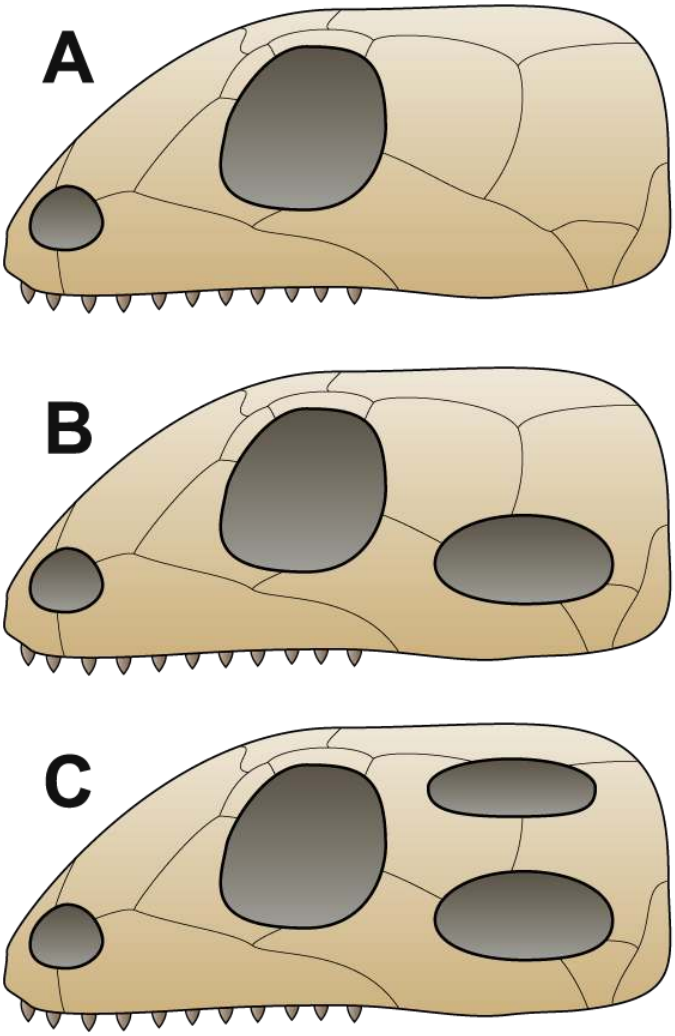


Figure 31.1.12: Major amniote skull types: A = anapsid, B = synapsid, C = diapsid. (Image by Preto(m) via Wikimedia.)

The amniotes had two major divisions in their early evolution: the synapsids, as discussed earlier, and the **sauropsids**, the branch that would eventually evolve into reptiles, dinosaurs, and birds. These early amniotes have been traditionally separated based on specific features of the skull. Two major groups make up the sauropsid branch: the anapsids and diapsids. Anapsids are reptile skulls with no openings in the rear of the skull, while diapsids have two such openings. As one might expect, the details of which of these three skull types was first and existed in the ancestral amniote (thus leading to the other types) is still controversial.



Figure 31.1.13: Mesosaurus reconstruction drawn (by Nobu Tamura via Wikimedia.)

Anapsids are a clade that is roughly equivalent to the parareptiles, a group mentioned before in the section on early reptiles. Examples include the pariasaurs and the mesosaurs, the first tetrapod group to evolve an aquatic lifestyle since the tetrapods left the water in the first place. *Mesosaurus* fossils were also used by Alfred Wegener to reconstruct Pangea, since they lived in terrestrial aquatic habitats during the Permian. Recent analysis of anapsids have completely removed anapsids as a valid clade, with even mesosaurs showing skull openings and skull variability within the group. This means that traditional anapsid skulls may be more of a result of convergent evolution than inherited traits. Humans, for example, technically have an ‘anapsid’ skull since our only openings are our orbits. Regardless of where the exact positioning of parareptiles or mesosaurs is, it is inferred that this lineage went extinct. The only possible exception is turtles, which will be discussed later.

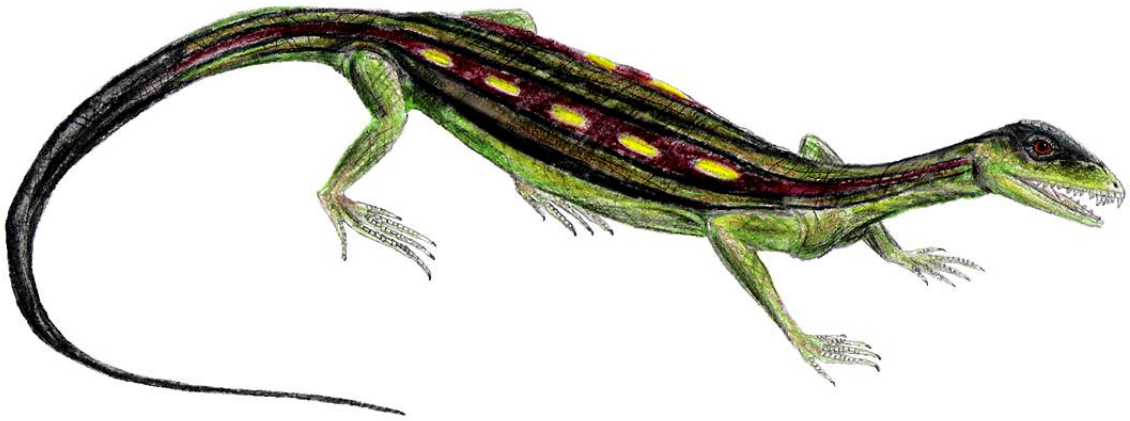


Figure 31.1.14: Araeoscelis reconstruction (by Smokeybjb via Wikimedia.)

The other major branch of the sauropsid family tree is the diapsids. Their name means ‘double arch,’ referring to the two holes toward the back of the skull. Diapsids are true reptiles, and the theoretical most recent common ancestor of all living reptiles is almost certainly a diapsid. The earliest diapsids were the araeoscelidans, which resembled modern lizards except for their elongated limbs.



Figure 31.1.15: Henry the Tuatara photographed at Invercargill, New Zealand (by KeresH via Wikimedia.)

A sister group to the araeoscelidans branched into two major divisions: the lepidosaurs and archosaurs. The lepidosaurs (‘scaled lizards’) are famous for their overlapping scales. Within the lepidosaurs, Rhynchocephalia (’beak heads’) are interesting in that they were highly diverse and common across the planet in the Mesozoic. Today, however, only one species has survived: the tuatara, a lizard endemic to small islands off the coast of New Zealand.



Figure 31.1.16: Mosasaurus skeleton displayed at the Natural History Museum of Maastricht, Netherlands. (Photograph by Ghedoghedo via Wikimedia.)

Squamata is the other group within lepidosaurs. First evolving in the Triassic, Squamata contains nearly all modern lizards and snakes, which is the most diverse order of land vertebrates that exists on Earth. One of the more famous groups of animals that are extinct within Squamata are the mosasaurs. They were the apex predators of the Cretaceous seas and became the largest reptiles ever found in the oceans of Earth (at over 50 feet long), but went extinct with the non-avian dinosaurs. Their closest living relatives are snakes.

Archosaurs



Figure 31.1.17: Protorosaurus reconstruction (by Nobu Tamura via Wikimedia.)

Archosaurs are the other major division within the diapsids. Their evolutionary history is complicated and has many extinct groups of importance besides the dinosaurs. The name archosaur means ‘ruling lizard,’ and throughout the Mesozoic, they were the dominant land animals. As with other groups, the definition of what an archosaur is and what is a close relative depends on the use

of the word. Traditionally, archosaur-like reptiles (*e.g.* not lepidosaur-like reptiles) could be considered archosaurs. However, the modern phylogenetic definition of an archosaur is the last common ancestor between crocodiles and birds. This means these archosaur-like reptiles are now considered archosauromorphs or archosauriforms. Which organisms were the first archosaurs is up for debate, but it may have been an animal like *Protosaurus*, a six-foot long lizard from the Permian.



Figure 31.1.18: Tanystropheus, and early archosaur, had a long neck like sauropod dinosaurs which came much later. (Photo taken at the Paleontology Museum of Zurich by Ghedoghedo via Wikimedia.)

In the Triassic, in the open landscape and ecospace following the Great Dying (AKA the end Permian Mass Extinction), the archosaurs diversified greatly. Many different lines of archosauromorphs that originated soon went extinct at the next extinction, the end-Triassic extinction. Interestingly, there was a large amount of convergent evolution with future forms that became more well known. An early example is *Tanystropheus*, which had a long neck like sauropod dinosaurs.

Why Triassic Animals Were Just the Weirdest



Figure 31.1.19: Drapanosaurus, the chameleon/bird-like reptile, had a large claw on its index finger and its tail! (Reconstruction by Nobu Tamura via Wikimedia.)

Another very strange and disputed early archosauromorph group are the drapanosaurs, which were tree-dwelling lizards that had a bird-like head, chameleon-like body, and a claw on the end of their prehensile tails.

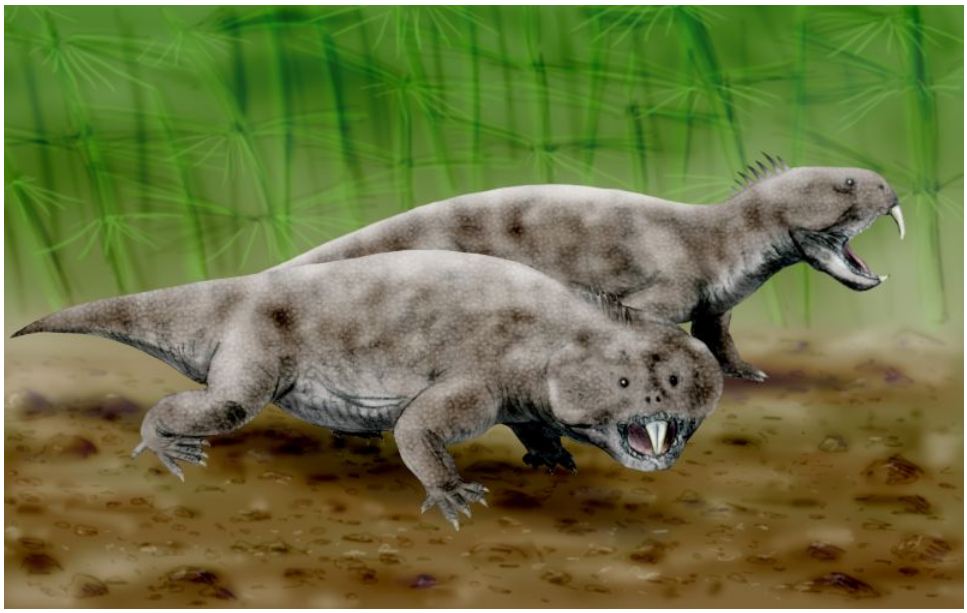


Figure 31.1.20: Hyperodapedon, a common rhynchosaur. (Drawing by Nobu Tamura via Wikimedia.)

Less in dispute, as far as being an early archosauromorph, are the rhynchosaurs. They had short, stocky bodies and were herbivores. They had the first complex chewing mechanism found on Earth, with rodent-like 'incisors' in the front of their mouths and large cheeks with densely-packed tooth batteries similar to a cow.

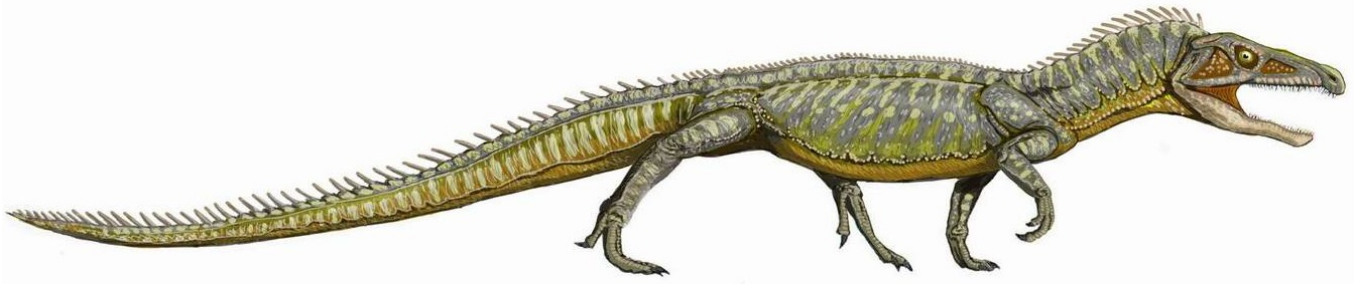


Figure 31.1.21: Proterosuchus reconstruction (by ДиБгд via Wikimedia.)

Within archosauriformes (the group more derived than archosauromorphs), but not true archosaurs, the most famous group would be Proterosuchidae. The crocodile-shaped animals had strange downward hooks on their upper jaws. Even though they are crocodile shaped, they are still only distantly related to modern crocodiles, and most likely looked similar due to them living in similar semi-aquatic habitats.

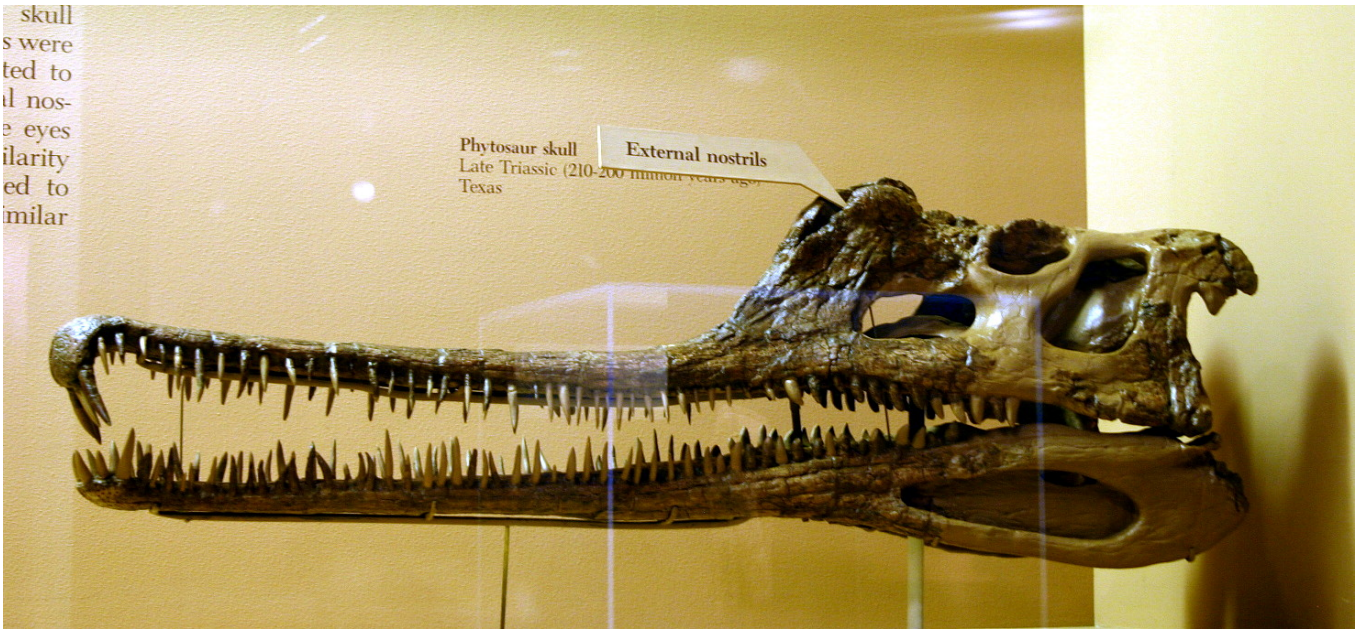


Figure 31.1.22: The skull of phytosaur. Note the nostril above the eye on the top of the skull. (Photo by Ryan Somma.)

Archosauria proper has two major divisions, nicknamed after the living animals that came from that group. Pseudosuchia, a name that means 'false crocodiles,' actually do contain modern crocodiles and therefore are called 'crocodile-line archosaurs.' Avemetatarsalia, which refers to the bird-like foot the animals have, are called 'bird-line archosaurs.' One important group which may be a sister group to both of these lines or may be on the Pseudosuchia line are the phytosaurs. They were also crocodilian in shape, but had noses in the center of their skull instead of at the end of the snout. Despite their name ('plant lizards'), they were actually apex predators in the Triassic.

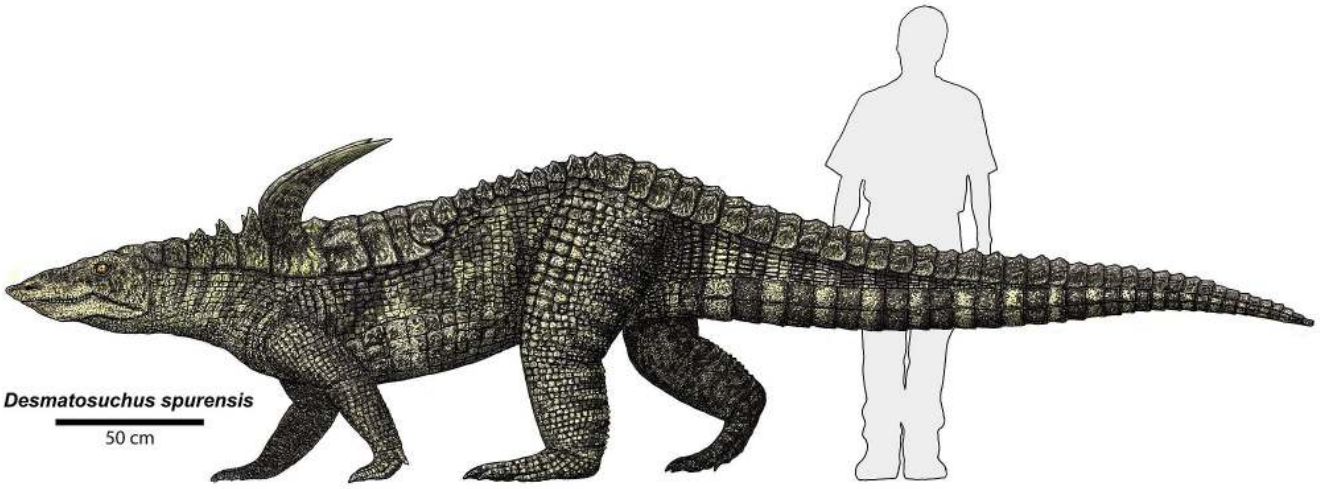


Figure 31.1.23: Desmatosuchus, an aetosaur with thick armor and spikes and a pig-like nose.

Within Pseudosuchia, there are many different groups, but one of the more interesting are the aetosaurs, another classic example of convergent evolution. They had snouts like pigs which aided in their terrestrial herbivore lifestyle. They were also covered in armor, superficially similar to the ankylosaurs which would come along more than 50 million years later.

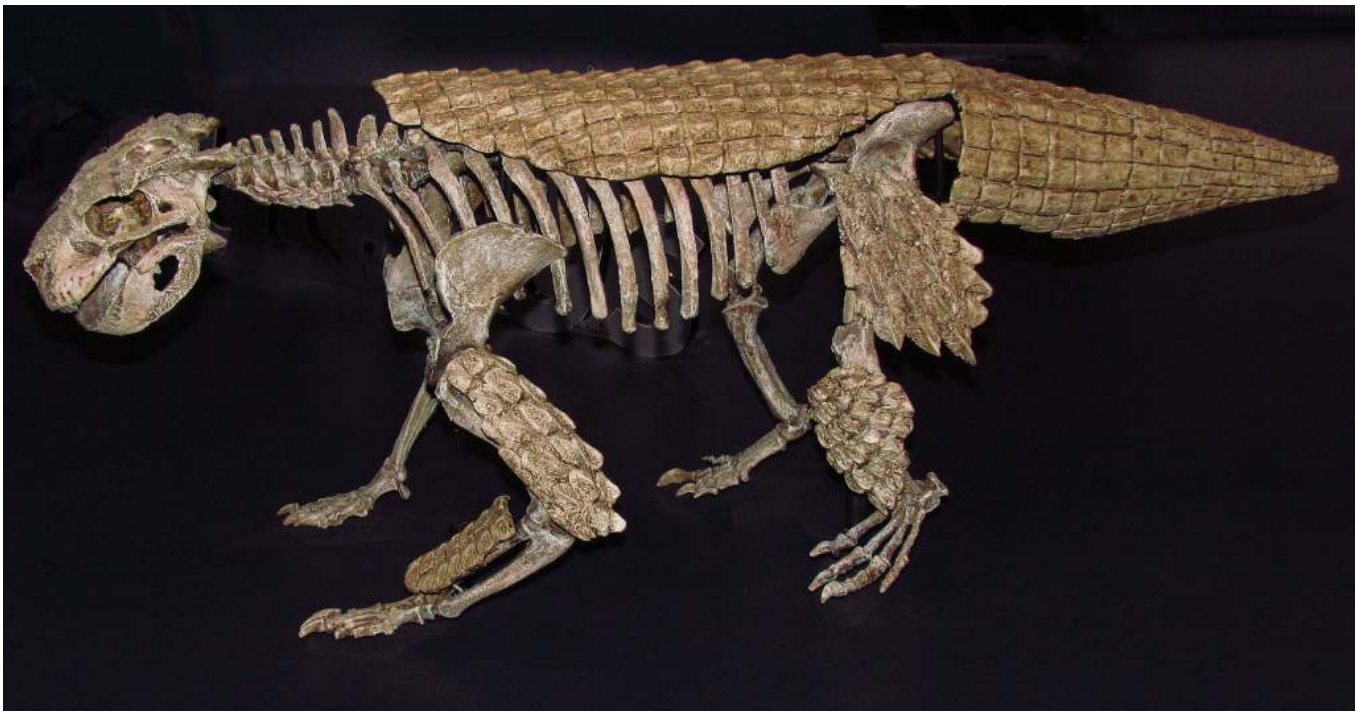


Figure 31.1.24: Simosuchus, an example of a strange early croc. (Photo of a cast found in the Royal Ontario Museum by D. Gordon E. Robertson via Wikimedia.)

Actual crocodiles, or at least the group Crocodylomorpha, first appeared in the late Triassic. Even within this group, there is great diversity of forms and lifestyles. This includes fully aquatic crocs, fully terrestrial crocs, crocs with hooves, armored crocs, and even vegetarian crocs. Many of these more unusual forms went extinct with the dinosaurs, leaving the crocs we know today. The first modern crocodilians appeared toward the end of the Cretaceous.

Sauropsids, Diapsids, Archosaurs, and Pseudosuchia: Did I Get It?

? Exercise 31.1.1

Which skull type is found in primitive groups like parareptiles and mesosaurs?

a. Synapsid

b. Euryapsid

c. Diapsid

d. Anapsid

Answer

d. Anapsid

? Exercise 31.1.2

Which animals are most closely related to the aquatic Mosasaurs?

a. Snakes

b. Tuataras

c. Plesiosaurs

d. Crocodiles

Answer

a. Snakes

? Exercise 31.1.3

Many archosaurs exhibit _____, which means distantly related animals can show similar forms in similar environments.

a. adaptive evolution

b. transform evolution

c. divergent evolution

d. convergent evolution

Answer

d. convergent evolution

Avemetatarsalia

Pterosaurs

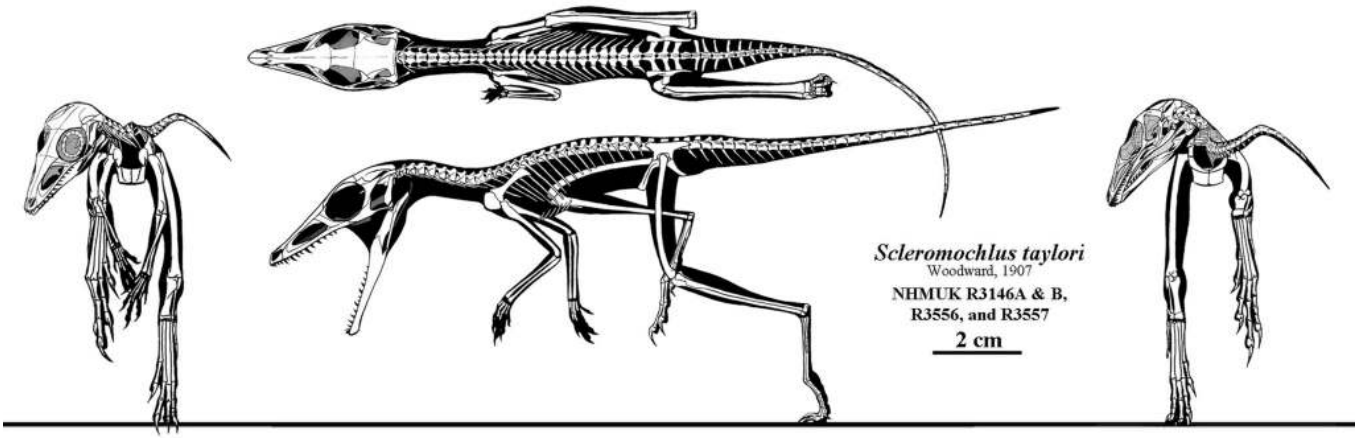


Figure 31.1.26: Scleromochlus, a possible early ancestor of pterosaurs. (Drawing by Jaime A. Headden via Wikimedia.)

Within the bird-line archosaurs, Avemetatarsalia, the most famous early group is the pterosaurs. In fact, a similar term that is used is almost as a synonym for Avemetatarsalia is Ornithodira, which is defined as the clade that contains the most recent ancestor of all dinosaurs and pterosaurs. These flying reptiles (the first vertebrates ever to evolve powered flight) have an even more complex and mysterious evolutionary history than other groups due to the thin, fragile nature of their skeletons. One possible ancestor to the pterosaurs is *Scleromochlus*, a Triassic animal that spent its time hopping. If this animal is not an early pterosaur, it may be an early avemetatarsalian, or even an early archosauramorph.



Figure 31.1.27: Photograph of a well-preserved Rhamphorhynchus fossil from the Natural History Museum in Brussels (by M0tty via Wikimedia.)

All pterosaurs had a long fifth finger that extended to form their wing. The pterosaurs have two major groups within their clade. The rhamphorhynchoids had toothed beaks and long tails. Though fish are traditionally seen to be the main diet of these flyers, some had sharp teeth for tearing terrestrial animal flesh. The namesake of the group, *Rhamphorhynchus*, also had a rudder-like tail.

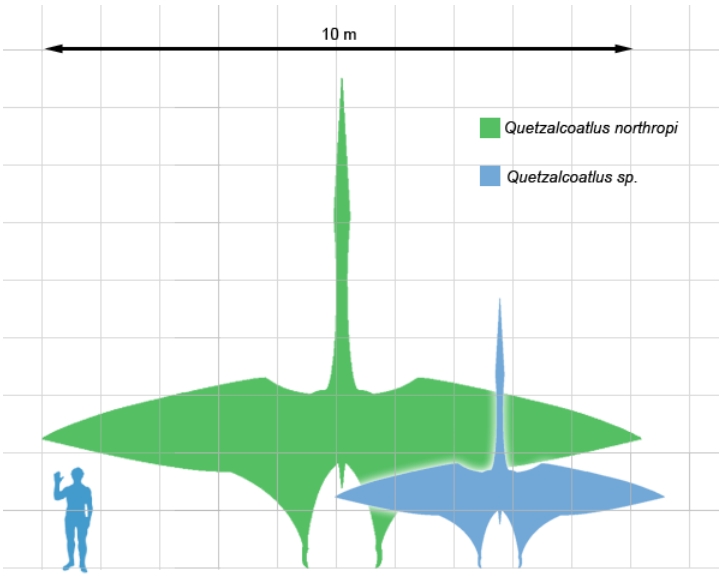


Figure 31.1.28: Size comparison between two Quetzalcoatlus species and a human. (Illustration by Matt Martyniuk (Dinoguy2), Mark Witton and Darren Naish via Wikimedia.)

The other group of pterosaurs are the pterodactyloids. This name derives from *Pterodactylus*, a famous member of the pterodactyloids. As an aside, the term ‘pterodactyl’ is commonly used in place of pterosaur to refer to all flying reptiles, though that is incorrect. It should only refer to *Pterodactylus* and nothing else. *Pteranodon*, another famous pterodactyloid, gives the name ‘pterosaur’ to the overall group. Pterodactyloids are known for a lack of tails and more elaborate crests on their skulls. The most

remarkable member of this family is *Quetzalcoatlus*. Named after the Aztec feathered serpent god, it is the largest animal ever known to fly at a wingspan of 36 feet!

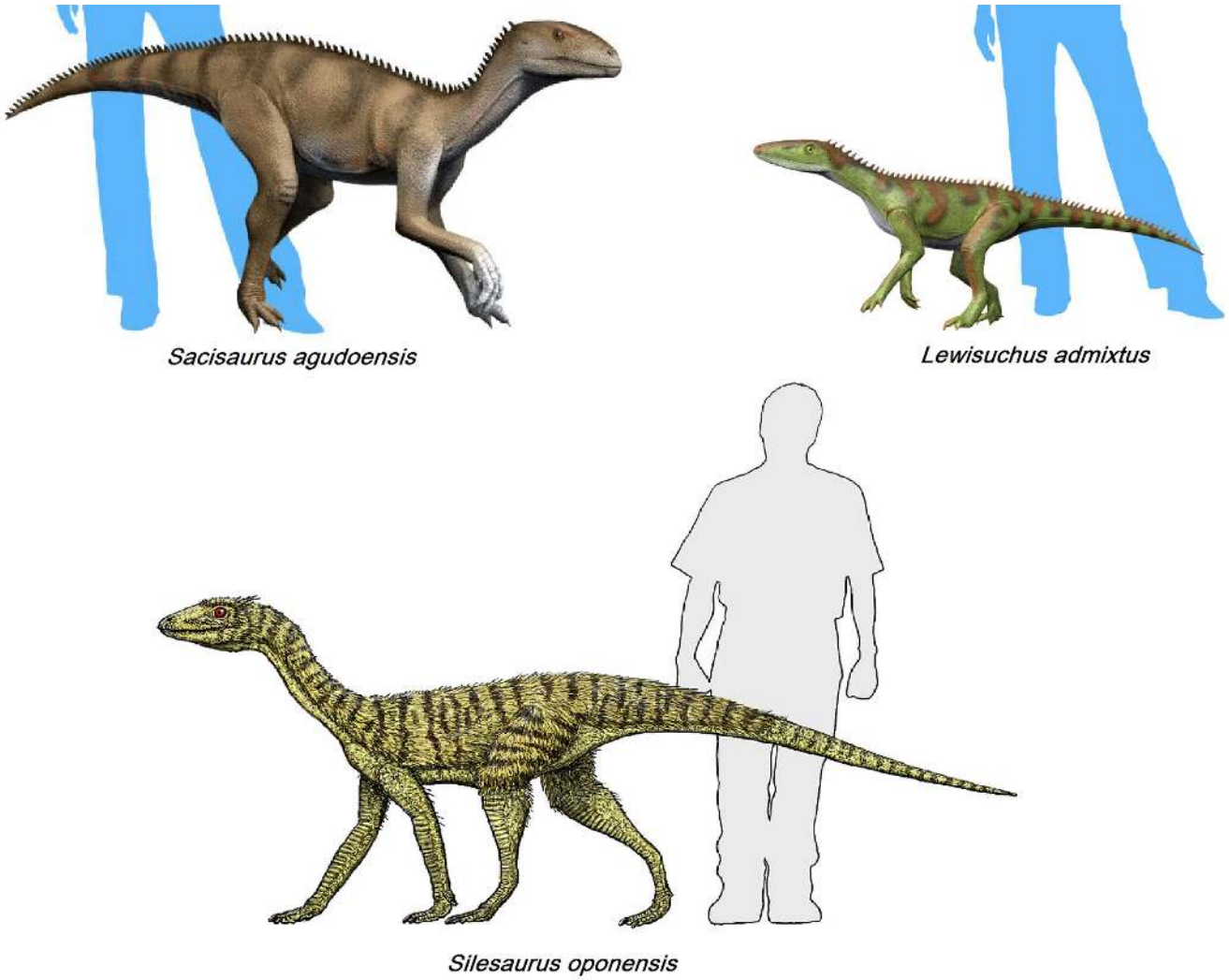


Figure 31.1.29: Different sileasaurs compared to humans. The interpretation of them being quadrupedal verses bipedal is debated. (Drawings by Nobu Tamura and US National Park Service via Wikimedia.)

The more evolved bird-line archosaurs are generally called dinosauromorphs. The animals that are dinosauromorphs but not dinosaurs are generally small carnivores. To a novice, they would certainly look like dinosaurs. The group that is closest to dinosaurs are the silesaurs. They are currently classified as a sister group to Dinosauria. For more details on dinosaurs, go to Dinosaur Classification.

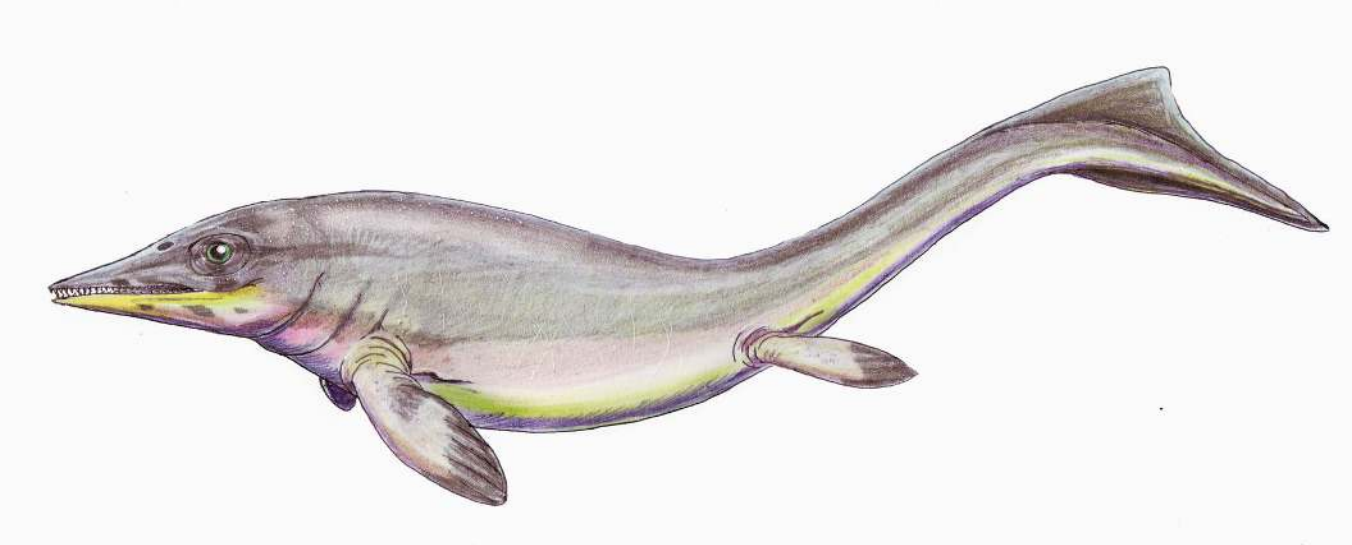


Figure 31.1.30: Grippa, an early ichthyosaur. (Drawing by Dmitry Bogdanov via Wikimedia.)

One group that has an uncertain placement within the diapsid lineage (but are still generally considered diapsids) are the ichthyosaurs. These marine reptiles were an early branch on the archosaur tree, because they were already prolific in the early Triassic. Part of the difficulty in their placement phylogenetically comes from the lack of transitional fossils from land-dwelling ancestors to fully aquatic ichthyosaurs. Because of this uncertainty, various paleontologists have placed them as basal diapsids, as lepidosaurs, or as archosaurs. There is no clear consensus.



Figure 31.1.31: Ophthalmosaurus, an advanced ichthyosaur, had the largest eyes (as a proportion of body size) ever seen in an animal. Its eyes were 9 inches long on a 13 foot animal, meaning the eye took up almost 6% of the body length! These large eyes are most likely used for hunting in deep, dark waters. (Picture by Captmondo via Wikimedia.)

Ichthyosaurs, much like the creatures whom they convergently share a body plan – sharks and dolphins – were fast swimmers that hunted fish and other ocean animals, like ammonites. They even gave live birth, with a remarkable fossil preserving this process. They were some of the top predators in the ocean and filled many niches. For example, many had large eyes, suggesting they dove to great depths to hunt. Even with all their success, by the mid-Cretaceous, they were largely extinct, possibly out-competed by the mosasaurs.

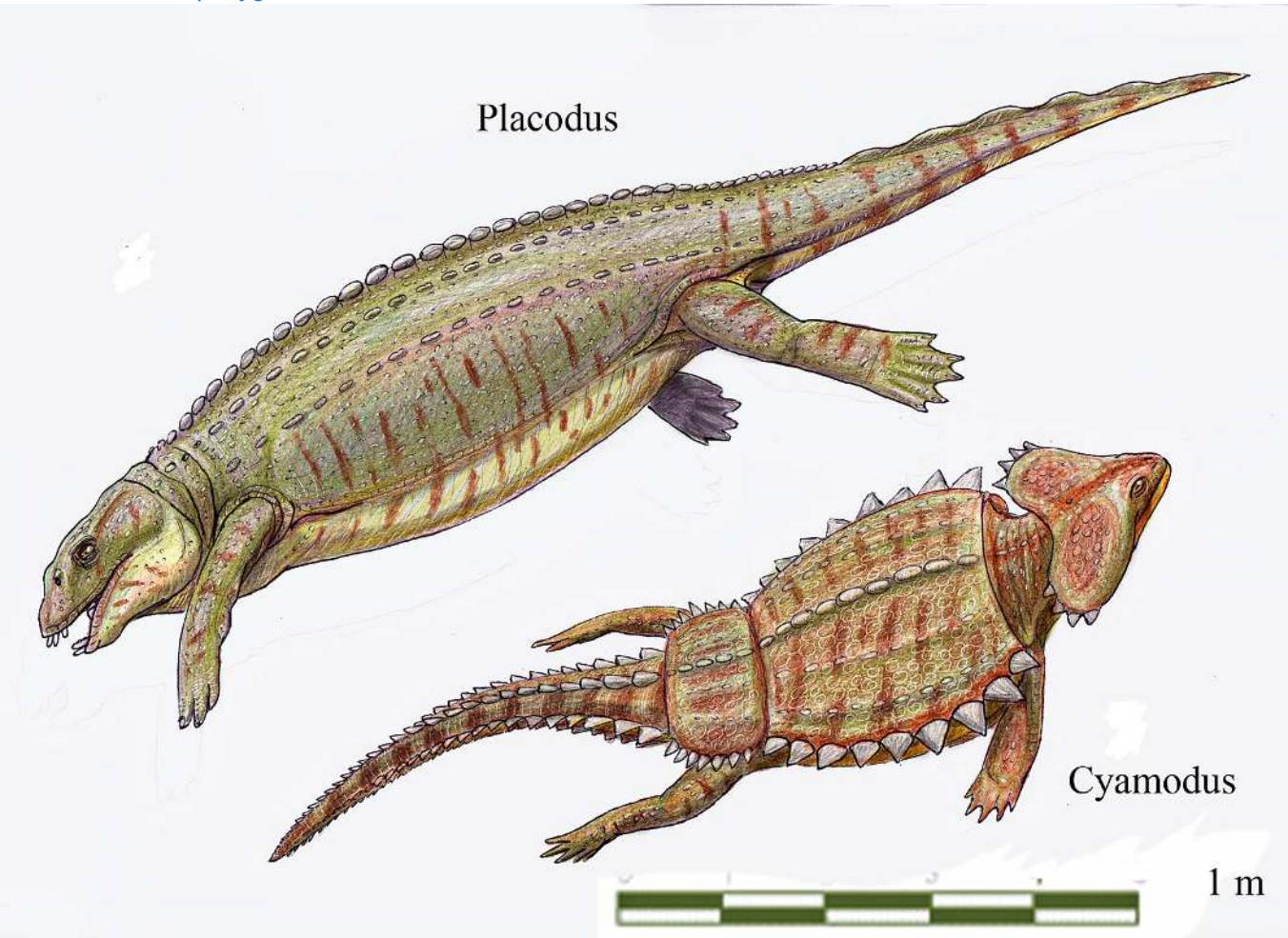
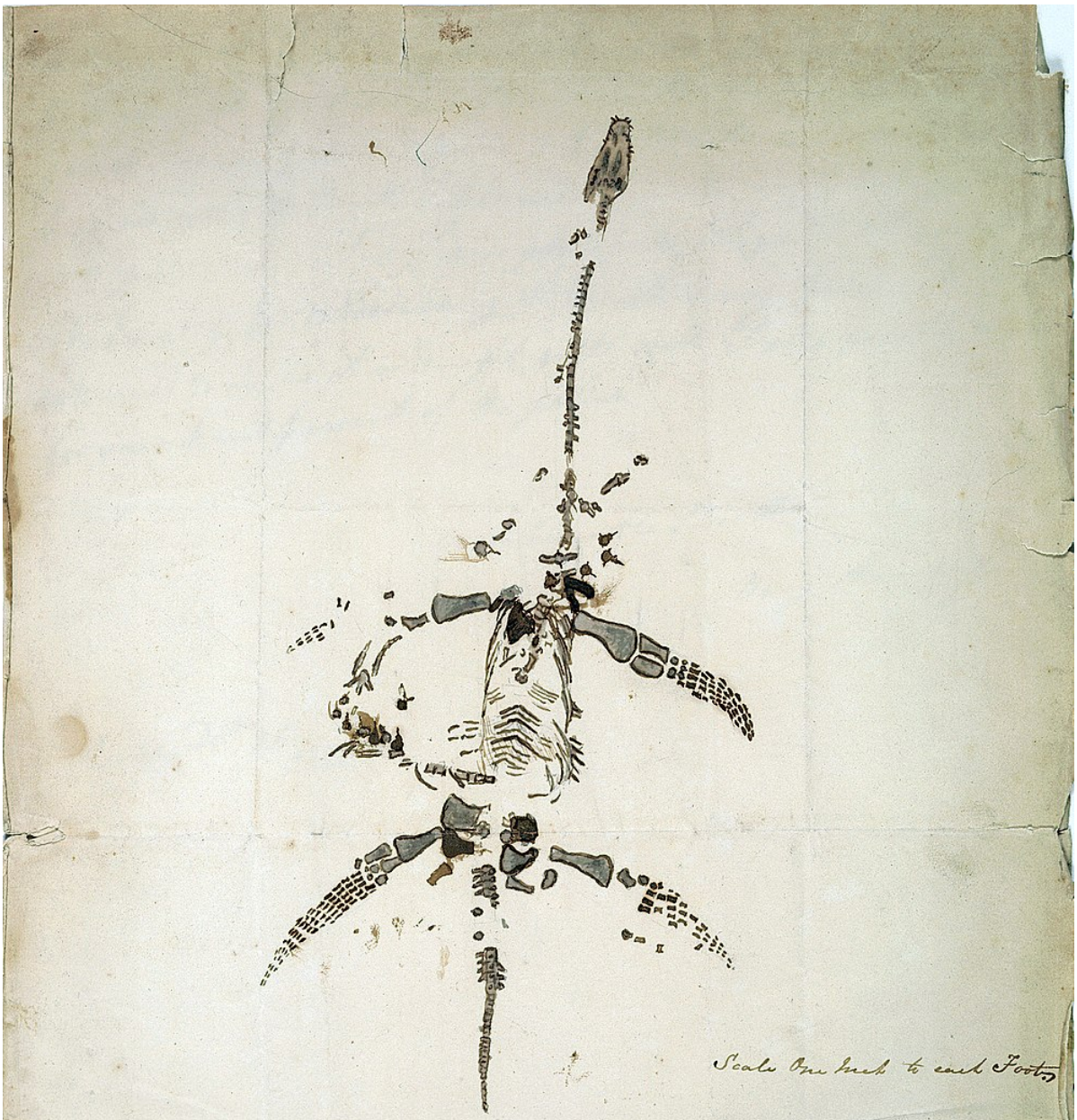


Figure 31.1.32: Examples of placodonts, drawn (by DiBgd via Wikimedia.)

There is one other major marine reptile group and one other major living reptile group that has not been covered yet in this case study, and it turns out they are related. These groups are plesiosaurs and turtles. However different these groups may seem, the current established phylogeny has them related. One early (Triassic) group that shows some crossover between the two were the placodonts. They were marine with tough jaws and body armor, in some cases being full shells.



Scale One Inch to each Foot

Sir

I have endeavoured in a rough sketch to give you some idea of what it is like. Sir you understood me right in thinking that I said it was the supposed plesiosaurus, but its remarkable long neck and small head, shows that it does not in the least resemble their congeners; in its analogy to the Ichthyosaurus, it is large and heavy but one thing I may venture to assure you it is the first and only one discovered in Europe, Colonel Birch offered one hundred [illegible] but your letter came one days post before [illegible]

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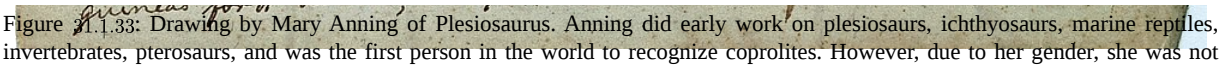


Figure 31.1.33: Drawing by Mary Anning of Plesiosaurus. Anning did early work on plesiosaurs, ichthyosaurs, marine reptiles, invertebrates, pterosaurs, and was the first person in the world to recognize coprolites. However, due to her gender, she was not taken seriously as a scientist. She not only was excluded from publishing her work but had her work published by others without credit. Only after her death did some deserved recognition begin to emerge. This includes the alleged inspiration for the tongue-twister 'she sells sea shells by the sea shore' and being named one of the top 10 female scientists in Britain's history by the Royal Society.

The group that includes the placodonts and plesiosaurs is called Sauroptergia, which means 'lizard flipper.' Along with the plesiosaurs, other groups like the nothosaurs and pachypleurosaurs lived along side the placodonts in the Triassic. At the end of the Triassic, all of the groups went extinct except for the branch of this group became the plesiosaurs, the only members of their superorder to survive the Triassic mass extinction. In the Jurassic and Cretaceous, they became very diverse and widespread in the oceans, alongside the ichthyosaurs (mostly earlier) and mosasaurs (mostly later). Their body plan featured a long neck and small head for grabbing fish, except for the subgroup known as pliosaurs, which had larger jaws and shorter necks. Like the other large marine reptiles in the Mesozoic, they went extinct at the same time as the dinosaurs 66 million years ago. They are probably most famous today for being the supposed creature known as the Loch Ness Monster. There are many reasons why to doubt this, including early evidence that turned out to be a hoax, but scientists have even attacked the specific reasons why it could not be a plesiosaur.

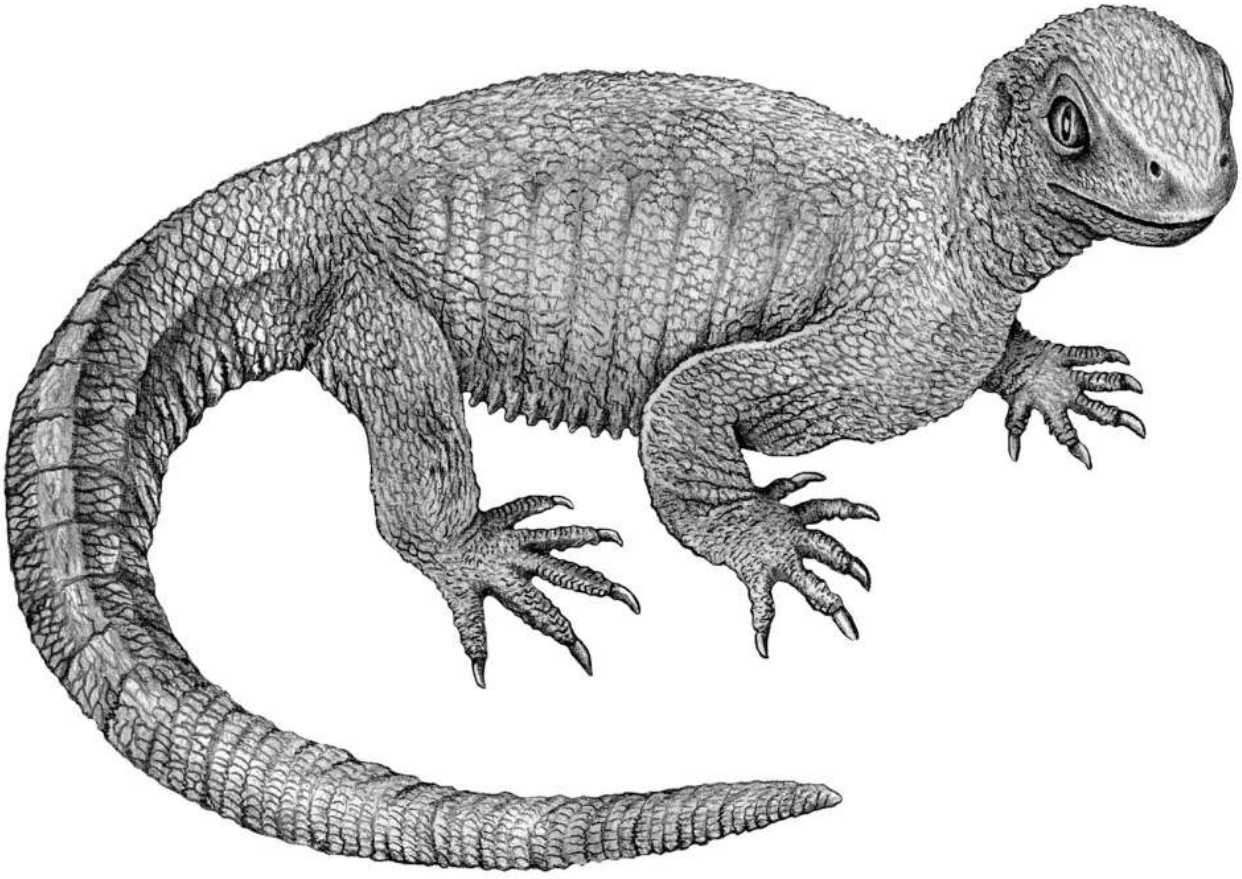


Figure 31.1.34: Pappochelys, possibly the first turtle. (Drawing by Rainer Schoch via Wikimedia.)

Knowing where turtles fit into the tetrapod family tree can help with the placement of Sauropterygia, but that is no easy task. Traditionally, turtles were considered anapsids. This comes from the fact that turtles show no hole openings in their fortified skulls. Also, turtles excrete urine more similarly to mammals, potentially pointing to a very basal position in their split from other reptiles, which would match anapsids well. However, early turtles like *Pappochelys* have been shown to be diapsids. They started out as land animals, potentially using their reinforced torsos to help with digging. Millions of years passed before they diversified into the multitude of forms seen today, including venturing back into the sea as the largest living marine reptiles. Genetic evidence, comparing them to living archosaurs and lepidosaurs generally favors them as more closely related to archosaurs, but the debate

certainly is still open. Some genetic clock studies place the divergence between turtles and other tetrapods all the way back in the mid Carboniferous, adding uncertainty to how primitive turtles might be.

Avemetatarsalia, Ichthyosaurs, and Sauropterygia: Did I Get It?

? Exercise 31.1.1

The two groups of pterosaurs can be separated based on the absence or presence of _____, head crests, and tails.

a. wing claws

b. teeth

c. feathers

d. beaks

Answer

b. teeth

? Exercise 31.1.2

Ichthyosaurs are known for their large ____.

a. necks

b. eyes

c. teeth

d. coprolites

Answer

b. eyes

? Exercise 31.1.3

Turtles are most likely most closely related to which group?

a. Plesiosaurs

b. Mesosaurs

c. Pariasaurs

d. Ichthyosaurs

Answer

a. Plesiosaurs

Final Thoughts

The evolution of tetrapods may be very complex and controversial, but understanding this process can not only give you a great understanding of the animals that inhabit our planet today, but also beautifully demonstrate many of the founding principles of evolution. While dinosaurs are the group that gets the most attention, there was a whole array of interesting reptiles that are not dinosaurs. Some of these non-dinosaur reptiles share similar features with dinosaurs because of their common amniote tetrapod ancestors–they all have four limbs and certain features of the skull. There are also many wonderful examples of analogous traits via convergent evolution in the reptilian family tree due to them living in similar ecological niches (e.g. pliosaurs and crocodilians). The diversification of reptiles is an extraordinary example of adaptive radiation. An adaptive radiation is when a group of

organisms diversifies to fill all potential ecological niches to fully take advantage of all available resources. The expansion of reptiles to land, sea, and air exemplifies why the Mesozoic was historically dubbed “The Age of Reptiles.”

Further Reading

Rich, T. H., *et al*. (2001) “Monotreme nature of the Australian Early Cretaceous mammal Teinolophos”. *Acta Palaeontologica Polonica* **46(1)**. Pages 113–118.

Zhe-Xi Luo; Chong-Xi Yuan; Qing-Jin Meng; Qiang Ji (2011). “A Jurassic eutherian mammal and divergence of marsupials and placentals”

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CHAPTER OVERVIEW

32: (Case Study) Dinosaur classification

Learning Objectives

After reading this chapter, students should be able to:

- Identify the two major groups (orders) of dinosaurs, separated based on their hips
- Describe major subdivisions within each order, like theropods, sauropodomorphs, thyreophora, ornithopods, and marginocephalia.

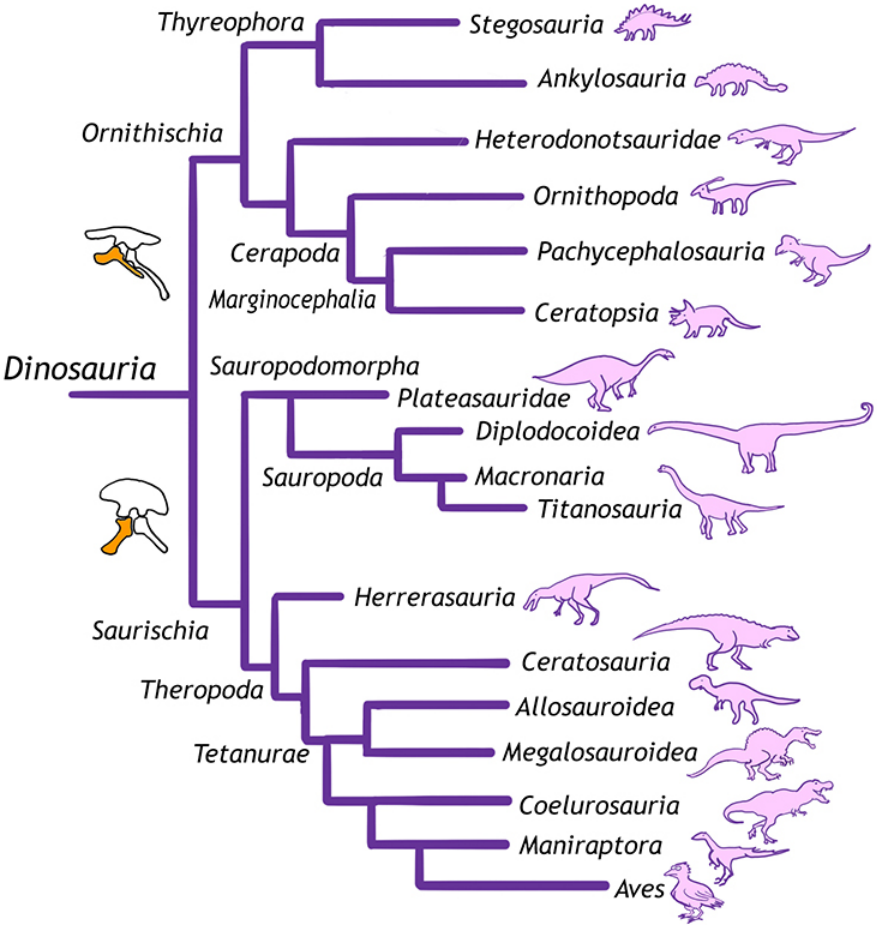


Figure 32.1: The major branches of the dinosaur family tree. (Image by Callan Bentley.)

Dinosaurs are a diverse group of reptiles and can be classified into various different groups. Understanding these groups can not only help in placing fan favorites like *Triceratops* and *Velociraptor* into a greater context, but also to show how different animals are related to one another and evolved from common ancestors. In this case study, we will talk about the seven major groups that include the vast majority of known dinosaurs and how those groups relate to one another.

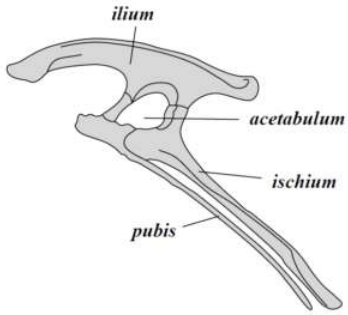


Figure 32.2: Ornithischian pelvis with a closed structure. (Image by AdmiralHood on Wikimedia.)

The most basic subdivision of dinosaurs is based on their hips. This division, first proposed by British paleontologist Harry Sheely in 1888, has traditionally been thought to be at the 'order' level in biological classification schemes, but modern research suggests that instead, it may merely be a clade. The hip has three bones: the pubis, ischium, and ilium. Some dinosaurs have a closed hip system in which the pubis and the ischium are close together. These dinosaurs are called bird-hipped dinosaurs, or Ornithischians. This style of hip is more similar to modern birds.

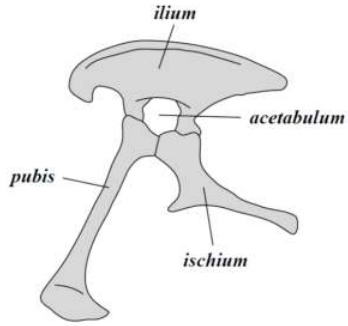


Figure 32.3: Saurischian pelvis with an open structure. (Image by AdmiralHood on Wikimedia.)

Other dinosaurs have an open structure to their hips with the pubis and ischium far apart and are known as Saurischian dinosaurs. Saurischian means 'lizard hip' because these hips are more similar to living lizards. Ironically, the dinosaurs that eventually evolved into birds are the Saurischian (lizard-hipped) dinosaurs whose descendents eventually adopted the closed-hip structure as well — a nice example of convergent evolution. These two branches of the dinosaur family tree are traditionally considered sister groups, but it should be noted that not all dinosaurs fall easily into one of these two groups (such as early dinosaurs like *Chilesaurus* and *Nyasasaurus*) and some paleontologists debate where to place other dinosaurs (such as the early meat-eating Herrerasaurs). Another complication is that not all scientists agree that with the premise that these are the two main groups. Most notable is a 2017 study by Baron et. al. (link to blog post) that says the traditional tree is all wrong. Since this new phylogeny is not widely accepted, below is the classic and prevailing interpretation of the members of each of the hip-based classifications.

32.1: Saurischians
32.2: Ornithischians
32.3: Final Thoughts

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32.1: Saurischians

Theropods

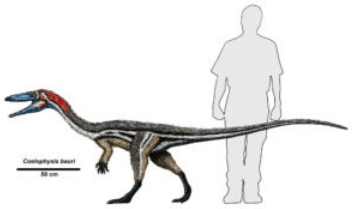


Figure 32.1.1: Drawing of Coelophysis compared to an adult human. (Image from Petrified Forest National Park.)

Saurischian dinosaurs have two major subdivisions: theropods and sauropodomorphs. The theropods are a diverse and complex group that are famous for meat eating. There is evidence of other dinosaurs being omnivores, and there are some theropods that strayed from a carnivorous diet, but all of the famous dinosaurs with big sharp teeth are theropods. Early (basal) theropods were swift and small like *Coelophysis*. Its name, which means 'hollow form', is a reference to an important trait that all theropods share: hollow bones. This made them lightweight, and therefore presumably more agile, which is helpful in chasing down prey. A less obvious but related trait they share is a series of air sacs used throughout the body used in respiration, at least partially filling the empty spaces in bones. This mechanism, seen in their living descendents today (birds), also helped allow a high level of activity while releasing heat more efficiently. Their other most notable shared trait is the three-toed feet with sharp claws.



Figure 32.1.2: Cast of Dilophosaurus skeleton, found in Arizona, from the Royal Ontario Museum. (Photo by Eduard Solà via Wikimedia.)

Evolutionarily, the theropods trended larger over time, a development first seen in *Dilophosaurus*. Contrary to its petite portrayal in the *Jurassic Park* movie franchise, it was over 20 feet long. As features such as a neck frill and the ability to spit poison are difficult to preserve in the fossil record, there is no direct scientific evidence for either. By the end of the Jurassic, these early large theropods diversified into famous forms like *Ceratosaurus*.



Figure 32.1.3: An illustration of Carnotaurus (by Andrey Atuchin via Wikimedia.)

Ceratosaurs would become truly massive in the Cretaceous, especially in South America, which was an isolated continent at the time. This is most exemplified by *Carnotaurus*, a name that means "the meat-eating bull," which had incredibly small vestigial arms. A sister group to the ceratosaurs, Tetanurae, was a more advanced branch of the theropod tree. This would have several branches, including the allosaurs (Allosauroidea), which lead to animals like *Allosaurus*, and the megalosaurs (Megalosauroidea), which had the crocodilian-like spinosaurs and the first dinosaur ever scientifically described, *Megalosaurus*. Interestingly, the allosaurs also got large and fierce in South America, with animals like *Carcodontosaurus* and *Giganotosaurus*, which were both estimated to be up to 40 feet long. These are arguably the largest land-dwelling theropods ever.



Figure 32.1.4: Yutyrannus tail fossil with feather-like impressions. (Photo by ★Kumiko★ from Tokyo, Japan via Wikimedia.)

The third branch off Tetanurae is Coelurosauria. This group is best known for the evolution of feathers. While skin and related features are rare to preserve, there is significant evidence that scientists have been able to estimate that all members of this group had some form of feathers. Some of the larger members may have shed them with age (like the birds do with down that keeps them warm as juveniles) or evolved them away, but those details are difficult to discern from the fossil record. Most of the largest members of the group, where skin impressions have been found, show no feathers at all. This was complicated in 2012 when *Yutyrannus* (meaning 'feathered tyrant') was described. It was clearly a coelurosaurian, and an early tyrannosaur, but was almost 30 feet long and covered in early feather-like structures.

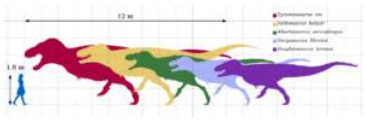


Figure 32.1.5: Largest Tyrannosaurs compared to a human. (Drawn by PaleoGeekSquared on Wikimedia.)

The earliest group of the coelurosaurians are the tyrannosaurs. Starting out small in the Jurassic, with many early members smaller than a full-grown human, their descendents became massive by the late Cretaceous. Famous examples include *Daspletosaurus*, *Albertosaurus*, *Tarbosaurus*, and *Tyrannosaurus*. All of the large members of the family had short arms with two digits, though evidence suggests that the arms still had some function. What that function is conjures much debate, from predation to mating. The group appears to have evolved in Asia, and although many of the larger, later members are famous from western North America, they also lived in Asia, on the east coast of the United States, and possibly even Australia.



Figure 32.1.6: Reconstruction of Deinocheirus (by FunkMonk via Wikimedia.)

Heading along the coelurosaur family tree, the next major group is *Compsognathus* and relatives, which were very small and delicate and probably ate insects and small animals like early mammals and lizards. Next are the ornithomimosaurs, whose name means “bird mimic lizards.” These dinosaurs had a body plan similar to ostriches, including a toothless beak. No teeth means a meat-eating diet is not an option, so these animals were mainly herbivores. They were only moderately sized, with the exception of *Deinocherius*, who was truly strange! It was 35 feet long, including a 3.5 foot long head with a duck-like bill, and the tail vertebrae which implies tailfeathers. It also had 8 foot long arms (the longest of any theropod) with giant claws and a sail on its back.



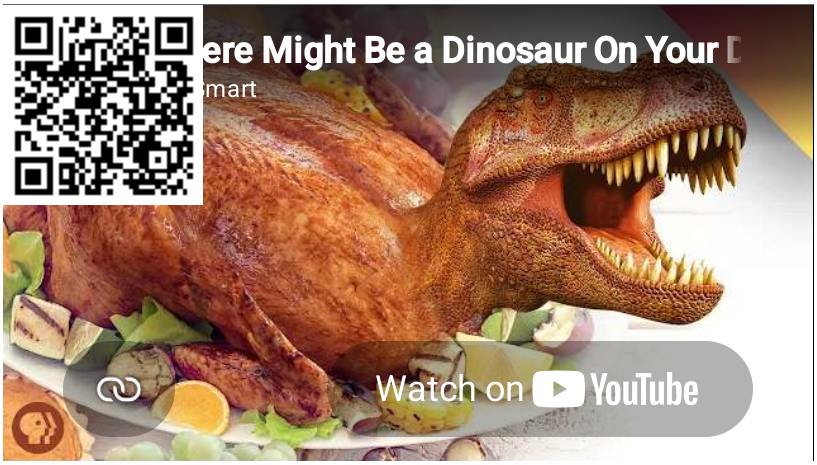
Figure 32.1.7: Therizinosaurus reconstruction (by Nobu Tamura via Wikimedia.)

A sister group to the ornithomimosaurs is the maniraptorans, or ‘thief hands.’ They have several adaptations that make them more bird-like, including the wishbone, which only they have. Dinosaur groups within Maniraptora include the short-armed, one-fingered alverezsaurs, the therizinosaurs who had the longest claws in the history of life (at over 3 feet), the poorly named (not egg eating or stealing but brooding) oviraptors, and the group paraves, which means ‘next to birds.’ Subdivisions of Paraves include Eumaniraptora and Deinonychosauria.



Figure 32.1.8: Various dromeosaurs (raptors) compared to humans. (Image by Fred Wierum via Wikimedia.)

Within Paraves, the most notable dinosaur group is the dromaeosaurs, colloquially known as the raptors. The trait that most identifies them is their second toe sickle claw, which was held above the ground. *Velociraptor* is perhaps the most famous member of the group due to the *Jurassic Park* films, but the animal in the movie is actually patterned after *Deinonychus*, a larger raptor. In fact, the animal that most matched the movie depiction is *Utahraptor*, which ironically was published and named just days after the film’s premiere. Deinonychus is also important because when it was described in 1969 by Jon Ostrum and Robert Bakker, it helped spark the Dinosaur Renaissance, which led to our modern ideas of dinosaurs as dynamic animals instead of the lumbering beasts of the early 20th Century.



Video by PBS that breaks down how dinosaurs are related to birds.



Figure 32.1.9: The famous Berlin Specimen of Archaeopteryx. (Picture by H. Raab (User: Vesta) via Wikimedia.)

A related group to the dromaeosaurs is avialae, which can be considered (depending on definitions) to be birds. It is assumed that all members of this group were capable of powered flight. *Archaeopteryx*, traditionally known as the first bird, would be an early member of this group. This group originated in the Jurassic and became quite prolific in the Cretaceous, and led to the group Neornithies, which is the clade that includes all modern birds. Current analyses consider the ratites, the group which includes large flightless birds like ostriches and emus, to be the most primitive in the modern bird phylogeny.

Theropods: Did I Get It?

? Exercise 32.1.1

Besides feathers, what unique adaptation do theropods have that was taken advantage by birds?

a. Sharp claws

b. Long Necks

c. Teeth

d. Hollow bones

Answer

d. Hollow bones

? Exercise 32.1.2

Which theropod group is closest to birds?

a. Dromeosaurs (raptors)

b. Tyrannosaurs

c. Alverezsaurs

d. Carnosaurs

Answer

a. Dromeosaurs (raptors)

? Exercise 32.1.3

Which theropod was earliest (evolutionarily oldest)?

a. Megalosaurus

b. Allosaurus

c. Carnotaurus

d. Coelophysis

Answer

d. Coelophysis

Sauropods



Figure 32.1.10: Skeleton of Plateosaurus. (Photo by Ryan Somma via Wikimedia.)

The other group of saurischian dinosaurs, the sauropods, are thankfully much less complicated than the theropods. Sauropods became the largest animals to ever walk the earth, but they started out relatively small like other dinosaurs. Early sauropods were traditionally called prosauropods, but that term has started to become unpopular. The preferred term, to encapsulate the entire sauropod clade (all true sauropods and their ancestors) is sauropodomorphs. Early sauropodomorphs (now called Plateosauridae) were bipedal and probably omnivores, with sharp claws on their hands that could have helped tearing vegetation or flesh. The best example of an early sauropodomorph is *Plateosaurus*. It had blunt teeth and was about twice as tall as a human. It also had an elongated neck, allowing it to reach leaves that most other species at the time could not access.



Figure 32.1.11: Restoration of Melanorosaurus (by Mario Lanzas via Wikimedia.)

Over time, the sauropodomorphs got larger and larger. The important step from early sauropodomorphs to true sauropods is when their weight became so large that they were permanent quadrupeds. There is debate as to which genus represents this transition, but an animal like *Melanorosaurus* or *Vulcanodon* are good candidates. It has more robust limbs, with specifically longer hind limbs with less-useful claws, to support its increasing weight. Interestingly, the teeth of these animals were not what you would expect for a tree browser. They kept their omnivore-like needle teeth and used gastroliths, small rocks they would swallow and store in their gizzards to help digest the tough plant material, instead of grinding and chewing their food in their mouths.



Figure 32.1.12: Diplodocus life reconstruction (by Fred Weirum via Wikimedia.)

Early sauropods include the club-tailed *Shunosaurus* and the incredibly-long-necked *Mamenchisaurus*. All sauropods still have a front toe claw, a legacy of their bipedal beginnings. More derived sauropods split into two major groups: The first group are the diplodocids (Diplodocidea), which had a long body plan. This group includes *Apatosaurus*/*Brontosaurus* (which may be the same animal or may be different animals depending on who you ask), the spiny-necked dicraeosaurs, and the namesake of the group, *Diplodocus*.



Figure 32.1.13: Life sized model of Argentinosaurus at Clore Garden of Science in Weizmann Institute of Science in Israel. (Picture by MathKnight and Zachi Evenor via Wikimedia.)

The other major group of sauropods is Macronaria. These are animals that had a more upright stature, most famously demonstrated by *Brachiosaurus*. A subgroup of macronaria are the titanosaurs. As the name suggests, these were generally very large animals that commonly had some armored skin coverings made from bone. Included in this subgroup is *Argentinosaurus*, which is the best candidate for the largest known dinosaur, based on decent fossil remains of a leg and spine. Upper estimates for its size is 130 feet long and 100 tonnes. Other titanosaurs, such as *Sauroposeiden* and *Patagotitan*, may be larger but are estimated from more incomplete remains.

Sauropods: Did I Get It?

? Exercise 32.1.1

Which group covers the most animals?

a. Prosauropod

b. Sauropod

c. Sauropodomorph

d. Sauropodiform

Answer

c. Sauropodomorph

? Exercise 32.1.2

What feature do all sauropods have that is a remnant of their early history on two legs?

a. Sharp teeth

b. Hollow bones

c. Long necks

d. Toe claw

Answer

d. Toe claw

? Exercise 32.1.3

What separates the major groups of advanced sauropods?

a. Meat-eating verses plant-eating

b. Heavy verses light

c. Long verses upright

d. Spikes verses no spikes

Answer

c. Long verses upright

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32.2: Ornithischians



Figure 32.2.1: Heterodontosaurus reconstruction with possible feather-like structures. (Drawing by FunkMonk via Wikimedia.)

The sister group to saurischia, the ornithischians are a diverse group of dinosaurs. Their hip, which is more flattened and pointed towards the tail, allows for a larger gastro-intestinal system which these animals used to process their almost exclusively plant-based diets. The first ornithischians were small bipedal runners, with *Lesothosaurus* claiming the title of oldest ornithischian and the Heterodontosaurs being the most famous members of this group. Like the name implies (hetero=different, dont=tooth, saurus=lizard), these animals had a unique trait that most reptiles do not, neither at that time nor even today: differentiated, specialized teeth. They had more rasping teeth toward the back of their mouths and more spiky teeth toward the front. They also had an extension of their jaw in front of the teeth. This would become a beak in its descendents, a feature that is common in many ornithischians.

Thyreophora

Stegosaurs



Figure 32.2.2: A, mounted skeleton in right lateral view and B, laid out with missing elements reconstructed before mounting. (Image by Susannah Maidment et al. & Natural History Museum, London)

There are two major subgroups within ornithischia: thyreophora and neornithischia. Thyreophora represent the group of armored dinosaurs, which had bony buildups protruding and covering the skin. The stegosaurs, one major branch of the thyreophora tree, were common in the Jurassic. Their bony plates on their back are still one of the biggest mysteries in paleontology. Various ideas, from thermoregulation, defense, and sexual display have been proposed, but none have been fully convincing. One thing to consider is related animals, like *Kentrosaurus* and *Miragaia*, had very different set ups. Any explanation of why *Stegosaurus'* had its plates has to explain why others did not have them.

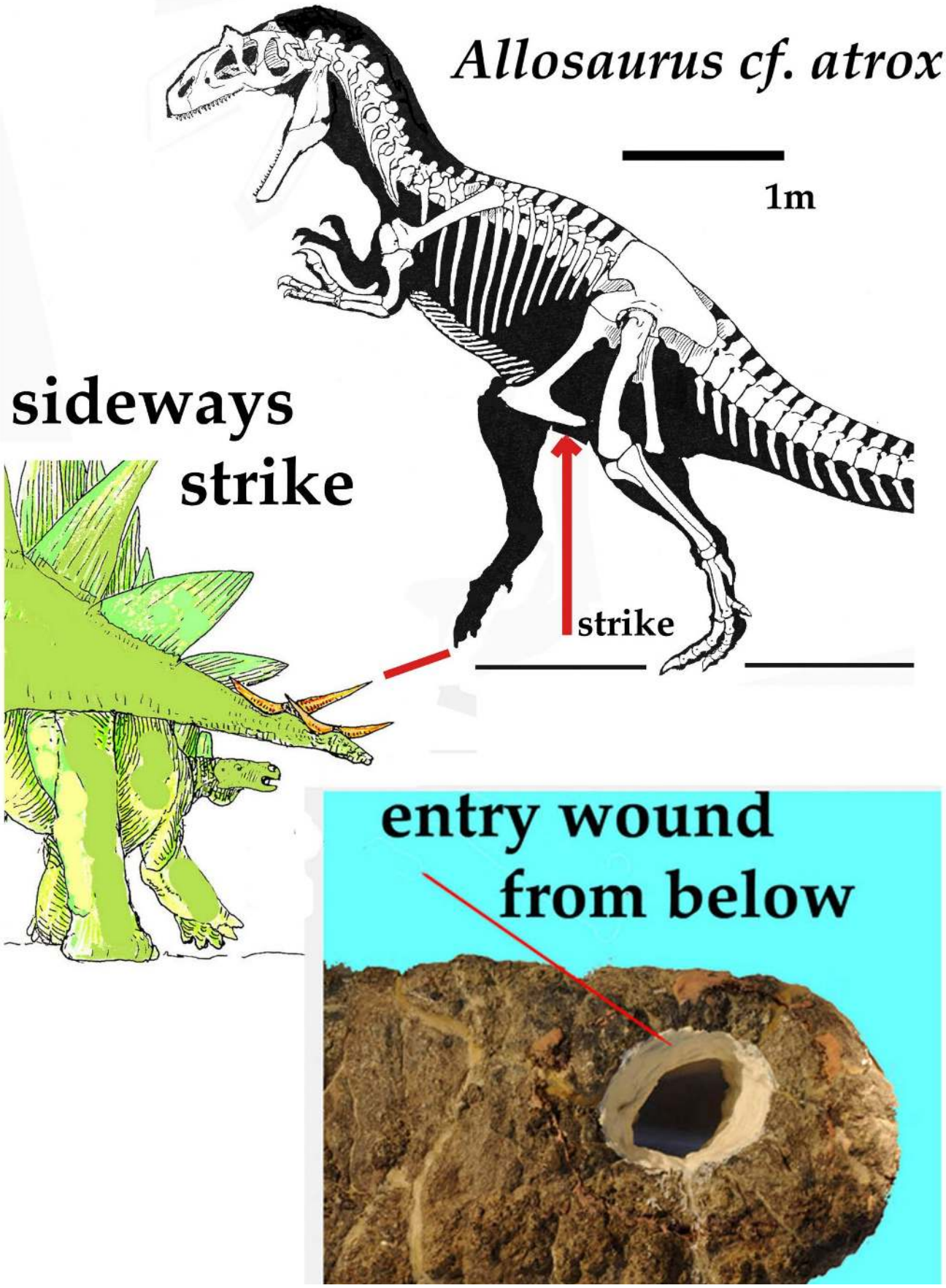

Allosaurus cf. atrox
1m
strike
sideways
strike
entry wound
from below

Figure 32.2.3: Cartoon and picture of a Stegosaurus/Allosaurus interaction. (Image care of Robert Bakker.)

Less confusing are the defensive tail spikes, which match holes in the bones of predatory dinosaurs at the time. Jokingly called a 'thagomizer' in a *Far Side* cartoon by Gary Larson in the 1980s, this name has now been somewhat accepted in the scientific community. One famous example of the thagomizer's impact was first presented in 2014 by Robert Bakker and colleagues in which an Allosaurus had a hole in its groin that abscessed and probably killed the animal. The most likely cause of the hole is a thagomizer spike.

Ankylosaurs



Figure 32.2.4: Ankylosaurus reconstruction (by Emily Willoughby via Wikimedia.)

The other major group of thyreophorans are the ankylosaurs (Ankylosauria). These were heavily armored tank-like animals common in the Cretaceous. There are at least two major subdivisions in this category: the tail-clubbed Ankylosauridae and the more spiky Nodosauridae. Common depictions of *Ankylosaurus*, as in the *Jurassic World* films, show an unrealistic hybrid that had the shoulder spikes with the tail club, but in reality it was one or the other.

Early Ornithischians and Thyreophora: Did I Get It?

? Exercise 32.2.1

What feature do early Ornithischians like Heterodontosaurus have that they passed down to their descendants?

a. Clubs

b. Beak

c. Fangs

d. Spikes

Answer

b. Beak

? Exercise 32.2.2

Stegosaurus' tail spikes are called _____, named by cartoonist Gary Larsen.

a. sharpinator

b. stabifier

c. thagomizer

d. allofier

Answer

c. thagomizer

? Exercise 32.2.3

Which of the following would a nodosaur (and only a nodosaur) have?

a. Tail spikes

b. Tail club

c. Shoulder spikes

d. Tail without club or spikes

Answer

d. Tail without club or spikes

Ornithopods

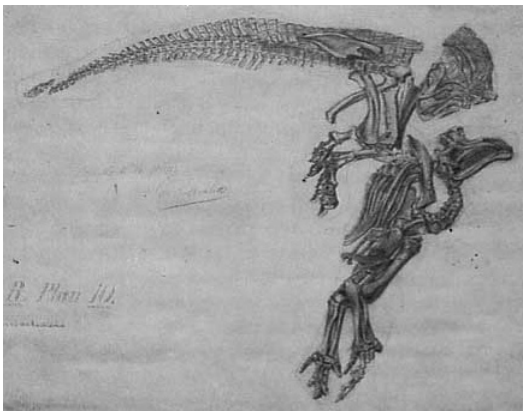


Figure 32.2.5: Drawing of an Iguanodon skeleton by G. Lavette as it was found in 1882, nearly articulated.

The other side of the ornithischian tree (*i.e.*, Cerapoda/Neornithischia) has three other major groups. The more complex, diverse, and prolific of the three are the ornithopods. Their name means 'bird feet' due to the three toes, but remember that the theropods (only distantly related) were the group that evolved into birds. Their success did not come from armor, speed, or size but something much more mundane: chewing. Their tooth battery would remind one of a modern horse or cow; highly complex (for the time) and effective at breaking down tough plants. Like other groups, they started small and bipedal (like the hypsilophodonts), but eventually grew larger. The large ornithopods are speculated to be somewhere between a biped and a quadruped like a chimpanzee or gorilla. A famous member of this group that is transitional between the early and later ornithopods is *Iguanodon*. Though quite large, its thumb spike and prehensile pinky must have had some purpose, suggesting it had to spend some time on two legs.

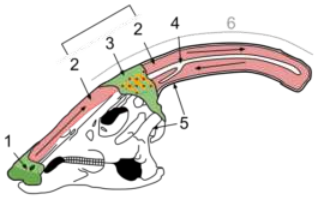


Figure 32.2.6: Parasaurolophus skull with a view into nasal cavity of the cranial crest. It is hypothesized that the Parasaurolophus pushed air through its long cranial crest to make low register sounds that could be heard for miles. 1. Nostril 2. Sinus cavity 3. Skin 4. Intra Cranial bone 5. Skull 6. Cranial crest

When the average person thinks of the animals in this category, typically the word they would use to describe them would be 'hadrosaur' or 'duck-billed dinosaur.' These terms are appropriate, but only for the more derived members of the group. So while *Iguanodon* is not a hadrosaur, other famous ornithopods like *Parasaurolophus* and *Edmontosaurus* are hadrosaurs. The hadrosaurs break into two major groups: the saurolophines (formerly known as hadrosaurine) which had no crest or solid crests and the lambeosaurines which had hollow crests. One famous member of saurolophines is *Maiasaura*, the first dinosaur found with definitive evidence of parenting, hence the name, 'good mother lizard.' The young of *Maiasaura* were found to have worn teeth before they had the ability to walk, suggesting the mother had to have brought food to the brood. *Parasaurolophus* is a famous member of the lambeosaurines. Their hollow crests have been speculated to allow them to make sounds, presumably to communicate with members of their own species.

Ceratopsians



Figure 32.2.7: Drawing of Protoceratops (by AntoninJury via Wikimedia.)

The sister group to the ornithopods is marginocephalia. These animals also had tooth batteries, but also had head ornamentation. Marginocephalia has two major subdivisions. The first is the ceratopsians, or horn-faced dinosaurs. It should be noted that despite the name, not every member of this group had a horn on their face. What was much more universal was a frill, a bony projection that extended from the skull back towards the neck. Early members of the group were from Asia and were small with tiny frills. A famous early example is *Protoceratops* from Mongolia. The name, which means 'before horned face', shows it had the neck frill and just the smallest bumps where horns evolved in others.



Figure 32.2.8: Triceratops skeleton from the Los Angeles Museum of Natural History. (Photo by Allie_Caulfield edited by MathKnight via Wikimedia.)

Within the ceratopsian family, there are two subfamilies. The chasmosaurs typically had more trapezoidal frills with larger brow horns. The most famous member of this group is *Triceratops*. The centrosaurs typically had larger nose buildups (horns and sometimes bulbous bony growths) and spikes on the frill. *Styracosaurus* is a famous example of a centrosaur. There is still much debate as to the purpose of the horns, spikes, and frills. Popular convention interprets these items as defenses from predators. While there is some evidence for this, not all scientists agree that they were defensive, especially the frills. Most ceratopsians' frills were weak and thin, especially in the centers of each half, making them useless against direct combat. Other ideas for their growth include sexual display and species recognition. This goes along with many fossil deposits of ceratopsians that show many individuals in the same place of different ages, which suggests herding or other social behavior.

Pachycephalosaurs

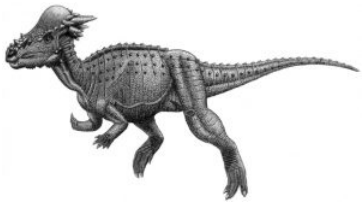


Figure 32.2.9: Sketch of a Pachycepholosaur from the Hell Creek Formation of South Dakota (by Jordan Mallon via Wikimedia.)

The other branch of marginocephalia is the pachycephalosaurs, a name that means 'thick headed lizards.' They were bipedal and smaller than many other dinosaur types, with the largest animals just reaching the height of a grown human. Their head buildups did not result in frills or true horns, but a dome head surrounded by small spikes. This dome is often suggested to be for intra-species combat over mates or territory, similar to what rams or deer do today. The evidence of this, like so many other things in the fossil record, is circumstantial and incomplete. Their domes could have been used for similar things as their ceratopsian cousins: mating and recognition. Another complication scientists have between the pachycephalosaurs is naming and differentiating them. It has been proposed that many individual genera may be invalid, with these variations proposed to instead reflect differences in age

or gender. *Stigymoloch* and *Dracorex*, for example, have been proposed as the subadult and juvenile versions of *Pachycephalosaurus*, respectively.

Cerapoda: Did I Get It?

? Exercise 32.2.1

Hadrosaurs are split based on their _____.

a. feet

b. teeth

c. crests

d. tails

Answer

c. crests

? Exercise 32.2.2

Where did ceratopsians start?

a. Asia

b. Eastern US

c. Africa

d. Western US

Answer

a. Asia

? Exercise 32.2.3

Many pachycephalosaurs may actually be what instead of separate species?

a. Lack of complete fossils

b. Different positions in the pack

c. Evolutionary changes over time

d. Different ages

Answer

d. Different ages

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32.3: Final Thoughts



Figure 32.3.1: Maybe the greatest fossil find ever, from Mongolia: a Protoceratops and Velociraptor engaging in battle. The Velociraptor's sickle claw is lodged in the side of its prey, while the Protoceratops grabs the Velociraptor's hand in its mouth. (Picture by Matt Affolter.)

Obviously, there is still much to learn about the dinosaur family tree, in how individual members are placed and how they are interrelated. Understanding the seven major groups, theropods, sauropods, stegosaurs, ankylosaurs, ornithopods, ceratopsians, and pachycephalosaurs, is a first step to help the novice place a new dinosaur in the greater context, and also understand the history of life a little bit better. As more discoveries are presented, old ideas have been abandoned in favor of better hypotheses, with inconsistencies and unknowns debated, both hallmarks of science. How these animals fit into the greater reptile family tree in the Mesozoic also demonstrates these changing ideas of evolutionary history, as well as great examples of convergent evolution. Even our understanding of modern animals, like birds and crocodiles, has been enhanced as we learn about their extinct dinosaur relatives. For more details on the debates within the dinosaur family tree, please read the dinosaur debate section of this text.

Further Reading

Baron, Matthew G.; Norman, David B.; Barrett, Paul (2017). "A new hypothesis of dinosaur relationships and early dinosaur evolution" (PDF). *Nature*. **543** (7646): 501–506.

Brusatte, S. The Rise and Fall of Dinosaurs: A New History of a Lost World

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CHAPTER OVERVIEW

33: (Case Study) Dino Debates - dinosaur ecological controversies

Key Concepts

- Dinosaurs have many lines of circumstantial evidence that implies social groups, like mass graves and footprints
- There is still no direct evidence of pack hunting, but there is direct evidence of other hunting strategies
- Dinosaurs have strong evidence that at least some species had intensive care of their young, including nests and parental help
- Many dinosaurs might not be separate species or valid due to incomplete remains, sexual dimorphism, and changes to animals as they age
- Famous examples include *Brontosaurus/Apatosaurus*, *Triceratops/Torosaurus*, and *Pachycephalosaurus/Stygimoloch/Dracorex*
- Though not every fossil shows impressions of skin or feathers, enough have been found to show how they developed in theropods
- Recent studies have shown a larger number of species had feathers, though large animals like *Tyrannosaurus rex* probably were still featherless
- Dinosaurs have significant evidence that they were not ectothermic (cold blooded), and may have been endothermic (warm blooded)
- Feathers and their size may have helped dinosaur thermoregulation
- At the very least, dinosaurs were agile and active animals who could move much more like mammals and birds than lumbering lizards as they were presented prior to the dinosaur renaissance

33.1: Dinobook or Saurogram? - Dinosaur Social Behavior
33.2: A Dinosaur by Any Other Name - Species controversies
33.3: To Feather or Not To Feather - Dinosaur Body
33.4: Movement and thermoregulation - Get that blood flowing!
33.5: Further Reading

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33.1: Dinobook or Saurogram? - Dinosaur Social Behavior

There is a lot of debate over the social-networking skills of dinosaurs. On one hand, the living descendants of dinosaurs (birds) are highly social. On the other hand, social interactions are inherently difficult to preserve in the fossil record, so direct evidence is hard to come by. The two main ways that social behaviors can be inferred is through trackways and mass mortality sites.



Figure 33.1.1: Cal Orcko, Boliva is known for its enormous limestone wall that is covered by one of the longest stretches of dinosaur footprints ever found. In this picture, these sauropod tracks are seen starting in the left middle of the image then trending to the top of the wall.

https://geo.libretexts.org/@go/page/22818

Dinosaur track sites are known in many places around the world. Lots of useful information can be gained from them, including speed and gait. Some of them have enough tracks to represent multiple individuals and even multiple species. When this happens, paleontologists can infer social behaviors that would never be preserved otherwise. Each dinosaur group has a characteristic style of tracks, though in many cases, the exact dinosaur that laid them down is hard to distinguish. Trace fossils like tracks are also given their own nomenclature that follows the same rules as naming a biologic organism, complete with geniuses and species.



Figure 33.1.2: Theropod trackway from the Ardley Quarry. (Photo by J.J. Day.)

A good example of a dinosaur track site which has a reasonable interpretation at behavior is the Ardley Quarry in the UK. It not only shows a large group of animals walking the same direction (including possible prey and predators), but even different ages all together, similar to what is seen in large mammal herds today. This does not prove that dinosaur herds and modern mammal herds had the same structure, but it is a good indication. It has even been used to make evolutionary inferences about the extent of radiation of different dinosaur groups.





Another famous example is Dinosaur Stampede National Monument in Australia. There is still debate as to what exactly happened at this site, where a lone large individual was walking through the mud with many different smaller individuals. Regardless of the exact circumstances, the large variation in the smaller tracks give an indication that the group was a herd with different-aged animals.

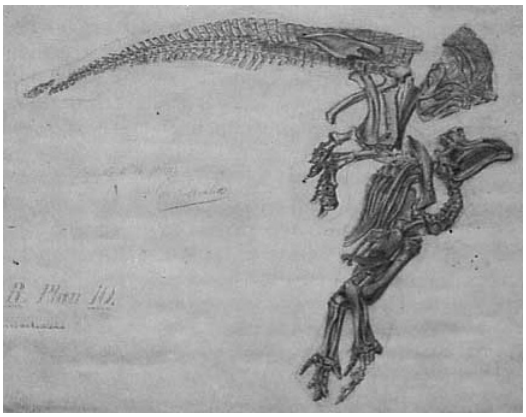


Figure 33.1.3: Drawing of an Iguanodon skeleton by G. Lavette as it was found nearly articulated.

Mass mortality sites also help get at the behavior of dinosaurs, with the general thought that if they lived together, they died together. The first of these was discovered in Belgium in 1878 when nearly 40 articulated Iguanodons were found. In 1974, at least 14 *Pachyrhinosaurus* individuals were found together in an ancient Alberta riverbed, most likely washed up together when trying to cross a flooded river. There are countless others, but just to highlight some important ones: In 2008, a find in Inner Mongolia, China was announced that had more than 20 *Sinornithomimus* animals that were all immature sub-adults, between one and seven years old. It is proposed that they were a social group, perhaps left to their own devices while adults cared for the infants.



Figure 33.1.4: Deinonychus reconstruction (by Fred Wierum via Wikimedia.)

Deinonychus, the dinosaur that the raptors from the Jurassic Park movies were based on, has a lot of interesting evidence of social behavior. North American fossil sites have found a lot of individual animals of varied ages in concentrated sites. Also, they have been found in close proximity to a possible prey animal, called *Tenontosaurus*. *Deinonychus* was about 1/20th the size at best, so how did they bring them down? Many have speculated that this is an indication of pack hunting. Certainly, their proximity in the fossil record could have been related to scavenging, but pack hunting is an intriguing interpretation.

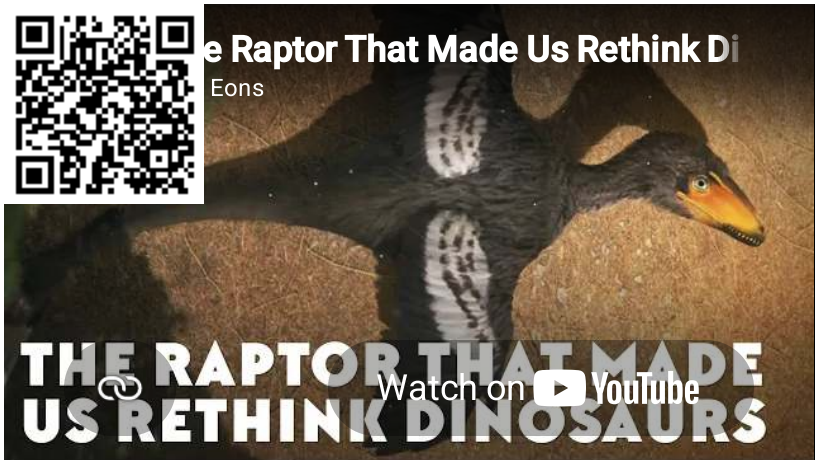




Figure 33.1.5: Maybe the greatest fossil find ever, from Mongolia: a Protoceratops and Velociraptor engaging in battle. The Velociraptor's sickle claw is lodged in the side of its prey, while the Protoceratops grabs the Velociraptor's hand in its mouth. (Photo by Matt Affolter.)

The most unambiguous dinosaur fossil depicting behavior is the famous fighting dinosaurs. Found in the Gobi in 1971, it is a fossil that has a *Protoceratops* and a *Velociraptor* (the smaller relative of Deinonychus) forever locked in combat. The Velociraptor has its killing claw lodged into the side of the Protoceratops, while it fights back by grabbing the attacker's arm in its beak. There is nothing yet found that comes close to depicting behavior like this fossil, which is why many consider it to be the greatest find ever.

Rearing behavior

Another aspect of social behavior is parental investment. Many reptiles are not known for being great mothers and fathers, with sea turtles famously laying their eggs on a beach and leaving. The closest relatives of dinosaurs (crocodiles and especially birds), however, are noteworthy parents in the animal kingdom. So where do dinosaurs fall on this spectrum?



Figure 33.1.6: Citipati fossil found in a brooding position over eggs. (Photo by Dinoguy2 via Wikimedia.)

The first noteworthy discovery on this front was nests. In the early 1920s, an expedition led by Roy Chapman Andrews in Mongolia found what is believed to be the first dinosaur nest and eggs. Near the nest was a skeleton, described and named by Henry Osborn as *Oviraptor*. Initially, the nest was thought to belong to a common dinosaur in the stratum *Protoceratops*. *Oviraptor*, which means "egg thief," was assumed to be stealing the eggs and eating them with its toothless beak. Even Osborn at the time cautioned the use of such an interpretation in its name, but it stuck. In the 1990s, more discoveries, like the related animal *Citipati* found fossilized in a brooding stance over a nest helped clear the image of *Oviraptor*, though it is too late to change the name.





Figure 33.1.7: Fossil reconstruction of a Maiasaura and hatchlings.

Maiasaura, the “good mother lizard,” with the femanine ‘saura’ used for “lizard,” was described by Jack Horner in the 1970s. Its discovery has done more to advance our understanding of parental behavior than any other dinosaur. Instead of finding just one dinosaur or one nest, an entire nesting ground was discovered, spaced closer together than the length of one adult, as some birds do today. The eggs were arranged in a spiral and placed on top of rotten vegetation to keep them warm (instead of using the parents’ warmth). The most important find was of hatchlings, which had underdeveloped legs which made walking unlikely, and yet, had worn teeth. It is estimated that the hatchlings might have spent a year in the nest, and although the mortality rate was high, it was the parents of these babies that fed them and protected them.

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33.2: A Dinosaur by Any Other Name - Species controversies

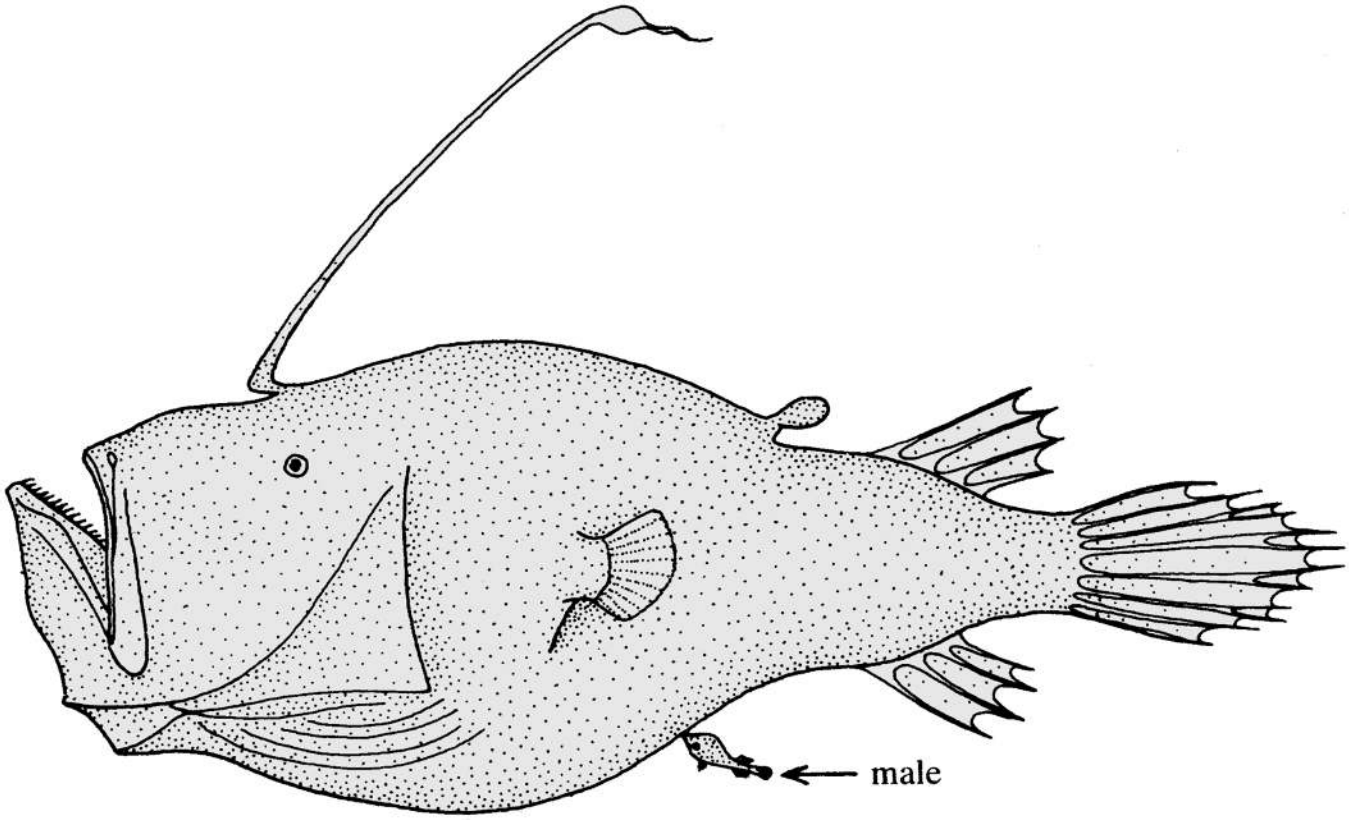


Figure 33.2.1: Angler fish famously have incredible differences between males and females. If they were found as fossils, would they be classified as the same species?

When a paleontologist finds a new fossil, one of the first things they analyze is the dimensions of the fossil to compare it to existing species. If the organism is different enough, then a new description is written about it, defining the new species. Seems easy, right? There is a problem though. How does the paleontologist know how representative the organism was to its species? Imagine you are a future scientist studying human fossils. If you came upon the bones of a male basketball player, and then the bones of a female gymnast, would you think they were the same species? How about an elderly person versus a juvenile?



Figure 33.2.2: Jane, previously assigned to Nanotyrannus, is now believed to be a young Tyrannosaurus. (Picture from the Burpeee Museum on Natural History by Zissoudisctrucker via Wikimedia.)

Of course, part of the training of any paleontologist is studying modern anatomy and populations of organisms so you have a basis for comparison for things like sexual dimorphism (changes between the genders of a species), gerontology (changes based on age), variability within a species, or even diseases that may alter the appearance of tissues. But when presented with an extinct species, certain extrapolations and assumptions are required. This has led to several controversies in paleontology, notably within dinosaurs. In fact, a very large number of dinosaurs, even some famous ones, may be nothing but a synonym of an existing species. In biologic nomenclature, the first name published gets precedent. This is why *Ultrasaurus*, *Nanotyrannus*, and *Monoclonius* are now considered invalid, and some scientists feel as many as one third of dinosaur species may join them in infamy. Some other famous examples are described below.

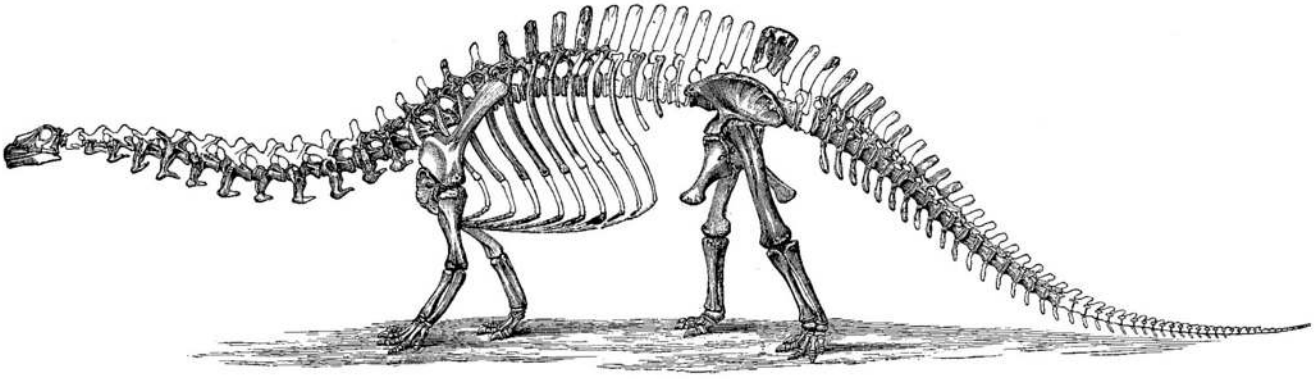


Figure 33.2.1: O.C. Marsh's Brontosaurus skeleton, with skull material that was later assigned to Brachiosaurus.

Perhaps the most famous dinosaur naming faux-pas comes from *Brontosaurus*. It is still one of the most famous dinosaurs ever, but it might not even exist! Toward the end of the 19th Century, scientists O.C. Marsh and Edward Cope were in a constant battle to describe and discover as many dinosaurs as possible, a rivalry known as the Bone Wars. This led to rushed, inaccurate, and sloppy science as they fought for prominence in paleontology, and *Brontosaurus* was one of the victims. In 1877, Marsh described a sauropod which he named *Apatosaurus*. Two years later, Marsh described a similar animal, but different enough in his mind to warrant a new genus. He named it *Brontosaurus*. One part of the skeleton was missing, however: the skull.



Figure 33.2.3: Modern corrected Apatosaurus skeleton. (Photo by Tadek Kurpaski of the Carnegie Museum via Wikimedia.)

In 1903, Elmer Riggs, a paleontologist from Chicago's Field Museum, said that *Apatosaurus* and *Brontosaurus* were similar enough to be the same genus, and since *Apatosaurus* was named first, the term *Brontosaurus* became invalid as a junior synonym. However, when the first mount was made for the American Museum of Natural History (the first sauropod mount ever displayed) in 1905, the name *Brontosaurus* was chosen for the skeleton. Since the skull was not known, a bulky skull similar to *Camarasaurus* was used. This mount may at least partially explain why the term *Brontosaurus* has persisted, even though throughout the 20th Century, almost all scientists agreed with Riggs that *Apatosaurus* was the proper name that should be used. In 1909, an *Apatosaurus* skull was found (which was more slender, similar to another sauropod, *Diplodocus*), but the knowledge of this skull remained limited until it was published in the 1970s.

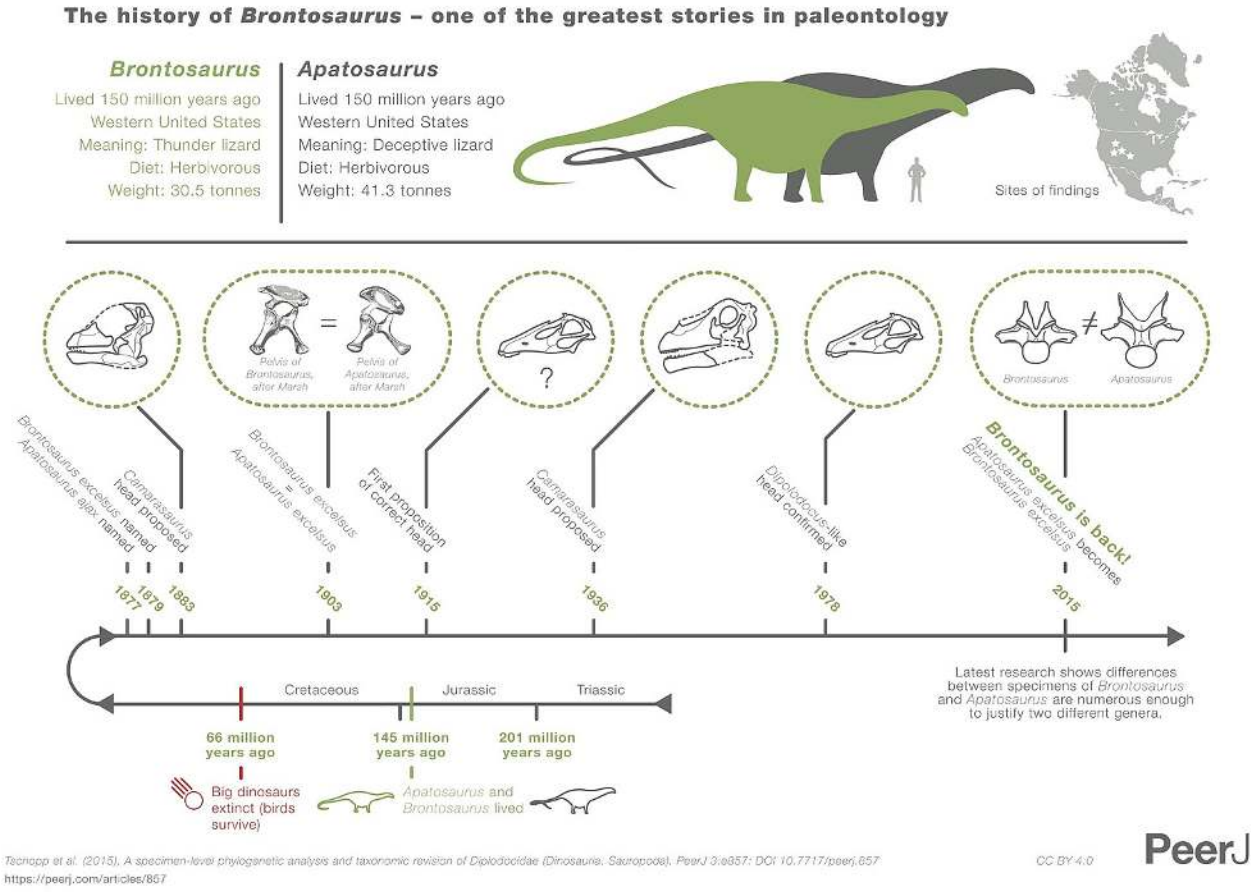


Figure 33.2.4: Infographic showing the timeline of Apatosaurus vs. Brontosaurus.

In 1998, Robert Bakker argued that different species of *Apatosaurus* should be split into multiple genuses. Then, in 2015, Emanuel Tschopp, Octavio Mateus, and Roger Benson released a study on *Apatosaurus* and related animals and their analysis also concluded that there is enough variation in the *Apatosaurus'* to justify a new genus name. Rather than create a new name, they decided to resurrect the *Brontosaurus* name for this new genus. While not all paleontologists agree, it may have solved the problem of the popular, but potentially incorrect, *Brontosaurus*.



Figure 33.2.5: Range of skulls of Triceratops, with youngest on the right and oldest on the left. There are several dramatic changes in the skulls with age, including the angle of the horns and shape of the frill.

One of the more infamous controversies is called the toromorph hypothesis. *Triceratops* is a famous ceratopsian dinosaur from the late Cretaceous of western North America. *Torosaurus* is from the same general time, place, geologic formation, and habitat. It is noticeably larger than *Triceratops*, with a larger frill with large holes in the center of each half of the frill. Starting in 2009, John Scanella and Jack Horner proposed that *Triceratops* was just a juvenile of *Torosaurus*. Other related animals, like *Nedoceratops* and *Ojoceratops* have also been suggested as growth stages of *Triceratops*. Evidence to support this position includes several juvenile-like features of *Triceratops*, dramatic changes in *Triceratops'* known young versus known 'adults,' and a lack of *Torosaurus* young.

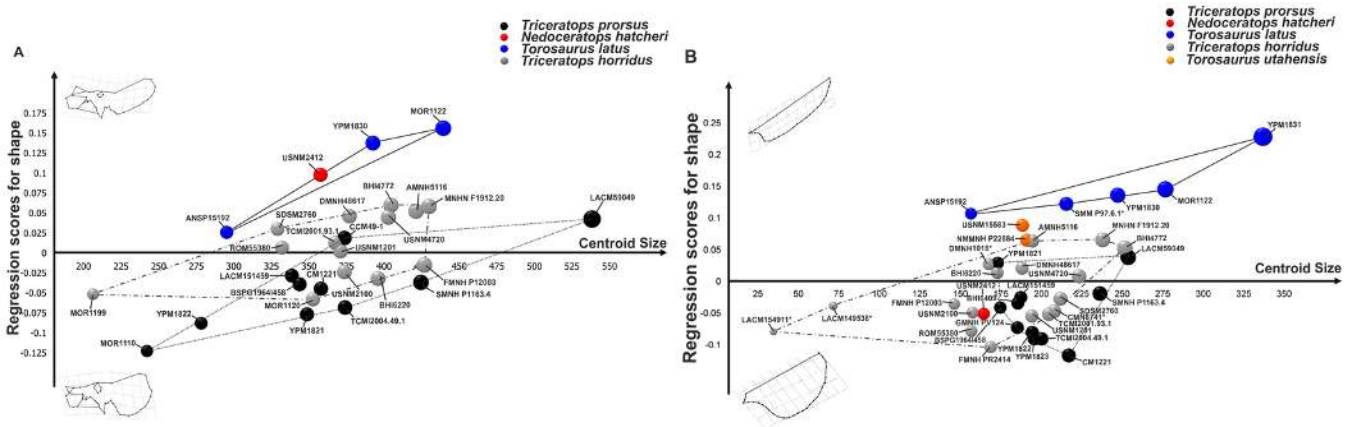


Figure 33.2.6: This analysis shows more overlap in skull features between Triceratops than other species that have been proposed to be synonyms.

This idea has come under a fair amount of scrutiny, most notably by Andrew Farke and Nicholas Longrich. They have analyzed features of the skulls of each animal and noticed some differences in the skulls. Even a computer analysis of morphometric statistics showed that there is a significant difference between the species. Also, subtle but important differences between the animals have been noted. For example, *Torosaurus* is much more rare than *Triceratops* overall, *Triceratops* specimens that seem old enough to be mature, slight differences in age and location of each species without exact overlap, and no other known ceratopsians which gain holes in frills with age.

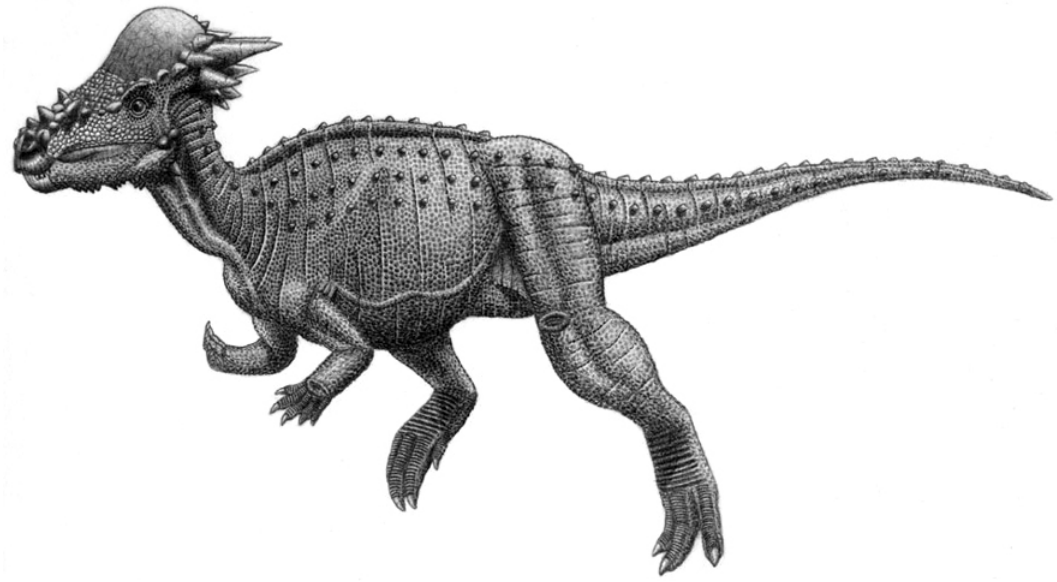


Figure 33.2.7: Sketch of a Pachycepholosaur from the Hell Creek Formation of South Dakota (by Jordan Mallon via Wikimedia.)

Pachycephalosaurs are dinosaurs famous for their thick skull domes. They are also some of the most poorly understood dinosaurs, as a complete skeleton has never been found of any member of the group. They are also related to the ceratopsians (within Marginocephalia) and have similar growth changes. This had led to significant species controversies.

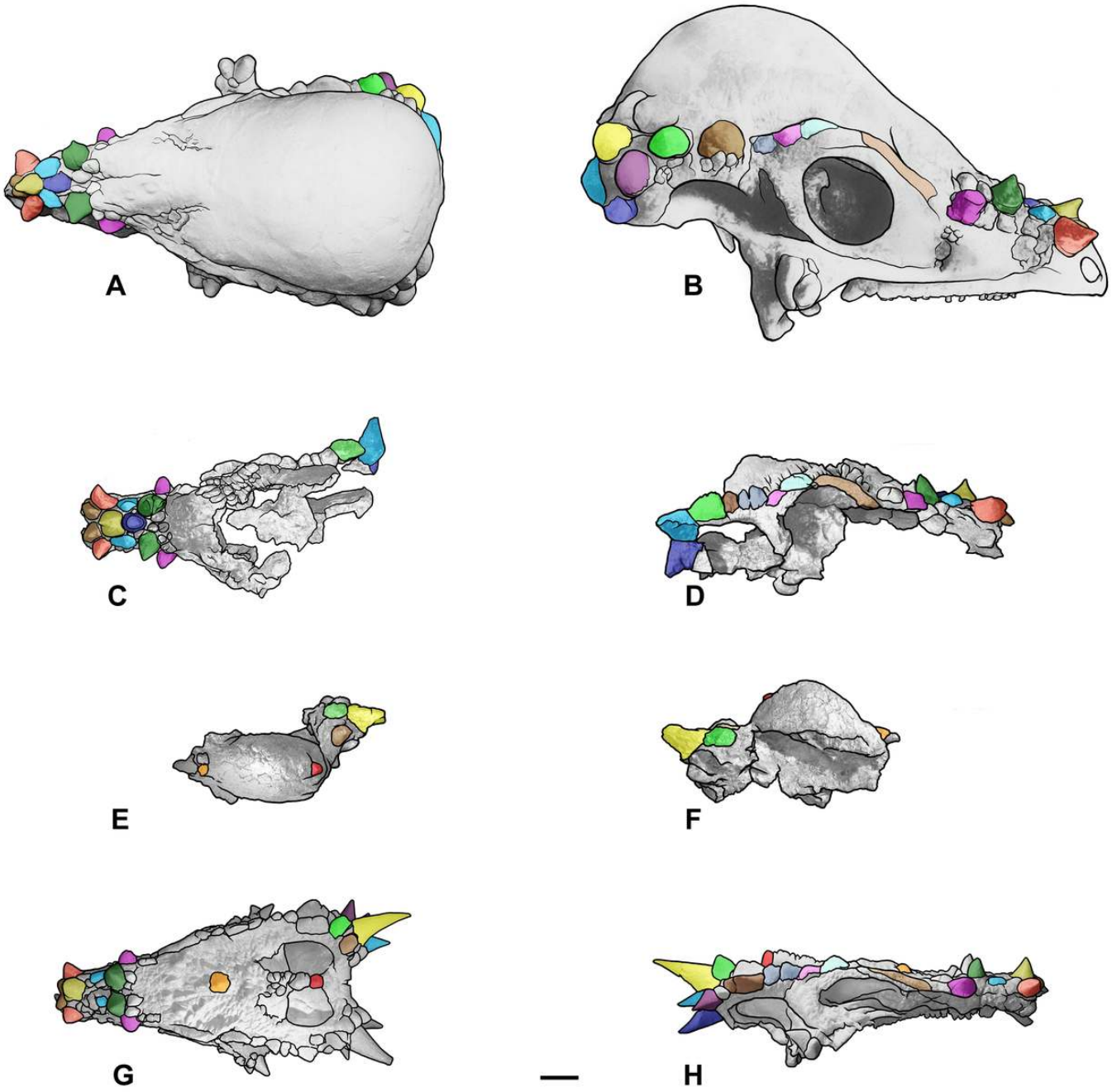


Figure 33.2.8: Analysis by Horner and Goodwin showing how many features of Dracorex and Stygimoloch can be seen in Pachycephalosaurus.

Pachycephalosaurus, the 'thick-headed lizard' as its name describes, was first identified in 1931. In 1983, a relative of *Pachycephalosaurus* named *Stygimoloch* ("Demon of the River Styx") was named, and in 2006, another relative named *Dracorex hogwartsia* ("Dragon King of Hogwarts") was named. Just a year later, scientists like Jack Horner had started to question these other species as growth stages of Pachycephalosaurus. In 2009 a study was published by Horner and Goodwin showing that specific points on the skull of each individual species can easily be correlated, supporting the idea that *Dracorex* being a juvenile, *Stygimoloch* being a sub adult, and *Pachycephalosaurus* being an adult and the proper name of the genus. This had spread to other members of the group as well; other flat-headed pachycephalosaurs (like *Goyocephale* and *Homalocephale*) have been proposed as juveniles of other genuses.

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33.3: To Feather or Not To Feather - Dinosaur Body



Early Feathers

Our interpretations about the coverings of dinosaurs have gone through many iterations over the years. The first dinosaur reconstructions were entirely scaly, in keeping with reptilian ancestry, even with finds like *Archaeopteryx* pointing to a feathered linage. Early fossil hunters reportedly only cared for the bones, and potentially hundreds of skin and feather impressions were lost to their carelessness. After the dinosaur renaissance and new finds in the later half of the 20th Century, feathers became more accepted. If anything, the feather debate has swung too far, suggesting all dinosaurs were covered in fuzz and fluff. Most dinosaurs have no remains that indicate feathers or not, so plenty is still unknown about dinosaur skin covering. But let us review what is known.

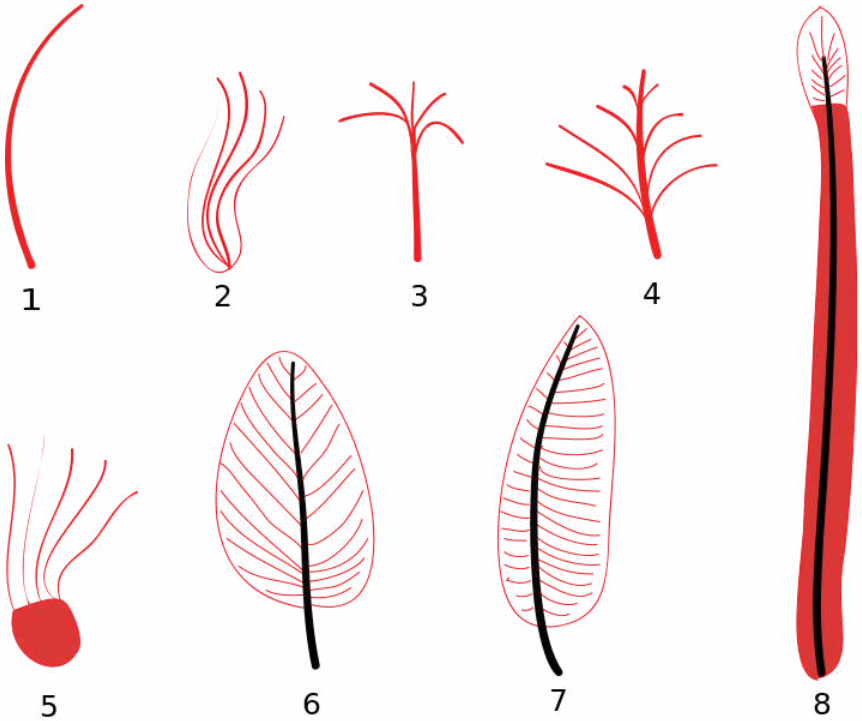


Figure 33.3.1: Diagram showing the stages of feather evolution as outlined by Xu & Gou 2009. 1. Single filament 2. Multiple filaments joined at their base 3. Multiple filaments joined at their base to a central filament 4. Multiple filaments along the length of a central filament 5. Multiple filaments arising from the edge of a membranous structure 6. Pennaceous feather with vane of barbs and barbules and central rachis 7. Pennaceous feather with an asymmetrical rachis 8. Undifferentiated vane with central rachis. (Drawing by Matt Martyniuk via Wikimedia.)

How and why did feathers evolve in the first place? Feathers are not fundamentally different from other vertebrate animal body coverings. The proteins found in feathers, mammal hair, and nails are keratin, similar to what is found in scales of reptiles. This means that feathers can be considered a homologous trait via divergent evolution from a common ancestor with modern reptiles. According to a recent overview of feather evolution (Xu and Guo, 2009), feathers first evolved as a single hair-like object and eventually became more complex. Most scientists infer that feathers would have started out for insulation, and eventually been adapted for service in new functions, such as display and flight. This is corroborated by the fact that smaller dinosaurs have been the most often found with feathers and other coverings. The smaller the animal, the higher the animal's body surface area to volume ratio, meaning heat loss can occur more quickly. Feathers could reduce this heat loss.

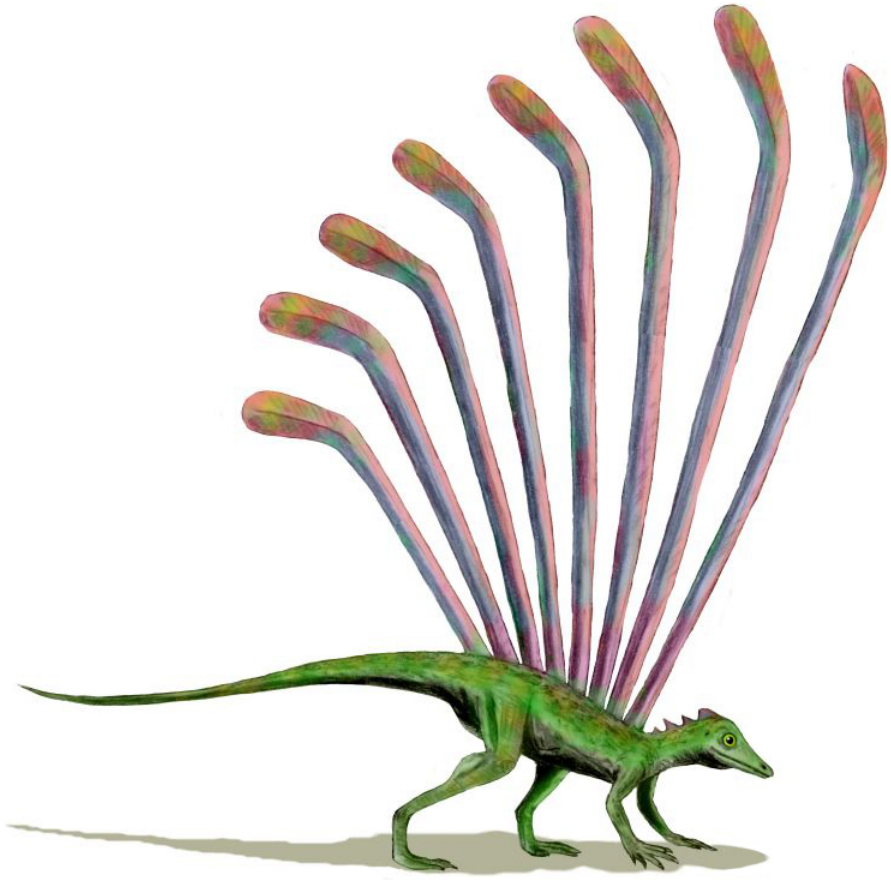


Figure 33.3.2: Sketch of Longisquama (by Nobu Tamura via Wikimedia.)

The fossil record does not give us many early examples of feathers, so the initial origin of feathers as a biological structure is hard to know. What is known is that almost all examples of feathers are found on dinosaurs, both avian and non-avian. One possible exception to that rule is found on the enigmatic fossil reptile *Longisquama*, known from the Triassic of Kazakhstan. Its name, meaning "long scales," refers to the large feather-like protrusions sticking from the animal's back. There is much debate over the purpose of these features, or if they are truly feathers in the first place. One animal that *Logisquama* may be related to is *Coelurosauravus*, an early reptile which has a membrane protruding from its sides, and presumably used for gliding. Could *Logisquama*'s long scales be used for gliding, too? Or were they for display to attract mates or identify members of its kind? More specimens and more discoveries of related animals may need to be found before we can know for sure.

Which dinosaurs had feathers?

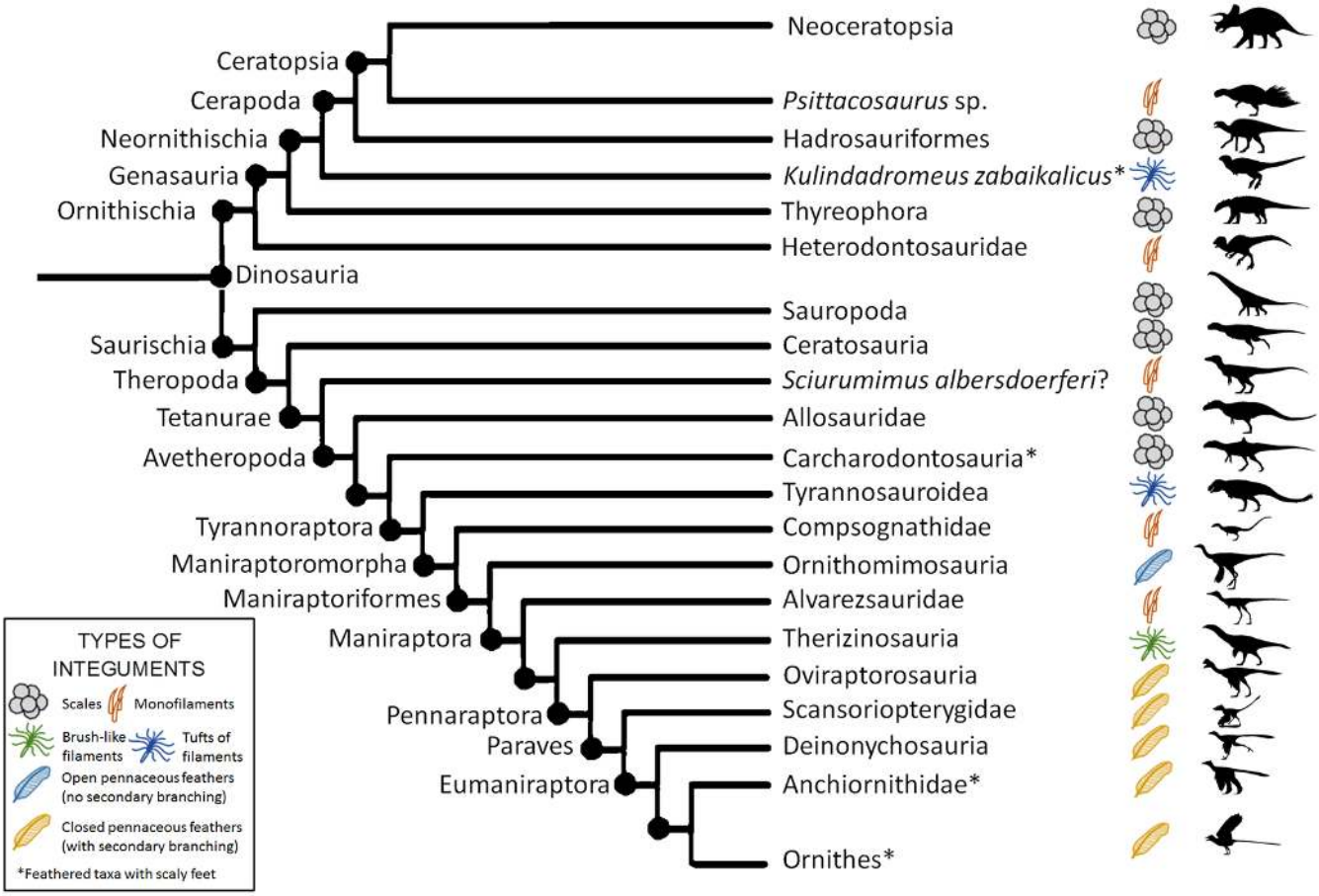


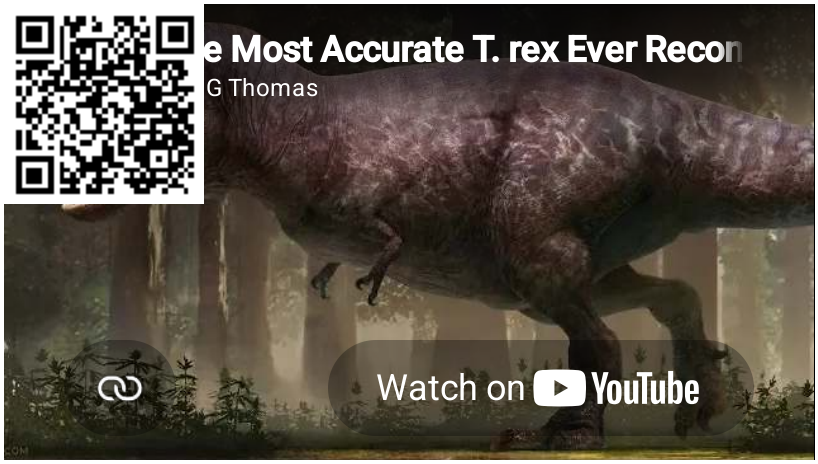
Figure 33.3.3: hylogeny and integument description based on Benton et al. (2019). Images of scales and feather types taken from File:Distribution of feathers in Dinosauria.jpg. Silhouettes from works by Slate Weasel, Matt Martyniuk, Audrey.m.horn, Tom Parker, Michael B. H. and Gareth Monger. (Cladogram by Kiwi Rex via Wikimedia.)

Dinosaurs with feathers or feather-like structures have been found on several branches of the dinosaur family tree. Interestingly, even some ornithischian dinosaurs had feather-like filaments. The pterosaurs had pycnofibres, which were dense bristle-like structures. While these were not exactly feathers, recent studies have shown more variety suggesting that some of these might correspond to dinosaur feathers. Another example, *Psittacosaurus*, a small early member of the ceratopsian lineage, has been found with a large number of filaments sticking out of its back, from the hip to the mid-tail. Full pennaceous (flying style) feathers, like those seen in modern birds, are found only in maniraptorans, with the most basal example from the ornithomimids. This means that the famous raptors of the *Jurassic Park* movies would almost certainly have been feathered. In fact, *Velociraptor* itself has been found with quill knobs, small impressions in the bone of the forearm where a feather would originate. Of course, complex feathers and significant coverings are only found on the coelurosaur branch of the theropods, with some paleontologists thinking all coelurosaurs had feathers.



Figure 33.3.4: Photo of Yutyrannus tail feathers (by Kumiko via Wikipedia.)

How about the *Jurassic Park* co-star, T*yrannosaurus rex*? In 2012, a dinosaur named *Yutyrannus* was described. Meaning "feathered tyrant," it is the largest dinosaur yet found with evidence of feathers. It has also been phylogenetically placed within the superfamily Tyrannosauridea, and thus could properly be called a tyrannosaur. *Tyrannosaurus* is considered further down the evolutionary tree with respect to *Yutyrannus*. Does that mean *T. rex* had feathers?



The most up-to-date information, based on both *T. rex* and closely-related large animals, has concluded that the feathers they had were scarce at best. The video game *Saurian* compiled this information based on published studies to make possibly the most accurate reconstruction yet. Despite this, a 2019 American Museum of Natural History exhibit shows *Tyrannosaurus* with a feathery body, making the argument that related animals indicate feathers were possible. The exhibit also shows a juvenile with feathers. This is another possibility: even if the adult version was bald, hatchlings may have needed the 'down' and could have lost it as they aged.

Dinosaurs to Birds

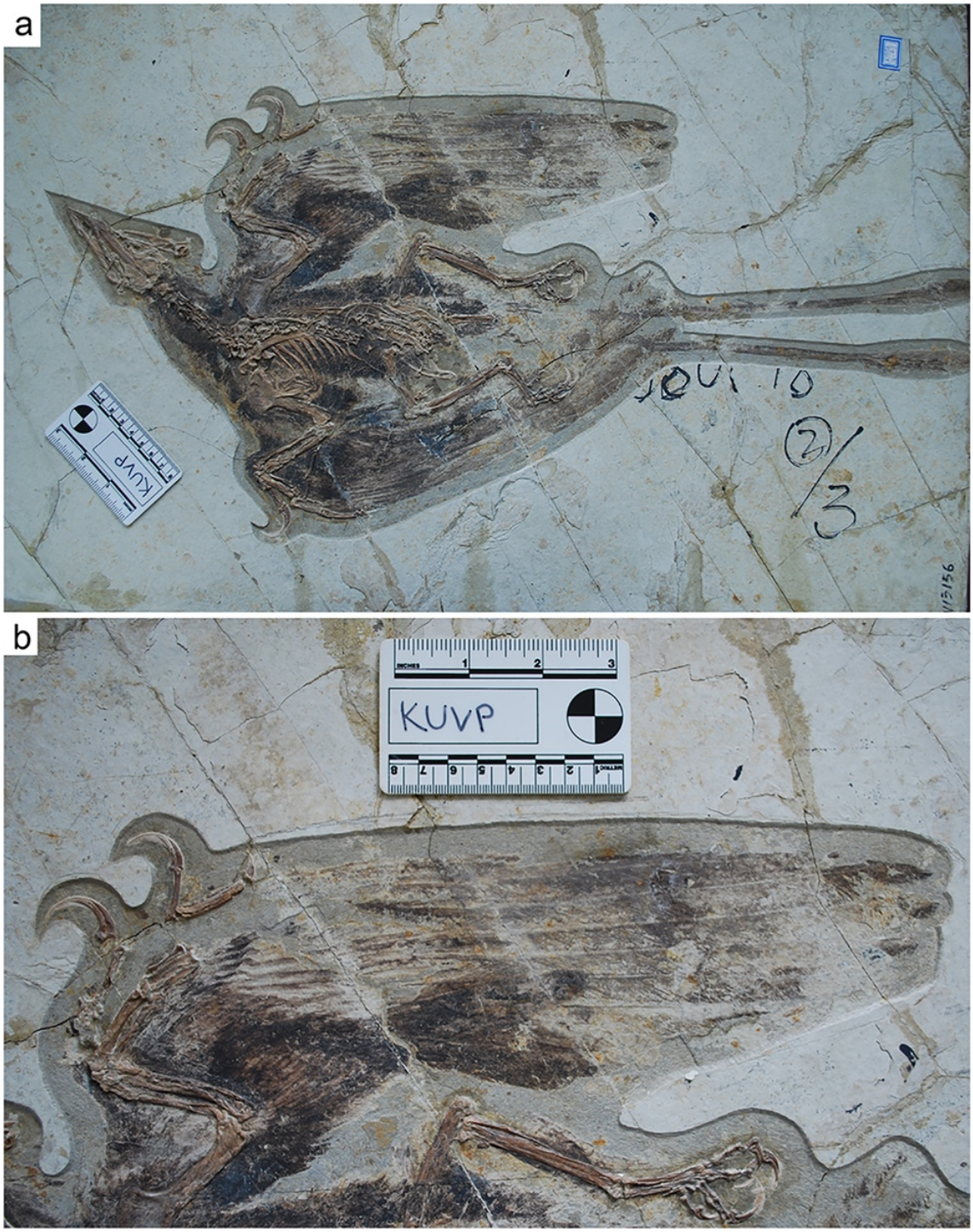


Figure 33.3.5: Photograph of Confuciusornis specimen (a) and close up of wing (b) IVPP V13156 (by Amanda R. Falk, Thomas G. Kaye, Zhonghe Zhou, David A. Burnham via Wikimedia.)

As evolution progressed, several non-avian dinosaur groups became more and more bird-like, with feathers being only part of the story. This group of animals belongs to the clade Parves, which means “next to birds.” The ornithomimids, for example, derive their name from *Ornithomimus*, whose name means “bird mimic.” This is due to their toothless beaks and overall body plan that resembled an ostrich. Other dinosaurs, like *Microraptor*, had an even more bird-like appearance, with long flight-like feathers on its arms… and also its legs! *Microraptor* was a dromaeosaur, in the same group as other famous animals like *Velociraptor* and *Deinonychus*. The dinosaur group that is most closely related to birds but not technically birds are the troodontids. The first non-pterosaur that could have flown was *Archaeopteryx*. It is part of a group called Avialae, “bird wings.” This is the group where the line between non-avian dinosaurs and birds becomes blurry. Euornithies is the clade that contains true birds, and a subgroup of that is called Aves, the crown group of modern birds.



Figure 33.3.6: Reconstruction of Archaeopteryx in a tree (by Nobu Tamura via Wikimedia.)

One of the biggest ongoing debates in bird evolution is how flight evolved itself. There are two main camps on this question. One idea is that feathers enabled these theropods to be more agile on the ground. As their agility increased, feather complexity also increased, eventually aiding their flight. The other idea proposes these animals already lived in trees, and were gliding from branch to branch. Their feathers assisted in gliding, and eventually led to flight. There is also an in-between, in which the wings helped with climbing, as some modern birds have been experimentally shown to do today. There are compelling arguments on all sides, so no one explanation has yet achieved widespread acceptance.

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33.4: Movement and thermoregulation - Get that blood flowing!

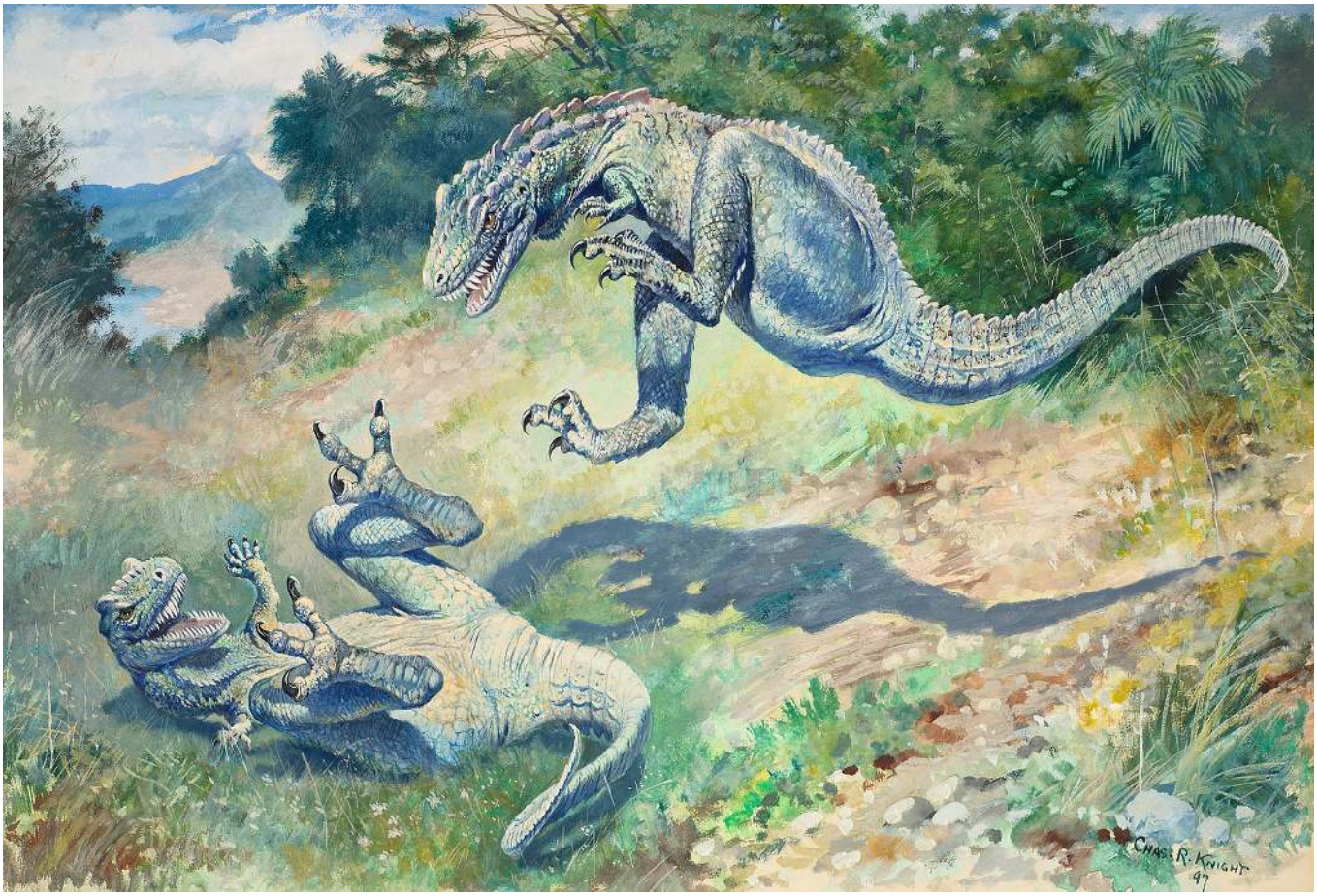


Figure 33.4.1: Leaping Laelaps, an 1897 painting by Charles R. Knight, might be the first depiction of active dinosaurs.

How dinosaurs moved and how dinosaurs regulated their internal temperature has been debated and controversial since their discovery. What may be less obvious is how much these ideas are interlinked. Dinosaurs' closest living relatives, crocodiles, are ectothermic (commonly known as cold blooded). This means they rely on their environment for warmth. This has advantages and disadvantages: less resources are needed, but they are less adaptable to cold climates and have limitations on endurance. Crocodiles, by the way, have several advantages over their ectothermic kin. They have a 4-chambered heart for better blood exchange efficiency, a secondary palate to allow breathing while eating, and the ability to walk upright (known as a high walk) for more efficiency in movement. These are adaptations that mammals share, and most scientists think dinosaurs had these adaptations as well.



Figure 33.4.2: An Emperor Penguin jumping out of icy water in Antartica. (Photo by Christopher Michael.)

Birds, the direct descendants of dinosaurs, are mostly endothermic (commonly known as warm blooded). That means birds (and the other endotherms, mammals) can live in many more climates across Earth and have a higher metabolism. This allows for faster movement but also requires more resources. It should be noted that there are also mesotherms, animals that are not quite warm blooded or cold blooded, but in between. Another complexity is homeothermy, the ability to maintain body temperature so that metabolic processes can be maintained. This is related to endothermy, but is distinct, because some endotherms do not have homeothermy. The question remains: where do dinosaurs fit into this spectrum? And how do we find out?

Early dinosaur movement



Figure 33.4.3: Outdated Megalosaurus model from Crystal Palace in London. (Photo by C.G.P. Grey.)

When dinosaurs were first described, they were thought to be related to modern reptiles, and were imagined as just larger versions of them. They were slow and lumbering in the minds of scientists at the time and were assumed to be ectothermic as well. They were mounted with sprawled legs and dragging tails. Evidence started to build that dinosaurs were related but distinct from living reptiles on Earth today. For example, dinosaurs have their legs positioned under their bodies. This is not just an aesthetic choice; a sprawled posture is much more energy intensive, so dinosaurs were more efficient than their reptile brethren. Also, sprawled posture interferes with breathing by flexing the lungs (called Carrier's Constraint), causing most sprawled animals to hold their breath while moving. Pontzer et al. (2009) argued that the posture of dinosaurs links them to endothermy alone. Even in 1869, Thomas Henry Huxley had concluded that dinosaurs were more closely related to birds as part of his defense of the newly-proposed theory of evolution by Charles Darwin. These ideas did not help the image of dinosaurs in the minds of the public, however. Even into the mid 20th Century, dinosaurs were nothing but big cold-blooded lizards.

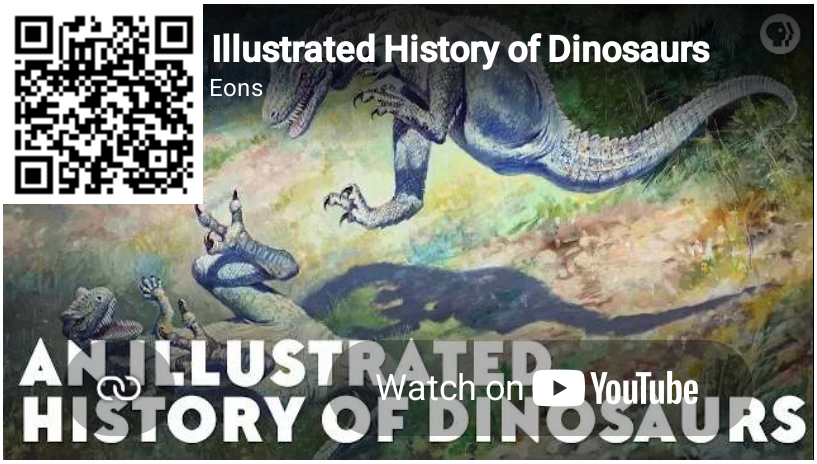




Figure 33.4.4: Deinonychus skeleton from the Field Museum in Chicago. (Photo by Johnathan Chen.)

Starting in 1964, John Ostrom, his student Robert Bakker, and others discovered and described the animal *Deinonychus*. With its small, light frame and bird-like posture, it was a great example that went against the slow and sluggish stereotype dinosaurs had in public opinion. This started what is now known as the Dinosaur Renaissance. Ideas that changed during this time include: a closer evolutionary connection to birds, a more nimble posture (without dragging tails), stronger ideas about herd behavior, and a more open mind on endothermic thermoregulation. The current idea about the extinction of the non-avian dinosaurs by a meteor impact also was developed during this time.

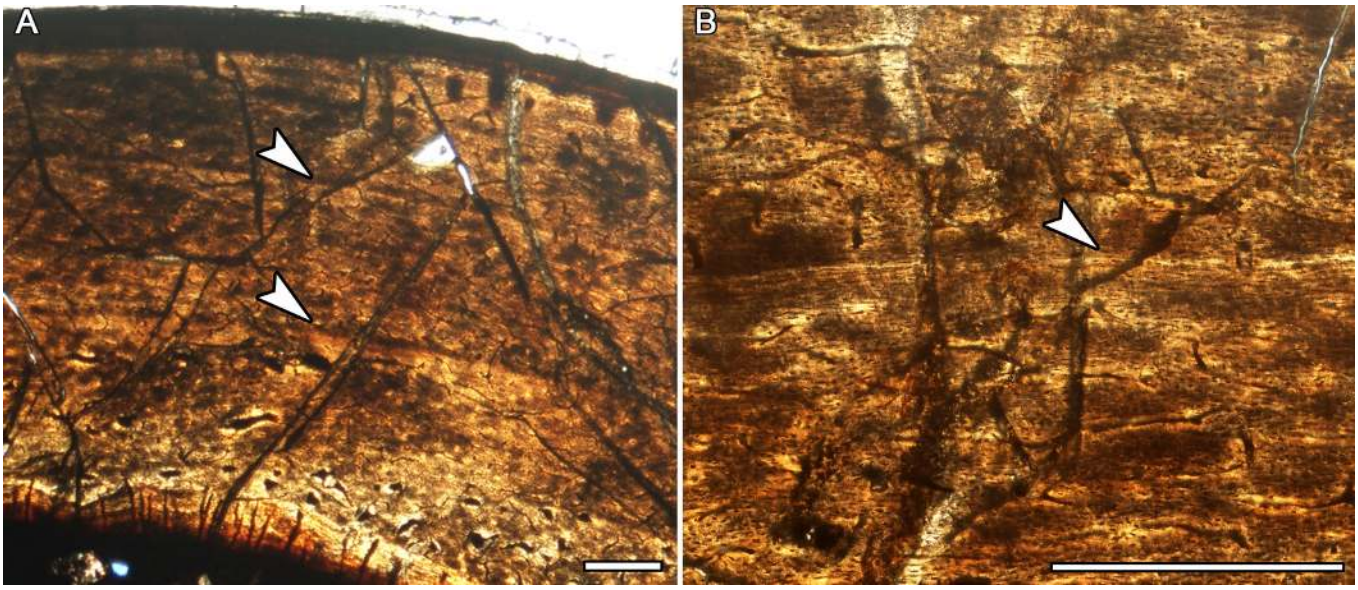


Figure 33.4.5: Bone microstructure of polar hypsilophodontid. Arrows indicate annuli, places where growth would have stopped in the winter. Scale bars, 500 µm. (A) is femur, (B) is zoomed in. (Photo by Holly N. Woodward, Thomas H. Rich, Anusuya Chinsamy, and Patricia Vickers-Rich.)

This revolution in the thinking about dinosaurs has basically continued to this day, but the debate over dinosaur thermoregulation is still ongoing, mostly because it is harder to decipher than some of the other debates. One practice that has increased our understanding of physiology of dinosaurs is histology: the microscopic study of tissues, namely bone. Several studies have found

that dinosaurs had fast growth rates. They also had fibrolamellar bone and Haversian canals, both of which are common in endothermic animals (but not exclusive to them).

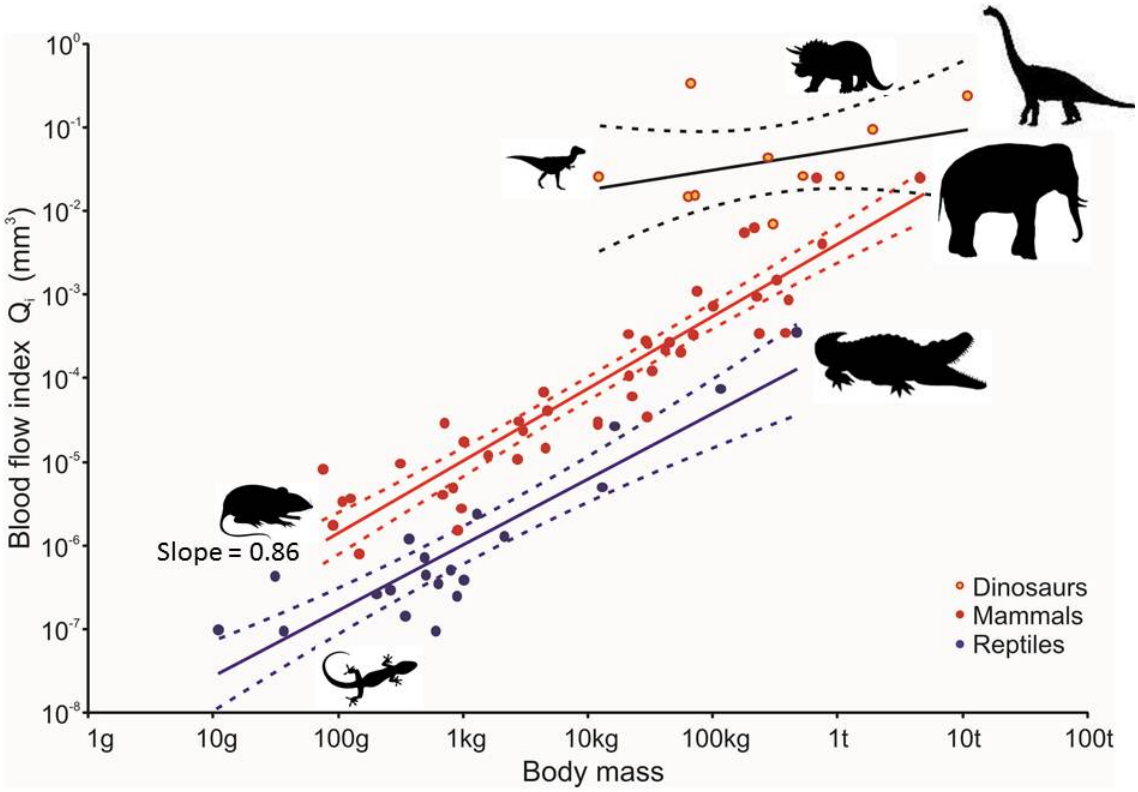


Figure 33.4.6: Graph of blood flow verses body mass, showing dinosaurs are closer to mammals than reptiles. (By Cervical67 via Wikimedia.)

Bone structures can also estimate blood flow, with high blood flow being attributed to endothermy. By looking at the small holes throughout bones, called nutrient foramina, and comparing the size of the holes to the size of the animal, you can estimate whether the animal is endothermic or ectothermic. As you can see from the figure, dinosaurs match much more closely with mammals, and may have been highly active, and possible endotherms as a result.

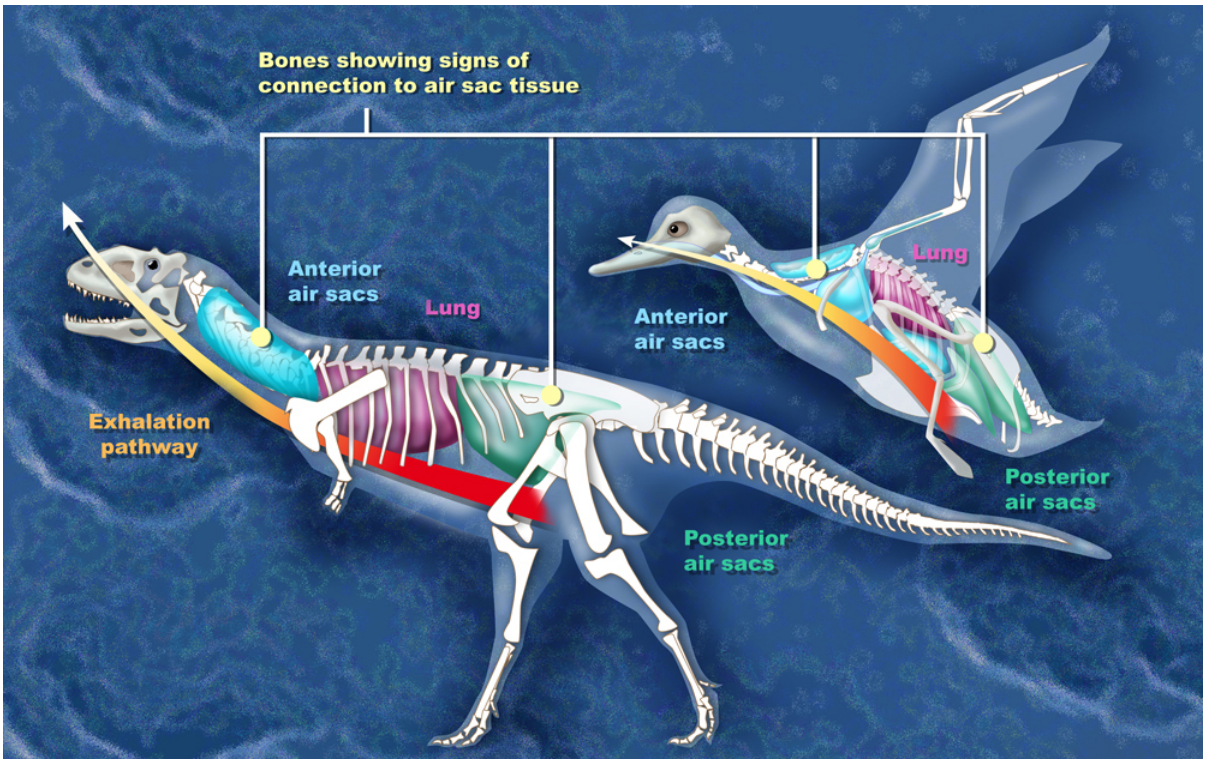


Figure 33.4.7: Image showing parallels between bird air sacs and proposed dinosaur air sacs. (By Zina Deretsky, National Science Foundation)

Another clue comes from breathing. Some scientists think dinosaurs had an air sac system similar to birds. The sacs are found mostly in their skulls and vertebrae, and would have allowed more efficient exchange of oxygen. They have also been pointed to as a weight-saving system in larger species. Of course, these features are almost impossible to fossilize. While there is no doubt that dinosaur descendants (birds) have them, where they developed in dinosaur's evolutionary history is very much up for debate. Theropods, the branch of the dinosaur tree that turned into birds, have the best evidence for possible air sacs in non-avian dinosaurs.

Endotherm without Endothermy?



Figure 33.4.8: Life sized model of Argentinosaurus at Clore Garden of Science in Weizmann Institute of Science in Israel. (Picture by MathKnight and Zachi Evenor via Wikimedia.)

The paragraphs above were just a partial list of evidence that could be used to explain if dinosaurs were endothermic or ectothermic, there is another factor to consider: dinosaurs could have been functionally endotherms without being actual endotherms. How? By keeping in their heat. This could have happened in two main ways. First, feathers (or similar proto-feather structures) may have evolved as a warmth structure first, then became used for flight later. Secondly is an idea that is called inertial endothermy, bulk endothermy, or gigantothermy. This suggests the large size of dinosaurs, with selective blood pumping taking advantage of surface area to volume ratio, could have had high blood temperatures without the food costs of endothermy. Of course, smaller dinosaurs and the young of larger dinosaurs would not have had this advantage, but feathers on the smaller animals or young could have helped.

Dino Debate – Did I Get It?

? Exercise 33.4.1

Which of these evidences for dinosaur social behavior is hardest to find?

a. Toothmarks on bones

b. Individuals interacting with each other

c. Nest sites

d. Many individuals found in close proximity

e. Footprints and trackways

Answer

b. Individuals interacting with each other

Exercise 33.4.2

Which of these dinosaurs is most likely not a valid species?

a. Apatosaurus

b. Brontosaurus

c. Triceratops

d. Torosaurus

e. Stygimoloch

Answer

e. Stygimoloch

Exercise 33.4.3

The first feathers developed in dinosaurs are most likely used for what?

a. Warmth

b. Climbing

c. Sexual display

d. Floating

Answer

a. Warmth

Exercise 33.4.4

The dinosaur renaissance most changed the way we think about:

a. dinosaur size

b. dinosaur art styles

c. dinosaur parental behavior

d. dinosaur food consumption

e. dinosaur movement

Answer

e. dinosaur movement

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CHAPTER OVERVIEW

34: (Case Study) Western Mesozoic orogenies

Learning Objectives

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

- Discuss changes to the tectonic setting along the western coast of North America throughout the Mesozoic
- Describe individual similarities and differences between the Nevadan, Sevier, and Laramide orogenies
- Evaluate changes in deformation style and location of deformation in these orogenies over time

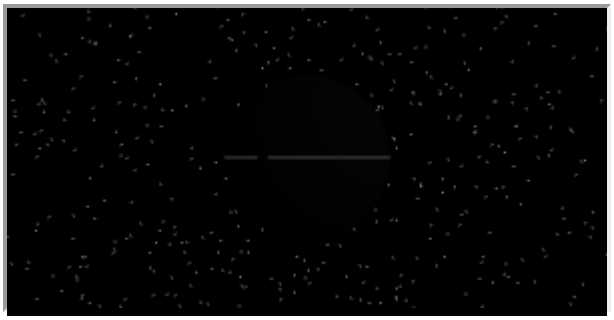
34.1: Western Mesozoic Orogenies

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34.1: Western Mesozoic Orogenies

Introduction

In the Paleozoic, after the breakup of Rodinia, Laurentia (the name for the continental core of North America) was an island continent near the equator. It was a time of great biological productivity, firstly in the shallow seas left behind by Rodinia's breakup, and later on land as the first swamps emerged and spread across the continent. Most of the tectonic activity was found in lands that make up the modern-day Eastern Seaboard of the United States, with the Taconian, Acadian, Alleghanian, and Ouachita orogenies helped to build North America and assemble Pangaea. Below is a look at the Earth around the start of the Taconian Orogeny.



Early in the Paleozoic, the western side of North America was a relatively quiet passive margin. The enigmatic Antler orogeny occurred in the Devonian and Carboniferous, but the cause and extent of it are disputed. That was followed by the even more controversial Sonoma orogeny which may have occurred as late as the early Triassic. Evidence for both the Antler and Sonoma orogenies is only found in Nevada. The only other Paleozoic-aged tectonic activity in western North America is the uplift of the Ancestral Rockies, deep-seated uplifts that formed inland, in present-day Colorado and New Mexico. These may be far-reaching byproducts of other orogenies that were going on at the time (Ouachita or Sonoma orogenies).

Nevadan Orogeny



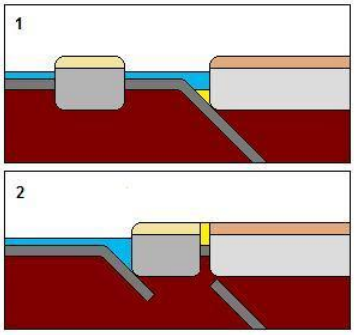


Figure 34.1.1: An illustration of a terrane docking. As an island arc comes toward the main continent and collides (docks), the subduction zone switches to the other side of the island. (By Actualist via Wikimedia.)

The first major orogenic event in the Mesozoic of North America was the **Nevadan orogeny**. This occurred on the other side of North America from the breakup of Pangaea and started later, in the mid-Jurassic. At some point prior to the Nevadan deformation, an active margin had developed with a volcanic arc along the west coast. These volcanic rocks and related sediments were deformed with the accretion of numerous terranes from the ancient Pacific Ocean, brought eastward as the Farallon Plate (along with the Kula plate) subducted beneath the North American Plate. The Farallon Plate used to exist between the Pacific Plate and the North American Plate.

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Figure 34.1.2: Location of the Klamath Mountains in California and Oregon. (By Alexrk2 via Wikimedia.)

The Farrallon's western boundary with the Pacific Plate was divergent: it was a spreading center (an oceanic ridge). The Farallon's eastern boundary (with the North American Plate) was convergent: a subduction zone. There, the Farallon was recycled into the mantle. Any island arcs, microcontinents, or oceanic plateaus it was carrying were too thick, buoyant, or protruding too much to subduct. When they reached the subduction zone, these terranes were scraped off and pushed onto the North American continent, adding to its volume. As they were pushed eastward, they acted like a bulldozer, scraping and shoving the volcanic and sedimentary rocks that were layered atop the continental crust. The deformation of the Nevadan orogeny is most easily seen in the Sierra Nevada mountains of California and the Klamath mountains along the coast of California and Oregon.

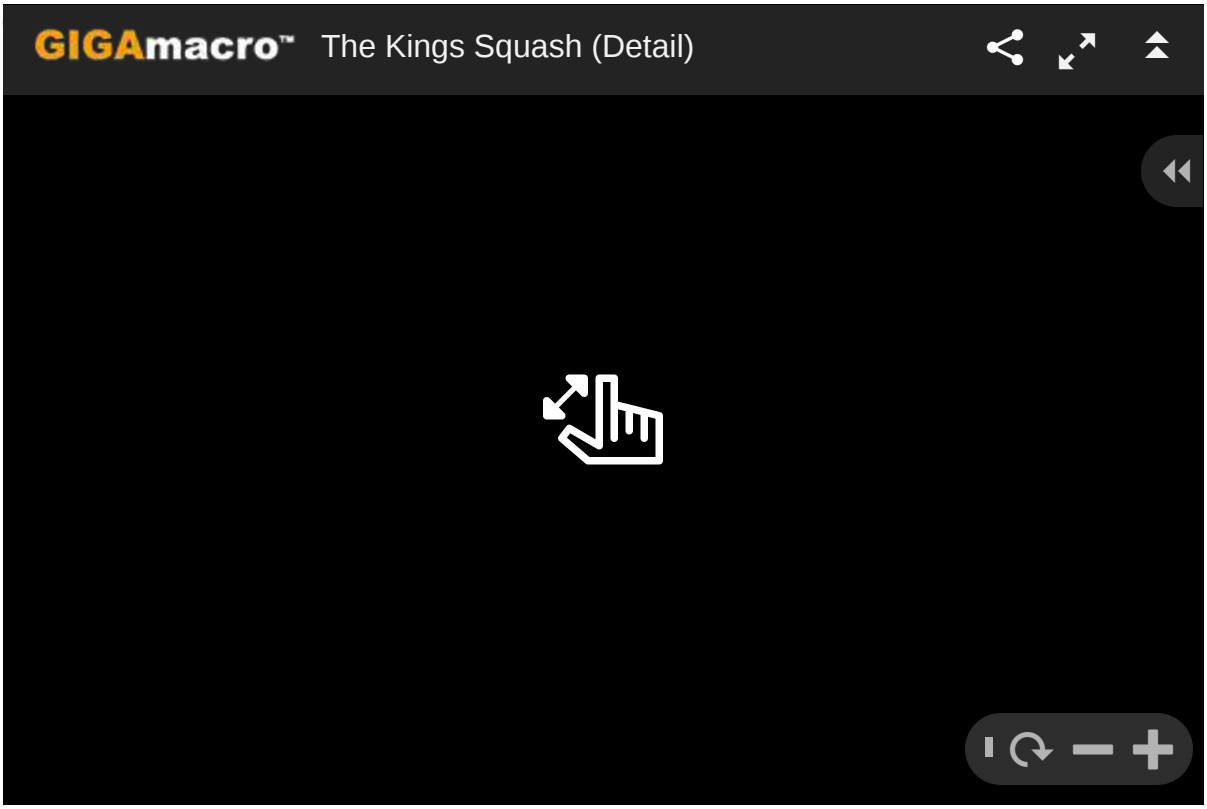


Figure 34.1.3: Marbles of the Boyden Cave Roof Pendant show the results of deformation culminating with batholith emplacement. In addition to the intricate folding displayed here, don't forget to appreciate the patterns that recent weathering has superimposed on the exposure.

Did I Get It? – Nevadan

? Exercise 34.1.1

Which of the following orogenies was not on the west coast of North America in the Paleozoic?

a. Sonoma

b. Antler

c. Taconian

d. Ancestral Rockies

Answer

c. Taconian

? Exercise 34.1.2

Where can the rocks which were part of the Nevadan Orogeny be found?

a. Montana

b. Oregon

c. Idaho

d. Utah

e. Washington

Answer

b. Oregon

Exercise 34.1.3

Which tectonic setting caused the Nevadan orogeny?

a. Continent-continent collision

b. Mid-ocean ridge

c. Subduction

d. Transform

e. Rift

Answer

c. Subduction

Sevier Orogeny





Figure 34.1.4: Generalized map of Sevier deformation. (Image by Quickdraw123 via Wikimedia.)

The next orogeny to occur in western North America was the **Sevier orogeny**. While terranes still played a factor in the deformation, the Sevier orogeny occurred much further inland and was more widespread than the Nevadan orogeny. The earliest events associated with the Sevier orogeny are as early as 160 Ma (even before the Nevadan) and the latest deformation was 50 million years ago. However, the majority of this deformation occurred between 120 and 80 million years ago. Subduction at this time changed due to the fact that the Farallon plate was hot and young, causing a Chilean-style subduction zone.

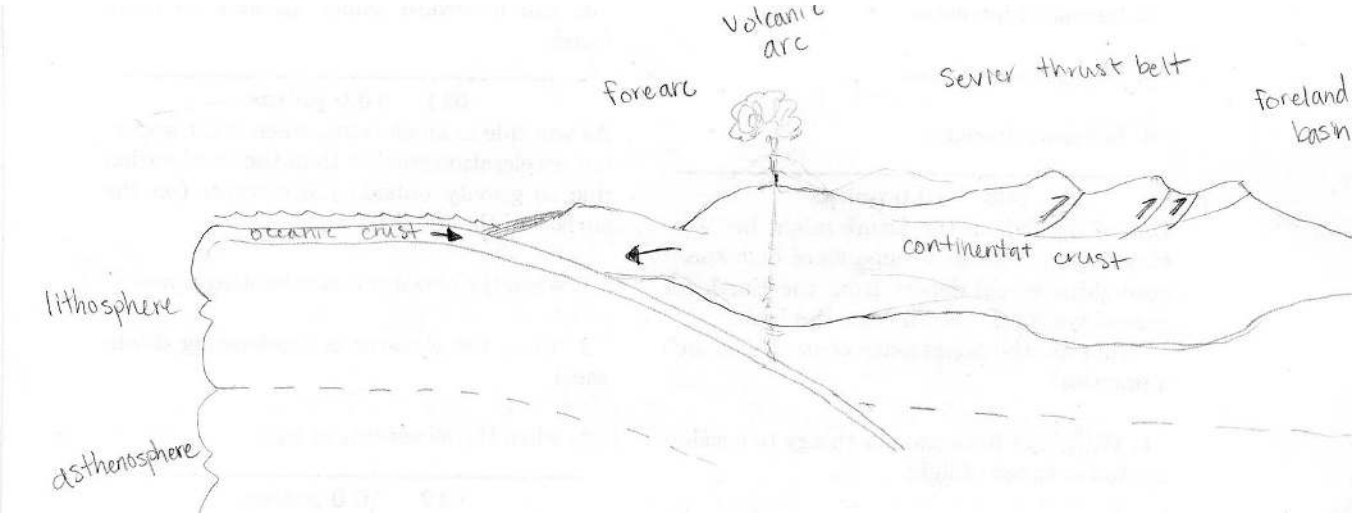


Figure 34.1.5: W-E Sketch of general features of the Sevier Fold-and-Thrust Belt. (Drawing by Pinkcorundum via Wikimedia.)

This means the resistance to subduction caused extra convergent forces in the overriding plate, mainly in the form of a fold-and-thrust belt, which extended inland. Much like the modern Andes, this created a two-ridged mountain chain, with the volcanic arc positioned closer to the trench, and a back-arc mountain range formed from compressional, low-angle thrust faulting. Between these two ridges today is a high plain, called the Altiplano in the Andes. While the exact geography of the mountains during Sevier time is unknown, it can be assumed that it was similar. Igneous rocks, closer to the Sierra Nevada, and shoved stacks of sedimentary rocks far from their original positions deep in the crust and further to the west back up this assertion. These displaced layers of rock arched upward and broke, with the slabs of older rock traveling along thrust faults up on top of younger rocks.

Consider the nascent mountain-building evident in western North America in this paleogeographic depiction from 105 Ma:



Those mountains were driven by compression from the west. As new terranes accreted to the continent, they pushed the pre-existing sedimentary layers out of the way, so that direction was to the east. This process initiated during the Jurassic period of geologic time, but reached its zenith during the Cretaceous. Most of the Farallon Plate was subducted during that time, but two remnants persist into the present day: the Cocos Plate, currently being subducted beneath Central America, and the Juan de Fuca Plate, currently being subducted beneath the Pacific Northwest states of Washington, Oregon, and northern California.

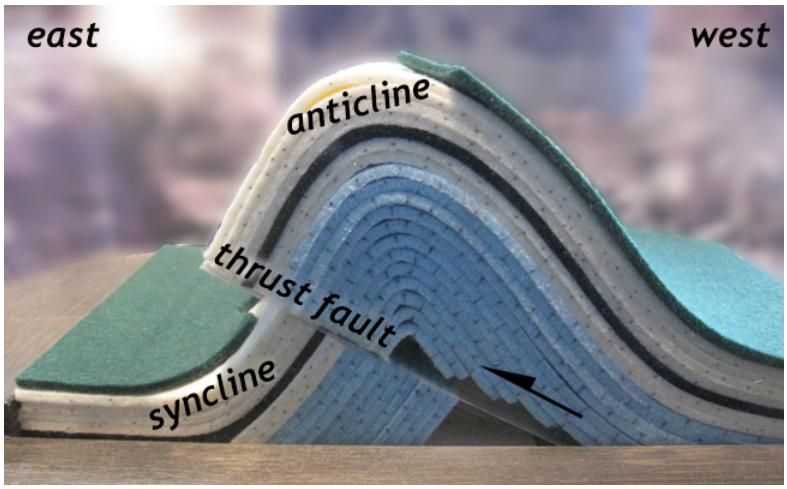


Figure 34.1.6: This felt model of Turtle Mountain, Alberta is on display at the Frank Slide Visitor Center. The structure shown is very similar to the structure of many faults and folds in the Rockies caused by Sevier deformation.

When the sedimentary layers were compressed, they buckled and broke, shearing to the east as they deformed. The result was an immense number of asymmetric folds with the short limb on the east and the long limb on the west, frequently breaking at the hinge of the fold, and initiating a fault. These distinctive structures are the typical structural ingredient in active and ancient compressional mountain belts around the world. This style of deformation is known as thin-skinned deformation. This is referring to these cover rocks, mostly sedimentary, that are involved in the deformation. Crystalline basement rocks are not typically involved in this process. A great example of this occurs in Glacier National Park.

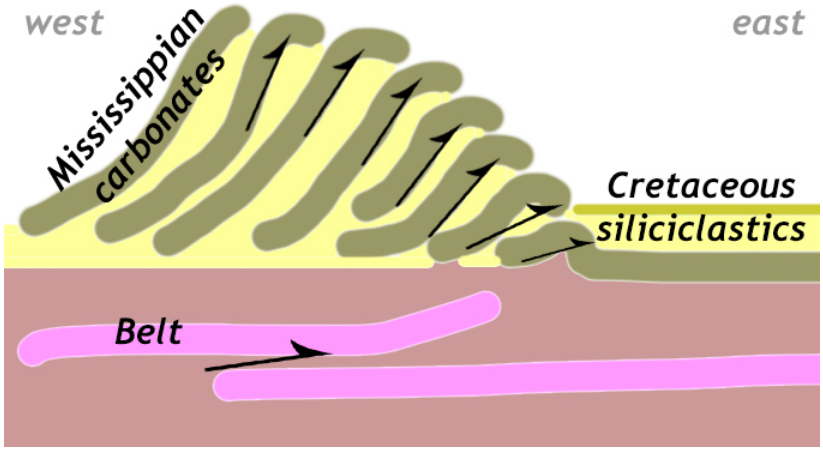


Figure 34.1.7: Cartoon cross-section through an imbricate stack of thrust sheets, as seen at Sun River Canyon, near Glacier National Park.

In general with faults, the footwall rocks are those below the fault, while the rocks above the fault are called the hanging wall. However, additional terminology is used when it comes to thrust faults, with their significant lateral offsets, and their orientations so close to horizontal. The block or slab or rock above the thrust fault is sometimes called a thrust sheet, or by the term nappe.

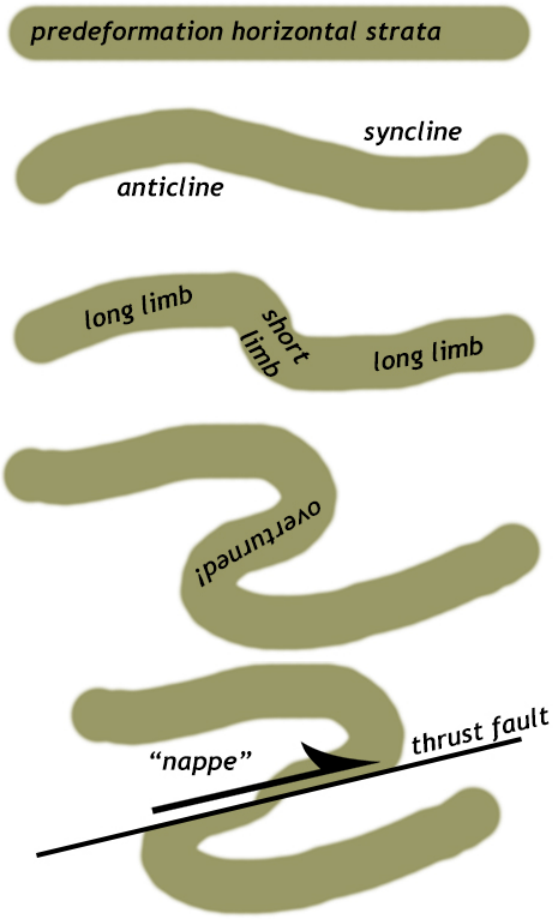


Figure 34.1.8: A cartoon cross-section showing 5 steps in the folding and thrust faulting of an asymmetric anticline to make a "nappe."

Nappes form when horizontal rock layers are compressed laterally (*i.e.*, from the side) and first buckle into asymmetric folds. The short "limb" of these anticlines and synclines forms in the transport direction (the direction toward which the strata are being thrust). Frequently, this short limb rotates back on itself and becomes tectonically inverted. It also frequently is the place where the strata rupture, and a new thrust fault initiates. Once the mechanical discontinuity of the fault exists, it's the "weakest link," and further deformation is accommodated mainly through sliding along that fault surface, not additional folding. Eventually, the fault stops moving, and a new fault forms "outboard" of the first (*i.e.*, further away from the source of the compression). As deformation proceeds, the older thrust sheets get shoved backward and out of the way as new thrusts form beneath them, rising up to the surface. The result is that the older, more western thrust sheets get rotated to steeper dip angles as deformation proceeds. This whole "stack" of nappes is said to be "shingled," or "imbricated."



Figure 34.1.9: Sandbox deformation experiment showing how layers of sand fold and fault due to lateral compression. This closely mimics the evolution of fold-and-thrust faults in the Sevier orogeny (except in this model, compression comes in from the east instead of the west). (Model and source video by Marco Martins-Ferreira.)

Watch the animated GIF here of a sandbox deformation experiment to get a better sense of this process. It's useful to watch it many times, focusing on a different region each time and watching how it changes with increasing deformation. Imagine that the perspective here is the same as you would see if you were (a) able to go back in time and (b) hover above the crust and look down the axis of the range front, and (c) watch geological processes happen at much faster rates than they really do, and (d) had a magical pair of X-ray glasses that allowed you to peer into the Earth and see these complicated 3D structures evolve in a 2D cross-section.

Laramide Orogeny

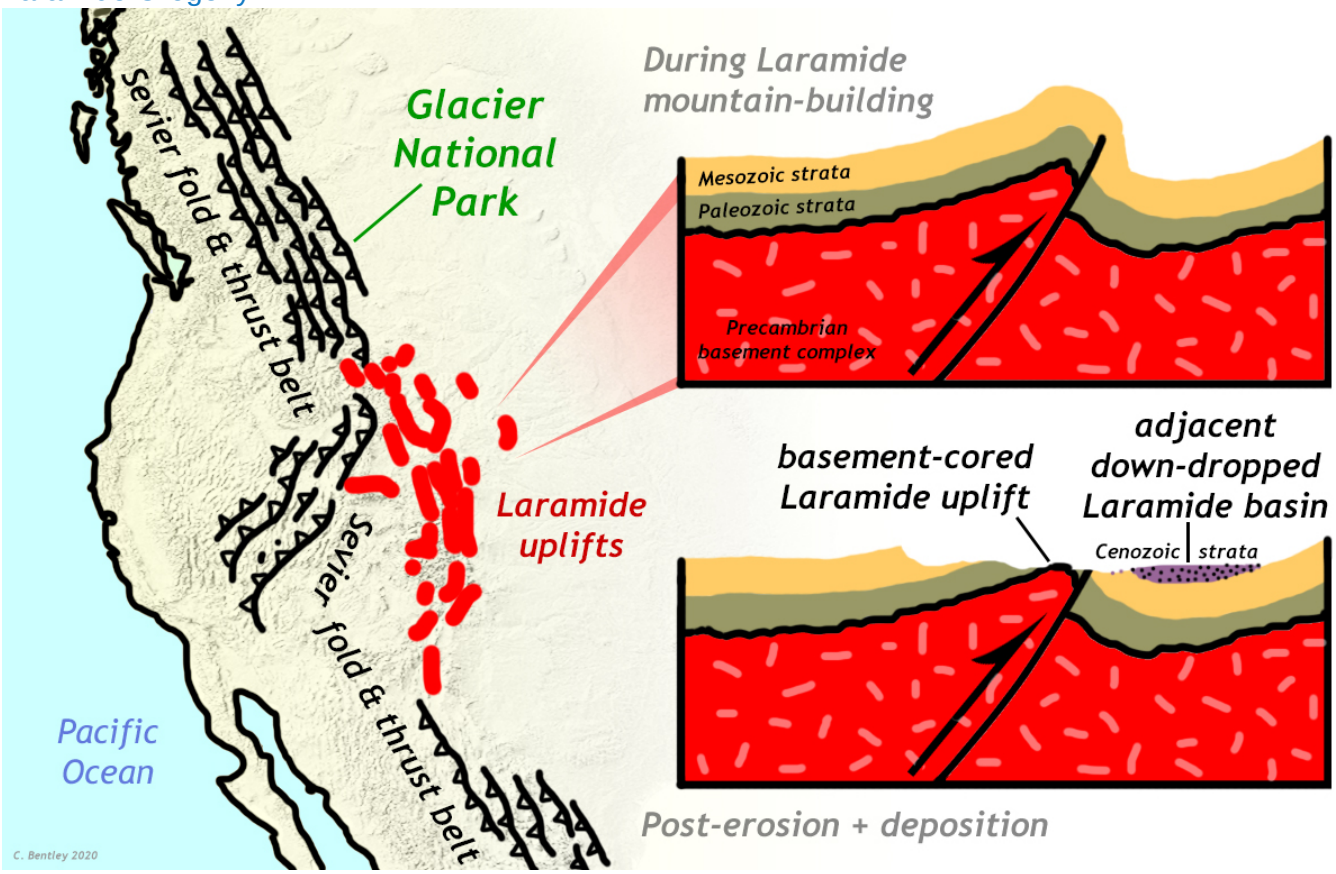


Figure 34.1.10: Map showing the geography of the Sevier fold and thrust belt relative to Laramide uplifts. Cross-sectional views show the structure of the basement-cored Laramide uplifts.

The third mountain-building event that took place in the western US during the late Mesozoic and early Cenozoic, overlapping with but distinct from the Sevier orogeny, is the **Laramide orogeny**. Ultimately, the Laramide, too, was caused by the subduction of the Farallon Plate. Compared to the Sevier, however, (1) the deformation style was different, (2) the deformation was typically more

recent, (3) the deformation took place further to the east, and (4) the deformation was deeper in the crust, involving the crystalline basement. These three variables allow the two orogenies to be distinguished.

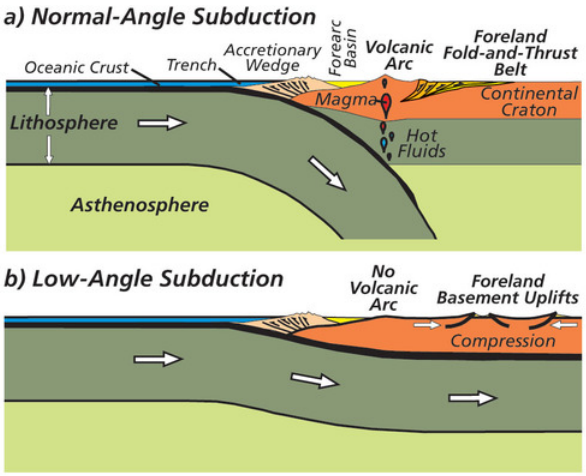


Figure 34.1.11: Normal (a) and flat-slab (b) subduction. (Picture by R. J. Lillie and Parks and Plates by the National Park Service.)

The Laramide Orogeny was driven from beneath. Continuing the trend from the Sevier orogeny, the subducting plate was even younger and hotter, causing the subduction angle to become more shallow over time, eventually getting so small that the slab was dragging along the base of the overriding plate. This is known as flat-slab subduction. This subducting but dragging slab is believed to put pressure on the overriding plate, causing warps in the crust to develop. The lack of a mantle wedge below the overriding plate and above the subducting plate also meant that no magmatism occurred during the Laramide orogeny, or at least none related to subduction as it had occurred in the previous orogenies. The onset of Laramide mountain-building was in the late Cretaceous, 80 to 70 Ma, with the activity wrapping up in the window of 55 to 35 Ma. So the two orogenies overlapped in time, but the Sevier started first and ended first.

Deformation style

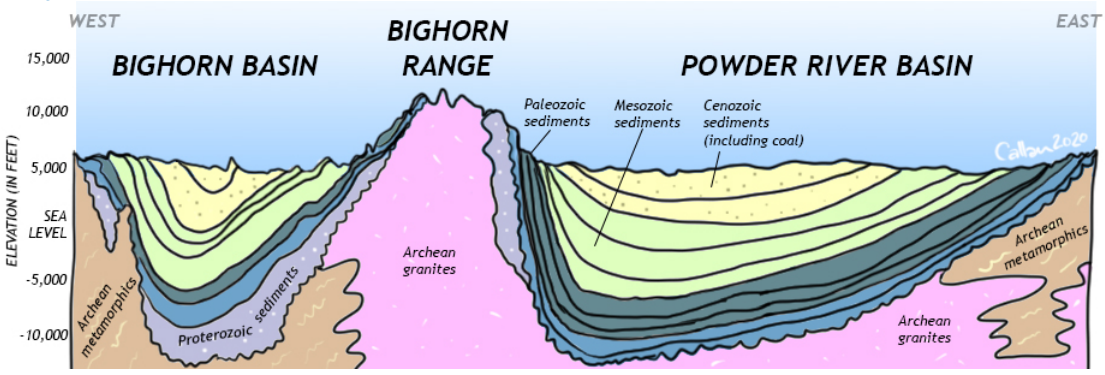


Figure 34.1.12: The Laramide Orogeny generated huge synclinal basins where sediments accumulated through the Cenozoic.

The Laramide orogeny's style of deformation is thick-skinned, which is a way of noting that the crystalline rocks deep in the crust are very much involved. Laramide deformation is characterized by very large anticlines and synclines that involve basement rock. The anticlines generated highlands and mountain ranges, which were weathered and eroded. The synclines were so significant that they warped the surface of the Earth, generating lowlands. These sedimentary basins that were filled with huge amounts of cast-off sedimentary detritus from the eroding mountains nearby, making new Cenozoic-aged sedimentary deposits that filled the depressions. In some cases, these young sedimentary layers reach thicknesses of 5000 meters! The economically important coal and natural gas deposits of Wyoming formed in Laramide basins, as did the fossil-rich lake deposits of the Green River Formation.

Location

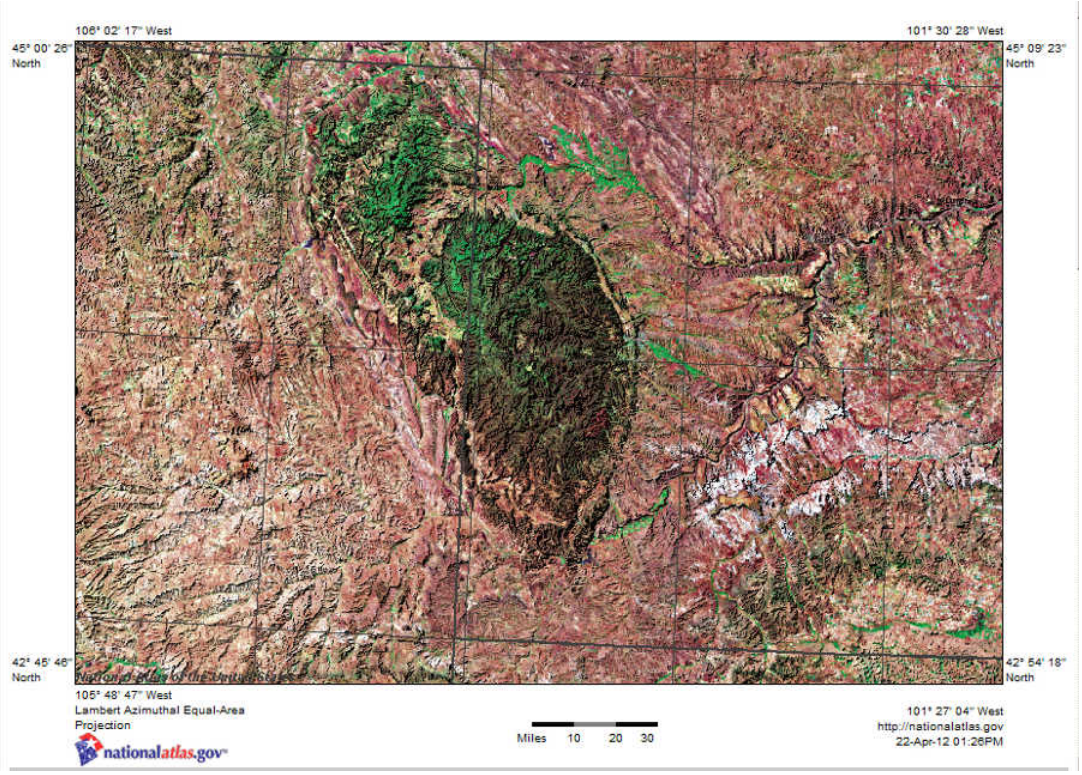


Figure 34.1.13: Shaded relief map of the Black Hills of South Dakota. This bulging dome-shaped uplift is common in the Laramide, in contrast to the much more linear Sevier orogeny uplifts. (Image by nationalatlas.gov in the public domain.)

Laramide deformation does not geographically overlap Sevier deformation. Generally, Laramide deformation is further east. The Black Hills of South Dakota, the Bighorn Mountains of Wyoming, and the Colorado Front Range are all examples of basement-cored Laramide deformation. In contrast, the Sevier fold and thrust belt is further west, in a sinuous arc that runs from the Alberta/British Columbia border through western Montana, Wyoming, Utah, and Nevada, and all the way to Mexico. The Sevier orogen is a much longer feature, too, running twice as long from Canada through the U.S. to Mexico, while the Laramide uplifts are all clustered in the mountain west of the United States alone.

Did I Get It? – Sevier and Laramide

? Exercise 34.1.1

What makes the Sevier orogeny different than the Nevadan?

a. More magmatism

b. More east

c. Deeper in the crust

d. Lasted for a shorter amount of time

e. More localized

Answer

b. More east

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? Exercise 34.1.2

What mountain range today has a tectonic setup most similar to the Sevier orogeny?

a. Lesser Antilies

b. Himalaya

c. Appalachian

d. Andes

e. Zagros

Answer

d. Andes

? Exercise 34.1.3

What makes the Laramide orogeny different than the Sevier?

a. Happened longer ago

b. Deeper in the crust

c. Only occurred in Nevada

d. More west

e. Occurred in a larger area

Answer

b. Deeper in the crust

Conclusion

Though the Nevadan, Sevier, and Laramide orogenies are long over, understanding their influence on the geology and topography of the western edge of North America is still crucial today. The Nevadan started as subduction allowed terranes to attach to Laurentia. The Sevier orogeny was an Andean-style subduction zone complete with back-arc thrust sheets. As subduction continued, the bouncy of the subducting plate caused the plate to descend at a shallow angle, causing a much-different Laramide orogeny. This then set the stage for the Basin and Range and the formation of the San Andreas fault.

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CHAPTER OVERVIEW

35: (Case Study) San Andreas Fault and the Basin and Range

Learning Objectives

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

- Explain the events that lead to the Basin and Range and San Andreas Fault System
- Identify the physiographic features of the Basin and Range, horsts and grabens
- Describe the extensional tectonics and volcanism of the Basin and Range
- Explain the complex tectonics of the modern San Andreas Fault System and its development

35.1: The History of the San Andreas Fault and the Basin and Range

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35.1: The History of the San Andreas Fault and the Basin and Range

Introduction

What happens when a mid-ocean ridge is subducted?

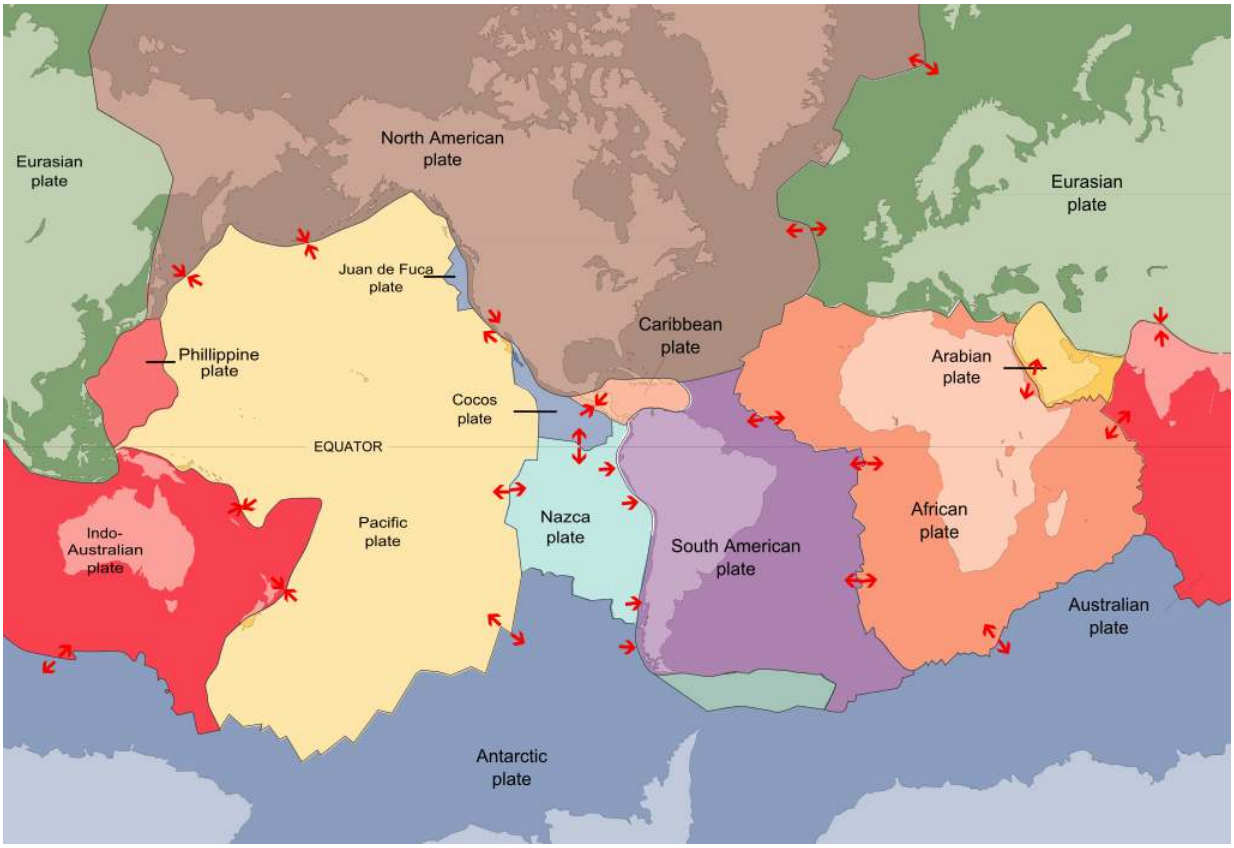


Figure 35.1.1: Plates and plate boundaries with a general sense of motion identified.

The basics of plate tectonics are essential to introductory geology. Two essential aspects are the idea that ocean basins grow via seafloor spreading at mid-ocean ridges and close via subduction. But what happens when these two processes combine? The answer to this is complicated, and is best exemplified by exploring the Cenozoic tectonic history of western North America. This was the first place on Earth where this process was understood, and it is rare enough that not many examples of this can be clearly described in the geologic record.

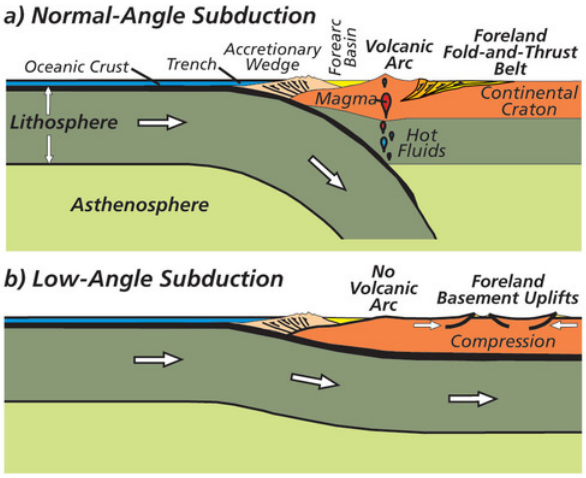


Figure 35.1.2: Normal (a) and flat-slab (b) subduction. (Picture by R. J. Lillie and Parks and Plates by the National Park Service.)

In the Mesozoic, the west coast of North America was dominated by subduction. This led to the Nevadan, Sevier, and Laramide orogenies. The last of these, the Laramide orogeny, was typified by flat-slab subduction. The reason the subduction was flat was that the subducting plate (called the Kula Plate and the Farallon Plate) was hot and young. The subduction zone was close to the mid-ocean ridge, and therefore, the plate was newly formed and still hot from mid-ocean ridge processes. As the plate subducted, the warm temperature lowered its density: warm rock is less dense than cold rock. It was buoyant, making the subduction happen with a shallow angle. Eventually, the mid-ocean ridge itself was subducted, starting around 30 Ma. This event caused two important tectonic features to form: the San Andreas Fault System and the Basin and Range Geologic Province.

Basin and Range

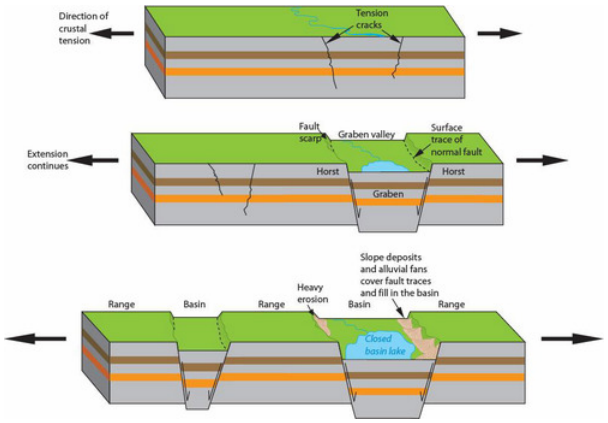


Figure 35.1.3: Generalized rift diagram. A wide rift has many valleys, a narrow rift focuses the extension in one main area. (Sketch by Trista L. Thornberry-Ehrlich, Colorado State University and the National Park Service.)

The Basin and Range is a singular geologic province that covers a large portion of the southwest United States and western Mexico. It is probably the best modern example of a wide rift; a continental divergent boundary where the deformation is spread out. Unlike typical (narrow) rifts, like the East African Rift, where the normal faulting due to tensional stresses related to rifting creates a central graben, wide rifts distribute the faulting over a series of horsts and grabens.



Figure 35.1.4: Definition of Basin and Range, Great Basin, and Great Basin Desert. (Image by National Park Service.)

Note that the Basin and Range is different from the Great Basin, which is defined as an area of drainage basins that do not flow towards the ocean but drain internally. The Great Basin is within the Basin and Range, and is made up of the northern arm of the Basin and Range geologic province.



Figure 35.1.5: Photo of the central Basin in Range (in Nevada). (By NASA.)

Wide rifting is the cause of the famous Basin and Range physiography: seemingly endless north-south trending mountain ranges bound by north-south trending valleys. Nevada, which is entirely within the Basin and Range, is actually the most mountainous state in the U.S. This is not because it contains the tallest mountains, but because it contains the most mountains. Each of these mountains are bound by normal faults.



Figure 35.1.6: Sketch of the Basin and Range Geologic Province (by Thatmaceguy via Wikimedia.)

This faulting has doubled the width of the state of Nevada over the last 50 million years. Most of the faults in the Basin and Range are either inactive in the modern day or have a low level of activity. Only the normal faults toward the eastern (*e.g.* Utah's Wasatch Fault near Salt Lake City) and western (*e.g.* eastern Sierra Nevada faults in eastern California) extremes are highly active.

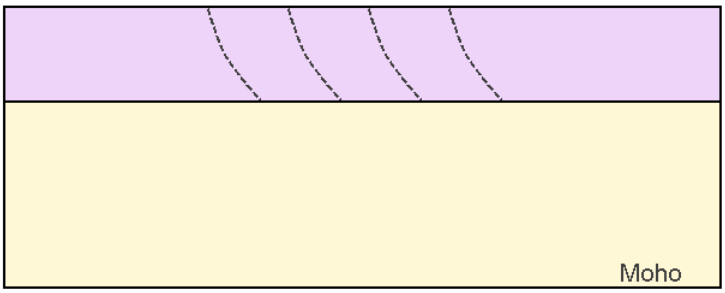


Figure 35.1.7: Animation of tilted-block faulting, one of the ways extension occurs in the Basin and Range (by Aturn4000 via Wikipedia.)

The faulting that started the Basin and Range is complex and not very uniform. In some places, extension started before 55 million years ago, and in some places, it is still active, as mentioned above. However, the majority of the extension occurred around 20 $\pm$ 10 million years ago. The extensional provinces in the northern part of the Basin and Range started generally earlier, while the southern portion (especially south of 40° north latitude) mostly started later, around 30 million years ago. On any single extensional fault system, it rarely lasted more than 10 million years (Axen el al., 1993).

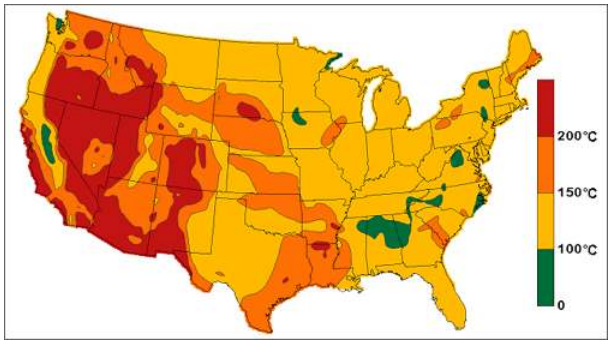


Figure 35.1.8: Geothermal resource map of the US (by the US Department of Energy.)

Another important side effect of Basin and Range extension is heat. A thin plate after stretching means an easier time for upwelling material in the asthenosphere to come up to the surface. This is shown in very high heat flows in the Basin and Range area. Even today, 98% of the geothermal electricity in the United States is produced in Basin and Range states, with California and Nevada being the most productive.



Figure 35.1.9: Wheeler geologic area, Colorado, showing the eroded remnants of the Fish Canyon Tuff, a 5000 cubic kilometer eruption, one of the largest ever measured. (Photo by G. Thomas via Wikimedia.)

Related to heat is the extensive volcanism across the Basin and Range. The details of how this volcanism occurred is interesting, especially since the preceding Laramide Orogeny is noted for lacking volcanism. The details are debated, but one way or another, subduction stopped, and a 'hole' developed in the shallowly-subducting slab which allowed material to come up from below. This event, known as the Mid-Tertiary Ignimbrite Flare-up, started in two swaths, one in Washington and Idaho at about 54 Ma, another in Arizona/New Mexico near the US border with Mexico at 43 Ma. The northern branch of volcanism moved south, and the southern branch moved north, converging near Las Vegas at 21 Ma (Humphries, 1995). This period is noted for producing gigantic explosive eruptions, including some of the largest ever known on Earth, arnd are recorded in extensive tuff deposits. Some of these eruptions are more than 5000 times volumetrically larger than the May 1980 Mt. St. Helens Eruption.

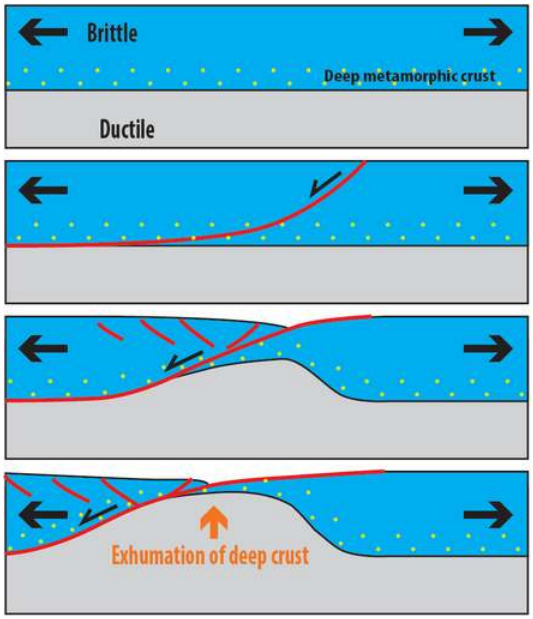


Figure 35.1.10: Sketch of metamorphic core complex (by ManuRoquette via Wikipedia.)

The importance of the Basin and Range to geology is vast. Not only did the extension cause many sedimentary formations, fossils, and resources to be exposed and easier to find, it also helped in the understanding of normal faulting. Two important and related features were first recognized as a result of studying Basin and Range extension: the detachment fault and the metamorphic core complex. Detachment faults, also called low-angle normal faults, were very controversial when they were first described. Most normal faulting occurs at high angles, more than 30° from horizontal, typically around 60°. These faults either started at high angles and rotated to lower angles with extension or form at low angles, which pushed the known limits of rock mechanics at the time. Some geologists and geophysicists felt that the detachment fault could not exist! But over the years, more and more evidence led the entire community to accept these faults, with examples found across the world and on the ocean floor. Near mid-ocean ridges that have a higher spreading rate than their ability to produce magma to form the plate, detachment faults have been known to form, like what is happening in the southwest Indian Ridge near Madagascar. There is even evidence of active detachment faults in Papua New Guinea, which was a hole in the theory previously.

Movement and detachment faults also cause the formation and exposure of metamorphic core complexes. These are places where the large amounts of extension (dozens of miles in places) has allowed deep crustal rock to bulge ductilely toward the surface. The detachment in these cases marks a boundary between highly different rocks: deeply seated, highly metamorphosed basement rocks in the footwall separated by more brittlely-deformed tilted rocks of the hanging wall. Through progressive deformation, the deep rocks 'pooch upward' and enter the brittle regime. Signatures of depth, such as mylonite, are overprinted by signatures of shallowness, such as brecciation. Perhaps the best studied and understood detachment fault and metamorphic core complex is in the Whipple Mountains near the border of California and Nevada (one of the ranges of the Basin and Range). Another example include the Black Mountains near Death Valley.



Figure 35.1.11: Geology students examine the Wasatch Fault outside Salt Lake City, Utah. The Wasatch may have been active as a thrust fault before the Basin and Range and as a normal fault during the Basin and Range. (Photo by Matt Affolter.)

Extension in the Basin and Range has two commonly-cited causes, which may in themselves be at least somewhat related or both occurring at the same time. One cause, which is partially favored as an explanation for earlier extensional provinces in and around Idaho, is extensional collapse. Starting in the middle Mesozoic, the western edge of North America had been experiencing compressional forces due to subduction and terrane accretion. This resulted in the Nevadan, Sevier, and Laramide Orogenies. By 60 Ma, there was a decrease in this compression, with a complete reduction in this force starting 30 million years ago. So many mountains had been built by thrust faulting over this time, and the lithosphere was overthickened. When the forces that built these mountains finally subsided, the topography was allowed to 'relax' and part of the crust spread out and thinned, causing at least some of the Basin and Range extension. One big piece of evidence that backs up this model is reactivated structures. These are found all over the west as former thrust faults turned into normal faults upon the transition to extensional tectonics.

San Andreas Fault System

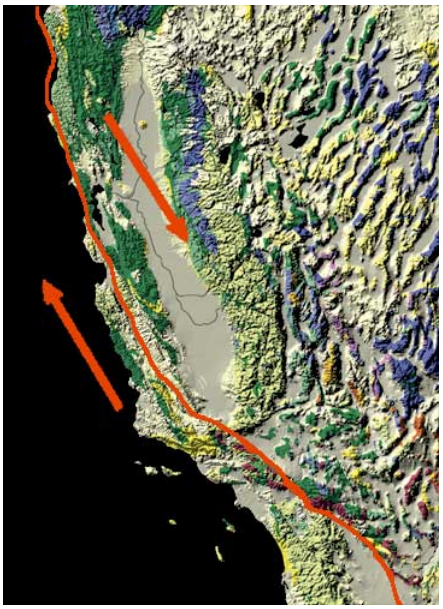
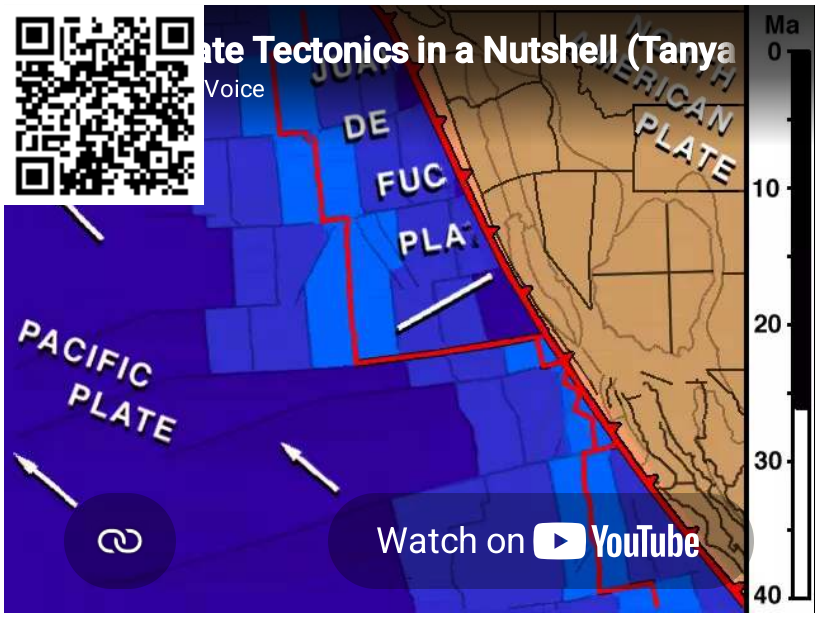


Figure 35.1.12: Map of the San Andreas Fault in California, showing plate motion. (Image by Kate Barton, David Howell, and Joe Vigil of the USGS.)

The other cause of extension in the Basin and Range has to do with the creation of the San Andreas Fault System, a complicated but vital piece of the story in understanding the modern plate boundary between the North American and Pacific Plates.

In 1970, still in the infancy of the development of the theory of Plate Tectonics, Tanya Atwater published a paper that explained the newly-discovered magnetic striping in the ocean crust off of the northern California coast. She used the offshore striping patterns in the ocean crust to reconstruct onshore tectonic history. Analysis of the striping implied that a significant event occurred around 30 million years ago: the mid-ocean ridge between the Pacific and (now mostly subducted) Farallon Plates started to subduct. Original Farallon Plate subduction had a convergent motion with respect to the North American Plate. However, once the mid-ocean ridge began subducting, the relative motion of the plates changed direction. This is because the new plate motion was between different plates, namely the North American and Pacific plates. The relative motion was no longer convergent, but instead mostly transform, with the Pacific Plate moving northwest with respect to the North American Plate. This change in transform motion started to form a strike slip fault at the plate boundary. This would eventually become the San Andreas fault system.



The current motion, and most likely motion in the past, is not exactly parallel to the fault system itself. There is a divergent component to the motion, and that divergent motion might be at least somewhat responsible for Basin and Range extension, particularly after 30 million years ago. It is also responsible for the opening of the Gulf of California (rifting of the Baja California peninsula from the mainland of Mexico) and the more than 90° rotation of the Transverse Ranges in Southern California (Ingersoll and Rumelhart, 1999). The video above, illustrated by Tanya Atwater, shows both of these events, in addition to the widespread extension of the Basin and Range, especially in Nevada.

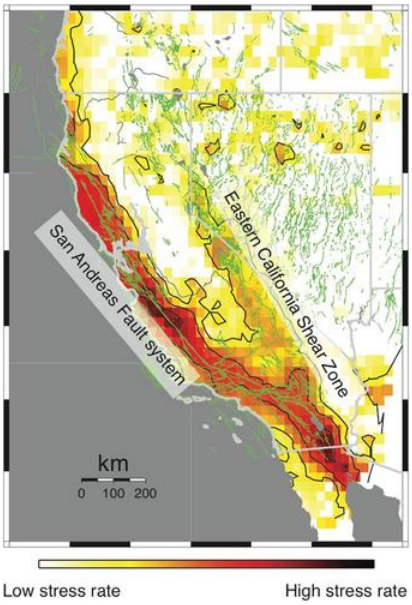


Figure 35.1.13: Stress map of California and nearby regions, showing the Eastern California Shear Zone. (Map by the USGS.)

The San Andreas Fault, at least the current, modern one, has not been the only transform boundary that has existed in this location. In the southern section, there have been at least two previous iterations (Powell and Weldon, 1992). More importantly, a 'new' San Andreas appears to be forming in eastern California and western Nevada. Called the Eastern California Shear Zone and Walker Lane Fault Zone, these are just part of the reason that the San Andreas Fault proper only accounts for about 75% of modern plate motion. The other 25% is spread over many different faults, including the faults in this probable future version of the plate boundary. This zone is responsible for famous and large earthquakes like the 1872 Mw ~7.6 Owens Valley Earthquake, 1992 Mw 7.5 Landers Earthquake and the 2019 Mw 6.4&7.1 Ridgecrest Earthquakes.

Exercise 35.1.1

Which orogeny overlapped in time with the Basin and Range?

a. Sevier

b. Nevadan

c. Laramide

d. Sonoman

e. Antler

Answer

c. Laramide

Exercise 35.1.2

Which type of faulting formed the majority of the Basin and Range?

a. Transform

b. Detachment

c. Thrust

d. Reverse

e. Normal

Answer

e. Normal

Exercise 35.1.3

Which geologic feature was discovered and first described in the Basin and Range?

a. Horst

b. Metamorphic Core Complex

c. Ophiolite

d. Graben

e. Ignimbrite

Answer

b. Metamorphic Core Complex

Exercise 35.1.4

Which event caused the San Andreas Fault System to form?

a. Subduction of a mid-ocean ridge

b. Terrane accretion

c. Slab rollback

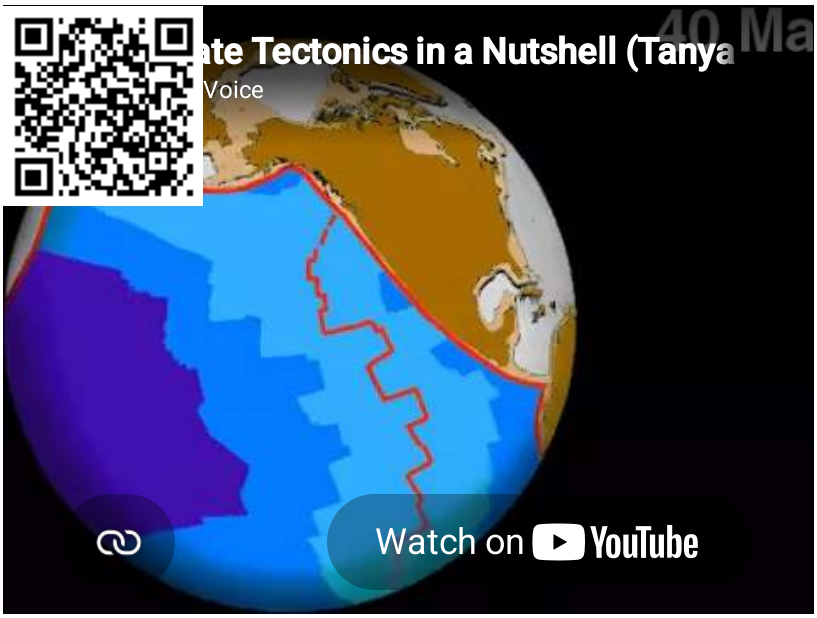
d. Collision within a rift zone

e. Flat-slab subduction

Answer

a. Subduction of a mid-ocean ridge

Final thoughts



For beginning geology students on the west coast of North America, the seemingly simple question “What plate boundary is on the western edge of the continent?” is a difficult one to answer. In California, the San Andreas Fault is the plate boundary, running east of Los Angeles and west of San Francisco, but in actuality the plate boundary is diffuse and spreads as far east as Salt Lake City, Utah as part of the Basin and Range. To the south, the San Andreas blends into a series of faults, like the San Jacinto Fault, and it eventually turns into a mid-ocean ridge as Baja California has rifted away from Mexico and become a peninsula. North of San Francisco, near the town of Eureka and Cape Mendocino (the westernmost part of the state), the plate boundary makes a westward turn into the ocean, ending the San Andreas Fault and transitioning into the Mendocino Fracture zone and the southern boundary of the Juan de Fuca Plate, a residual piece of the Farallon Plate. North of Cape Mendocino, the plate boundary between the Juan de Fuca and North America is convergent, forming a subduction zone. This stretches toward British Columbia, forming famous volcanoes like California’s Mount Shasta, Oregon’s Mount Hood, and Washington’s Mount St. Helens and Mount Rainier. In Canada, north of the Juan de Fuca Plate, the boundary between the North American Plate and the Pacific Plate is again a strike-slip fault called the Queen Charlotte Fault before curving around to generate a subduction zone again near Juneau, Alaska. The difficulty in answering the plate boundary question becomes even more complex when travelling back in time, when a mid-ocean ridge being subducted around 30 million years ago started the process of creating a transform that would eventually create the San Andreas Fault.

Further reading

https://www.nps.gov/articles/basinrange.htm NPS site on the Basin and Range

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CHAPTER OVERVIEW

36: (Case Study) The Parallel "Roads" of Glen Roy

36.1: The Parallel "Roads" of Glen Roy

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36.1: The Parallel "Roads" of Glen Roy

For this case study, let us consider a puzzle in the Scottish landscape.

Compare these two photographs, taken in the same valley:



Figure \(\PageIndex{1}\): Callan Bentley photo



Figure \(\PageIndex{2}\): Callan Bentley photo

What do you notice about the background of these two scenes, the valley called Glen Roy?

Let’s clear away the foreground distractions and take a good look:



Figure \(\PageIndex{3}\): Callan Bentley photo

The U-shape of the valley is a straightforward initial observation. Many geology students will be happy to tell you that indicates this valley has been glaciated. More subtle though are the horizontal lines on the hills bracketing Glen Roy.

Take a closer look, and compare the near hill to the far hill:



Figure \(\PageIndex{4}\): Callan Bentley photo

These lines are perfectly horizontal, and there are three of them. In the photo below, I'm standing at the elevation of the lowest of the three, as you can see the position of the line on the near hill lines up with the position of the line on the far hill:



Figure \(\PageIndex{5}\): Callan Bentley photo

The elevations of these lines are at 260 m, 325 m, and 350 m:



Figure \(\PageIndex{6}\): Callan Bentley photo

So what do they mean? What is the origin of these features?

They've been dubbed "the parallel roads of Glen Roy," but they aren't really "roads." One idea that might occur to your modern mind, following on from the glacial U-shape observation, is that they are glacial striations, but this doesn't work because they don't just go in one linear direction; instead they wrap around the landscape like contour lines on a topographic map, never deviating in their elevation. Glaciers don't do that. Glaciers flow, and they flow *downhill*.

So what else is there?

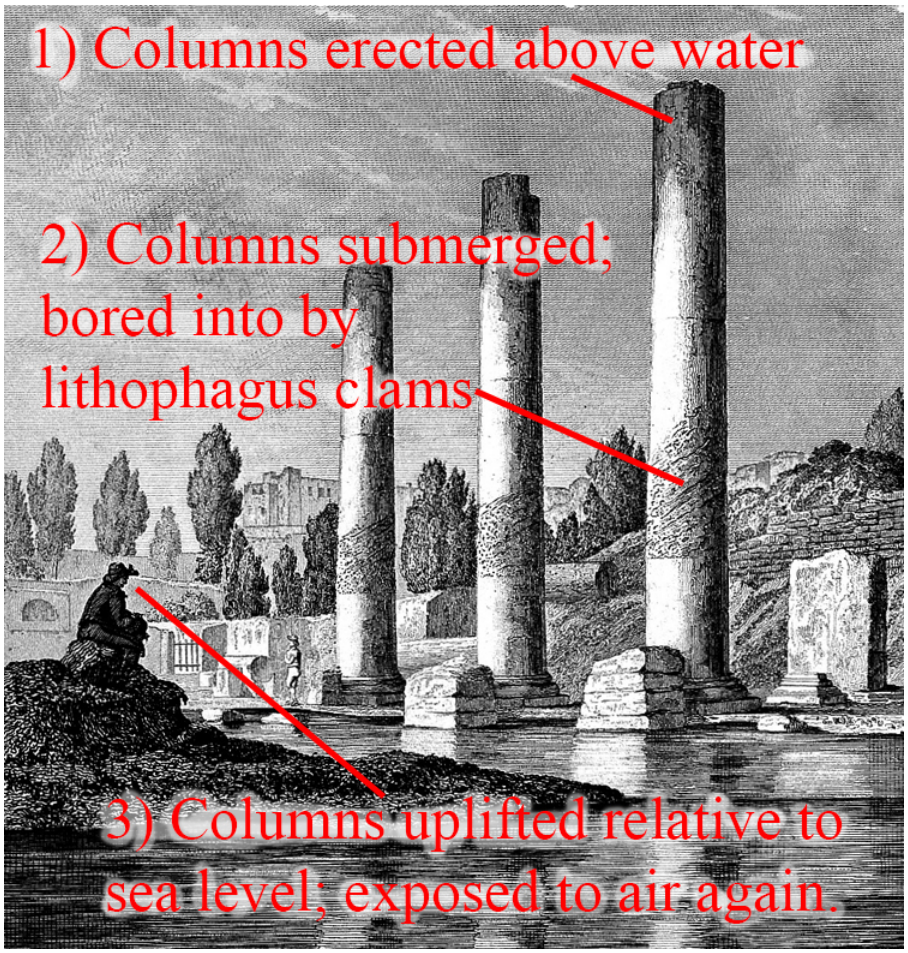


Figure \(\PageIndex{7}\): An annotated view of the lithograph of the Temple of Serapis in Alexandria, Egypt, that served as the frontispiece of Charles Lyell's influential text "The Principles of Geology."

Charles Darwin, that plucky young geologist freshly back from his voyage on the *Beagle*, took his ideas about crustal uplift and subsidence to Glen Roy in 1838. Darwin had been deeply impressed in South America by evidence of tectonic uplift: marine terraces at Coquimbo, Chile, in addition to his personal experience of an earthquake at Santiago. Of course, he had also been reading Charles Lyell, and knew that Lyell had extracted a surprising story of recent subsidence and uplift from the Temple of Serapis in Alexandria, Egypt. Lyell had used a lithograph of three columns from the temple as the frontispiece for the first several editions of *Principles of Geology.*

Once you learn a compelling or useful idea, it's easy to see it everywhere. Darwin had some success by using the idea of subsidence to explain coral atolls, for instance. Now, he applied these notions of vertical motions of the crust to Glen Roy. He thought Glen Roy was a former fjord that had been uplifted, and the "parallel roads" were ancient seashores. The next year (1839), he presented this idea formally to the scientific community.

Darwin was conscious of another hypothesis, pitched in the 1810s by John MacCulloch and Thomas Dick Lauder, who suggested these were not ***sea***shores, but ***lake***shores. The issue Darwin had with MacCulloch and Lauder's idea was that he could not imagine with a good way to dam Glen Roy and thus impound the water. Without a dam, there could be no lake. And with no lake, there would be no way to make a lakeshore. And Darwin knew the crust moved up and down elsewhere, so that was what he considered the best explanation for Glen Roy.

Enter Louis Agassiz, who in 1840 published *Studies on Glaciers*, the expression of ideas he had been promulgating since about 1837. Now there was a ready answer for the source of the requisite dam: **it was made of ice!** Indeed MacCulloch and Lauder had been correct, and Darwin was wrong. Glen Roy had been dammed by a glacier (the Spean Glacier), and as that glacier advanced, it blocked the flow of fresh water coming down the valley. The impounded water (a lake) found its way out of the valley at a series of

three different spillways, one at 260 m (a valley east of neighboring of Glen Spean), one at 325 m (a spillway connecting an arm of Glen Roy to Glen Spean), and finally one at 350 m (the head of the Glen Roy valley itself).

So this scene...



Figure \(\PageIndex{8}\): Callan Bentley photo

...now looked like this in the minds of time-traveling geologic thinkers:

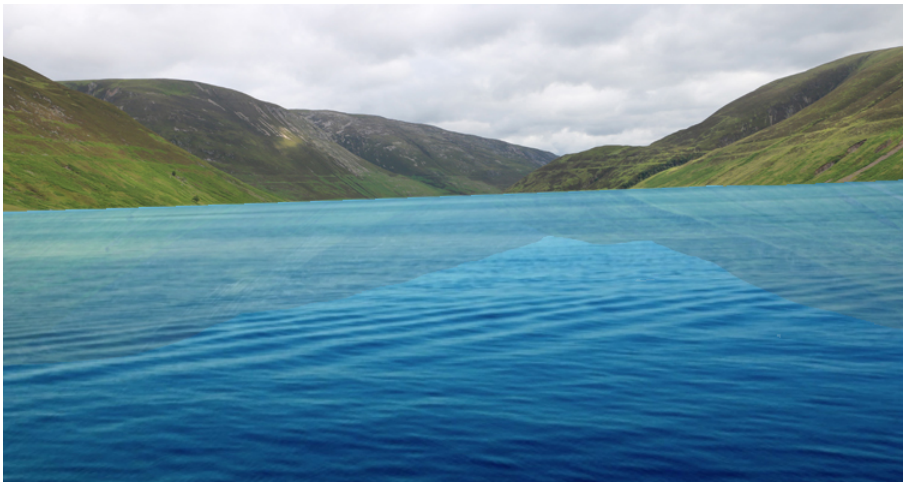


Figure \(\PageIndex{9}\): Callan Bentley photo illustration

Here is a set of three panels from an informational sign at a Glen Roy overlook to show the sequence of events:

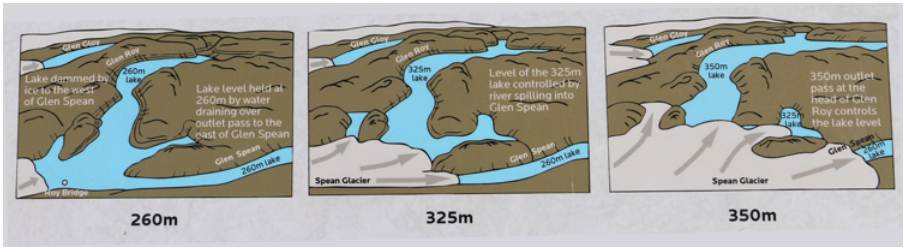


Figure \(\PageIndex{10}\): Glen Roy lookout diagrams

The “lake shore” hypothesis for explaining Glen Roy’s “parallel roads” was an important insight for later geologists examining similar features in places like Missoula, Montana. So the scientific idea was not only specifically important for this one valley in Scotland, but broadly applicable to similar situations in other locations, too – a manifestation of uniformitarianism.

The parallel roads of Glen Roy will not last forever. These subtle etchings on the hillsides are being torn down by modern fluvial erosion. The tracks of gullies and debris flows now overprint the Pleistocene glacial geomorphology, and the sharp definition of the

ancient lakeshores will degrade more and more through time:



Figure \(\PageIndex{11}\): Callan Bentley photo

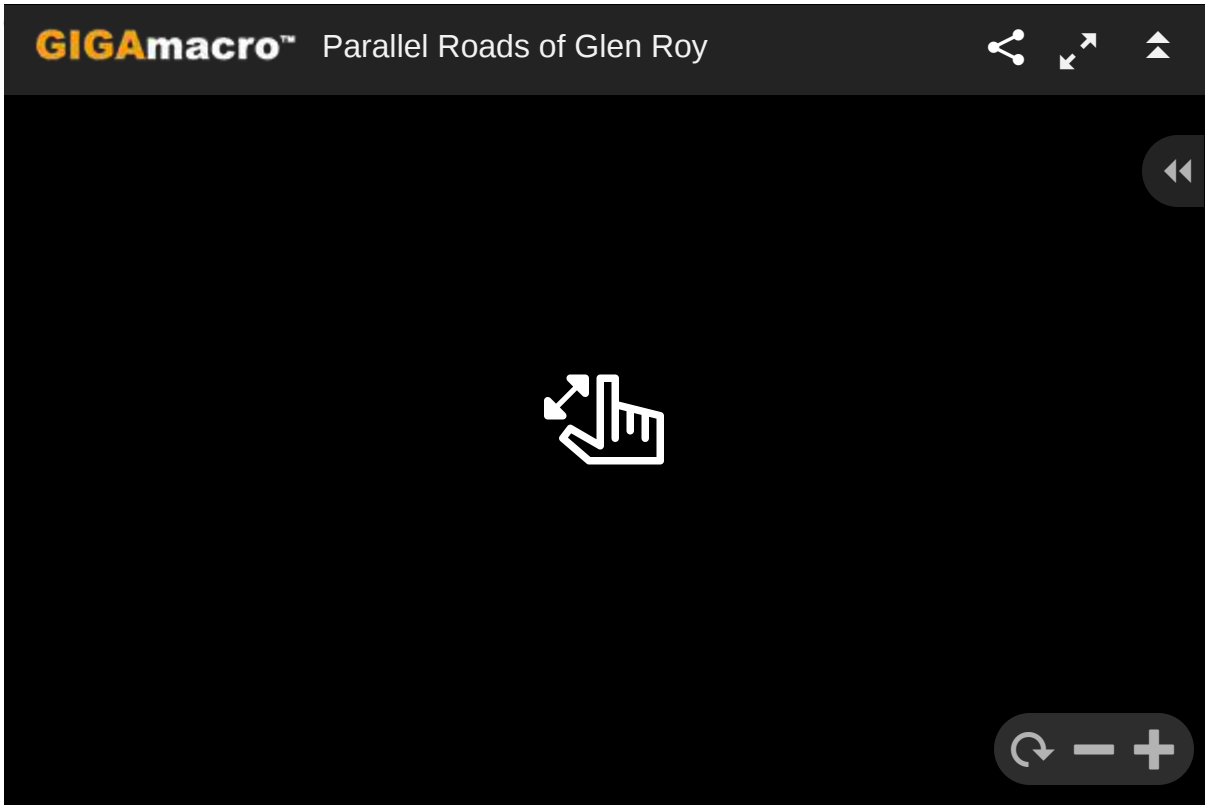
It won't be long (a few centuries, a few millennia) before this subtle landscape signal is lost to direct observation.

Glen Roy is a terrific place to contemplate geoscience controversies in general, and an excellent place to make the point that celebrated scientists such as Charles Darwin didn't get everything right. Darwin proposed multiple hypotheses during his career. Some of those (such as evolution by natural selection) have stood the test of time and scientific scrutiny. Others (like his "seashore" explanation for the parallel roads of Glen Roy) have been shown to be inadequate; a poor match for the available evidence. In science, we value truth above all – more than the celebrated personalities who bring that truth to light. Good ideas survive to fight another day; flawed ideas get tossed in the dustbin. Charles Darwin was not a god; he was a scientist, a generator of ideas. Some of his ideas have been validated; others discredited. This is one of the latter.

If you want to explore Glen Roy on your own from the comfort of your laptop, here are four GigaPans of the valley and its "roads." Each can be explored with your mouse/cursor:


GIGAmacro™ Parallel Roads of Glen Roy, Scotland


GIGAmacro™ Parallel Roads of Glen Roy, Scotland





Further reading

"Darwin and Glen Roy." The Darwin Correspondence Project, University of Cambridge, accessed 2020.

"Parallel Roads of Glen Roy," Wild About Lochaber, accessed 2021.

This page was modified from a blog post by CB originally published in 2019 on the *AGU Blogosphere*.

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CHAPTER OVERVIEW

37: (Case Study) The Channeled Scablands of eastern Washington



Figure 37.1: Location of the Channeled Scablands in Washington state. (Modified by CB from an original by Frank Ramspott – **republication permission currently being requested**.)

In eastern Washington state, there is a vast landscape with unusual landforms, the Channeled Scablands. Here, a dedicated geologist meticulously documented data that challenged his colleagues to accept a violent flood as a viable interpretation of past events. His discovery contradicted prevailing orthodoxy, but he was ultimately vindicated when another scientist documented evidence of a vast glacier-dammed lake in neighboring Montana.

37.1: Uniformitarianism vs. Catastrophism
37.2: J Harlen Bretz
37.3: Channeled Scablands geomorphology
37.4: Bretz floats his hypothesis
37.5: Glacial Lake Missoula
37.6: Catastrophes aren't against the rules
37.7: Further reading

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37.1: Uniformitarianism vs. Catastrophism



Figure 37.1.1: A rectangular canyon, called a ‘coulee,’ in the Channeled Scablands. (CC-by; Photo by Pictoscribe, via Flickr)

We have to start our story with the backstory. This was a philosophical debate, the tension between two competing ideas: uniformitarianism and catastrophism.

Before geology was established as a scientific discipline in western Europe in the mid- to late-1700s, the prevailing intellectual paradigm was based on a literal reading of the Bible, specifically the book of Genesis. Many people will be familiar with the story of Noah’s flood, one of many similar flood myths in different cultures. A global flood would be a great catastrophe —indeed, that was the point— and the thinking went that it might be responsible for moving a lot of sediment around from one place to another. Attributing geologic change to a flood (specifically to “The” Flood) was the ideology called **catastrophism**. This was palatable to those thinkers who anchored themselves firmly in the Christian faith, which was almost everyone in western Europe during the 1700s. Because of this, catastrophism is frequently invoked alongside **creationism**, the idea of a divine origin for the world. It is a favorite go-to explanation today among young-Earth creationists, those who think the Earth is far younger than most geologists do. For our purposes, the key thing to note is that broadly construed, catastrophism argued that geologic change was sudden and violent.

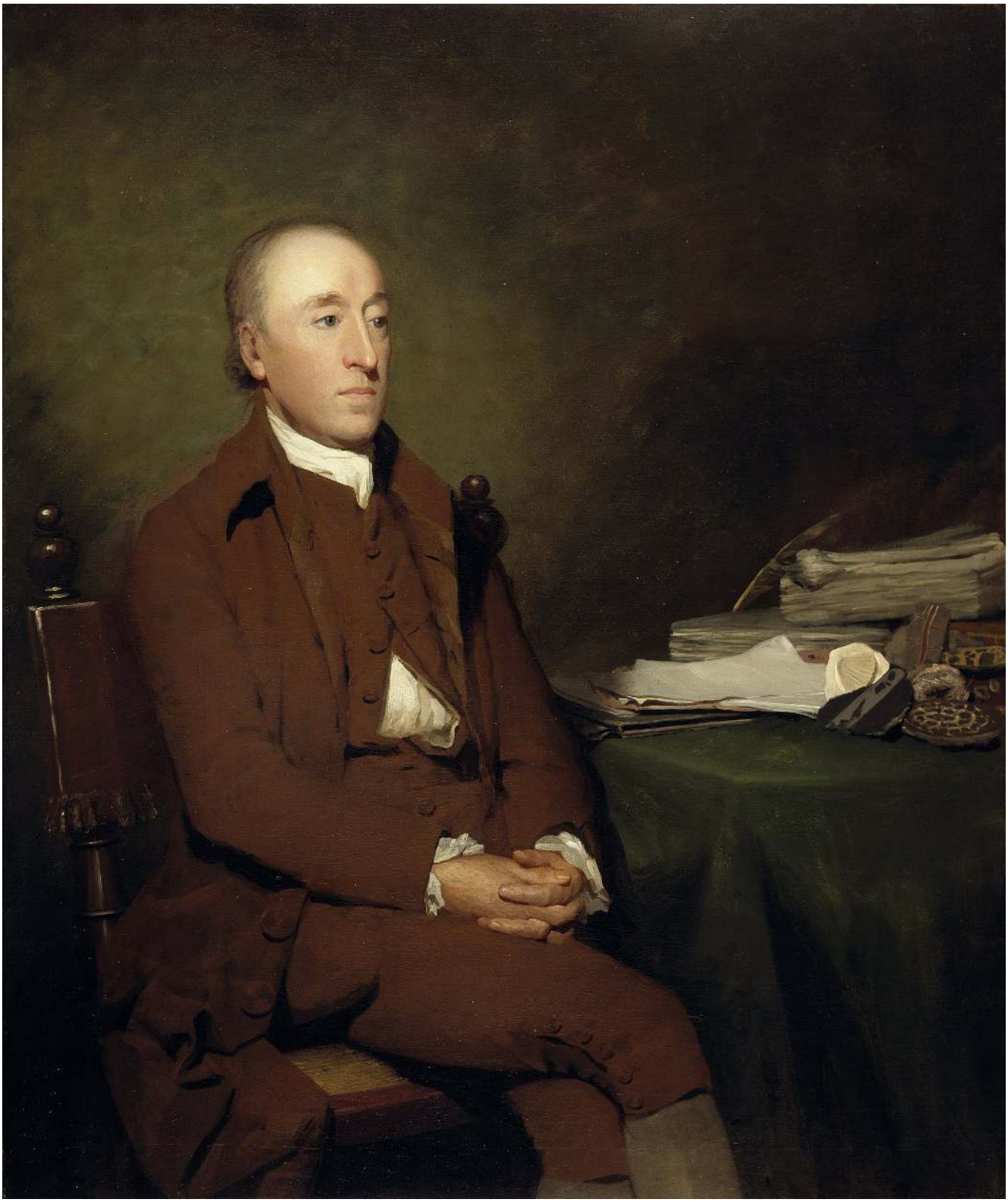


Figure 37.1.2: Portrait of James Hutton by Sir Henry Raeburn (Public domain via Wikimedia Commons.)

Then came the new discipline of geology. Early geologists found evidence that was inconsistent with all change being catastrophic. Foremost among them was James Hutton, a Scottish intellectual and farmer. Hutton saw the creeks draining his farm turn brown and turbid after a heavy rain, and inferred that some of the soil on his fields was being lost to the sea. He envisioned what must

happen to that sediment after it coursed out of sight. When the creek flowed into the sea, that sediment would settle out and make a thin layer over the seafloor. Repeated thousands or millions of times, this simple, inefficient process could eventually denude the land and carpet the bottom of the ocean with vast layers of sediment. He inferred that the many layered sedimentary rocks must have formed in just this way, a little bit at a time, in enormous horizontal sheets. The present day's slow and gradual change wasn't just the 'calm after the ~~storm~~ Flood,' it was the way that change had happened over eons, producing and erasing the geologic record through slow, gentle, gradual processes playing out over vast amounts of time. Repeated cycles of uplift and erosion at a given location were interspersed with times of deposition, burial, and lithification.



Figure 37.1.3: The angular unconformity at Siccar Point shows vertical Silurian graywacke strata overlain by Devonian redbeds, themselves at a non-horizontal angle. (Callan Bentley photo.)

Taking the next step from this insight, Hutton articulated the significance of the angular unconformity at Siccar Point, Scotland, where two packages of stratified sandstone are juxtaposed at two different orientations. Hutton realized that it would take six steps to make this distinctive outcrop: (1) **deposition** of gray mud and sand strata horizontally on the seafloor, (2) lithification of those muddy sands into graywacke followed by **rotation** of the layers into a vertical position, (3) **erosion** of some of those layers — presumably above sea level where erosion is easiest to accomplish, (4) a new round of **deposition** of horizontal sediment, this time highly oxidized red mud and sand — but still horizontally, (5) lithification of those new sediments into solid sandstone and shale, followed by **rotation** of those layers slightly, and (6) modern **erosion** of both rock types. That's a lot of detail to extract from one outcrop, and Hutton reckoned that going through all those steps must have taken a good long while. He thus established that geologic time must be fairly vast, fairly "deep." Siccar Point holds a special place in the hearts of geologists, since it was the place where "deep time" was first fully appreciated.

Since Hutton's insight, geologists have successively pushed our sense of the age of the world back further and further. William Whewell dubbed Hutton's big idea **uniformitarianism**, and Charles Lyell promulgated uniformitarianism far and wide in his best-selling textbook *The Principles of Geology*. Charles Darwin read Lyell's book, and used this newfound sense of an immensely old Earth as a necessary pre-condition for his own ideas about natural selection driving evolutionary change in living organisms.

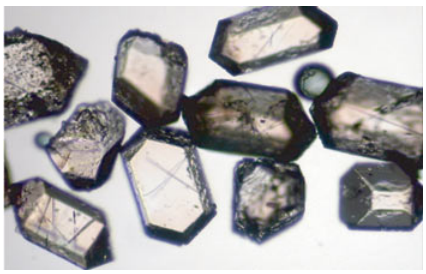


Figure 37.1.4: Uranium/lead ratios contained in tiny crystals called zircons are used to date ancient rocks. (Photo: Michael John Cheadle, University of Wyoming)

Later still, the discovery of radioactivity finally gave geologists a "clock" they could use to quantitatively determine when certain minerals formed. Before this technique was a few decades old, it had put an age on the planet that exceeded a billion years. Modern

estimates place the planet's formation at 4.566 Ga: plenty of time for slow, gradual change to dominate the geologic record. Uniformitarianism now had all the wiggle room it needed to explain Earth history. Geologists were now almost unanimously uniformitarian in their outlook. Catastrophic hypotheses smacked of young-Earth creationism, and that was plainly unacceptable to professional geologists. (An excellent exploration of flood mythology, creationism, and uniformitarianism may be found in the book *The Rocks Don't Lie*, by David Montgomery.)

This was the situation that J Harlen Bretz walked into, and he had some unwelcome news for his colleagues: he had found geologic evidence of an actual catastrophic flood.

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37.2: J Harlen Bretz

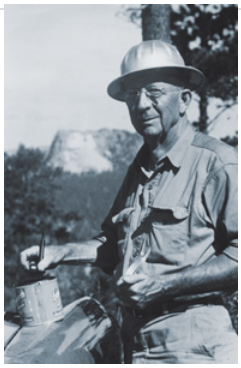


Figure 37.2.1: J Harlen Bretz, photographed by Julian Goldsmith in 1949.

J Harlen Bretz is the hero of this story, a geologist who followed the evidence in the landscape to his own conclusion, even though it flew in the face of the prevailing orthodoxy: uniformitarianism. He began his career as a high school teacher in Seattle, but became interested in geology, and soon quit his job so he could pursue graduate school. With a PhD in hand, he began studying a region in eastern Washington state where the farming was so poor that local farmers had dubbed it the "Scablands." After a spell working back in Seattle at the University of Washington, Bretz settled in at his *alma mater*, the University of Chicago, in Illinois. Summer field campaigns brought him back repeatedly to the Scablands, where he put his students to work mapping the strange landscape, which was unlike anything they had ever seen.

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37.3: Channeled Scablands geomorphology



Figure 37.3.1: Typical Palouse geography, seen near Walla Walla, WA. (CC-by; Photograph by Lynn Suckow, via Wikimedia.)

The area of the Scablands has a three-part geological history. The third part is the one that Bretz elucidated. We will get to that in a moment, but here's what came first. The area is part of the Columbia Plateau, a vast flood basalt province that mostly formed between 17 Ma and 15 Ma. Mafic lava flows are stacked up like a batch of pancakes. On the very top *(like powdered sugar on the pancakes?)* is a layer of loess, which is windblown silt. The loess makes for very fertile soils, and it makes for a region of prime agricultural land called the Palouse.

But as we shall soon see, the nurturing soils of the Palouse (and much of the basalt they overlie) were blasted away by the Missoula floods. That's the third part of the region's geologic history; the one discovered by J Harlen Bretz. So what did Bretz see instead of the usual basalt and loess? His great contribution to geology was meticulous documentation of a series of superlative landforms. The first of these to consider are the coulees.

Coulees



Figure 37.3.2: A portion of Grand Coulee in Washington holds water impounded by a modern dam (Banks Lake). This highlights the coulee's shape: long and narrow, with steep cliff-like valley walls. (CC-by license; Photo by Doc Searles, via Flickr.)

Local place names are one of the delights of traveling to different parts of the world. For instance, a small current of running water might be called a stream, a creek, a run, or a kill, depending on what part of the country you find yourself in. An amphitheater-shaped glacial valley head might be called a cirque, a corrie, or a cwm, depending on whether you're in the Alps, Scotland, or Wales. A valley may be a hollow (or "holler"), a draw, an arroyo, gully, a ravine, a dell or a glen.

In eastern Washington state, some valleys are called **coulees**. These are not the classic V-shaped valleys with a stream running along the bottom, however: they are rectangular or "boxy" in cross-section, thanks to the architecture of the basalt lava flows they cut through.

As Bretz explored the coulees of eastern Washington, he noticed several things: (1) many of the coulees were bone dry, despite presumably having been formed through erosion by running water, and (2) there was a pronounced common orientation to the coulees, trending from Spokane in the northeast off to the southwest. Bretz and his students documented these landforms across the Scablands, and when presented in a map, the pattern is striking:

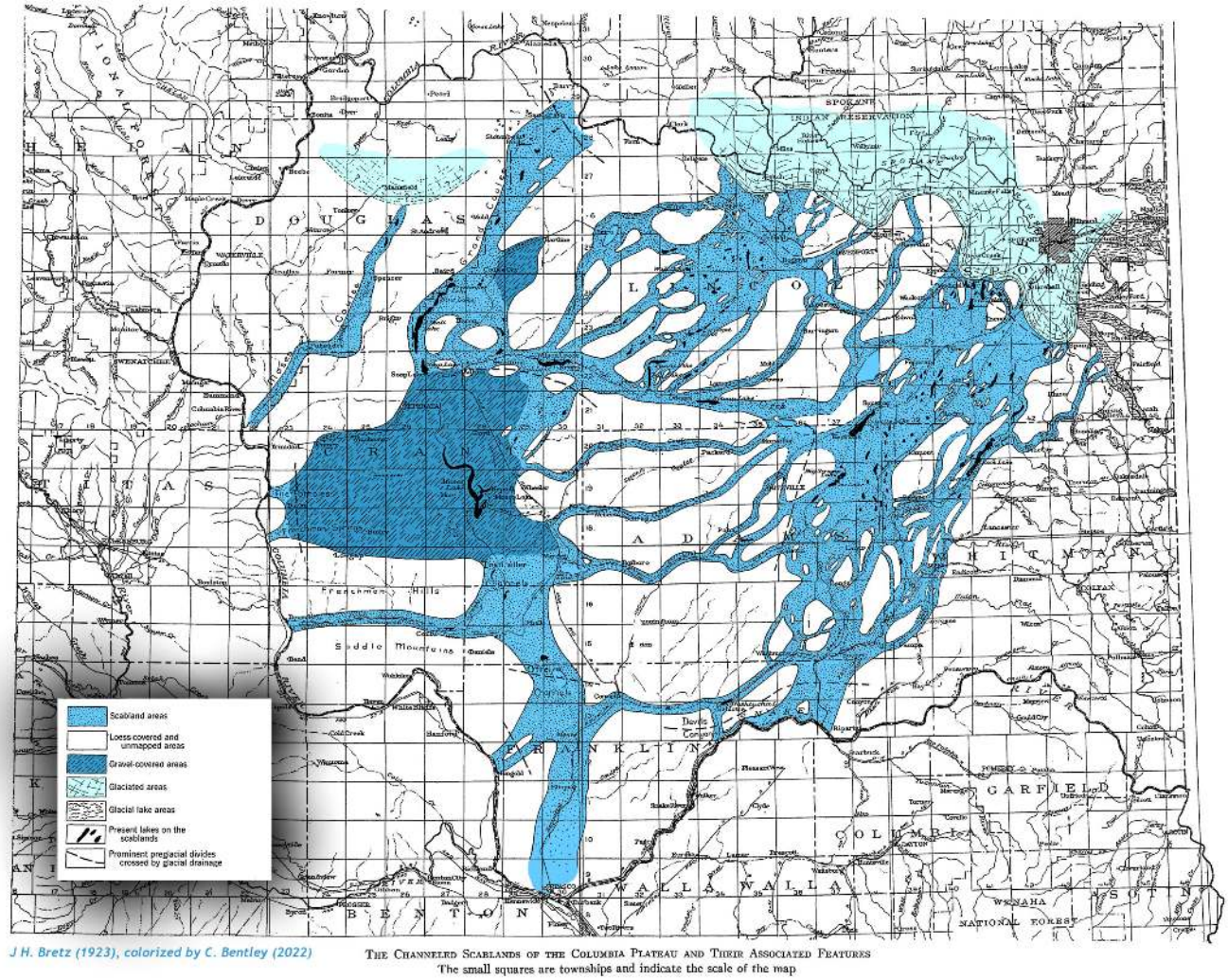


Figure 37.3.3: Bretz's 1923 map of the Channeled Scablands shows the prevailing northeast-to-southwest orientation of dozens of coulees.

The coulees branch and merge, wrapping around isolated mesas of basalt and loess. This is why Bretz dubbed the region the ***Channeled*** Scablands. The "Channeled" part of the name refers to these coulees.

Dry waterfalls

Sometimes the coulees featured dramatic downward steps along their downhill course. These "steps" would be dramatic waterfalls if water was flowing over them, but instead they are bone dry, often with dramatic plunge pools below the cliffs. The GigaPan below shows one of these sites; explore it for a sense of the tremendous scale of these features.



Bill Richards GigaPan

The Dry Falls complex separates upper Grand Coulee from lower Grand Coulee. It is 400 feet high and about three miles wide. That is about ten times the size of modern Niagara Falls in New York.

Eroding out these downstream channels and plunge pools would have produced a lot of new clastic sediment, principally basaltic gravel. The size of the clasts of basalt varied from “normal” sizes you could hold in your hand, to truly jawdropping monsters.

Big boulders

The biggest boulders are most worthy of our attention, because they establish an upper bound on how powerful a current of water it must have taken to move them. One giant is found near Soap Lake, Washington, part of a vast deposit of gravel called the Ephrata fan. It is a substantial distance from its source outcrop, but it is Columbia River basalt, so it seems unlikely to have been ice-rafted.

Check it out:



Figure 37.3.4: An exceptionally large boulder of columnar-jointed basalt that was tumbled along in the Missoula outwash floods. It was ultimately deposited near Soap Lake, Washington as part of an underwater fan complex, the Ephrata fan. (Callan Bentley photo.)

That is an enormous boulder, and it must weigh a lot. Though glaciers regularly transport erratics of this size, there is no evidence that glacial ice ever flowed over this part of the country. The nearest glaciers were dozens of miles off to the north... we'll come back to them in a moment.

Giant ripple trains



Figure 37.3.5: Normal-sized asymmetric (current) ripples in sand. Virginia, USA. (Callan Bentley photo)

As if this weren't enough, the Channeled Scablands also included vast trains of **giant ripples** – like the cm-scale ripples that form in sand on a stream bottom, but these were just gargantuan. Each ripple stood 35 feet tall, with a wavelength of 300 feet between ripple crests. They were arranged in large clusters called "trains." Over one hundred giant ripple trains have been documented in the region.

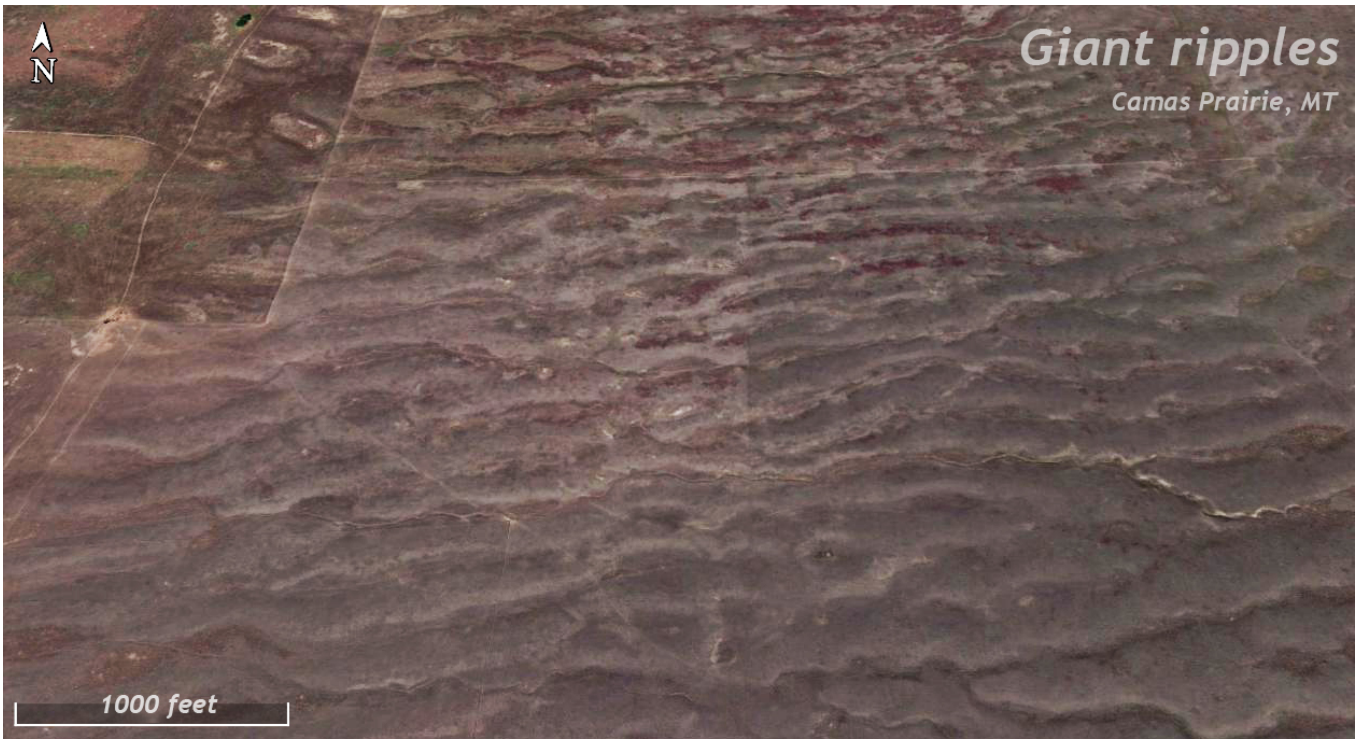


Figure 37.3.6: Camas Prairie, Montana, has exemplary examples of giant ripples. (Google Earth image.)

Consider this scene from Google Earth, which shows a satellite view of the giant ripples that cover the floor of the Camas Prairie in northwestern Montana. Many places in the Channeled Scablands also feature vast arrays of these ridges. To make ripples this large out of such large sediment would seem to require a lot of very fast moving water. At least, that's how it seemed to Harlen Bretz.

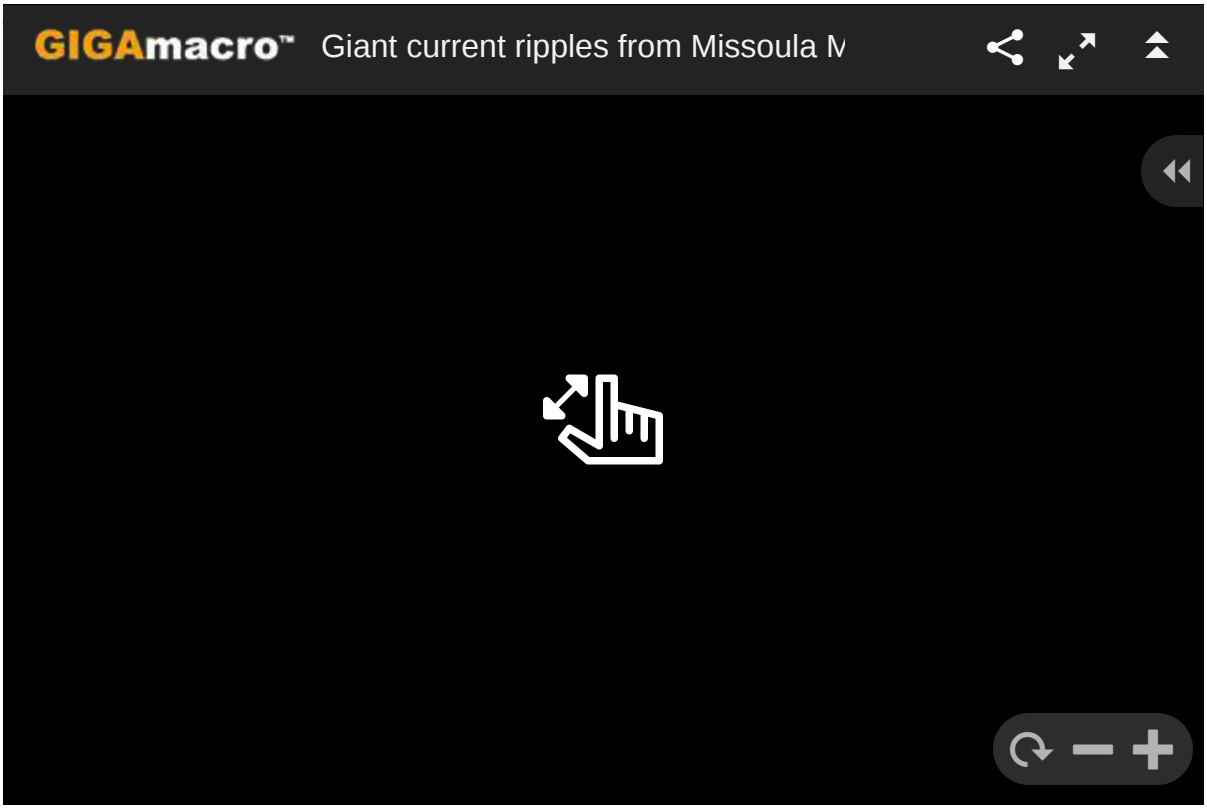
Instead of being made of sand, like normal ripples, these giants were constructed from boulders and cobbles – sedimentary particles that take vastly more energy to move than sand grains. Here's a cross-section through one:



Callan Bentley GigaPan

It would be difficult to identify that as a ripple from ground level! This is a case when backing away from the subject yields the critical perspective that gives insight.

Now, let's zoom out again, and see what these ripples look like from far away. Compare the giant current ripples in the GigaPan below (tan rolling hills above the modern river) to the sheer cliffs of basalt that define the edges of the Columbia River's gorge:



Callan Bentley GigaPan

These giant ripples could only be deposited by raging currents of immense power and depth.

Potholes



Figure 37.3.7: A typical bedrock pothole in a modern river, with a spiral arrow showing interpreted water flow path (vortex). Field of view is about 1 meter. (Callan Bentley photo.)

When fast-moving water flows over a tough bedrock surface, drag slows the current immediately adjacent to the rock. Meanwhile, less-impeded water flows freely in the main channel, and that sets up a differential rate of flow. Spiral shaped vortices develop, and if these vortices of water have sediment suspended within them (such as sand and silt, under normal circumstances), then they can focus abrasion on the bedrock at the point of the vortex. This creates a cylinder-shaped hole to be bored into the rock: a **pothole**.

In the Channeled Scablands, there are potholes, too – as you might expect from such violently-rushing waters. The difference between the "normal" situation and that found in eastern Washington is merely size. The potholes in the Channeled Scablands can

be a quarter-mile in diameter. Here's one such location, Potholes Coulee, shown in Google Maps. Zoom in and explore the substantial round holes in the ground. These are potholes – but incredibly enormous potholes!

In summary, Bretz had documented a suite of features that were both very unusual and could all be explained by a common hypothesis. The dry coulees and falls of raw basalt, lacking the mantle of fertile Palouse loess, the giant boulders, potholes, and ripple marks — they all suggested the same thing: the Channeled Scablands had been ravaged by a **giant flood** beyond anything recorded in human history.



Figure 37.3.8: Icebergs were left perched on dry land after a jökulhlaup drained Iceberg Lake in Wrangell-St. Elias National Park, Alaska. (NPS photo)

Bretz thought the flood might have been similar to big glacial outburst floods documented in Iceland, when volcanic activity melted a glacier from beneath. These *jökulhlaups* ("yokel-ahps") were impressive indeed, though not quite on the same scale as

what Bretz envisioned for the Channeled Scablands. The jökulhlaups were powerful, but perhaps not quite “catastrophic.” Bretz had cut his teeth working on the glacial geomorphology of Puget Sound, and knew that there was substantial evidence for glaciers not too far away – up by Spokane. Could a jökulhlaup have carved the Channeled Scablands?

Quiz

? Exercise 37.3.1

Which of the following landforms was *not* documented by Bretz in the Channeled Scablands?

a. giant potholes

b. dry waterfalls of vast size

c. giant ripple marks in vast "trains"

d. multiple parallel coulees

e. giant boulders

f. an angular unconformity

Answer

f. an angular unconformity

? Exercise 37.3.2

What event did Bretz infer must have carved the Channeled Scablands?

a. A divinely-induced flood, intended to drown the unworthy sinners of the world

b. A volcanic eruption

c. A tsunami

d. A very high tide

e. A jökulhlaup flood

Answer

e. A jökulhlaup flood

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37.4: Bretz floats his hypothesis

In 1923, Bretz finally put his audacious ideas into print. Like any careful scientist, he considered potential objections and attempted to disarm them before they could be deployed. He knew what he had to say was going to be controversial, because a giant flood was anathema to the uniformitarian paradigm that governed professional geoscience. If "the present is the key to the past," Bretz had a big issue: A flood of the size he was proposing had never been seen in human history. He was venturing into catastrophist terrain! But what choice did he have? He couldn't account for the singular collection of landforms in the Scablands any way except for the big flood hypothesis.

It should probably be noted here that Bretz's proposed flood was very much a regional ***(not** global)* affair. It would be a big mistake to leap from the conclusion that there was a big flood in the Channeled Scablands during the Pleistocene to anything on the continental or planetary scale. Bretz's flood was big, but it wasn't *that* big.

In January 1927, Bretz formally presented his idea that the Channeled Scablands were best explained by an enormous flood. He spoke at the Geological Society of Washington (that's D.C. this time, not the state), and it was a legendary case of scientific ambush. The professionals there chewed him up and spit him out, especially those who were older and professionally established, teaming up to vocally attack his 'flood' interpretation. But sitting quietly in the room was J.T. Pardee, a young and thus relatively vulnerable new recruit at the USGS. Pardee didn't want to run the risk of irritating his new colleagues and superiors by speaking out. But he knew that something no one else did – just to the east was a glacial lake that could have provided sufficient water to supply Bretz's flood: Glacial Lake Missoula.

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37.5: Glacial Lake Missoula

Above the town of Missoula, Montana are a series of subtle notches in the hillside. Like the giant ripples we mentioned earlier, they are hard to see if you're standing on them. They show up most photogenically when lightly dusted with snow, and viewed from high above.



Figure 37.5.1: Strandlines from Glacial Lake Missoula wrap around Mount Jumbo above Hellgate Canyon. (Image modified from a still from a video produced by the Montana Natural History Association.)

Each of these notches is **perfectly horizontal**, wrapping around the landscape without deviating in elevation even by an inch. Thomas Crowder Chamberlin noted these in 1886, correctly interpreting them as ancient lakeshores by comparing them to what he had read of the so-called "Parallel Roads" of Glen Roy, Scotland. The idea is that waves on the shore of a now-vanished lake must have lapped against the hillside with sufficient vigor to etch out a notch. The valley was apparently filled by lakes of many different levels in the past. These lakeshores are called **strandlines**.

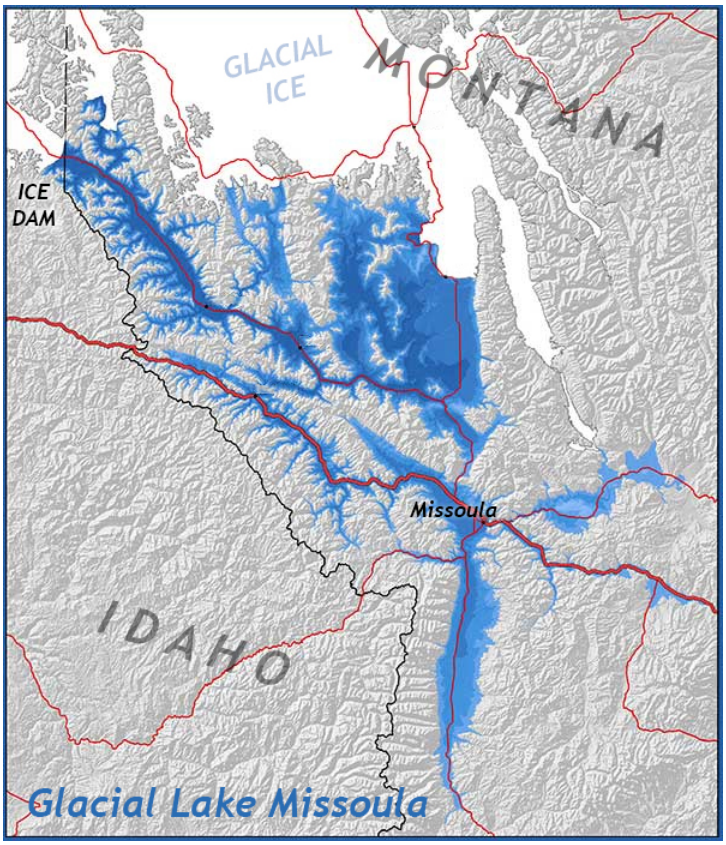


Figure 37.5.2: Map of Glacial Lake Missoula at its largest extent. Red lines are major roads. (Modified by CB from an original by the Montana Natural History Association.)

The Missoula strandlines imply an ancient lake once filled the valley where today the University of Montana is situated adjacent to the Clark Fork River. What formed that ancient lake? J.T. Pardee thought he knew. Pardee was a native Montanan who pursued a career doing geology for the United States Geological Survey. He traced out the strandlines across the landscape, heading north and downstream along the Clark Fork River. His mapping led him to Sandpoint, Idaho (on the Idaho/Montana border), where there was no evidence of a landslide dam, nor a lava dam, nor any other kind of dam that was still there. What impounded the Clark Fork to make the lake, then? If the Parallel Roads of Glen Roy were to act as any guide, perhaps it was **glacial ice** that dammed the river.

Pardee postulated that the Purcell Lobe of the great Laurentide ice sheet nudged south into the valley and made an ice dam 23 miles (37 km) wide, spanning the relevant drainage, and rising to half a mile (0.8 km) high above the valley floor. The impounded waters of the Clark Fork River pooled to make Glacial Lake Missoula. This huge glacial lake extended up to 200 miles upstream from the ice dam.

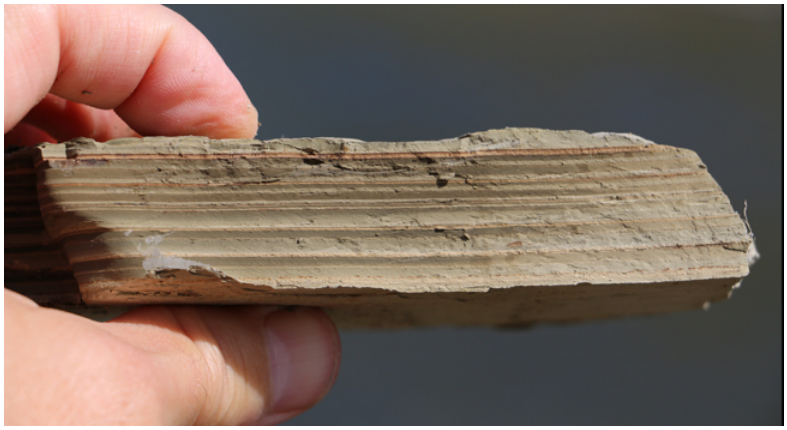


Figure 37.5.3: Varves in laminated mudstones deposited on the floor of Glacial lake Missoula. (Callan Bentley photo.)

In places, the lake may have been as deep as 200o feet. At the bottom of this vast volume of water, depositional conditions appeared calm. In many places, laminated strata record lake-bottom sedimentation that alternated rhythmically between silt and clay. These regular oscillations are interpreted as **varves**.

Why did the dam eventually break? The answer lies with the fact that ice is less dense than liquid water. So when the voluminous waters of the lake got deep enough, this giant ancient lake simply floated its icy dam away. Released by the now-absent dam, where did the waters of Glacial Lake Missoula go? In a word, **downhill**.

The path downhill — to the Pacific Ocean, via the Columbia River — ultimately led across eastern Washington. Released in an instant, the violent draining out of the lake unleashed 500 cubic miles of water across the landscape. With a discharge estimated at 60 times the size of the Amazon River, the flood waters gouged in and spread out, and reconvened, and pushed boulders, deposited giant ripples, tumbled over waterfalls, and drilled out massive potholes.

And then the ice flowed back in, rolling over the same ground, backing up the same river again, regenerating a new iteration of Glacial Lake Missoula, and etching a new strandline into the surrounding valley walls.

This 2015 video, by Nick Zentner and Tom Foster, shows many of the key details:



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37.6: Catastrophes aren't against the rules

So J Harlen Bretz ended up having an impact on geologic thinking that went far beyond the dissected canyons of the Columbia Plateau. He pushed the limits of our imagination, forcing geologists to consider that maybe they hadn't seen it all. Nature had tricks up her sleeve that might be powerful, singular, and of vital importance, but these processes are still well within the bounds of the laws of physics. If we want to be good at interpreting the past, we have to consider the potential for unique violent events to periodically occur. If not for Bretz, would Walter and Luis Alvarez have dared to invoke an extraterrestrial impact for the end-Cretaceous extinction event? Would the Snowball Earth hypothesis ever have been proposed?



In 1979, the Geological Society of America (GSA) unambiguously celebrated Bretz's willingness to follow the field data where it led him, even though that path led to the conclusion of a vast catastrophe well beyond the human experience. In accepting GSA's Penrose medal, Bretz humbly summarized, "Perhaps I can be credited with reviving and demystifying legendary catastrophism and challenging a too rigorous uniformitarianism."

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37.7: Further reading

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CHAPTER OVERVIEW

38: (Case Study) The interplay of carbon and oxygen cycles in Earth history



Figure 38.1: This forest is busy photosynthesizing, an act with implications for the atmosphere, climate, and animal life. (CC-BY; Photo by Roland Sims, via Flickr.)

38.1: Photosynthesis and respiration

38.2: Carbon burial

38.3: Impacts on the cryosphere and biosphere

38.4: Isotopes

38.5: Fossil fuels

38.6: Anthropogenic climate change

38.7: Further reading

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38.1: Photosynthesis and respiration

Photosynthesis and its sister reaction respiration are key biogeochemical processes in the Earth system. The simplified chemical equation for photosynthesis is:

$$CO_2 + H_2O \rightarrow C_6H_{12}O_6 + O_2$$

(Requires a lot of **energy input**, from the Sun)

In other words, the Sun's energy can cause carbon dioxide to react with water, producing glucose (a sugar) and free oxygen. This reaction is facilitated by living organisms, principally cyanobacteria, algae, and plants **{LINK TO PLANT CASE STUDY}**. The glucose is the point of the reaction for these organisms, while the oxygen is just an incidental waste product.

For respiration (*or its inorganic equivalent, combustion*), just flip the equation around and run it in reverse:

$$C_6H_{12}O_6 + O_2 \rightarrow CO_2 + H_2O$$

(**Releases** a lot of energy)



Figure 38.1.1: Photo by Rosmarie Voegtli. (CC-BY; via Flickr.)

Every green plant you have ever seen is actively doing photosynthesis while sunlight shines on its leaves, and the same is true for primordial cyanobacteria and floating phytoplankton in the oceans. Collectively, these organisms are **powerful agents of geochemical change**: they extract CO_2 from the atmosphere, shucking off its oxygen as waste, and use its carbon to build their bodies.

Today, this effect can be measured in real time: photosynthesis in the northern hemisphere summer pulls about 6 ppm of CO_2 out of the air. That's **about 1.5% of the total atmospheric pool of CO_2!** There is not as much land in the southern hemisphere (and much

of it is desert), so the CO_2 drawdown effect is not as large during the austral summer. It is most pronounced between April and September of each year.



Figure 38.1.2: The 'leaf sheep' is an exception to the rule: a photosynthesizing animal. This 1-cm-long sea slug hosts symbiotic algae within its flesh. (CC-BY; Photo by prilfish, via Flickr.)

When animals eat algae or plants, they aren't doing it to be mean. They are doing it because they will die if they don't. Unlike autotrophic plants, animals are heterotrophs – a word that means **they need to eat**. Eating is an act whose intention is to grab some of the plants' stored energy and use it to drive the animal's own life: moving, growing, reproducing. As a general rule, animals are not capable of photosynthesis. Eating is their only option. Curiously though, in multiple lineages, some animals have figured out how to host photosynthetic cyanobacteria or algae within their bodies as endosymbionts.

Let's circle back now to photosynthesis. Driven by the energy in sunbeams, plants suture carbon dioxide to water, making glucose, and ejecting much of the oxygen as waste. But trees are not made of glucose. Instead, they bulk up by converting glucose into cellulose via this reaction:

$$C_6H_{12}O_6 + C_6H_{12}O_6 + C_6H_{12}O_6 + C_6H_{12}O_6,\ etc.\ \rightarrow (C_6H_{10}O_5)_n + H_2O$$

Cellulose is a much more durable compound. Non-woody plants such as grasses are capable of projecting upward from the ground because cellulose in their cell walls provides structural support. Along with lignin, cellulose makes up the bulk of trees and other woody plants. **It's wood.** Wood makes up the bulk of the planet's biomass (450 Gt of carbon, about ~82% of the total).

Wood is capable of being oxidized by organisms, but not especially rapidly. Fungi and bacteria can break it down, given enough time. Depending on the species of wood and levels of moisture and ambient oxygen, this decomposition can take a year or several centuries. Herbivores such as termites and cows host symbiotic bacteria in their guts which can break down cellulose more rapidly, but that's not a trick that humans can pull off! If we decide to oxidize wood quickly, we ***burn it.***



Figure 38.1.3: Bonfires are hot because of stored solar energy. What you can't see in this photo is the CO_2 and H_2O vapor that's being released. (Callan Bentley photo.)

We have probably all experienced fire. When you enjoy a bonfire, campfire, or the radiant heat of a woodstove, you're reversing the photosynthesis reaction via the secondary product of wood. All that warmth, all that orange and yellow light, all those loud crackling noises – these are various forms of **energy** that are released as the bonds holding together cellulose and lignin are broken. For most wood available to most people, that energy was captured in the past few years or decades.

The same is also true when we burn **ancient** plant biomass – though not "wood" or "cellulose" any more, ancient plants can provide long-term stores of ancient carbon. Let us next examine how these "fossil fuels" form, and what the implications are of their formative processes, specifically for the atmosphere and biosphere.

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38.2: Carbon burial

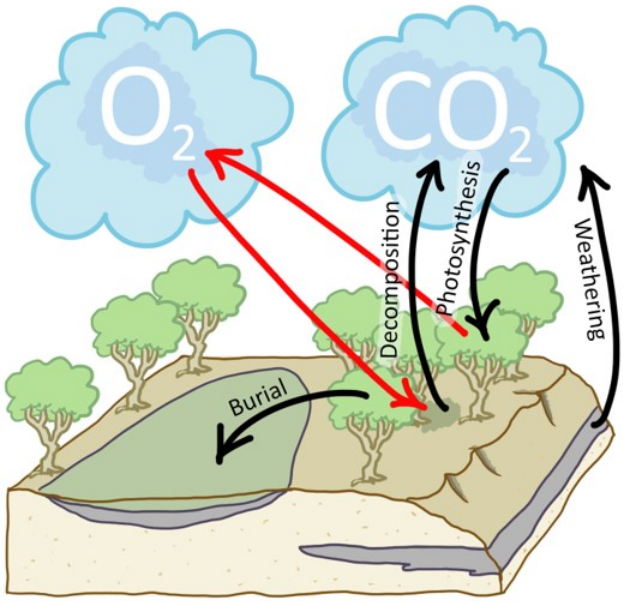


Figure 38.2.1: Carbon and oxygen fluxes between plants, the atmosphere, and the geosphere. (Callan Bentley cartoon.)

What must happen next is for plant biomass to be built up through photosynthesis and then **buried** under sediment before it gets a chance to rot or burn.

Consider this diagram, which highlights some of the fluxes (flows) between the atmosphere, plants, and layers of sedimentary rock. Plants pull CO_2 from the atmosphere and release O_2 as a waste product. The O_2 will eventually react with the plant carbon through decomposition, unless the plant matter is buried. Once buried, it is isolated from free oxygen; **it cannot decompose**. Its carbon then becomes part of sedimentary rocks, and is locked in the geosphere until uplift (often, mountain-building) brings it back into contact with atmospheric oxygen. This allows it to be weathered (re-oxidized).

What is the impact on the atmosphere of this natural carbon sequestration? It all depends on the timescale over which the carbon remains isolated from the atmosphere, deep in the crust. For our purposes, it makes sense to consider timescales of **tens to hundreds of millions of years.**

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38.3: Impacts on the cryosphere and biosphere

Let us consider the impacts of carbon burial on three phenomena: fire, climate, and animal physiology.

Fire



Figure 38.3.1: A 2007 wildfire in Idaho. (US Fish & Wildlife Service photo.)

We listed a few examples of anthropogenic fire above, but fire is also a natural part of the modern Earth system. **Wildfires** can be sparked by lightning strikes or volcanic eruptions. They are more likely to ignite and to spread in areas that are temporarily warm and dry, but have seen wet enough conditions in the past to build up a supply of plant fuel. Combustion occurs when the plant tissues react with atmospheric oxygen, and release the carbon in the plants back into the atmosphere as CO_2.

So **oxygen is a pre-requisite** for wildfire, and oxygen levels have changed throughout Earth history. How much oxygen is necessary? It turns out that fire requires at least **13%** O_2 in the surrounding atmosphere. Over the long-term, our planet has bulked up the free oxygen content of its atmosphere, from an initial 0% to the current ~21%. It is estimated that the critical 13% threshold was crossed in the Silurian. Two lines of evidence support this interpretation:

1. The oldest terrestrial plant fossils are spores that date to the Ordovician, suggesting colonization of the land surface by organisms akin to club mosses. These plants provided fuel to burn, but also shelter for animals. The oldest terrestrial animal fossils date to shortly thereafter. As plants colonized the land, they led to more of the planet's surface being devoted to (incidental) oxygen production. It is thought that oxygenation of the atmosphere would also have led to the development of an ozone (O_3) layer, which would have enabled animals to spread across the land due to lowering the influx of harmful ultraviolet radiation from the Sun.



Figure 38.3.2: Charcoal clasts in Silurian sandstone from Wales. (Modified from Figure 3 of Glasspool & Gastaldo (2022).)

2. By 420 Ma (**Silurian**), **charcoal** starts showing up. Charcoal is a carbon-rich version of what used to be wood, but all the water and volatile components have been driven off through heating it in the presence of limited amounts of oxygen. It is our first evidence that fires had started burning on Earth.

With the exception of a "gap" in the mid-Devonian, charcoal has been a common constituent of terrestrial sediments for the rest of the Phanerozoic Eon. Thanks to the rise in oxygen levels, fire has become a standard phenomenon on the surface of Earth. So the burial of plant biomass preserves some carbon for the future, but leaves behind oxygen at the surface, and that oxygen that makes it more likely for unburied plants to burn.

Ice

The greenhouse effect needs a quick recap in order to understand what happens next.

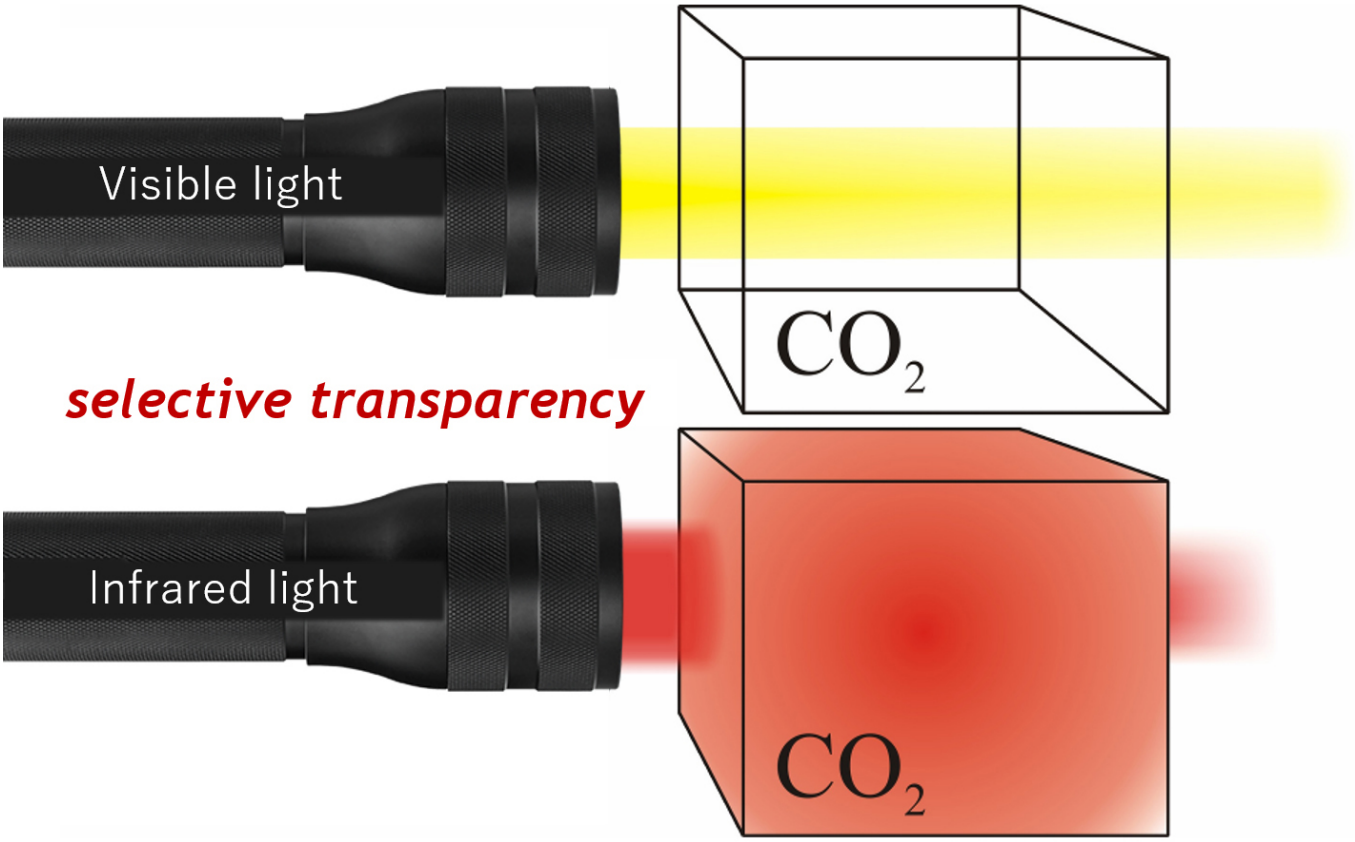


Figure 38.3.3: An illustration of the property of selective transparency in CO_2 gas. (Callan Bentley cartoon.)

CO_2 is an invisible gas. But it's only invisible to the visible wavelengths of light. Under other wavelengths of light, it shows up as a dense foggy cloud. Specifically, infrared light cannot pierce through CO_2 – the infrared light's energy is absorbed and re-radiated in a random direction. Examine the cartoon illustration of this idea at right: When a beam of visible light is directed through a glass box containing CO_2, it all passes through to the other side. But when infrared light is shone through instead, it gets stopped, and so less than 100% makes it across to the far side of the box. So CO_2 is **selectively transparent** to different wavelengths of light.

This matters intensely for the regulation of Earth's climate. Though the planet is irradiated by visible wavelengths of light, Earth materials absorb that light, and warm up. They then re-radiate some of the energy in the infrared wavelengths. Hot objects, like the pavement of a street on a summer's day, give off a lot of infrared radiation. If this infrared light were unimpeded, the hot street would shoot it straight out into space. But thankfully, that isn't what happens. Instead, greenhouse gases like CO_2 catch some of those infrared photons and spit them back out again. As a result, Earth doesn't drop to the frigid temperatures of outer space every night. If you want a planet where liquid water can exist, you need to equip your planet with greenhouse gases to modulate its temperature. Greenhouse gases like CO_2 are therefore a prerequisite for life.



Figure 38.3.4: The Neoproterozoic 'Snowball Earth' was probably caused by a drawdown in CO_2 levels.

CO_2 is a tiny proportion of our atmosphere by volume (we measure its presence in **parts per million**, not percent or even parts per thousand), but it is very, very important for regulating heat loss. Please do not confuse "small amount by volume" with

"unimportant." Because of the way atmospheric CO_2 interacts with infrared light, its level is ***the*** main Earth-derived control on climate.

Now let's think about how planetary-scale carbon burial impacts the greenhouse effect.

In the mid- to late-Paleozoic, plants evolved and spread across the land area of our planet. These plants extracted CO_2 from the atmosphere, but if they burned or rotted over short timescales, that carbon went straight back into the atmospheric reservoir as CO_2. That's an example of the short-term carbon cycle. On the other hand, if the plants did their photosynthesis, captured the carbon from the CO_2 and flung out O_2 as a waste product, and **then the plants were buried** in swampy sediments, then the plant carbon no longer can interact with the oxygen in the atmosphere. It has been **sequestered**. Now we're discussing the long-term carbon cycle.

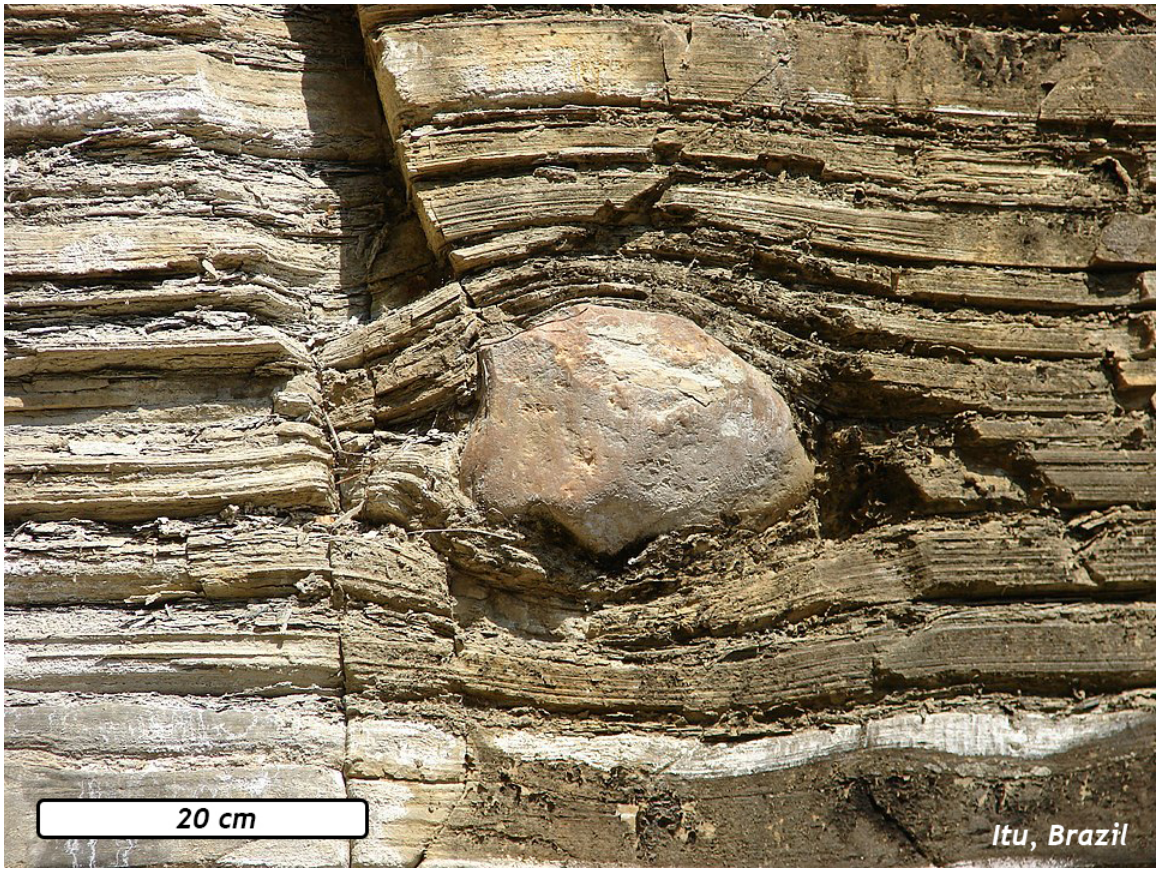


Figure 38.3.5: A dropstone of quartzite in layered rhythmite at Itu, Brazil. Note the faceted shape of the clast (induced by glacial scraping). (Photo by Eurico Zimbres, via Wikimedia.)

There is evidence of the reduction of atmospheric CO_2 at several points in Earth history. During the late Devonian period, excessive formation of black shale (full of carbon captured by marine plankton) led to an ice age and one of the big five mass extinctions. During the ensuing Carboniferous period, repeated transgressive burials of coastal swamps led to cyclothems including the coal seams that powered the Industrial Revolution. The Carboniferous gets its name from the carbon-rich nature of the rocks laid down at that time in England. Please realize that all that carbon in the rock is carbon that ***wasn't*** in the atmosphere. In the Neoproterozoic Era, the planet froze over when CO_2 levels got zeroed out due to excessive weathering of the continental crust. It wasn't organic carbon burial in that case, but reaction of CO_2 with silicate minerals that caused the CO_2 drawdown and cooling.

In summary, high rates of carbon burial lead to lower CO_2 levels, which causes a reduction in the greenhouse effect, and thus global cooling. Sometimes this global cooling is severe enough to cause a mass extinction.

Big bugs

But burial of lots of organic carbon doesn't just lower global CO_2 levels. It also means that the waste product of photosynthesis builds up and has a lesser amount of carbon available to react with. That waste product, of course, is O_2. With less organic carbon

available for oxidation, atmospheric oxygen levels would be expected to go up.

In the planet's earliest years, there was no free oxygen in its atmosphere. Gradually it built up through time, largely thanks to the photosynthetic work of cyanobacteria. By the Paleoproterozoic/ Mesoproterozoic boundary, it had accumulated to something like 1% of the atmosphere by volume. This has been dubbed the Great Oxygenation "Event." Currently, our atmosphere has an atmosphere that consists mainly of nitrogen (N_2) but also includes ~21% free oxygen (O_2).



Figure 38.3.6: The Goliath beetle is the biggest insect that the modern atmosphere can support. (CC-BY; Photo by Raul654, via Wikimedia.)

However, it hasn't been a straight path from the GOE to our present oxygenation. During the Carboniferous period, rates of carbon burial were high, so that means atmospheric CO_2 was low and therefore atmospheric O_2 was high. Estimates are that it got as high as **35% of the atmosphere**. In such an atmosphere, you wouldn't have to inflate your lungs as often in order to get the oxygen you need.

Insects don't have lungs. Instead, they have a network of branching tubes that runs throughout their body, and through which air circulates. The maximum size of insects today is controlled by the efficiency of this tube network in delivering sufficient oxygen to the tissues that need it. The Goliath beetle is the current record-holder as "biggest bug." It can get up to 4.5 inches (11 cm) long. While that's certainly impressive, it doesn't hold a candle to this:



Figure 38.3.7: Large fossilized *Meganeura* dragonfly. (Creative Commons Attribution-Share Alike 3.0 Unported, 2.5 Generic, 2.0 Generic and 1.0 Generic license.)

That's a dragonfly ***the size of a hawk***. Its wingspan is 28 inches (70 cm). It could never survive today, but during the Carboniferous, with an atmosphere of 35% oxygen, such a huge insect was plausible. Other examples of giant arthropods include *Arthopleura* and *Megarachne*.

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38.4: Isotopes

Isotopes are atoms of a given element that vary in their weight. Atoms are made of protons, neutrons, and electrons. The number of protons in an atom determine its identity as an element. For instance, all oxygen atoms have eight protons. All carbon atoms have six protons. Usually, the number of protons (which carry a positive charge) is matched by the number of electrons (which carry a negative charge). This balance makes the atom electrically neutral. If there are more protons than neutrons, the atom is a cation. If there are more electrons than protons, it makes it an anion. We'll ignore ions for the present discussion.

It's the number of neutrons that matters for determining isotopic identity. In addition to its six protons, a given atom of carbon can contain six neutrons, seven neutrons, or eight neutrons. When you add the number or *protons* to the number of **neutrons**, you get the ***atomic mass*** of that atom.

$$6 + \mathbf{6} = \boldsymbol{12}$$

$$6 + \mathbf{7} = \boldsymbol{13}$$

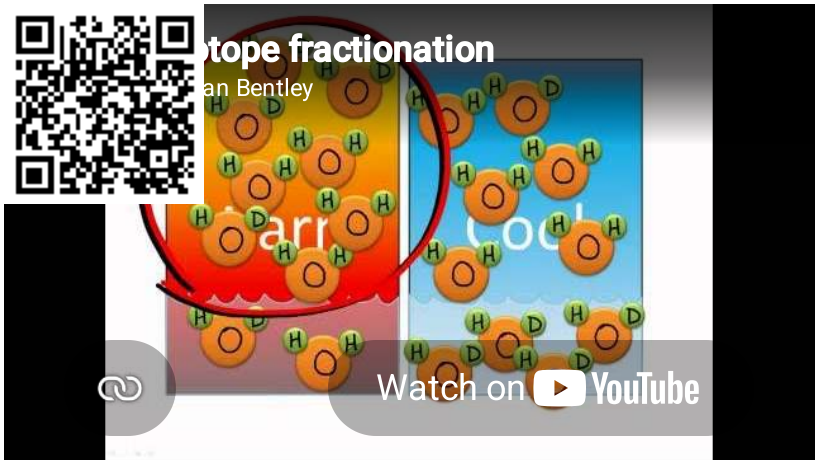
$$6 + \mathbf{8} = \boldsymbol{14}$$

So that means carbon atoms come in three isotopic varieties: ^{12}C, ^{13}C, and ^{14}C. You've probably heard of ^{14}C (carbon-14) as the basis of dating recent organic remains, such as wood, bone, or tissue. It is radioactive, meaning the atom is unstable. Eight is one neutron too many, and the forces that hold the atom together are incapable of maintaining its structure over the geological long-term. As a result, it's not useful to us for the current discussion. It's also a tiny amount of all carbon, about one part in a thousand. So let's set ^{14}C aside and focus now on ^{12}C and ^{13}C instead.

^{12}C and ^{13}C are not radioactive; the forces that hold their atoms together are durable over geological spans of time. Hence, we call them **"stable" isotopes.** ^{12}C is "light carbon." ^{13}C is "heavy carbon." Most carbon atoms are the lightweight kind, ^{12}C. It makes up about 99% of all carbon atoms. But there is also a measurable amount of heavy ^{13}C, comprising about 1% of all carbon on Earth.

It turns out that various natural processes separate light carbon from heavy carbon. These acts of separation, called isotope fractionation, enrich some parts of the Earth system in ^{12}C, while depleting other parts. For instance, photosynthesizing plants prefer ^{12}C-based CO_2 molecules, meaning that plants (and the animals that eat the plants) are **enriched** in ^{12}C, but that act of removing ^{12}C from the atmosphere makes it relatively **depleted** in ^{12}C. Or if you consider it from the perspective of the less-common ^{13}C atoms, then it's the other way around: life is the reservoir that's depleted in heavy carbon, and the atmosphere is enriched in heavy carbon. The **ratio** between these two isotopes is something we use as an indicator of past conditions on Earth, such as its climate. The ratio of ^{12}C and ^{13}C gets skewed between the major reservoirs in the Earth system as climate shifts. The same is true of heavy oxygen and light oxygen (^{16}O vs. ^{18}O), or light hydrogen and heavy hydrogen (^{1}H vs ^{2}H, the latter of which is also called D for "deuterium"). Critically, with all these isotopes, we can measure that skew.

This video discusses three different elements (hydrogen, oxygen, and carbon) in terms of natural fractionation of their isotopes, and relates that fractionation to Earth's prevailing climate:



Because plants prefer ^{12}C over ^{13}C, during times when there are high rates of carbon burial happening on Earth, much of the ^{12}C ends up underground, being converted to coal, oil, or natural gas. That means the atmosphere will be relatively enriched in ^{13}C instead. Because the ocean exchanges gases with the atmosphere, the waters of the ocean will also be enriched in ^{13}C instead of ^{12}C.

When climate is cool, there is less energy available to drive evaporation, so ocean water tends to be enriched in heavier isotopes: water molecules containing deuterium or ^{18}O. Evaporated water molecules are more likely to be isotopically lighter, and so will the precipitation that they drop. During ice ages, the snow that falls to make glacial ice will be a reservoir of isotopically lightweight water, both in terms of hydrogen and oxygen.

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38.5: Fossil fuels

Buried carbon does not necessarily stay buried. Our species has found it useful to extract carbon-rich gas, liquid, and solid compounds from the crust and set them on fire. By oxidizing that carbon, we release energy originally bound up in chemical bonds as much as ~350 million years ago. We refer here to and coal, petroleum, and natural gas: the "fossil fuels." They get this name because all the carbon in each fuel came originally from living things extracting CO_2 from the atmosphere. Fossil fuels would not exist without life. (Fossils are the remains of ancient life.) Specifically, ancient bog and swamp plants were the precursors for coal, and ancient phytoplankton were the precursors for petroleum and natural gas.

Coal



Figure 38.5.1: Bituminous coal. (Ralph & Cheryl Dawes photo.)

Coal is a solid fossil fuel that forms from compressed plant matter called **peat**. Frequently coal layers ("seams") still retain the impressions of leaves, stems, seeds, spores, and other botanical details. Coal comes in several "grades" that reflect varying degrees of compression. The lowest grade is **lignite** (brown coal), which is just a step up from peat, the original mix of plant bits. If it is more compressed still, it will become **bituminous coal**. And the highest grade of coal, **anthracite**, forms only when a coal seam is subject to metamorphism under the influence of mountain-building.



Figure 38.5.2: A peat slide in Shetland, U.K. This cool, wet climate supports the growth of peat, but not trees. Note that the hillside is not a “swamp.” (Callan Bentley photo.)

The plants that are the progenitor of coal live on the land, in wet, aquatic, or swampy settings. It’s traditional to think of coal as coming from swamps, since swamp water is usually stagnant, and thus poorly oxygenated. Carbon-rich plant fragments falling into that water are therefore more likely to avoid oxygenation long enough to be buried. The rate of new carbon being added as plants shed debris outstrips the rate at which oxygen diffuses across the calm surface of the water.

But peat can also form in cool, wet climates such as Ireland or Scotland. There, a layer of vegetation clings to the topography of the land. The damp weather means that sphagnum and heather, the main peat-building vegetation, grow on slopes and hilltops as well as in the valley bottoms, with underlying peat layers averaging a meter or two thick. The entire landscape is essentially blanketed in bog. It grows upward at about 1 cm/year.

Oil



Figure 38.5.3: Distillate of ancient algae, good for powering a civilization. (Callan Bentley GIF.)

Oil has a different origin story. Here, the photosynthesis was conducted by marine or lake algae, **phytoplankton** floating as single-celled organisms in the uppermost, sunlit portion of the water column. When they died, their bodies rained down to the bottom. This transferred the carbon they contained to the dark depths. Often, the bottoms of lakes or the ocean have far less dissolved oxygen than waters closer to the surface. As a result, carbon entering these anoxic waters tends to be buried before it can oxidize. As with swamp coal, it’s a case of the carbon burial rate exceeding the flow of dissolved oxygen to the same place.

Once buried, the algae have to be warmed up to a critical temperature – **about 100°C**, the temperature of a freshly made espresso. If it’s too cold, their little dead cells won’t transform into petroleum. If it’s too hot, the mixture will turn into natural gas (described below) and graphite. The temperature has to be just right.

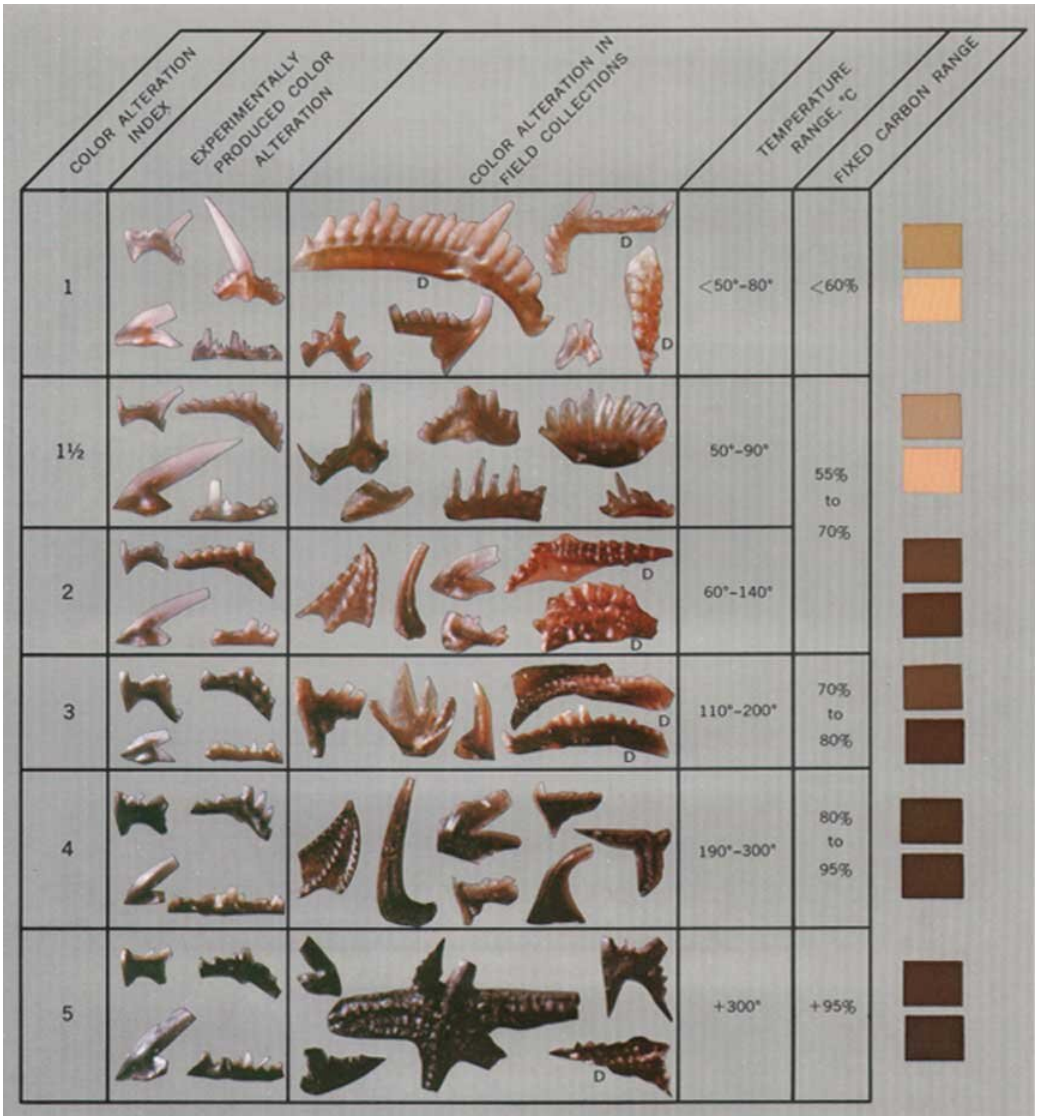


Figure 38.5.4: The Conodont Alteration Index of Anita Harris (then Anita Epstein), et al., 1977. (USGS image.)

Interestingly, one way geologists can make a quick but accurate assessment of the temperature to which potentially-oil-bearing rocks have been warmed is by examining fossils. Specifically, the spiky “elements” of the conodont animal. These are made of hydroxylapatite, a phosphate mineral. The key thing to know about hydroxylapatite is that it **changes color when heated**, like a piece of toast. Geologist Anita Harris developed a vital tool with the Conodont Alteration Index, a 6-point color scale that uses the various “toastings” of conodonts to see if their host rocks warmed up to the appropriate temperature for oil generation.

Conodonts can then be collected from numerous sites in a region, their color assessed using the Index, and then the presence of each color can be contoured. Here’s an example for the sedimentary strata of the Appalachian Plateaus + the Valley & Ridge provinces:

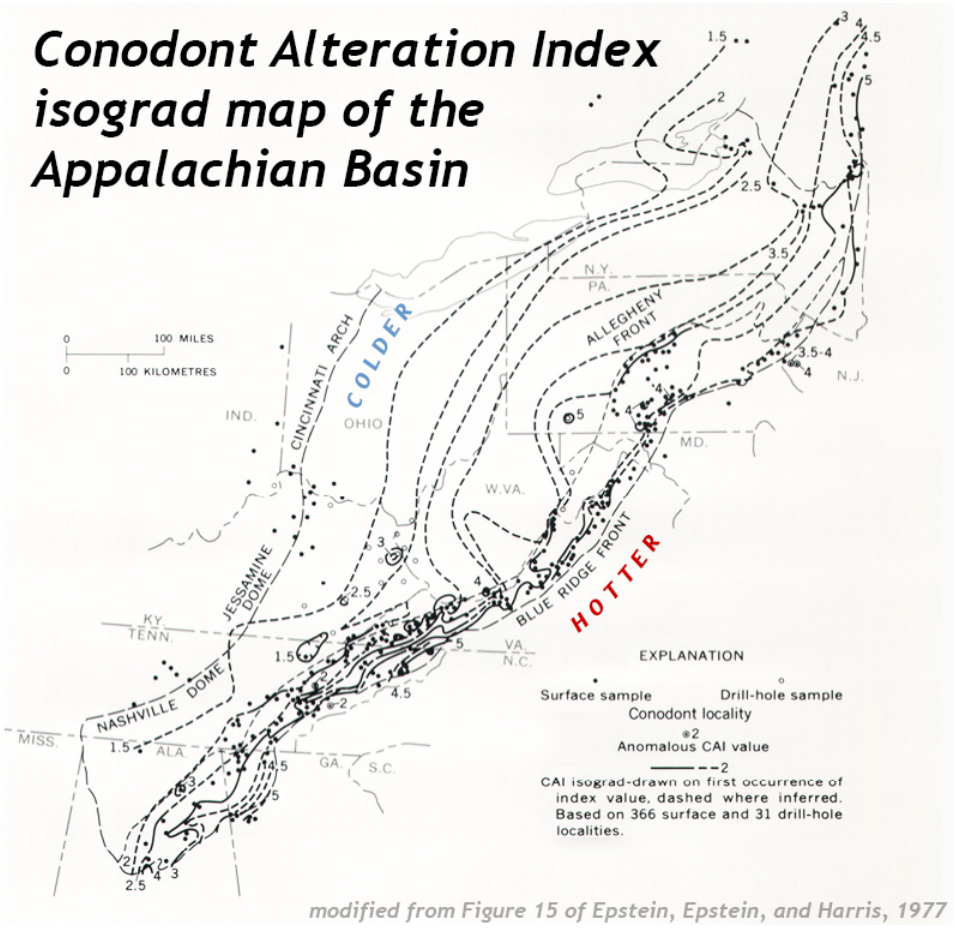


Figure 38.5.5: Conodont Alteration Index color distribution across the Appalachian Basin. (Modified from a USGS figure.)

Once produced, oil must migrate to a spot where it can collect in amounts where humans will find it economically worthwhile to drill it out and extract it. There are plenty of places where oil naturally seeps out at Earth's surface, such as the legendary tar pits of Rancho La Brea in Los Angeles, California.

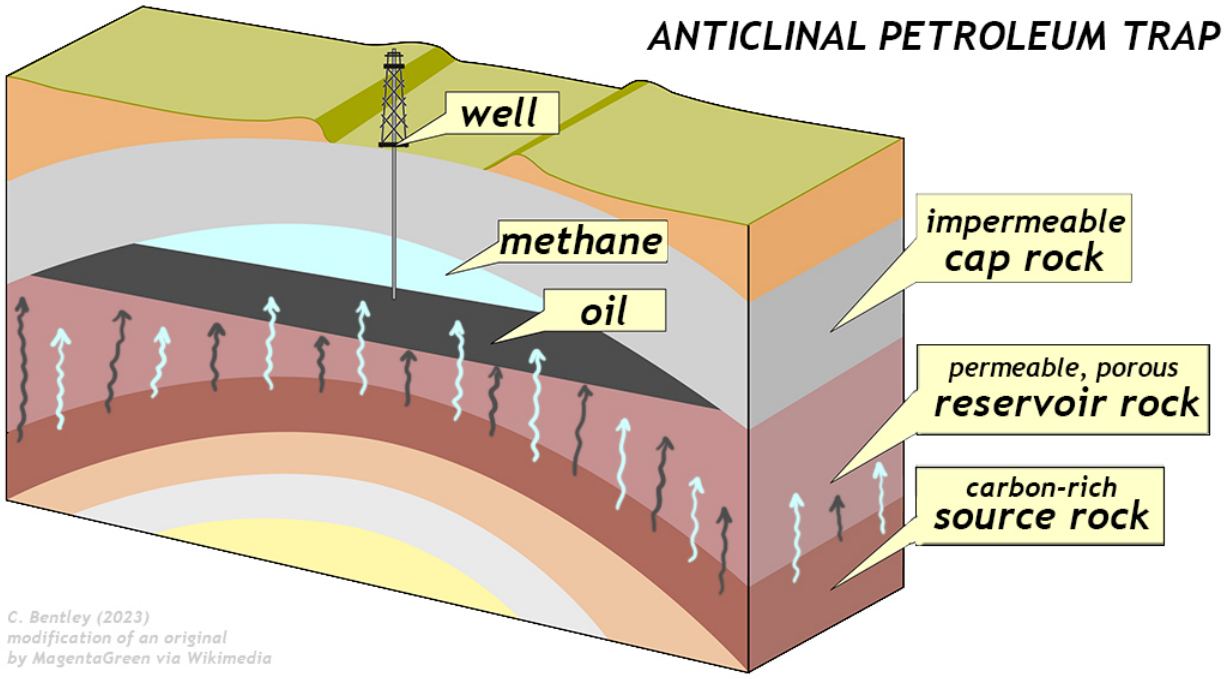


Figure 38.5.6: An anticlinal petroleum trap. (C.Bentley modification of an original diagram by GreenMagenta via Wikimedia.)

But the oil you burn in your automobile found a stable residence underground in some variety of subterranean oil trap. These have various shapes due to various geological circumstances, but one of the easiest to grasp is that of an anticline. If the arch-like shape of an anticline consists of a stack of layers with carbonaceous source rocks at the bottom (such as black shale), impermeable cap rock strata (such as shale) at the top, and a spongey reservoir layer (such as sandstone) in between, that makes for an ideal **petroleum trap**. Petroleum geologists are charged with finding these traps and directing the company's engineers to tap into them.



Figure 38.5.7: Outcrop of the Devonian-aged Marcellus Shale in Pennsylvania. Yellow field notebook for scale. (Modified from a Terry Engelder photo.)

Our third and final fossil fuel is neither a solid like coal, nor a liquid like oil. Instead, it's a mixture of gases, principally methane (CH_4). Like oil, its progenitor is phytoplankton. While these algae live, they photosynthesize. When they die, their bodies rain down into anoxic deep waters, imparting a high carbon content to the sediments deposited there. Typically, the resulting rock is a **black shale**.

Black shales are more common in the geologic record during times of sluggish ocean circulation or widespread stagnation, such as the end-Permian or the mid-Cretaceous. Black shales can also be deposited in lakes; they are not necessarily marine in origin.

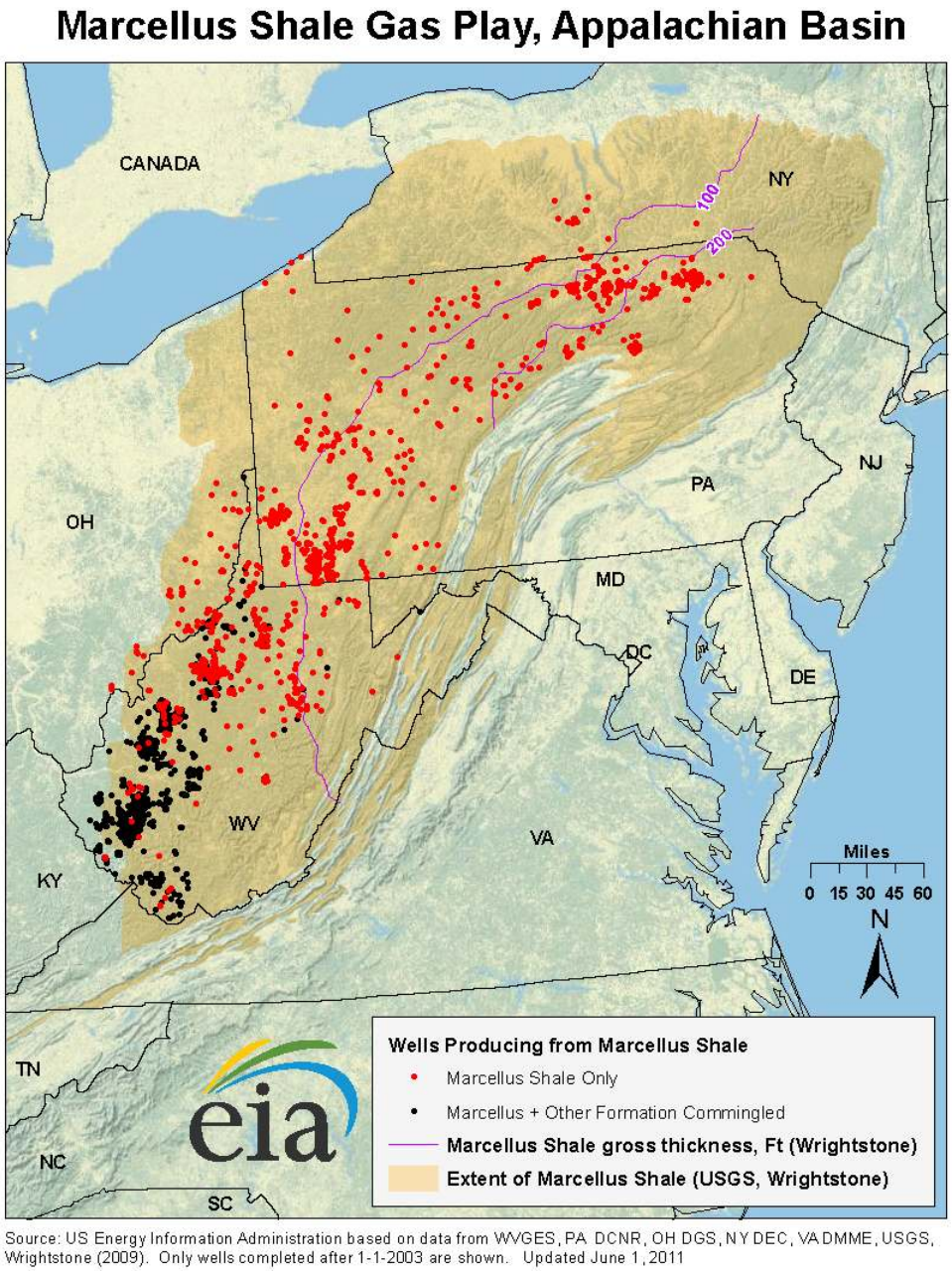


Figure 38.5.8: Wells producing natural gas from the Devonian-aged Marcellus Shale. (US Energy Information Administration map.)

The Marcellus Shale of the Appalachians and the Bakken Shale of the Dakotas are both Devonian-aged black shales with lots of natural gas trapped inside them. In the past decade, the advent of horizontal drilling has allowed fossil fuel companies to drill into these shales and intentionally shatter them, allowing the gas to flow out, up the drill hole, where it can be collected, transported, and sold. The shattering process is achieved by (a) injecting fluids under very high pressures, along with (b) sand, which props open the fractured shale once the cracks open up. This process is called **hydraulic fracturing**, or "fracking."

Natural gas is also produced in coal seams. This variety is referred to as "coal-bed methane." Coal miners have been very conscious of coal-bed methane since coal mining began – because it is invisible gaseous reduced carbon, one spark from mining equipment could trigger an explosion. Far too many miners lost their lives to this hazard. They called it (and any other inflammable gas) "firedamp."

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38.6: Anthropogenic climate change

Because of the rich energy content of fossil fuels, humans have gotten in the habit of burning them, and depending intensely on the burning. The modern global economy and the lifestyles of us modern humans are powered in large part by the combustion of fossil fuel — solids, liquids, and gases that derive their energy content from sunlight that shone on Earth tens to hundreds of millions of years ago. As plants do with oxygen waste, we toss our CO_2 waste into the atmosphere. Because it's invisible, we've been able to ignore it for a long time. Out of sight, out of mind.

CO_2 rise

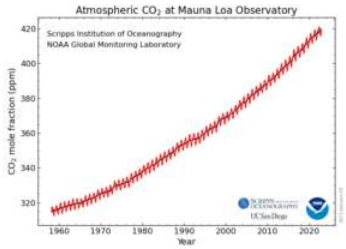


Figure 38.6.1: The record of the past half-century of changing CO_2 levels in Earth's atmosphere. (Scripps / NOAA graph.)

Since humanity began burning fossil fuels in earnest (the start of the Industrial Revolution), CO_2 levels have increased in the atmosphere of our world. The seasonal signal of photosynthesis is still there, but every year, CO_2 ends up at a higher level than the previous year. From a pre-Industrial level of ~280 parts per million (ppm) to the current ~420 ppm, it has increased by about 50%.

It would have been an even larger increase, had not the Earth system been so intimately inter-connected. Much of the CO_2 that humanity has emitted through the burning of fossil fuels has been re-extracted from the atmosphere: some has been taken out by plants doing photosynthesis more efficiently, and some has been soaked up by the oceans.

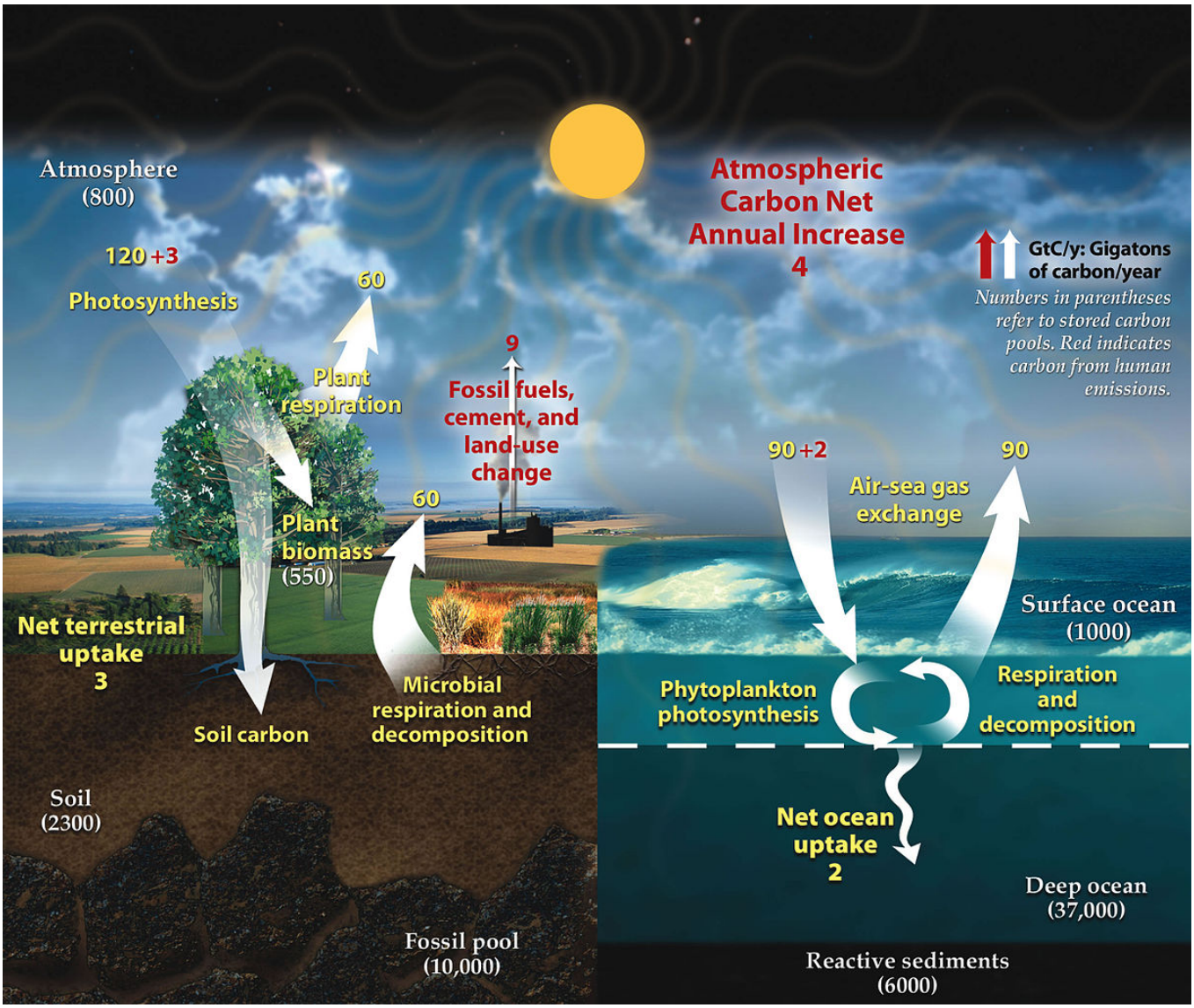


Figure 38.6.2: Thanks to CO_2 absorption by the oceans and photosynthesis by plants, the human-induced carbon loading of the planet's atmosphere isn't as severe as it might otherwise have been. (NASA graphic.)

Estimates of the flow of carbon between reservoirs in the Earth system suggest that human activities release about nine billion tons (gigatons) of carbon into the atmosphere per year, of which three is then pulled out again by plant photosynthesis, and two is dissolved in the oceans. That leaves a **net atmospheric gain of four gigatons per year**. Examine the graphic at left to explore the complexities of carbon fluxes between reservoirs, but follow the red numbers, which show the changes that result from the human economy's dependence on burning ancient trees and algae. This imbalance is why the graph above shows a relentless climb: the rates of carbon **absorption** can't match the rates of carbon **emission**.

Oceanographers are well aware of one of the consequences of this natural carbon sequestration: those extra two gigatons of carbon soaked into the oceans each year produce carbonic acid, which is making the ocean overall more acidic. Already surface ocean pH has dropped by 0.1 unit, with a forecast total of -0.7 pH units change over the next 750 years. This potentially has enormous effects for the viability of life in the oceans. Many of our greatest episodes of mass extinction are correlated with times of intense **ocean acidification**.

But the bigger issue is probably **global warming**. Because of the selective transparency of the CO_2 molecule, the higher level of this greenhouse gas has resulted in Earth retaining more energy than previously. This has caused the average global temperature to increase.

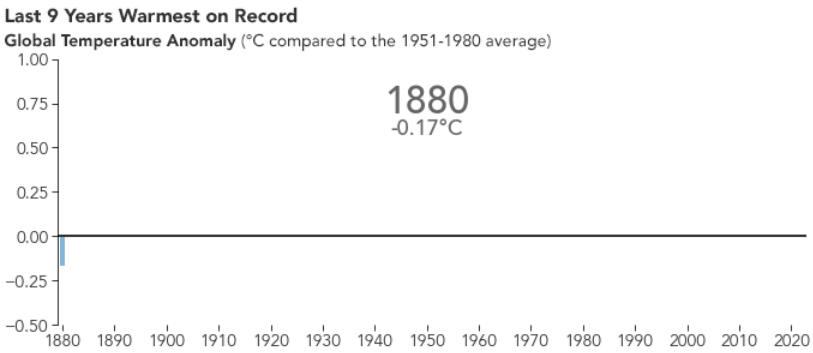


Figure 38.6.3: Ranking the last 142 years of global average temperatures in chronological order. (NASA animation of NASA data.)

There are many **consequences** of this change in the planet's climate, including: melting glacial ice and permafrost, rising sea levels, intensification of tropical storms, increasing desertification, and a spur to the migration of species toward more pole-ward latitudes and higher elevations. It is an interesting time to be alive and scientifically aware on planet Earth.

O_2 fall

There is another aspect of the burning of fossil fuels which is also noteworthy. The basic combustion reaction of $C + O_2 \rightarrow CO_2$ generates CO_2 of course, but it also consumes O_2. For every atom of carbon we oxidize to power our energy grid or heat our homes or propel our vehicles, we lose one molecule of free oxygen from the atmosphere. Thus, as the past six decades of measurements show a rise in CO_2, we should also expect a commensurate **decline in** O_2.

Indeed, that is what we have observed:

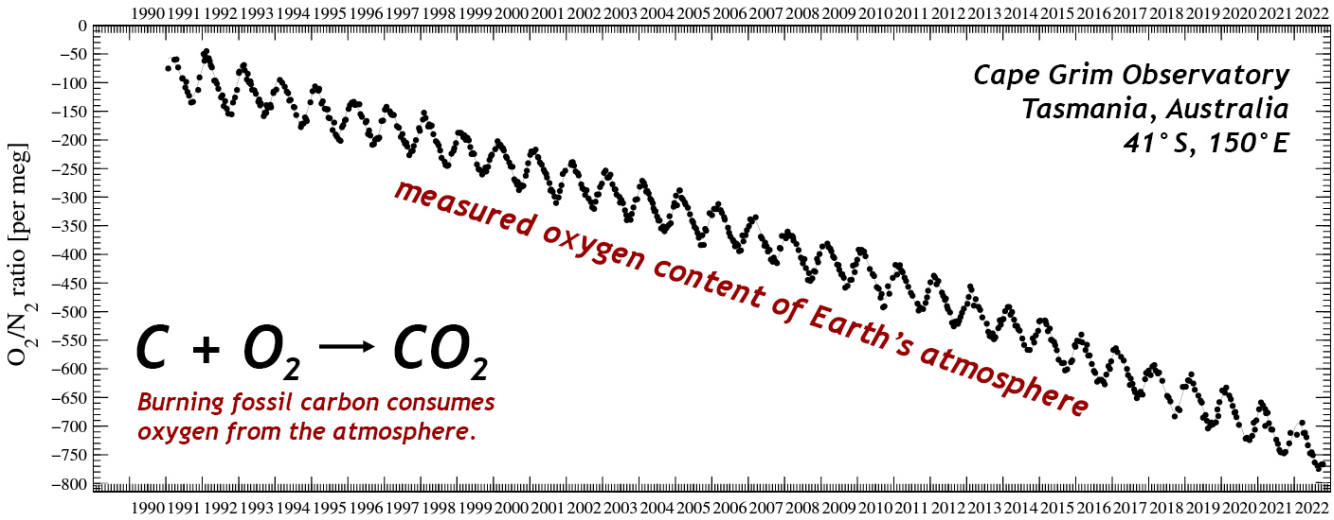


Figure 38.6.4: Measurements of atmospheric oxygen levels show a long-term decline. (Graph by Scripps O_2 program, annotated by C. Bentley.)

At first glance, this graph might seem alarming to readers who breathe oxygen, but remember that this is a parts-per-million level decline in a very large reservoir of oxygen: ~21%. It represents a drop of a few ten-thousandths of the total amount of oxygen. So **you don't have to worry** about running out of breathable air. We will exhaust our supply of fossil fuels long before atmospheric oxygen levels change by even 1%.

Instead, save your worrying for the impacts of global warming.

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38.7: Further reading

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CHAPTER OVERVIEW

39: (Case Study) Quaternary Quandaries - defining the geologic "now" and explorations of the Anthropocene



Figure \(\PageIndex{1}\): Castle Bravo atmospheric nuclear test, 15kT, at Bikini Atoll, March 1, 1954, Pacific Proving Grounds (Source: Wikimedia Generic CC). Nuclear tests like these have left behind radioactive materials that may end up defining the Anthropocene, beginning in 1950.

Key Concepts

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

- Describe how geoscientists define start and end dates for periods on the geologic time scale.
- Differentiate between the different epochs and ages within the Quaternary Period.
- Differentiate between the different ages within the Holocene Epoch.
- Describe how recent changes to the Holocene were made.
- Discuss the significance of the Meghalayan Age and its associated GSSP.
- Explain the current debate and discussions surrounding the potential defining of a new Anthropocene Epoch.
- Identify what a GSSP, or “Golden Spike” is in the geologic record.

How do we define the time in which we live? Can it be done in a word or an idea? Or, is it through the sum of our material past, the objects we leave behind to tell our story? Do we get to define our time, or is it something left to those who follow us later on?



Figure \(\PageIndex{2}\): A classic AM-FM radio dial (Source: Public Domain).

The band of time can be modeled metaphorically using the old AM/FM radios you may have seen a car, the kind with knobs you can use to tune it. Radio tuners have a starting frequency and ending frequency which, in the case of the FM band, runs from 87.5 to 108 MHz. If you wish to find a particular station, or frequency, all you needed to do was turn the dial. Each frequency was already defined, depending upon your location, by a given radio station. If you wanted the rock station, you tuned it to perhaps 100.7 MHz, and so on. Because it was already there and defined before you used it, you knew what to expect and what the parameters of a particular location on that radio dial was like. Radio stations are not really boundaries in the FM spectrum, but metaphorically they can be viewed as little waystations, locations of significance, amidst the staticky stuff in between, with nothing of distinction to set it apart from any other position on the spectrum.

Geologic time is very similar. How can we use it to give some kind of meaning to the expanse of time?

The Geologic Time Scale is a model that has been developed by lots of people working over lots of time. Their work that helps us answer these difficult questions. As is the case with all models, the Geologic Time Scale is wrong in many ways. However, it is still very useful. As time and knowledge accumulate, we improve the model. Such improvements can include everything from simple date changes to larger-scale changes like the introduction of new periods or epochs. However change progresses, the time scale is a model that represents our best understanding of the band expanse of time as we understand it. The geologic time scale is not static, but dynamic. It is managed by the International Commission on Stratigraphy (ICS), an affiliate of the International Geological Union. The work of the ICS is to compile ongoing research into the various portions of the time scale in order to refine the boundaries of the Eons, Eras, Periods, and Epochs we use to define time. To do this, **Global Stratotype Sections and Points** (GSSPs) are identified that provide a consensus marker in time for any given boundary. These GSSPs are often referred to as “Golden Spikes”, a reference to the golden spike used to complete the Trans-Continental Railroad, as they serve as clearly-defined time markers that allow researchers (and professors and students!) to share in common discussion.

But, how hard is it exactly to define such boundaries in time?

Over the course of this case study, we will explore three critical questions:

1. How do we use the geologic time scale as a model to make sense of time?

2. How do we make changes to this model as we learn more about time?

3. Can we use this model to help us define our own time in the present and recent past?

The geoscientists working with the Subcommission on Quaternary Geology are wrestling with these questions. Their subcommission is a part of the International Commission on Stratigraphy (ICS). Ultimately, this case study will explore a proposed new epoch, the Anthropocene Epoch. Before we get there, we need to spend some time exploring how the ICS has gone about defining our most recent era, periods, epochs, and ages. Hopefully, we can then revisit the above questions and have some satisfying answers.

39.1: A Quaternary Quandary

39.2: The Anthropocene

39.3: Concluding Thoughts

39.4: Further Reading

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39.1: A Quaternary Quandary

Our modern geologic periods, epochs, and ages have been a state of flux. Currently, we live in the Meghalayan Age of the Holocene Epoch of the Quaternary Period of the Cenozoic Era of the Phanerozoic Eon. However, there have been many changes to this hierarchy over the last few hundred years, with significant changes recently.



Figure 39.1.1: Giovanni Arduino (Source: CC BY 4.0; By Istituto Veneto di Scienze, Lettere ed Arti, https://commons.wikimedia.org/w/inde...curid=38573367).

Our modern sense of geologic time began in northern Italy. In 1759, a man by the name of Giovanni Arduino introduced three terms to help make sense of the geology of the Alps where he worked. He used the term “Primary” for the ancient metamorphic cores of the mountains, “Secondary” for the hard sedimentary rocks on the flanks, and “Tertiary” for the sedimentary rocks in the foothills, which were not quite as hard. *Primary* and *Secondary* were eventually dropped by scientists because they did not always describe what was seen in regions in other areas. *Tertiary* remained. In 1829, French geologist Jules Desnoyers added the term “Quaternary” to describe the most recent geological materials, including lava flows and loose sediments. The terms Tertiary and Quaternary are still with us, though embattled in different ways, their dates and how they are viewed having changed over time. As recently as 2000, these two periods comprised the entire Cenozoic Era, with the Tertiary Period lasting from 65 Ma until 2.58 Ma and the Quaternary Period running from 2.58 Ma to the present.

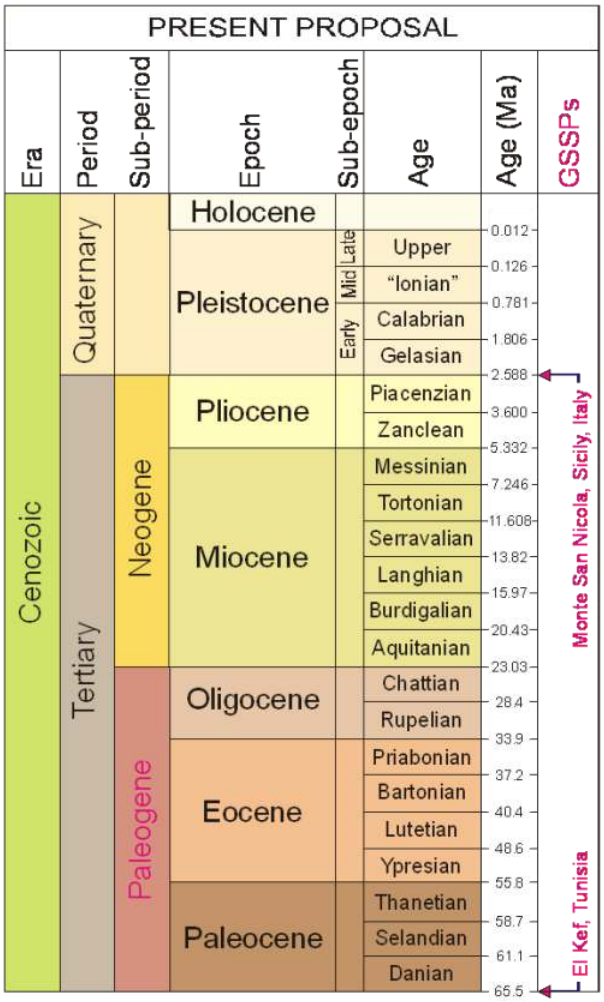


Figure 39.1.2: The Cenozoic Era. The Tertiary Period has been formally removed and the Paleogene and Neogene made Periods. This chart reflects the former situation prior to these changes (Source: The Geological Society).

In 2004, the keepers of geologic time, the ICS (International Commission on Stratigraphy) completely dropped the terms Tertiary and Quaternary from the time scale, replacing them with two divisions, the Paleogene Period (66 Ma to 23 Ma) and the Neogene Period (23 Ma to 2.58 Ma). This was done in order to standardize the Cenozoic Era time periods to be consistent with the rest of the time scale. Marine geologists felt that the Neogene should also encompass all of the most recent portions of time. A major disagreement arose, however, because terrestrial geologists felt that the most recent geological materials on land required that the old term Quaternary Period be retained. These sediments and formations included all of the ice age materials, which are different from Neogene materials prior to 2.58 Ma. Terrestrial workers won the argument and in 2009, the ICS voted to reinstate the Quaternary Period.

The base of the the Quaternary is marked by the beginning of what is known as the Gelasian Stage. The lower boundary of this chronostratigraphic unit is marked by a GSSP location near Gela, Italy, and is primarily a magnetostratigraphic boundary, but secondarily marks the extinction of a number of calcareous nanofossils. Perhaps most importantly, the Gelasian Age ("Stage" is a chronostratigraphic term equivalent to the lithostratigraphic term "Age") marks the start of the growth of northern hemisphere glaciers and ice sheets. You may know this time by another name: the Pleistocene Epoch. Perhaps images of mammoth, mastodon, saber-toothed cats, and early humans perhaps just popped into your mind!

Currently, the Quaternary Period is broken up into two Epochs, the Pleistocene Epoch (2.58 Ma to 11.7 Ka) and the Holocene Epoch. There is a move to add another time division, something perhaps called "Anthropocene," but it is not finalized yet. We will come back to that later.

Exercise 39.1.1

Geologic "Periods", like the Quaternary Period, are not necessarily static. What are some of the characteristics of a unit of geologic time that cannot change as new information arises?

a. Fossils that define the period

b. End dates

c. Name

d. Anything can change with new information

Answer

d. Anything can change with new information

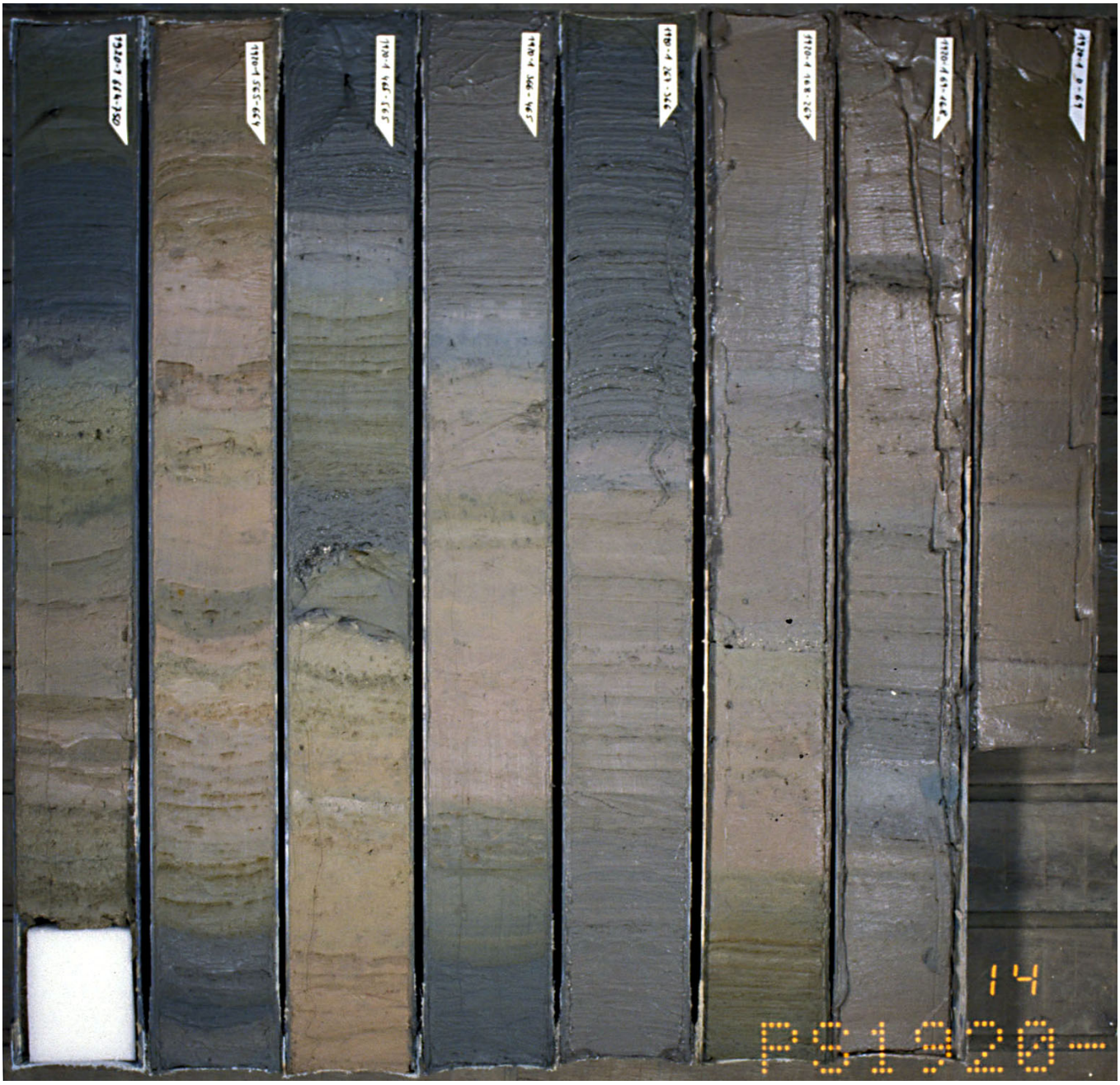


Figure 39.1.3: Sections of sedimentary deep sea cores from off the coast of Greenland (Source: CC BY 3.0; By Hannes Grobe (talk) – Own work, https://commons.wikimedia.org/w/inde...?curid=6829604).

The Quaternary Period differs from the Paleogene and Neogene Periods because it is defined by ice age cycles, known as greenhouse (interglacial) and icehouse (glacial) periods. An entire glacial/interglacial cycle typically lasts about 100,000 years and is driven by Milankovitch orbital changes. And, there are enough of these that the record is quite complicated. When you look at the climate record of the Pleistocene Epoch, however, you see many, many more up and down swings in temperature. These smaller, or shorter, changes in temperature during a glacial period are “stadials” and “interstadials”.

In addition to ice cores, researchers have collected extensive deep sea sediment cores from off the coast of places like Greenland. These sediment cores record critical data, oxygen isotopes in particular. From these, a very fine record of oxygen isotope fluctuations is recorded, which in turn reflect periods of warming and cooling. These alternating periods are called Marine Isotopic Stages (MIS). These stages are numbered from the start of the Holocene Epoch and going backward in time. Even numbers were assigned to cool periods (stadials) and odd numbers to warm periods (interstadials). For example, our current Holocene Epoch was

given the number MIS-1, since it is an interstadial warm period. The last major 100,000 year long glacial period, the Wisconsinan (so called in North America), encompasses several isotopic stages, MIS-2,3, and 4.

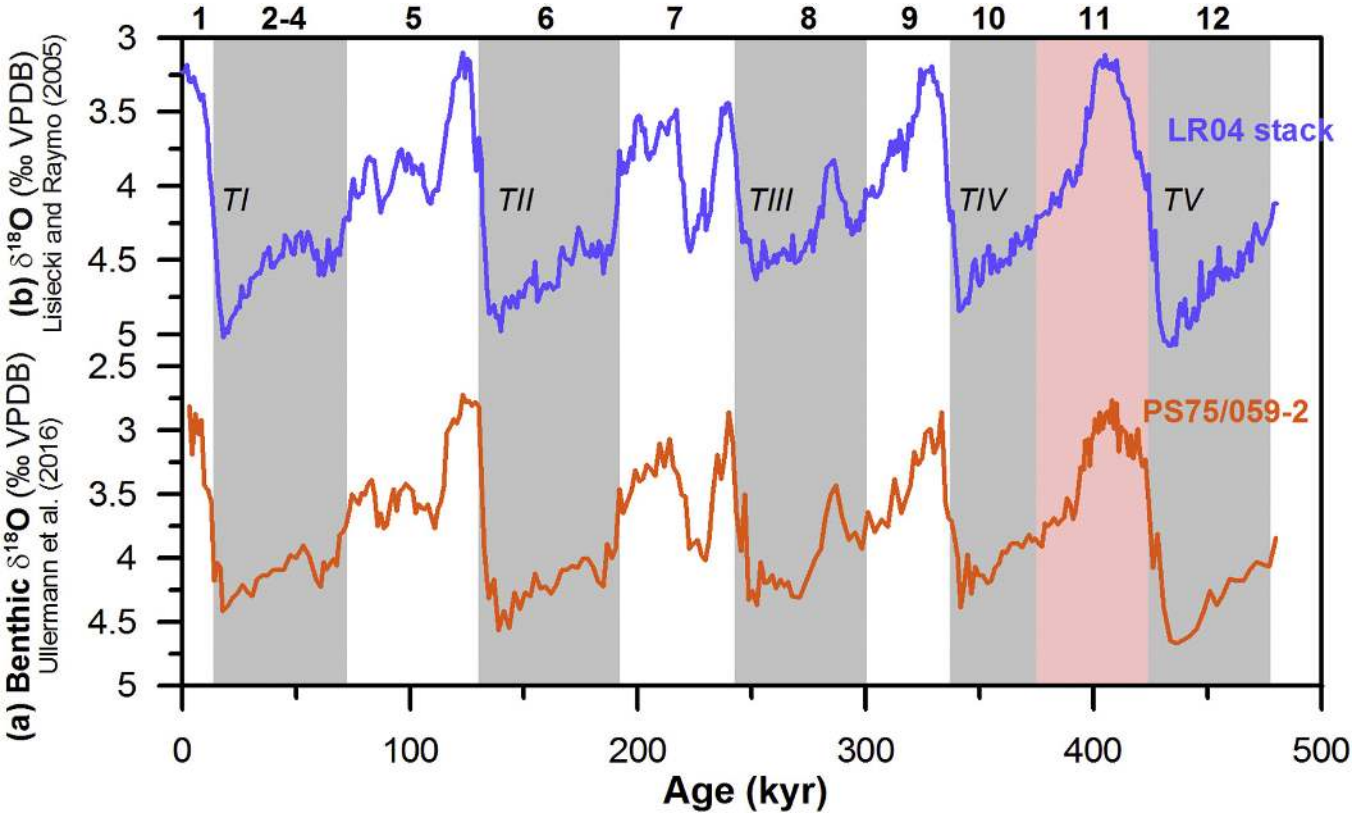


Figure 39.1.4: Marine isotopic stages 1-12. Gray areas are glacial intervals, white areas are interglacial intervals. Note that each glacial interval contains many smaller fluctuations, or stadials/interstadials. The two records here, both from benthic marine sources, agree very closely. The Holocene is marked by MIS 1, MIS 2-5 make up the rest of the last glacial/interglacial cycle (Source: Saavredo-Pellitero, 2017).

In all, over 100 Marine Isotopic Stages have been identified. These record 2.6 million years of climatic change throughout the Pleistocene and Holocene Epochs. It is worth asking why the Holocene, as MIS-1, is separated out as somehow different from the rest of the marine isotopic stages of the Pleistocene. In short, Holocene climatic records indicate that this interstadial period has not only now exceeded the average length of these (~1000 years) at 11,700 years, but it has also seen a much more stable temperature throughout. This is different than past interstadials during the Pleistocene. Coupled with an aberrant climatic shift at its start called the “Younger Dryas” (discussed later in the chapter), these attributes make the Holocene fairly unique from the Pleistocene.

Today, the Cenozoic Era consists of three well-defined and agreed upon periods, the Paleogene Period, Neogene Period, and Quaternary Period. Defining time is clearly challenging and requires detailed high-quality data. Now that we have firmly established the three Periods of the Cenozoic Era, we need to dive more deeply into the Quaternary Period to explore its epochs...and to see the changes that have been and are afoot there.

What’s in an epoch?

Geologic eons are broken up into eras and eras into periods. Geologic periods are broken up into epochs and epochs into ages. During the Cenozoic Era, the Paleogene Period is broken up into the Paleocene, Eocene, and Oligocene Epochs. The Neogene Period is broken up into the Miocene and Pliocene Epochs. Each one of these earlier Epochs is marked by important evolutionary and geologic changes that define these bands of time. The final Period of the Cenozoic, the Quaternary, is divided up into two epochs, the Pleistocene Epoch and the Holocene Epoch. These are all lithostratigraphic terms, defined by sedimentary rock layers. In terms of time, or chronostratigraphic nomenclature, the rock layers that define an epoch are referred to as a series.

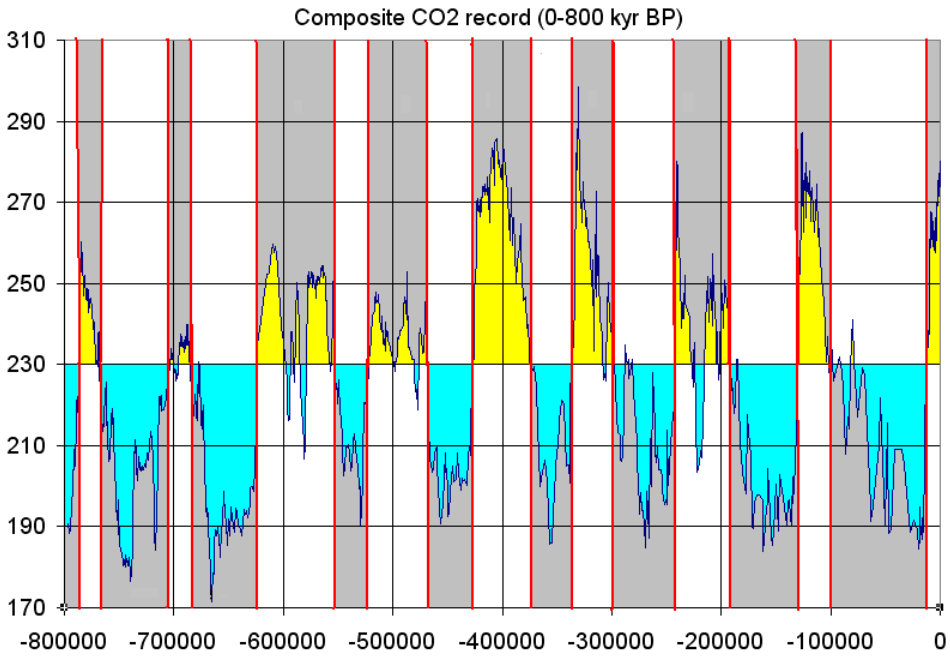


Figure 39.1.5: Icehouse and greenhouse periods as recorded in Antarctic ice cores from the last 800,000 years. Greenhouse periods are yellow, icehouse periods are blue. Note that the highest carbon dioxide concentration recorded during this time is around 280pm, versus today's 415ppm (Source: CC BY-SA 3.0; By Tomruen – Data from ncdc.noaa.gov, and this looks best: Composite CO_2 record (0-800 kyr BP), marked up with 230 ppm transition between glacial and interglacial periods, https://commons.wikimedia.org/w/inde...curid=16147504.).

The Pleistocene is an epoch of wicked climatic seesaws. At times, there is a layer of ice a mile thick over southern Canada. At other times, there is nearly nothing of that left. We also tend to define the Pleistocene by its fauna, including mammoths, mastodons, saber-toothed cats, and more. The name Pleistocene comes from the Greek and means "most new," a term given it by Charles Lyell, back in the 19th century. As mentioned above, the base of the Pleistocene is marked by the start of the Gelasian Age, a time when large ice sheets began to grow in higher latitudes around the world. Based upon there being 40 marine isotope stages identified during this time, there were likely at least 20 glacial cycles during this age. The age ends at 1.8 Ma with the extinction of the nannofossil *Discoaster brouweri*.

Subdivisions of the Quaternary System			
System	Series	Stage	Age (Ma)
Quaternary	Holocene		0–0.0117
	Pleistocene	Tarantian	0.0117–0.126
		Ionian	0.126–0.781
		Calabrian	0.781–1.806
		Gelasian	1.806–2.588
Neogene	Pliocene	Piacenzian	older

Figure 39.1.6: The Pleistocene Series of the Quaternary System (Source: ICS).

The story of the Pleistocene continues stratigraphically upward into the Calabrian Age (1.8 Ma to 0.774 Ma) into the Chibanian Age (0.774 Ma to 0.129 Ma) and finally to the "Upper" Age (proposed "Tarantian Age," 0.129 Ma to 0.0117 Ma). The boundary between the Calabrian and Chibanian is marked by the Earth's last magnetic reversal, but the boundary between the Chibanian and "Upper" Age is marked by the onset of the Eemian interglacial, which is the last warm period before the onset of the Holocene Epoch. Currently the "Upper" Age of the Pleistocene is under consideration for a new name, the Tarantian Age. However, as of just a few years ago from the time of this writing (June 2020), neither a lower or upper GSSP had been defined. In 2018, the upper boundary of a potential Tarantian Age was defined by the onset of the Greenlandian Age of the Holocene Epoch, as defined by a layer within a core taken from the Greenland ice sheet by the Greenland Ice Core Project. The very top of the Tarantian Age occurs

at the end of a brief return to icehouse conditions after a significant warm up. This event is referred to as a the "Younger Dryas" and its end is the beginning of the Holocene Epoch and the Greenlandian Age.

This all may seem complicated, full of funny names and obscure dates. There is truth to this on the surface. However, when you peel away all of this complexity, we have described a model that does something conceptually simple: **the geologic time scale helps us define time through the environmental change**. The model is useful because it is so adaptable.

A Clima(c)tic sidebar – The Younger Dryas

What is the "Younger Dryas," you might ask? The event itself was a brief return to icehouse conditions at the very end of the Pleistocene Epoch. This kind of funny-sounding name actually references a particular indicator plant, *Dryas octopetala*, an alpine-tundra wildflower that became more prevalent during this time. We know this because its pollen became more common in lake sediments.

It is called "Younger" because it is the last of three similar sudden cooling phases during a much longer, gradual, warming from the last glacial maximum. The others are called the "Oldest Dryas" and "Older Dryas." The Younger Dryas begins after a period of gradual warming that had lasted from 24,000 years ago to 12,900 years ago. For about 12,000 years, the Earth had been slowly warming, emerging from a long icehouse period as had been the cycle since the start of the Pleistocene 2.58 million years ago. Then, quite suddenly, actually in the course of just a few decades (a human lifetime or less), the global climate reverted to an icehouse period for another 1,000 years, until 11,700 years ago.

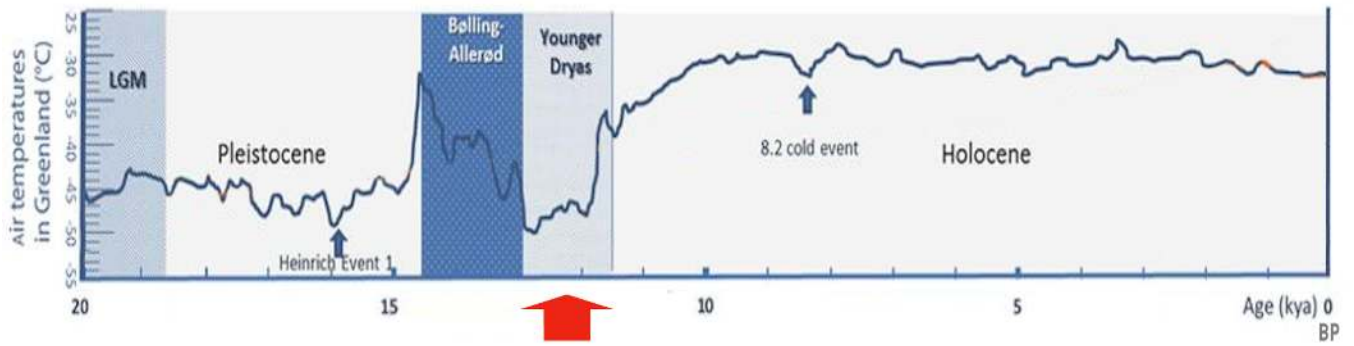


Figure 39.1.7: Temperature proxy record from oxygen isotopes for the period surrounding the Younger Dryas event. Note the rapid warming that occurs at the end of the Younger Dryas. This rapid warming, during most interstadial periods, is followed by several hundred years of gradual cooling. This has not been the record of the Holocene Epoch (Source: By Daniel E. Platt et al., https://commons.wikimedia.org/w/inde...curid=79219152).

What caused this particular climatic event is complicated and not fully understood. It is particularly interesting because of the rapidity of the cooling, which has its counterpart in today's similarly rapid warming at the hand of humans. Humans, however, did not cause the rapid cooling of the Younger Dryas. But, what did? There are three hypotheses worth exploring. None of them are conclusive, but because the question is open, science is at work.

Hypothesis 1: One hypothesis for the onset of the Younger Dryas is the Vela Supernova explosion that took place at that time. In the constellation Vela, a massive star exploded. Energy from this event is suggested as the cause of the observed evidence for the destruction of the ozone layer, changes in nitrogen levels at Earth's surface and in the atmosphere, megafauna extinctions, and changes in some stable isotopes. However, there is little understanding of how such events could cause these things. So, while you can read about this hypothesis, there is little support for it.

Hypothesis 2: A second hypothesis suggests that a volcano in southern Germany, Laacher See, erupted. An unusual volcano, a "maar lake," it is today a low-relief crater about 2km in diameter. It is calculated that the eruption at the time would have been large enough to send enough material into the atmosphere to account for the degree of cooling during this period. However, it has not been studied thoroughly enough.

Hypothesis 3: A third hypothesis proposes a bolide impact large enough to cause the climate change observed. A wide variety of evidence exists from sites around the world for their having been such an impact at that time. These include grains of melt-glass and nanodiamonds. What was missing was a crater of the right age. In 2018, under a glacier called "Hiawatha" in northern Greenland, researchers found a massive crater that may fit the bill. At 31km wide, it is not as large as Chicxulub (the dinosaur

extinction crater), but represents the effects of what was likely an iron asteroid 1.5km across slamming into the Earth at the time. Its timing is still up for debate, as it could be as old as 100,000 years or as young as 12,900. More research will bring answers.

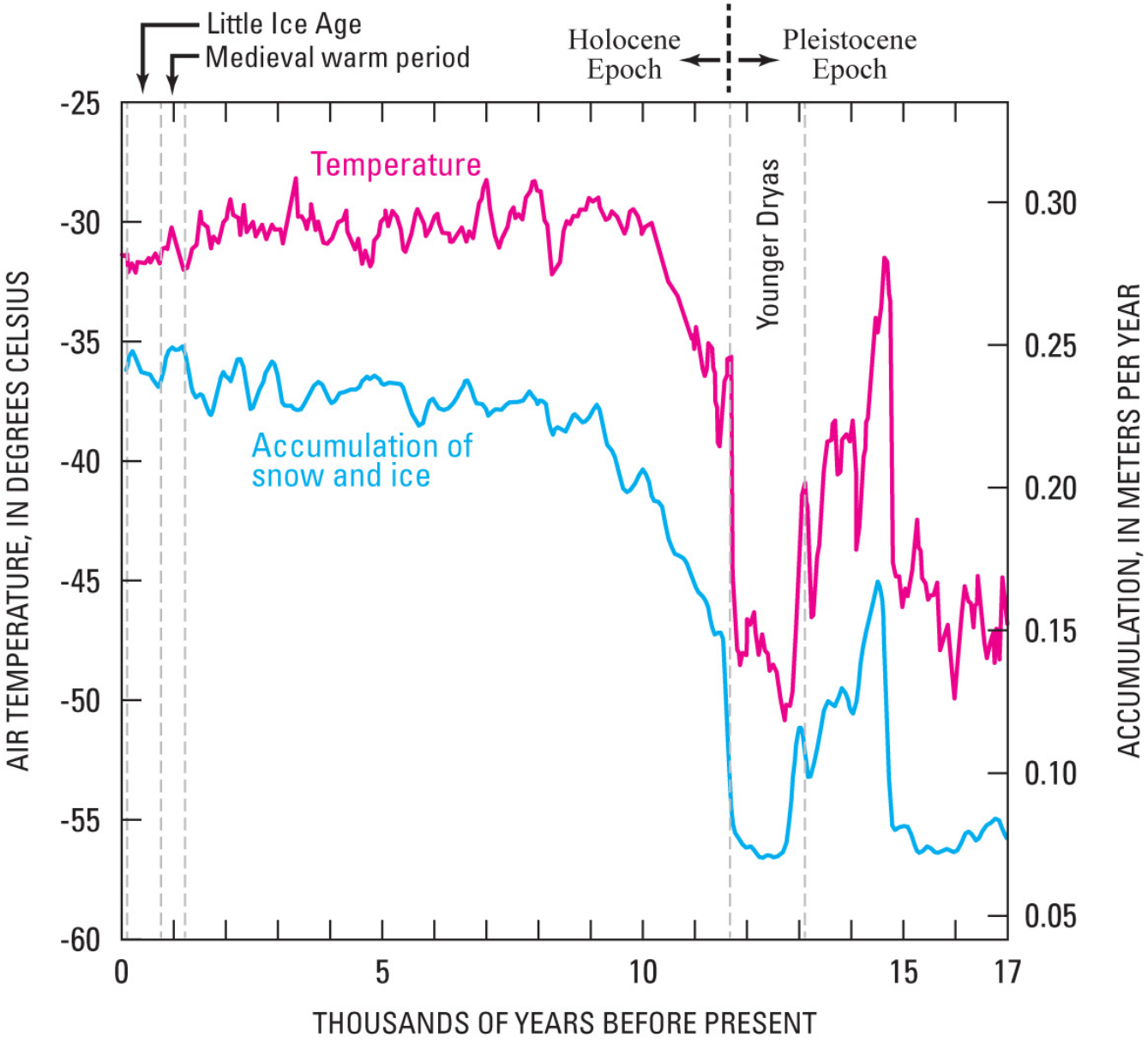


Figure 39.1.8: The Holocene Epoch is directly preceded by the Younger Dryas climate event at the end of the Pleistocene Epoch. Whatever caused the Younger Dryas, it was a short-lived cold excursion (Source: Public Domain, https://commons.wikimedia.org/w/inde...curid=73612526; By United States Geological Survey – https://pubs.usgs.gov/pp/p1386a/gallery2-fig35.html).

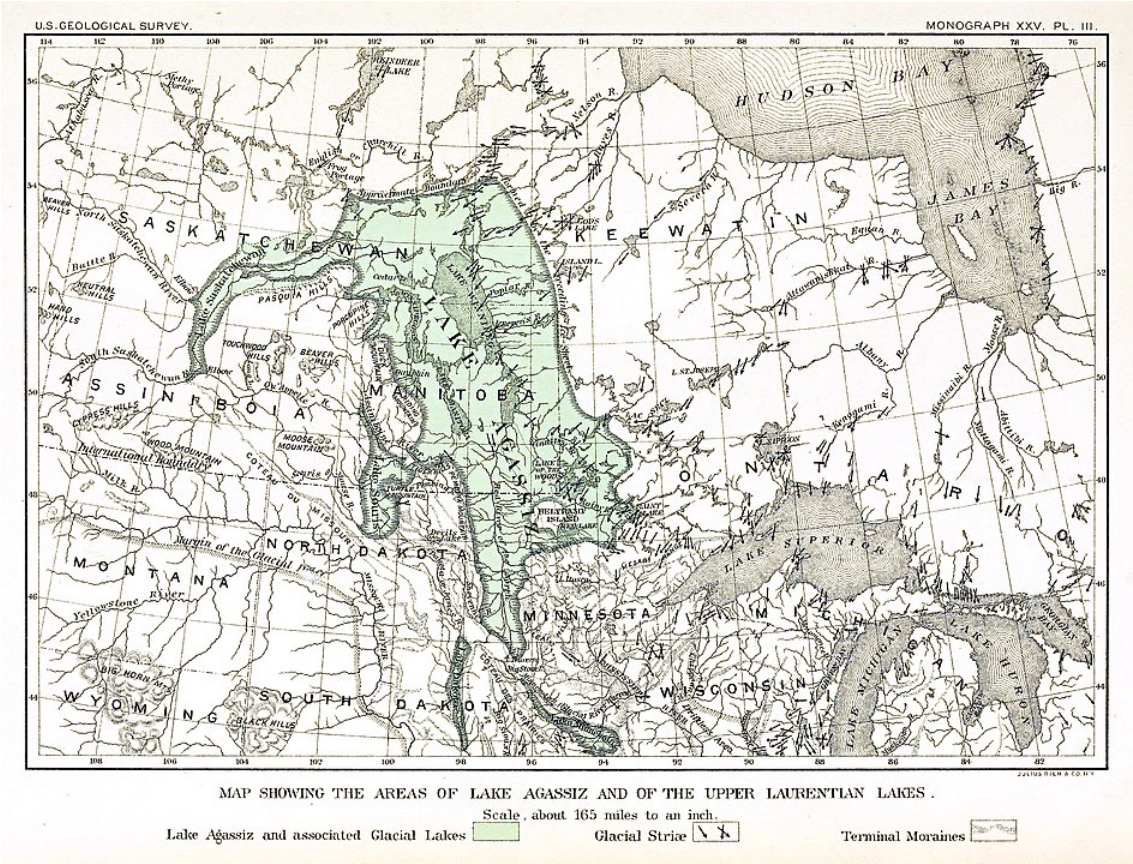


Figure 39.1.9: 19th century map of the extent of post-glacial Lake Agassiz created by geologist Warren Upham. The actual lake's boundaries were likely much greater than the area indicated here (Source: Public Domain).

Hypothesis 4: A fourth hypothesis for this abrupt onset of climate change is the emptying of glacial Lake Agassiz in southern Canada into the North Atlantic via the St. Lawrence seaway. Remnants of this post-glacial lake still exist today and include Lake of the Woods, Lake Winnipeg, and Lake Manitoba, among others. Periodically, this lake would be dammed up by ice, allowing fresh meltwater to accumulate over a very large region. There are five identified phases in all in the record. When the glacial dam would burst, massive volumes of freshwater would enter the North Atlantic and disrupt the current thermohaline circulation and significantly alter global climate.

Did I Get It? - Quiz

? Exercise 39.1.1

In chronostratigraphic terms, an Epoch is equivalent to a _____. Rocks contained in such units can be directly linked to a particular geologic Epoch, independent of geographic location.

a. Series

b. Stage

c. Unit

d. System

Answer

a. Series

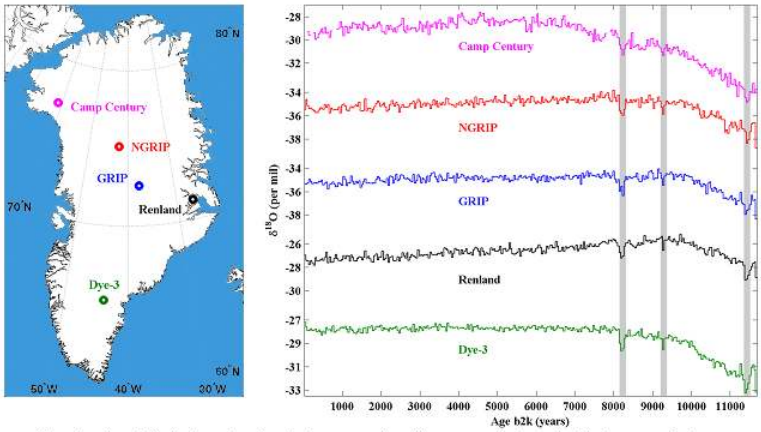
Eonothem & Eon	Erathem & Era	System & Period	Series & Epoch	Subseries & Subepoch	Stage & Age	GSSP
Phanerozoic (pars)	Cenozoic (pars)	Quaternary	Holocene	Upper & Late	Meghalayan	present
				Middle	Northgrippian	4250 years b2k
				Lower & Early	Greenlandian	8236 years b2k
			Pleistocene	Upper & Late	Unnamed	11,700 years b2k
				Middle	Unnamed	~126 ka
				Lower & Early	Calabrian	~0.773 Ma
					Gelasian	1.80 Ma
						2.58 Ma

Figure 39.1.10: Quaternary System subdivisions, including the Holocene and its newly accepted Ages (Source: ICS).

Holocene, from the Greek, means “Entirely New.” For 11,700 years (since 11.7 ka), the climate on our planet has been remarkably stable. Because of this stability, our own species has been able to not only flourish, but to spread well beyond the geographic ranges and biomes to which it was limited during the Pleistocene. While our hunter-gatherer ancestors within the genus *Homo* have been wielding fire as far back as 1.2 million years, they relied only on what the land provided them for all of that time when it came to food and shelter. That meant hunting game, gathering fruits and plants, and more. The long stability of Holocene climate, combined with our species’ cultural development up to that point, made technologies leading to agriculture, via the Neolithic Revolution, possible. This revolution would provide new sources of food, through the artificial selection of grains and subsequent plants (and animals, eventually), that no longer required consistent or even just seasonal movement. It was possible to settle in one place, develop towns, eventually cities, and the specialization and division of labor. It may have also made it already possible for our species to alter the planet’s climate. More on that in a moment...

Unusual Stability – Sort Of...

Is the Holocene’s climate unusually stable? When compared to other spans of time between ice ages, including the last interglacial (the Eemian), there are many similarities in terms of temperature and duration. The Holocene Epoch is one of the longest interstadial periods on record. This being said, there are some noteworthy departures from this stability that are worth exploring. As of 2018, two of these have been used to further divide our epoch into ages.



Map showing drill site locations for five ice cores that all penetrated the entire Holocene and the corresponding holocene d18O records (reflecting temperature) for the five ice cores. b2k means before A.D. 2000. Three cold events that are detectable in all five ice cores are shaded gray.

Figure 39.1.11: Oxygen isotope records for five Greenland ice cores. Note the isotope excursions at 8.2 ka, 9.3 ka, and 11.4 ka. Otherwise, the Holocene climate has been very stable (Source: University of Copenhagen Centre for Ice and Climate).

Subdividing the Holocene: Climatic Ages

While there have been one or two attempts in past decades to further subdivide the Holocene Epoch, nothing stuck. The Quaternary, in general, is overwhelmingly defined by ice ages. Because Quaternary stratigraphy is overwhelmingly defined by deep sea marine sediments and ice core markers, it is reasonable then to assume the task of subdividing the Holocene Epoch would be no different.

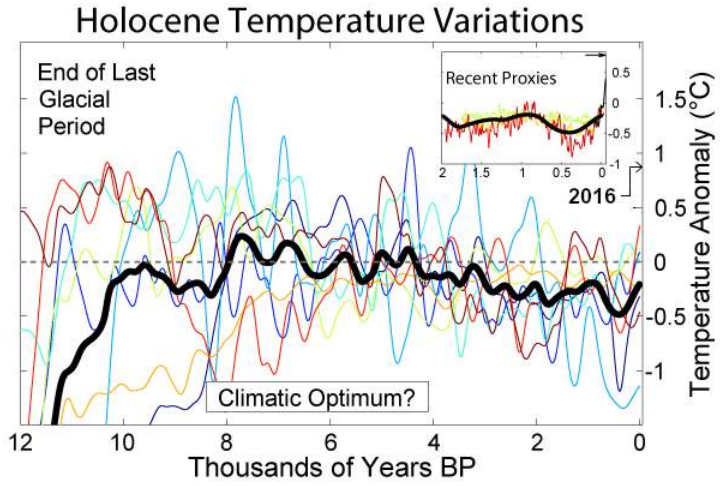
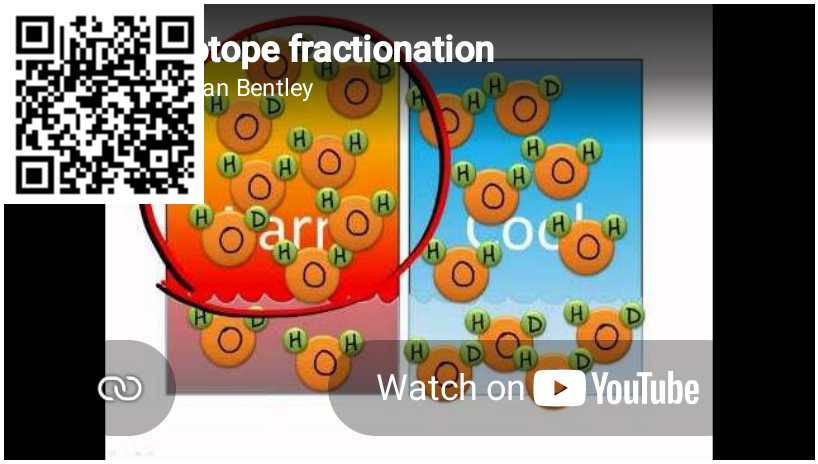


Figure 39.1.12: Holocene temperature reconstructions taken from 8 published sources (Source: Wikimedia Creative Commons).

However, the Holocene is not punctuated by long-lasting changes in climate. A three-part division was proposed based upon ice core and cave speleothem data. The base of the Holocene Epoch, the Greenlandian Age, and the subsequent Northgrippian Age, have GSSPs marked in ice cores, NGRIP2 and NGRIP 1, respectively. Our current age, the Meghalayan Age, is recorded in a speleothem record from a cave in Meghalaya, India. Within the boundaries of the Holocene Epoch, two major climatic events of note can be picked out during the 11,700 year long Holocene, one at 8.2ka (ka = thousands of years ago, or *kiloannum*. "ka" and "ky" are historically interchangeable) and the other at 4.2ka.

The 8.2 ka event – The Greenlandian Age ends and the Northgrippian Age begins

The base of the Holocene begins at 11.7ka with the Greenlandian Stage GSSP recorded in the NGRIP2 ice core. But what happened? Why is it noteworthy? It is located at a depth of 1492.45m where a shift to heavier $\delta^{18}O$ values is recorded. In stable isotope science, we can use the environmental fractionation, or separation, of isotopes of some elements to tell us something about temperature, moisture, or other variables. Because ^{16}O and ^{18}O move around in the environment in different and predictable ways, we can use this particularly isotopic pair as a paleothermometer. to understand the Greenlandian/Northgrippian transition, oxygen isotope thermometry is a key. Below is an excellent discussion of the basic science of these methods.



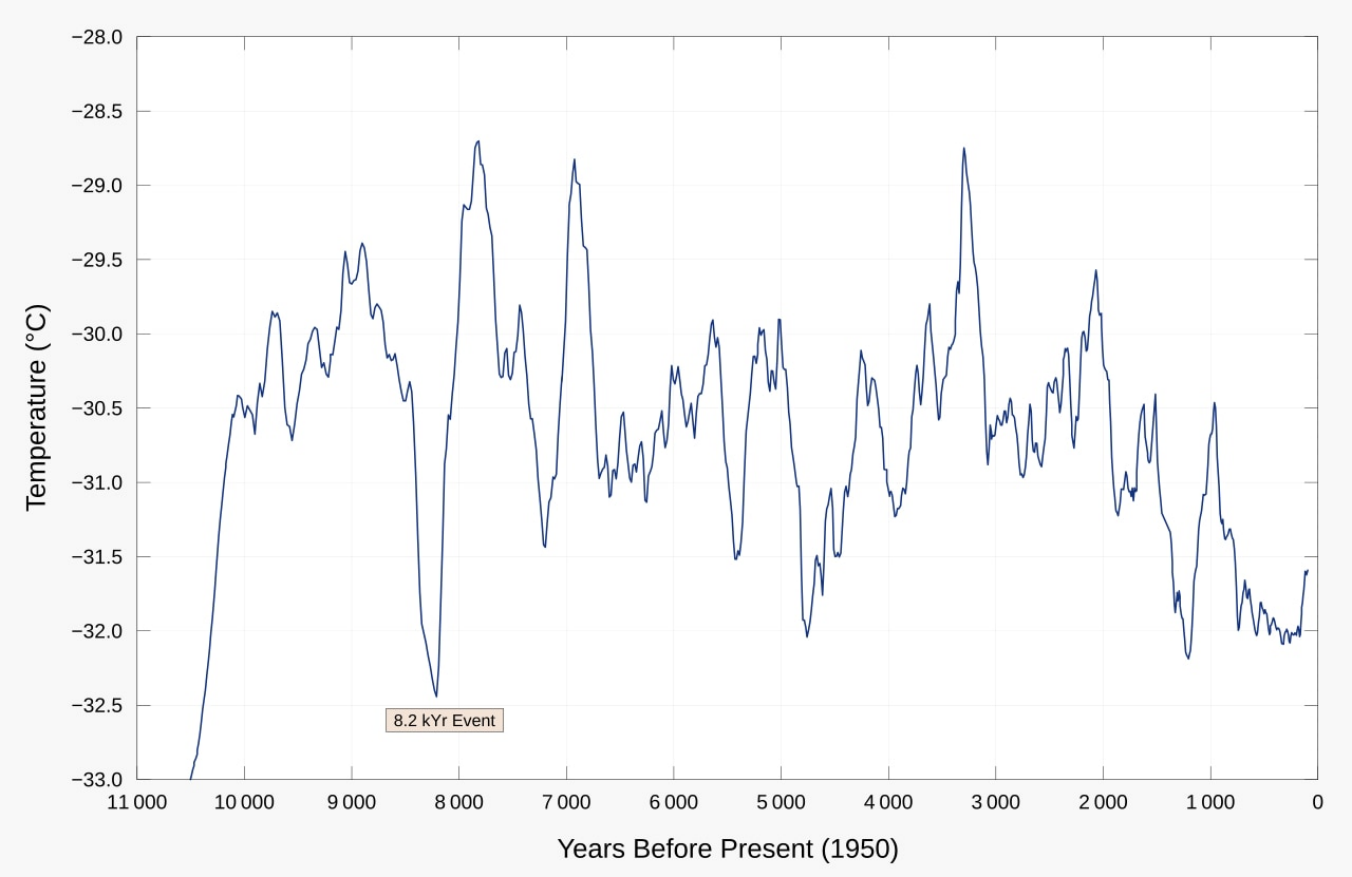


Figure 39.1.13: Central Greenland reconstructed temperature from GISP2 data. Note the temperature departure at 8.2ky (Source: CC BY-SA 3.0; By Giorgiogp2 – Own work, https://commons.wikimedia.org/w/inde...curid=13309952).

In the ice, according to Walker, *et al.* (2018), the change is quite rapid, taking place in just a few years. Accompanying the changes in $\delta^{18}O$ were similarly abrupt declines in deuterium values, a reduction in dust concentrations, and a reduction in Na (sea salt) values. This evidence denotes a sudden cooling, very likely on a global scale, as recorded in central Greenland ice core reconstructions.

At 8.2ky, the Greenlandian Age ends, giving way to the Northgrippian Age. This GSSP is defined by a boundary in the NGRIP1 Greenland ice core at a depth of 1228.67m. Walker, *et al.* (2018) describe this interval as representing a clear signal of climate cooling during an otherwise warming interval of the Holocene. Recorded in a wide range of proxy records around the globe, it is significant because it is relatively short-lived.

The cause of this event is not entirely certain, though it is typically attributed to the release of fresh, cold glacial meltwater from Lake Agassiz and Ojibway into the north Atlantic. These proglacial lakes were quite large, located between today's Lake Superior and the Hudson Bay in southern Canada. They are survived today by a wide variety of landscapes, including Lake Winnipeg, Lake Manitoba, Lake of the Woods, and other bodies of freshwater from northern Minnesota well into the Canadian provinces of Saskatchewan, Manitoba, and Ontario.

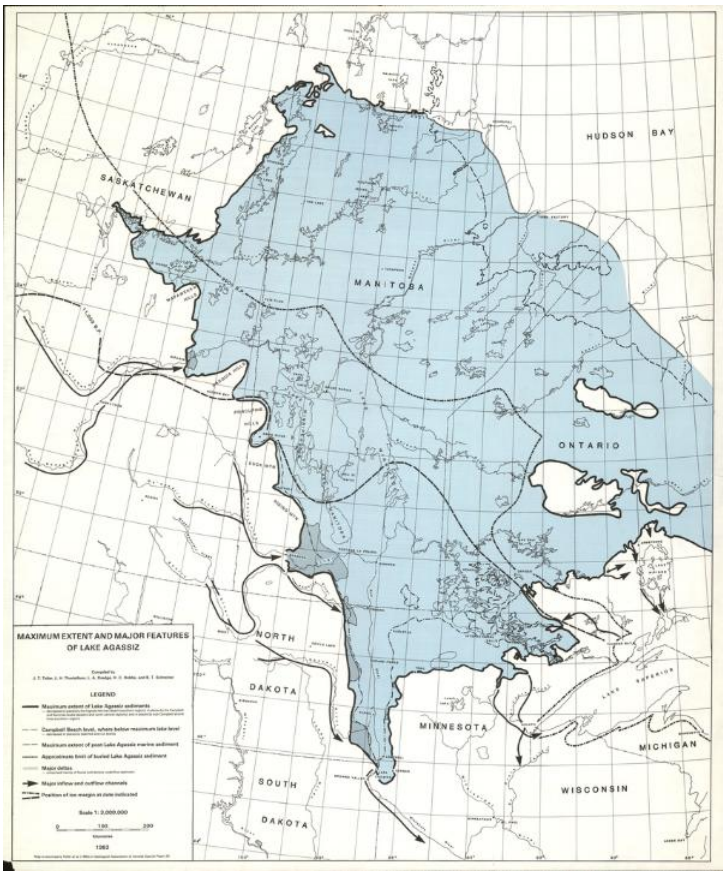


Figure 39.1.14: Maximum extent of proglacial Lake Agassiz (Source: University of Manitoba Libraries Map Collection).

Water likely spilled through the Great Lakes and St. Lawrence Seaway in a great rush, released by a glacial dam. This type of event is called an outburst flood. The sudden infusion of so much cold freshwater into the north Atlantic would have upset the thermohaline circulation in the ocean. This circulation moderates northern temperatures and the shutting off of the current would have caused a significant cooling event, particularly in the northern hemisphere, but with global importance. This cooling could have been as rapid as 3.3°C in just 20 years. Spotty records in tropical regions such as Indonesia also bear records that show a similar cooling, indicating that it was not just an Arctic event. Globally, atmospheric carbon dioxide dropped by 25 ppm over the next 300 years.

The 4.2 ka Event

The Northgrippian Age would come to an abrupt end with a another global cooling event, this time accompanied by significant aridification. In 2018, the International Commission on Stratigraphy officially identified the GSSP for the event in a speleothem formation from Mawmluah Cave in Meghalaya, India. As a result, the new age is called the Meghalayan Age. It is the current geologic age in which we live. There is also an auxiliary GSSP contained in an ice core from Mt. Logan, Canada.

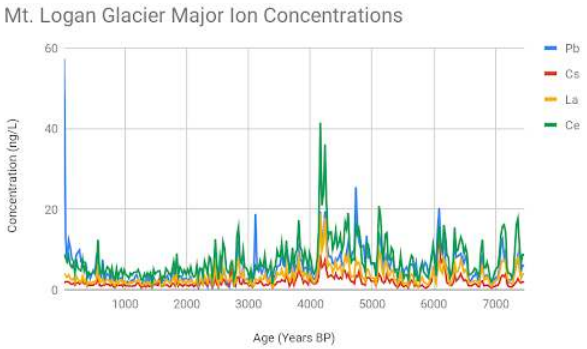


Figure 39.1.15: Major ion concentrations from Mt. Logan glacier, the auxiliary GSSP for the Meghalayan Age, Alaska and the Yukon Territory The major ion spikes around 4200ky are consistent with increased airborne particulates due to widespread aridification around the globe.

At the outset of the Meghalayan, many places around the world became much, much drier. This includes large regions of every continent. Changes in east Indian and Chinese monsoons were triggered, moving patterns further to the south than they had before. With changes in weather patterns that would last well over a century, human civilizations at the time were impacted. The degree of this effect is debated among archaeologists and climatologists, but the 4.2ky event coincides with the end of the Old Kingdom in Egypt, the Akkadian Empire in Mesopotamia, the Lianzhu Culture in China, and the Indus River Civilization in India.

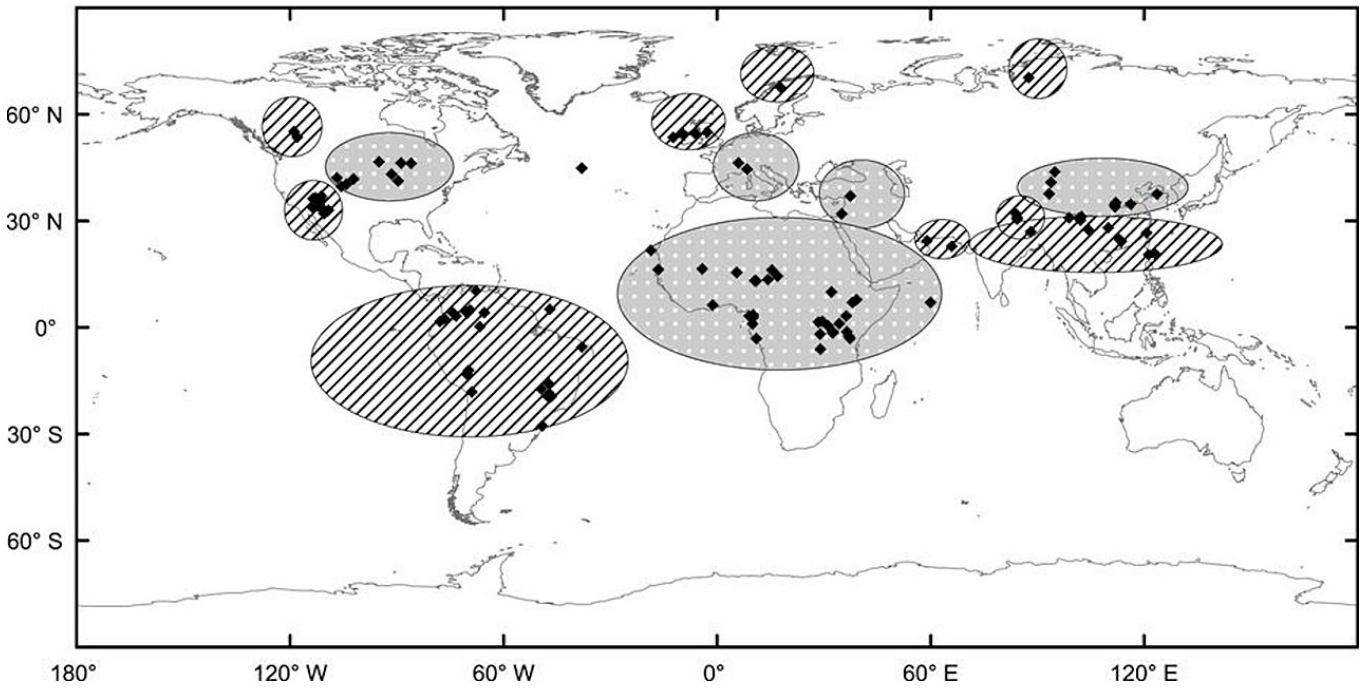


Figure 39.1.16: Global distribution of the 4.2ky event, as shown by hatched (wet conditions) and dotted (dry) areas. Source: By Jianjun Wang, Liguang Sun, Liqi Chen, Libin Xu, Yuhong Wang & Xinming Wang – The abrupt climate change near 4,400 yr BP on the cultural transition in Yuchisi, China and its global linkage. (CC BY-SA 4.0.)



Figure 39.1.17: Speleothem from Mawmluah Cave, Megahalaya, northeast India. The position of the 4.2ka event is shown. This specimen is 308mm long. (Source: Walker et al., 2018).

The GSSP itself is marked in the speleothem by a marked reduction in rainfall at 4200 years BP. Evidence from other regions suggests major oceanic and atmospheric changes, particularly in circulation patterns, sometimes referred to as the "Holocene Turnover." The event appears to have been global. Evidence for this period of aridity is recorded in many locations around the world, including north Africa, the Middle East, and around midcontinental North America. Glaciers record the event in locations like Kilimanjaro, the Andes, and flowstone in Italy. In the northern hemisphere, glaciers re-advanced due to cooling. In mid- to low latitudes, it was marked by aridification in many areas along with disruption of the westerlies, the Indian Summer Monsoon, and the East Asian Summer Monsoon. The resulting megadrought seems to have lasted about 250 years.

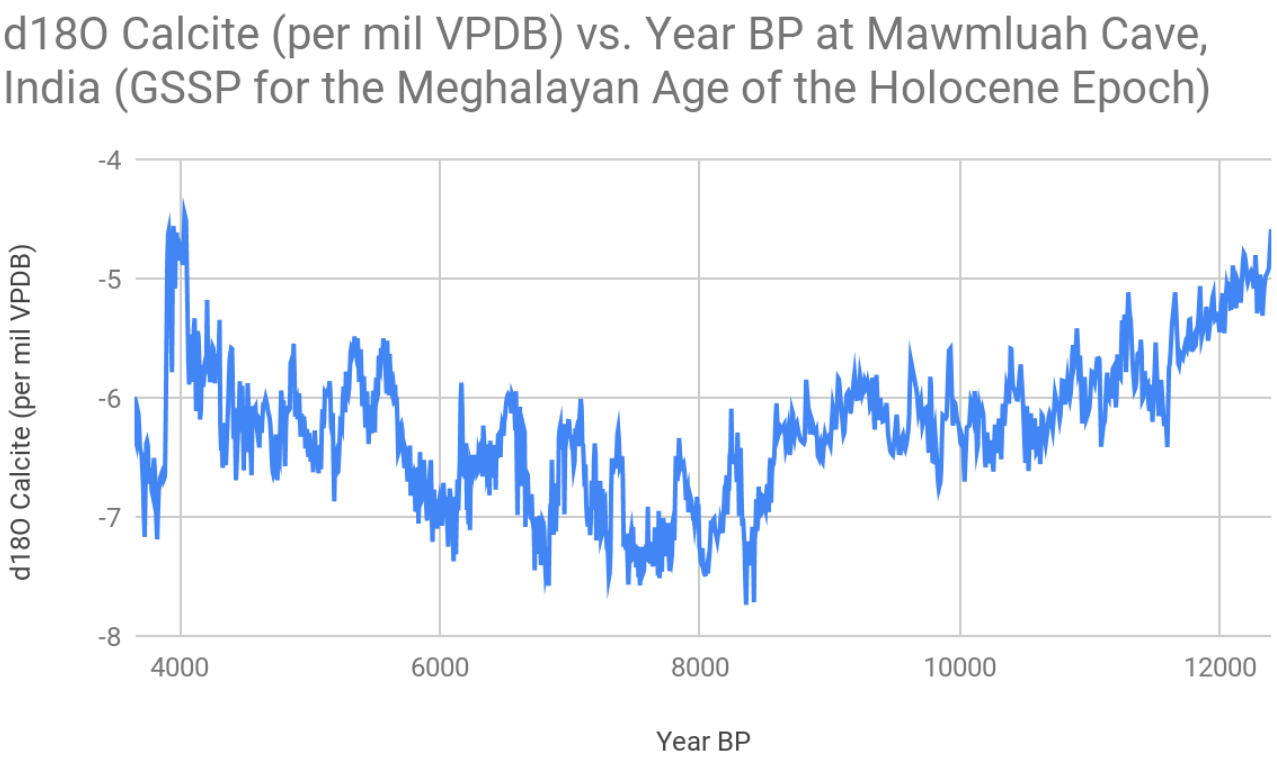


Figure 39.1.18: Oxygen isotope record from the speleothem in Mawmluah Cave, Meghalaya, India. The spike that occurs around 4200ky is the GSSP for the Meghalayan Age and marks a severe drying event.

This ends the major official divisions of the Holocene. However, the 8.2ky and 4.2ky events are not the only climate "blips" during the epoch. Two other minor ones occur much later on. Ironically, these have received the most attention over the past few decades, so it is possible you have heard of them.

Did I Get It? - Quiz

? Exercise 39.1.1

Geologic Ages are the shortest lithostratigraphic unit and are equivalent to Chronostratigraphic Stages. The Holocene Epoch now has three Stages. Which of these is not one of them?

a. Greenlandian

b. Northgrippian

c. Meghalayan

d. Calabrian

Answer

d. Calabrian

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Figure 39.1.19: Hvalsey Church, Greenland, part of a Viking settlement constructed during the Medieval Warm Period (Source: Wikimedia Commons).

The Medieval Warm Period lasted between the years 900 and 1300 CE. It is also sometimes referred to as the Medieval Climate Optimum or Anomaly. For the north Atlantic region, the climate was marginally warmer than the period following up to the mid-20th century. It can be seen in the GISP-2 ice core from Greenland. Possible causes of this anomaly include changes increased solar activity, decreased volcanic activity, or even changes in ocean circulation. There is not strong evidence that it was a global event.



Figure 39.1.20: A reconstruction of what the Viking settlement at L'Anse Aux Meadows, Newfoundland, Canada, may have been like. (CC BY 2.0; Photo by Dylan Kereluk.)

In western culture, the focus on this period is largely due not only to the cultural effects it had on mainland Europe, but as a cause for trans-Atlantic exploration, particularly by the Norse. During this time, groups of Norse colonized Iceland, then Greenland, and eventually reached North America (Newfoundland's L'anse Aux Meadows) around 1000AD under Leif Erickson. Beginning around 985 CE, Greenland became an important settlement, getting large and stable enough for the Catholic Church to send a Bishop. Walrus tusk, collected for its ivory, was a major export.

Even during this time, however, the Greenland climate was quite harsh, requiring that cattle be contained within stone and earthen barns with walls 10 feet thick for nearly eight months of the year.

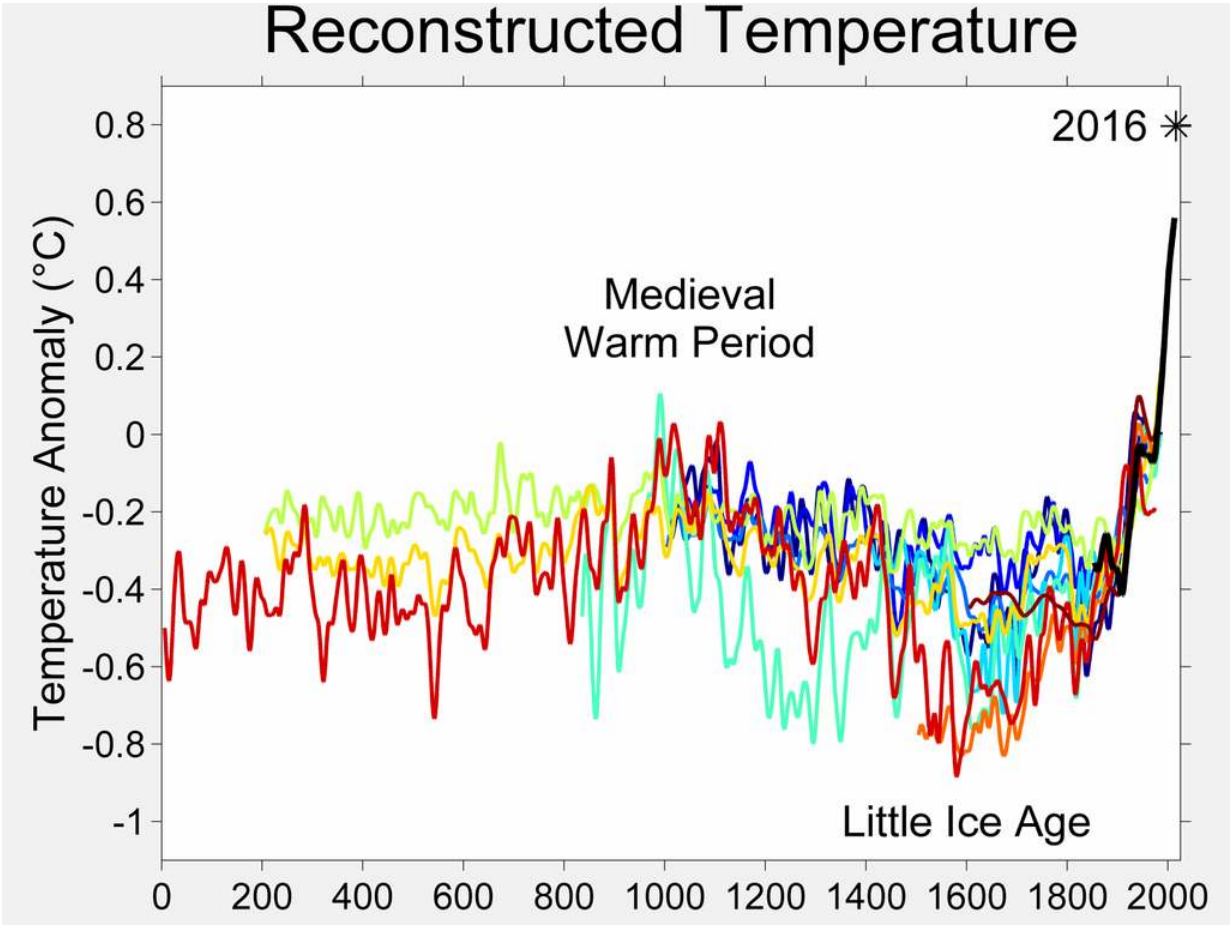


Figure 39.1.21: Data showing 11 different published reconstructions of temperature during the last 2000 years globally. Redder colors are more recent reconstructions, bluer are older. Black shows instrumental record data. (Source: CCAS 3.0 Unported; Global Warming Art)

The Little Ice Age

Eventually, it became much harder to make a living in Greenland. This was not just due to economics, but also to the onset of the Little Ice Age. As shown in the temperature reconstructions above, this period of cooling occurred directly after the Medieval Warm Period. Again, it was not necessarily global and may have been a regional event in the north Atlantic. It lasted from about 1300 to 1850 CE, the traditional start of the industrial revolution in Britain.

Possible causes have been proposed include the usual suspects: changes in solar radiation, increased volcanic activity, and changes in ocean circulation, among others. Even decreases in human population has been proposed as a cause, due to the Black Death plague that killed millions of Europeans during the 14th century. In any case, life in Europe during this time was a mix of hardship, cultural Renaissance, and exploration. The latter was largely motivated as a search for resources to supply a now very resource-poor continent. The cultural and climatic ramifications of this exploration echo into the present day.

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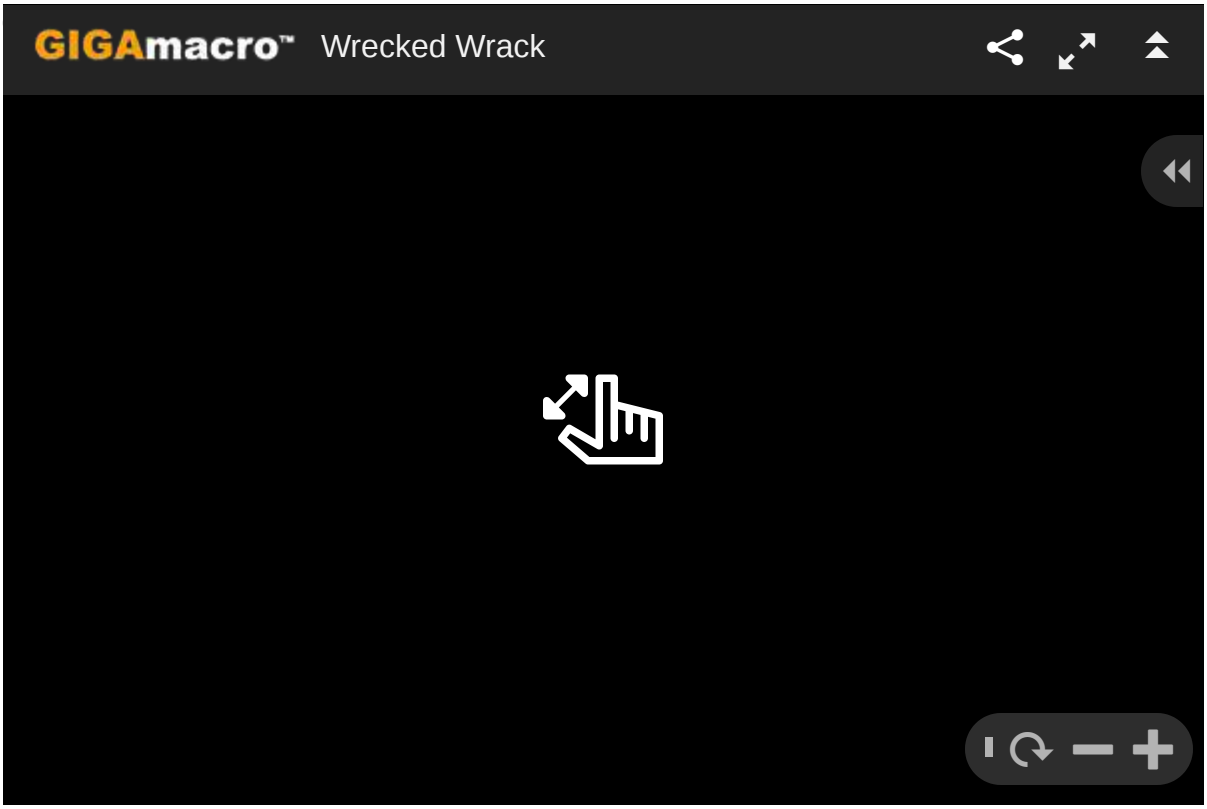
39.2: The Anthropocene

Officially, this should be the end of the story. Up to this point, we have explored how the the Quaternary Period, its epochs and ages, have been subdivided using GSSP markers from the past that are clear and can be agreed upon by specialists (more or less). The distinction between GSSPs housed in ice marking boundaries in the Quaternary versus more traditional stratigraphic markers in rock that mark the rest of the time scale does not require a great leap of understanding. But is this the end of the story of our modern geologic time?

Humans are a force of nature. As a single species and in a very short period of geologic time, we have altered the climate, fragmented the rest of the biosphere, and are likely the cause of a sixth mass extinction. Here is a short list of anthropogenic (human-caused) effects on our global environment that, since at least the mid-20th century, are amplified with enough intensity that some have been led to argue that we are in a new geologic moment:

- Humans have been and are altering the composition of our atmosphere. This has been going on since at least 1750 with the onset of the Industrial Revolution in Great Britain, but perhaps as far back as the Neolithic Revolution around 10ka. This includes greenhouse gas emissions such as carbon dioxide (CO_2) and methane (CH_4), but also chemical and particulate emissions from other industrial, commercial, and household sources. These include black carbon, chlorofluorocarbons (CFCs), HCFCs, *etc*.;
- Humans are changing the chemical physical composition of the hydrosphere. Again, this has been occurring in increasing intensity since 1750. It includes direct pollution of marine settings chemically and through sewage and litter, as well as indirectly through seawater absorption of excess CO_2 from the atmosphere (ocean acidification). Surface freshwater and groundwater resources have also been impacted, sometimes catastrophically;
- Humans are changing the layout and composition of the terrestrial and marine biosphere. This alteration includes overhunting of select large game over time, fragmentation of forest and habitat more generally, alteration of landscapes from wild to cultivated for agricultural purposes, and selective elimination of “pests.” It also includes the rapid spread of non-native species to areas due to human commerce. This has led many to hypothesize that we have entered a sixth period of mass extinction, akin to the “big five” (discussed in other chapters);
- Humans are altering the surficial makeup of the geosphere. We are altering coastlines, slopes, and other geomorphology. Our agricultural practices contribute to soil loss, desertification, and the drying up of surface water bodies;
- Humans are altering the exosphere through the addition of space debris;
- Secondary, often amplifying feedback effects, result from all of these. That is, a direct alteration of one environmental variable often leads to a indirect change in another that is not being directly affected by humans. Many times, such effects amplify the negative impacts of the initial human activity through system feedbacks. An example is the human addition of CO_2 to the atmosphere that leads to warming, which leads to the melting of ice, which leads to lower polar albedo, which leads to more absorption of solar radiation, which leads again to greater warming.

In the history of our planet, it is possible that no single type of organism has ever had this magnitude of an impact (One possible contender is the oxygenation of the atmosphere by cyanobacteria during the Proterozoic). It is definitely the case that the impact of a single species has ever led to change occurring at the ***rates*** the planet is now experiencing. For a small taste of this, check out the “Wrecked Wrack” image below. This represents the trash picked up along a 0.8 km (0.5 mile) stretch of the beach on South Padre Island, Texas. These materials could be described variously as anthropogenically-unique materials. You might one day call them technofossils, as they are some of the objects likely to end up as a part of our modern “fossil” record.



These realities are alarming, but also scientifically verifiable. The material pollution documented above is being deposited in sedimentary basins today. It will likely enter the geologic record as a distinct indicator of humanity's time on the planet.

Because of this, in 2000, atmospheric chemist Paul Crutzen and limnologist Eugene F. Stoermer proposed the creation of a new period of time, the Anthropocene, to mark the age of humans and their effect on the planet. This idea was not entirely new. Soviet scientist Vladimir Vernadsky, in 1938, wrote of human scientific thought itself as a geologic force, using the term Anthropocene for our modern time. The term had also been used by later Soviet scientists as a way to refer to the Quaternary as a whole. Stoermer had been using the term himself since the 1980s, as a way to refer to human impacts on the Earth. The modern use of the term as we will read about it below goes back to a story Paul Crutzen tells of being at a conference where at one point someone mentioned something about the Holocene. Feeling that using this term to describe our current time was wrong, he suddenly stated, "No, we are in the Anthropocene." By 2008, geologist Jan Zalasiewicz wrote in *GSA Today* that defining an Anthropocene Epoch is now appropriate. Since then, the ICS has created an "Anthropocene Working Group" which, in 2016, voted to proceed toward identifying a GSSP to define the beginning of an Anthropocene Epoch.

This would mean an end to the Holocene Epoch and Meghalayan Age sometime in our recent past, perhaps within the last 100 years. Indeed, while various start dates have been proposed, some of which will be discussed below, the greatest consensus is forming around a start date of 1950, a date marking a peak in radionuclide fallout from global nuclear testing as well as the start of a time period referred to as "The Great Acceleration." Since 1950, this acceleration is seen in data wide-ranging enough from the Earth system to cultural and economic trends. Overall, human impacts have led to exponential increases in these trends due to the force of human population growth and increasing development through the use of fossil fuels.

Did I Get It? - Quiz

? Exercise 39.2.1

The human effects on the global environment are well documented. Still, to define a new Age or Epoch, it is necessary to meet certain standards. Which of the following is not a standard for the establishment of a new geologic Epoch?

a. Public opinion

b. A global event that leaves behind a geologic record of itself.

c. A clear stratigraphic marker.

d. A GSSP that can be easily accessed by researchers.

Answer

a. Public opinion

Challenges and Implications

After reading this case study, you may already have identified what may be the greatest philosophical hurdle in defining a new epoch based upon human activity. That is: if our model of geologic time is based upon defining GSSPs using traditional stratigraphic markers (including in ice), how can we define an entirely new geologic age of any kind using markers that may not yet be a part of the geological record? This may seem to go against our model of geologic time and the rules we have used to create it. Does this new designation follow the model and its rules closely enough to be accepted as valid by the scientific community?

A second issue is one of vanity. While certainly the vast majority of the scientific community understands the magnitude of the human impact on the environment, is the definition of a new epoch truly useful in the traditional geologic sense? Or, is it more an exercise in defining our age as a way of encouraging political or social movement toward solving major environmental challenges? How will geologists make use of this epoch – an epoch that may have so far only lasted about 70 years – a blip barely registering in the whole of geologic time? Even if the standard of comparison is the whole Quaternary Period, 70 years is not much.

These are tough academic — even epistemological — questions. The issue is being debated right now among geoscientists. Here, we can only ask the questions rhetorically, but they will beg an answer at some point in the future.

FINDING A START TIME

In a practical sense, the Anthropocene Epoch would need to be defined by a GSSP just like any other boundary on the geologic time scale. This means, succinctly, a geologically-distinct stratigraphic marker that is temporally distinct, clear in the geologic record, easily studied, accessible, and eventually can be ratified by the ICS through agreed upon rules and policies. This has led to lively debate within the ICS's Subcommission on Quaternary Geology, its Anthropocene Working Group, and among other scholars whose fields of work directly inform the debate. Due to the unique nature of the proposed Anthropocene Epoch and the focus on human impacts, debates over this change have extended beyond just geoscientists. Archaeologists, anthropologists, and specialists in other fields have tossed their ideas and contentions into the fracas. While this has certainly made the identification of a GSSP for an Anthropocene Epoch a whole lot more exciting than the usual act of defining a GSSP, it has also muddied the waters.

The big question is, when do you mark the start of anthropogenic effects? When so much of the Holocene is marked by the story of our species, how do you choose where to draw the line?

The Industrial Revolution: 1750

Because of the clear effects of fossil fuel combustion and their clear, well-documented effect on our changing climate, the start of their widespread usage by humans presents an obvious "go to" time candidate. Data, particularly regarding greenhouse gas CO_2 emissions since then would seem to support this start date as a clear favorite.

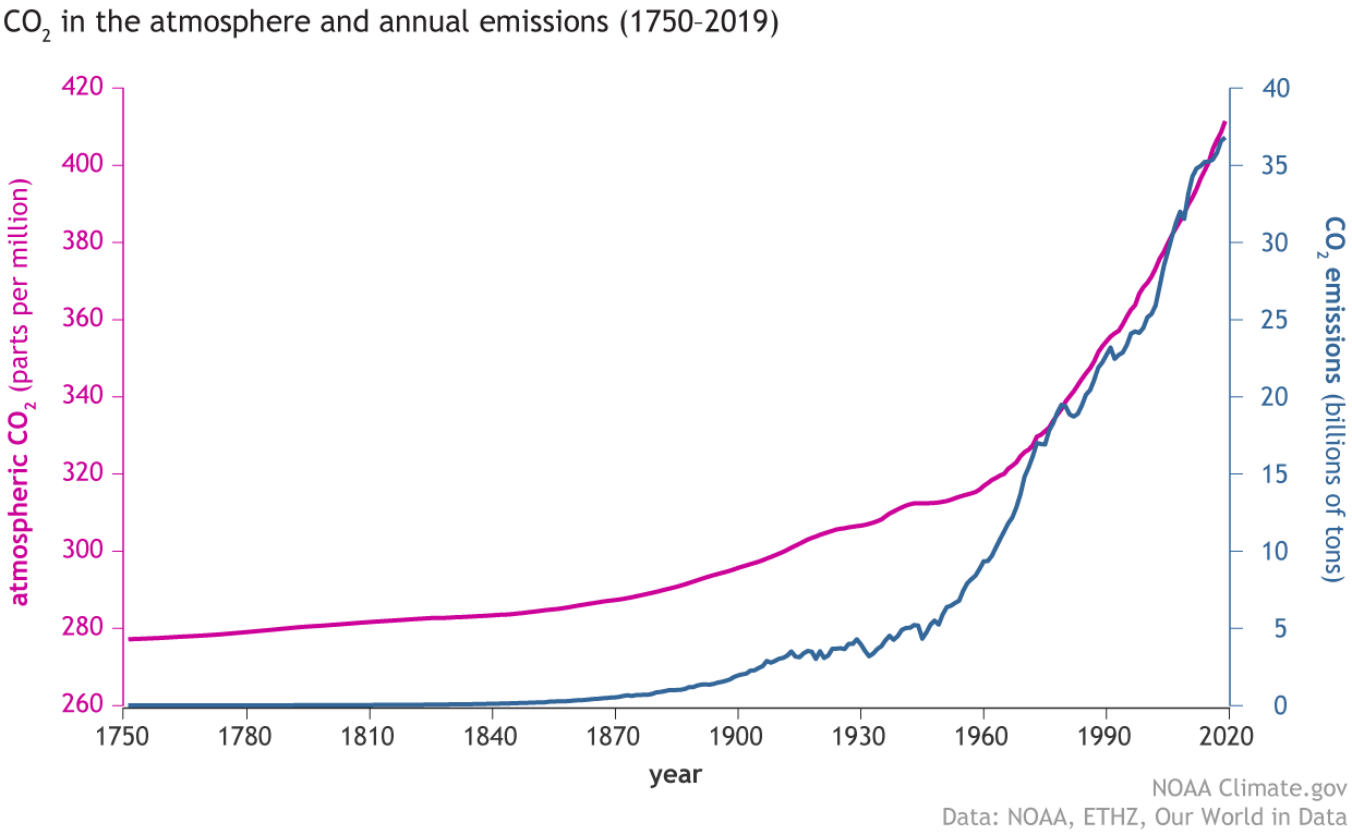


Figure 39.2.1: Carbon dioxide (pink line) has increased along with human emissions (blue line) since 1750. (Source: Climate.gov, adapted from original by Dr. Howard Diamond (NOAA ARL). Atmospheric CO_2 data from NOAA and ETHZ. CO_2 emissions data from Our World in Data and the Global Carbon Project.)

Instrumental carbon dioxide records from locations like NOAA's observatory atop Mauna Loa, Hawaii, have only been collected for a few decades. The graph above is reconstructed from other proxy data, as is the data for human emissions. Still, this would seem to be pretty convincing, particularly if we extend carbon dioxide emissions a bit further back in time.

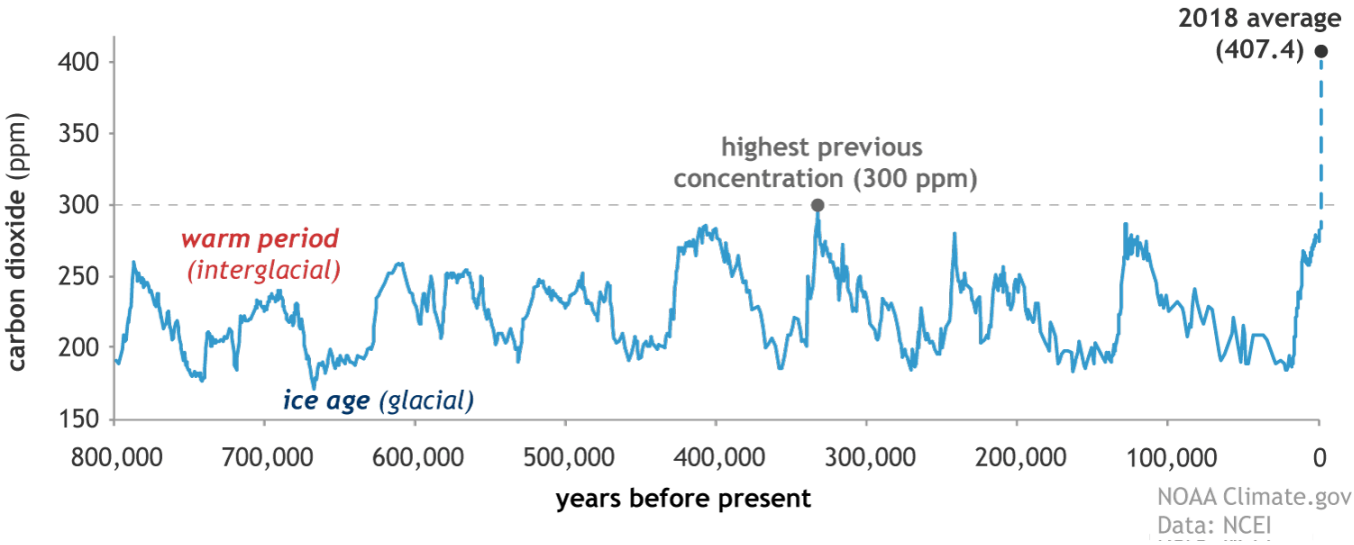


Figure 39.2.2: Global atmospheric carbon dioxide reconstructions (ppm) for the last 800,000 years to the present from ice core records. Quaternary glacial (valleys) and interglacial (peaks) periods are well recorded here. Throughout the duration of the graph, the highest recorded concentration is 300ppm, at a moment in time over 300,000 years ago, well before there are records of our hominid ancestors' ability to utilize fire as a tool. (Source: Climate.gov, based on EPICA Dome C data (Lüthi, D., et al., 2008) provided by NOAA NCEI Paleoclimatology Program.)

When examining the chart above, the case for 1750 as a start date for an Anthropocene Epoch seems quite clear.

However, there are a few problems. One of these is that the main data of record is atmospheric carbon dioxide, and that alone. While this is not in itself unusual, identifying a GSSP using the usual rules is challenging. For one thing, it is not necessarily recorded in ice at this point, or reliably, particularly due to the practical issue of global warming having melted so much of the recently deposited ice, and likely to melt a lot more of it in the future. But if the GSSP cannot be identified in ice, it needs to be found in some other solid Earth record that is well-preserved and accessible. While it is possible that an anthropogenic peak in black carbon (such as soot from power plants and factories) could be potentially identified, the sediments that contain it are not yet rock.

There are other problems, but just these few illustrate the challenge of arguing for 1750 as a start date. The start date makes sense culturally and in some limited data, but a GSSP is really hard to identify.

Besides – why start at 1750 when it can be argued that humans have been altering the atmosphere for much, much longer?

The Ruddiman Hypothesis: The Neolithic Revolution at ~10,000 BP

In 2005, William Ruddiman, then a Professor of Environmental Sciences at the University of Virginia, published a book (*Plows, Plagues, and Petroleum: How Humans Took Control of the Climate*) that argued for a much earlier start date to the Anthropocene. Ruddiman's idea is known as the "Early Anthropocene" hypothesis, since he argued that human-induced changes in atmospheric greenhouse gases actually began with the advent of agriculture, the Neolithic Revolution.

The idea is very similar to the case for 1750. Human technological innovation has led to major changes in the Earth system. It is just that these effects were from a switch from hunter-gatherer nomadic lifestyles to more sedentary agricultural lifestyles. Agricultural innovations associated with the Neolithic Revolution, such as the domestication of cereal crops beginning around 10,000 BP in the Near East "Fertile Crescent" region containing the Tigris and Euphrates Rivers (modern Iraq, parts of Iran and Turkey, Syria, and Israel/Palestine) and parts of the Mediterranean Sea coast. This particular innovation would develop independently in several other locations, including central China, Egypt, Mesoamerica, central North America along the Mississippi River, and coastal South America. While cereal grain domestication occurred earliest in the Near East, independent development of agriculture in other areas allowed the sharing of technological innovations around the globe.



Prehistoric agriculture, as is still done today, involves the widespread clearing of land and burning of debris. The clearing of land removes vegetation, a natural carbon sink, and its burning then releases carbon back into the system, pumping carbon dioxide into the atmosphere. Some of this carbon had been stored for thousands of years. Such burning for agriculture can still be seen today in places like Indonesia, where massive areas of land are being "slashed and burned" for the planting of palm oil plantations, very often illegally. This burning of debris and accumulated peat is a major source of anthropogenic greenhouse gas emissions today also.

However, widespread burning of debris is not only associated with agriculture. Humans have been using this technique for much longer to burn away underbrush in forests to aid in hunting. When early Europeans arrived in the Americas, for instance, it was observed that it was possible to drive a wagon with four horses side by side with little trouble. Indigenous Americans were able to efficiently clear large swaths of forest in this way. When European disease and warfare drove them out of their lands or killed them

altogether, Europeans simply cut down forests or, in the case of forests left standing, did not have a cultural practice of burning underbrush. Thus, it accumulated. Human alteration of the environment over time has turned biomes into anthromes.

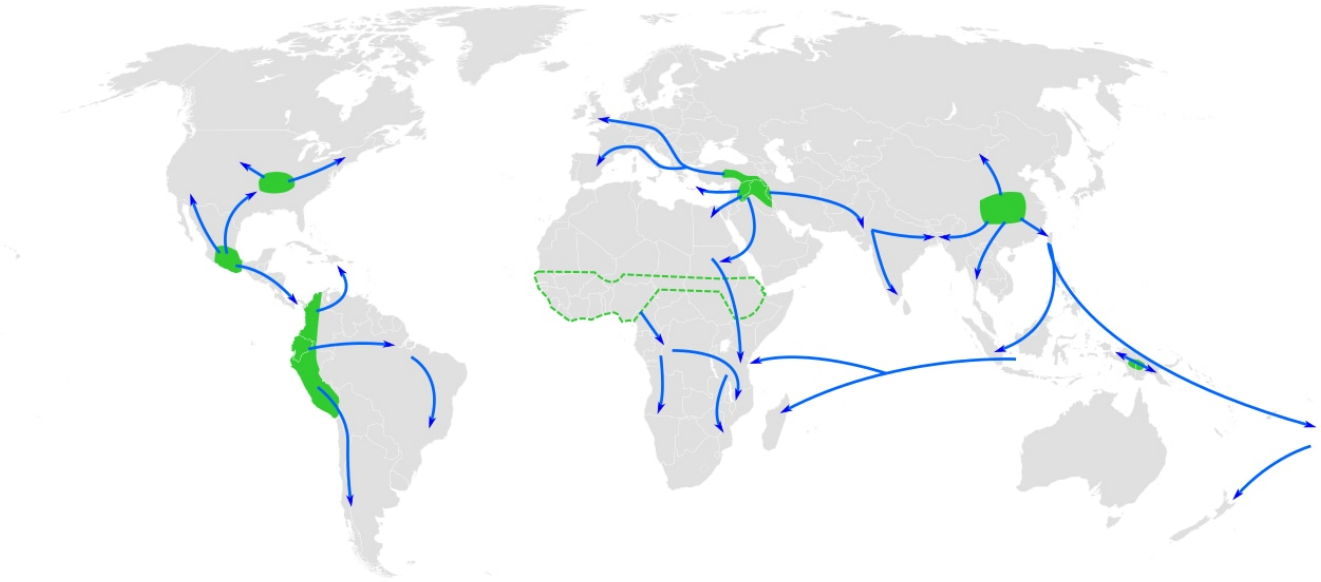


Figure 39.2.3: Global centers of origin for agriculture and its spread in prehistory (Source: CCBY-SA 3.0; Joe Roe)

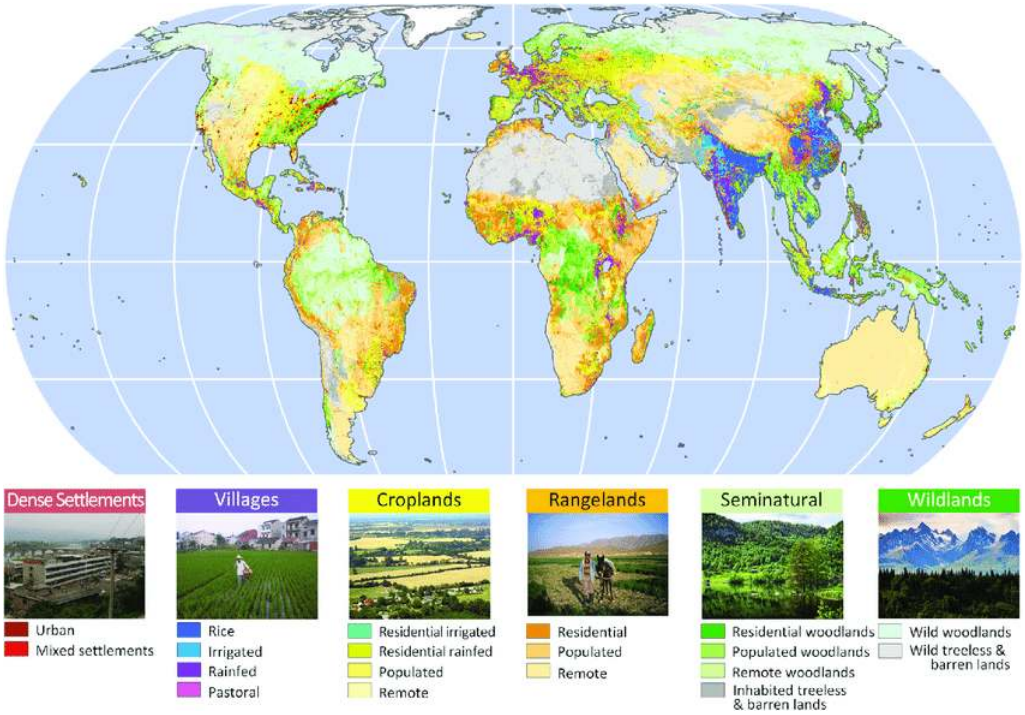


Figure 39.2.4: Anthromes, or Anthropogenic Biomes, dominate the Earth's landscape (Source: Ellis, 2019). *Do we have permission to use this?*

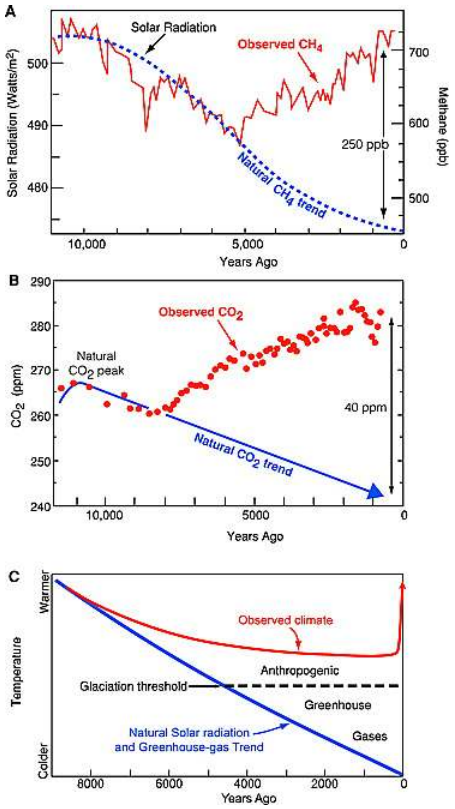


Figure 39.2.5: The “Early Anthropocene Hypothesis”. In this hypothesis, it is argued that human land clearance for plant agriculture, including the burning of debris and forest understory, increased carbon dioxide emissions. Shortly after this cultural development, animal husbandry, particularly of ruminants like sheep and cattle, increased methane emissions. These two effects, limited as they were geographically, may have been enough to forestall the natural progression into another glacial period (Source: Ruddiman, 2003). *Do we have permission to use this?*

Around 8,000 years ago, humans domestication of animals began in earnest. Ruminant animals, in particular, the contributed large amounts of methane to the atmosphere. These animals are still a major source of anthropogenic methane today. Within the rumen of these animals, a group of Archaea known as “methanogens” (Euryarcheota), produce methane from the chewed-up vegetation the animals have eaten. The process is known as enteric fermentation. Contrary to popular imagination, these animals pass this gas into the atmosphere through belching, rather than flatulence. Cumulatively, the amount of methane is very significant.

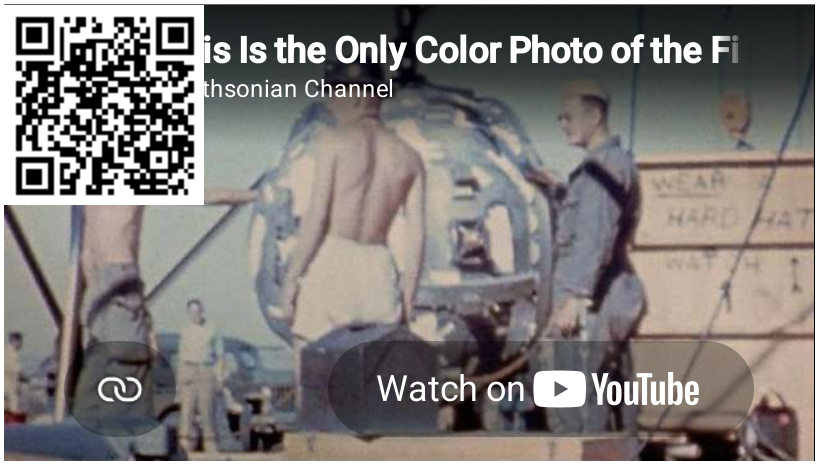
Both of these gases are potent greenhouse gases. The “Early Anthropocene” argument uses climate modeling to make the case that the collective developments associated with human agriculture were enough to forestall entering a glacial period. Further, this would mean that the climate changes we are experiencing today are simply a more rapid manifestation of climate changes resulting from technological innovations our ancestors made thousands of years ago.

How would this work as a GSSP? This would place the start date for the Anthropocene Epoch at 10,000 BP. With the Holocene Epoch GSSP identified at 11,700 BP, this would either reduce the length of the Holocene to a mere 1,700 years, or make a good argument for either renaming it or just re-imagining how we view the Holocene altogether, thus tossing out the Anthropocene Epoch concept entirely. As for a physical stratigraphic boundary in the rock record, archaeological data could be used to mark a cultural progression, but this would be a significant departure from convention. In reality, we are left with the same problem with using 1750 as a start date. It is nearly impossible to identify a physical location where a rock formation accuratey records this transition.

Is there a time when humans really made their mark in a more explosive manner?

The Atomic Age: 1950 and the “Great Acceleration”

21 days before the first atomic bomb was dropped on the Japanese city of Hiroshima, the first test of an atomic bomb took place near Alamogordo, New Mexico. On July 16th, 1945, the Trinity Test resulted in a 22 kiloton explosion. Below is some of the only footage of that test that exists. The explosion itself can be seen near the end of the clip.



The explosion itself had a yield equivalent to 15,000 to 20,000 tons of TNT. In the years following World War II and until 1963, eight countries developed and tested nuclear weapons. Post-1963, further tests were conducted by Pakistan, India, and North Korea, but these are much more limited in scope. Still, the radionuclide fallout from these early tests, beginning with Trinity, are verifiably recorded in the geologic record.

These nuclear tests represent a technological innovation with power beyond anything else humans have harnessed. Because of this, and because of the global distribution of the radioactive fallout, the emerging consensus is that this period of nuclear testing, beginning in 1945 and extending into the early 1950s, is the ideal temporal horizon to use for the start of the Anthropocene Epoch.

Radionuclide accumulation in the geologic record is not the only significant anthropogenic signal to accumulate from this point. In 2015, scientists with the International Geosphere-Biosphere Programme released their latest data set for what they deem "The Great Acceleration" (Steffen *et al.*, 2004). The idea here is that the rapid and massive growth of socio-cultural trends and Earth system trends beginning around 1950. The advent of the nuclear age and a new source of energy, helped usher in a global industrial revolution that continues its rapid expansion to the present. Socio-economic trends such as human population growth, GDP, fertilizer consumption, urbanization, etc. have put major pressure on the environment. Resulting environmental trends include increases in atmospheric carbon dioxide, nitrous oxides, methane, stratospheric ozone, lower ocean pH (acidification), and forest loss. When graphed, data for all of these groups of variables produce exponential curves, largely beginning around 1950. While not all of these trends will make ideal geologic records, they certainly argue that a rapid increase in anthropogenic population, activity, and energy production has made a lasting impact on the planet.

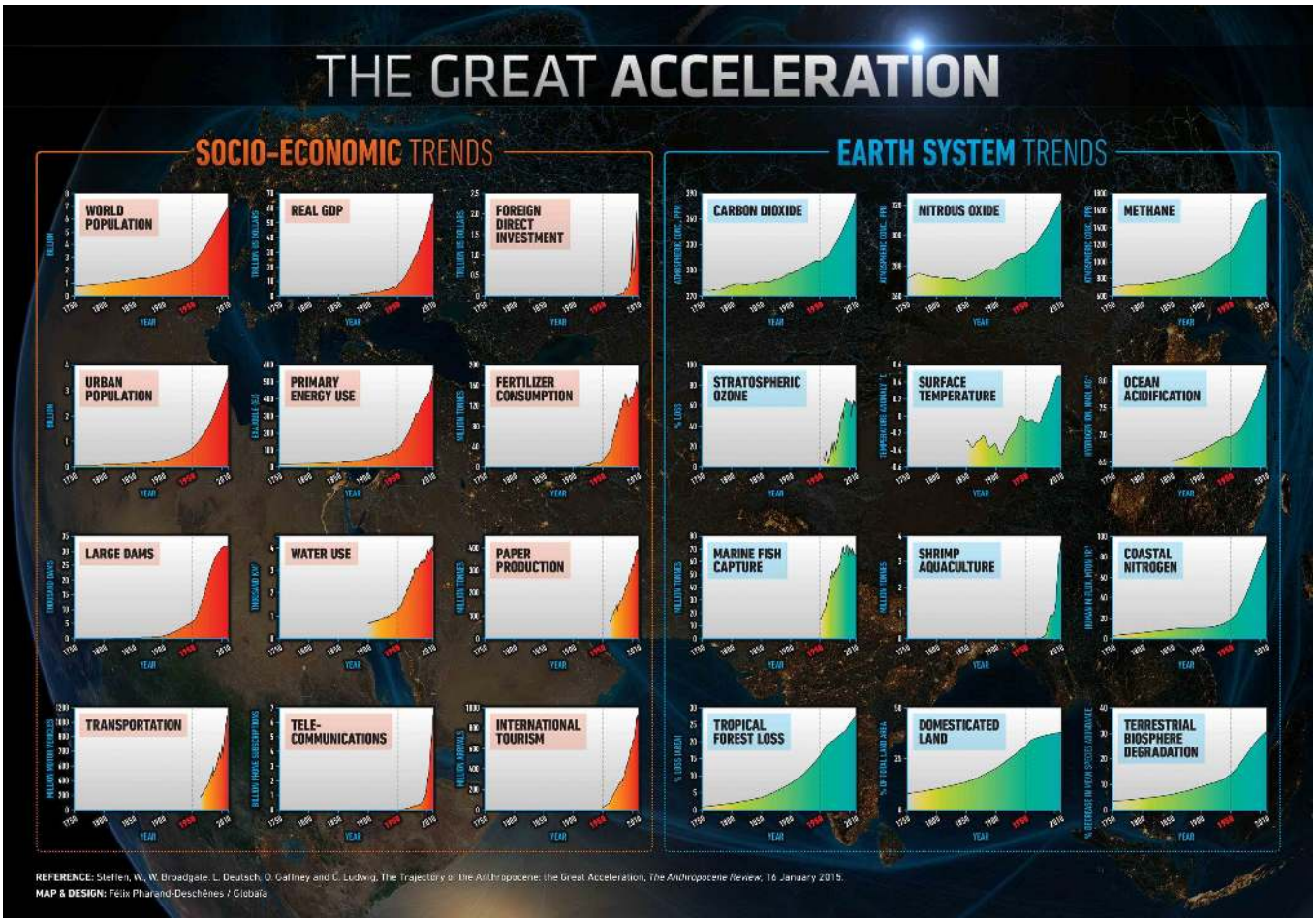


Figure 39.2.6: Socio-economic and environmental trends included in disucssions of “The Great Acceleration” (Source: IGBP, adapted from Steffen et al., 2004).

The case for an Anthropocene Epoch start date around the year 1950 is compelling. There is a clearly different trend of change in the global natural environment with human activities as the sole or primary driver.

As of this writing, the hunt is still on for a GSSP. It is not unusual to identify a moment in time before a GSSP marker is found, as there are still boundaries on the geologic time scale that completely lack a GSSP to physically define them. Nevertheless, there are a handful of locations being actively pursued as having the potential to contain the key marker for an Anthropocene Epoch. One candidate is Crawford Lake, Ontario, Canada, where its depth precludes complete mixing of layers, allowing its bottom sediments to lie undisturbed. Other sites include an Antarctic ice core, cave deposits in Italy, coral reefs in the Caribbean and Australia, and a peat bog in Switzerland. If present, the signal at all of these locations is a mix of chemical markers consistent with radionuclide fallout, such as carbon-14 and plutonium-239, but also including other potential markers like organic pollutants, microplastics, and fly ash.

RATIFYING A GSSP

The currently agreed-upon temporal marker for the end of the Holocene Epoch and the start of the Anthropocene Epoch is the year 1950. As data documenting the so-called “Great Acceleration” discussed above shows, anthropogenic effects on the environment have increased by orders of magnitude over background rates, been very sudden, progressed very rapidly, and have altered the natural environment enough to be recorded in the geologic record, beginning with radionuclide fallout. The Anthropocene Working Group documents >+1 order of magnitude increases in sediment transport due to urbanization and agriculture; marked and abrupt perturbations in biogeochemical cycling of carbon, nitrogen, phosphorus, and various metals; global warming; global (eustatic) sea-level rise; ocean acidification and proliferation of “dead zones;” rapid biosphere change on land and sea; habitat loss, hyper-predation/hunting, increase of domestic animal populations, and widespread expansion of invasive species; and the abundance and

spread of new human-designed substances such as fly ash, plastics, concrete, and technofossils (Anthropogenic Working Group, 2020).

Because these alterations of the Earth system will persist for many millennia (some will be permanent), the Anthropocene Working Group has established that these materials are sufficient to produce durable geologic strata that are sufficiently unique to be separate from Holocene stratigraphy. Because of all of this, the Anthropocene Working Group voted on May 21, 2019 and answered in favor of the following questions at the International Geological Congress in Cape Town, South Africa:

1. Should the Anthropocene be treated as a formal chronostratigraphic unit defined by a GSSP? **29 members voted in favor, 4 voted against.**

2. Should the primary guide for the base of the Anthropocene be one of the stratigraphic signals around the mid-twentieth century of the Common Era (*i.e.*, circa 1950)? **29 voted in favor, 4 voted against.**

These questions now guide the future work of the ICS's Anthropocene Working Group. Current work is focused on identifying a GSSP for the Anthropocene Epoch. Candidate locations will certainly be proposed, vetted, and voted upon. The progression from merely suggesting the idea in 2000 to voting on going forward with a formal proposal has only taken 18 years. It is very possible that, within the coming decade, you may find yourself in a new geologic epoch.

Did I Get It? - Quiz

? Exercise 39.2.1

We are living in the Anthropocene Epoch

a. No. A new unit and its GSSP have yet to be ratified.

b. Yes. The new unit has been ratified. A search for the GSSP is ongoing.

Answer

a. No. A new unit and its GSSP have yet to be ratified.

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39.3: Concluding Thoughts

Before we began this case study, we defined three important questions about how we define geologic time:

1. How do we use the geologic time scale as a model?

2. How do we make changes to this model?

3. How can we use this model to help us define our own time?

You should be able to provide your own answers to these questions. As we have explored some Quaternary quandaries regarding how to apply the model of geologic time, GSSP development, and more, it should be clear that it is not a perfect model. Like all good models we develop to explain what we observe and test in nature, it is flawed at present, but nonetheless quite useful. Keeping this open-minded philosophy about time is important. We can continue to improve the model only if we are aware of its shortcomings or failures. However, the geologic time scale and the techniques we use to create it provide us with powerful resources to help us test important questions about our planet's past and our own humanity.

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39.4: Further Reading

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Steffen W., Sanderson A., Tyson P.D., Jager J., Matson P.A., Moore B. III, Oldfield F., Richardson K., Schellnhuber H.J., Turner B.L. II, and R.J. Wasson (2004). Global Change and the Earth System. IGBP Global Exchange.

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CHAPTER OVERVIEW

40: (Case Study) Human Evolution



Figure 40.1: Clovis points from the Rummells-Maske Cache Site, Iowa, USA (Image: CC BY-SA 3.0; Billwhittaker, Wikipedia).

Student Learning Outcomes

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

- Discuss *Homo sapiens sapiens* in the larger context of hominid evolution;
- Highlight some of the critical splits in the human phylogenetic tree;
- Discuss some of the features of hominid skull anatomy that can be used for comparative genotype analysis between hominid groups;
- Describe how humans migrated from Africa, including the timing and pathways;
- Describe how we can trace these migrations using paternal and maternal haplogroups;
- Identify some of the key technological innovations during the evolution of hominids, particularly among anatomically modern humans and Neanderthals;
- Understand how human evolution is very similar to evolution within other taxonomic families of organisms.

40.1: Humans as a species
40.2: Paleoanthropology
40.3: Highlights of our family tree - Australopithecina and Homo
40.4: Further Reading

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40.1: Humans as a species

The evolution of humans began 13.8 billion years ago with the Big Bang itself, which set in motion the forces and energy that would make our history all possible (though, not inevitable!). Elements created and compounds formed after this event would eventually form our planet, Earth. At some point after 4 billion years ago, living, self-replicating materials formed that would become our first ancestors. In order for this to happen, these early life forms had to use resources from their environment to survive and thrive.

So it is with humans today. All life extracts energy and nutrients from its environment for respiration, digestion, and procreation. Our species, and some of our ancestors, took this extraction to a new level. They developed tools. These tools made resource extraction much more effective, leading to new evolutionary pressures, more adaptations, and eventually, you and me.

This case study examines the development of our species. The story is a long one, so this is merely a summary. Ultimately, it serves to highlight several aspects of humans. Two threads of hominid evolution are emphasized. We explore the evolution of tool use among hominids as a way of exploring the development of human intelligence and our species' ever-increasing ability to extract resources from its environment. We also explore some of the particular anatomical features of hominid skulls as a way of showcasing hominid physiological development.

Here are some of the key threads that are woven together in this chapter:

1. Like other organisms, anatomically modern humans evolved in a manner consistent with evolutionary theory.

2. Hominid evolution can be traced using fossils and comparative anatomy.

3. Like other organisms, humans extract energy and alter their environment to survive.

4. Human ingenuity with tools began deep in hominid history, and is not unique to anatomically modern humans.

Before you dive into the specifically human part of the story, you are encouraged to begin with a larger view of our mammalian past as a review. Watch the video below:



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40.2: Paleoanthropology

If you read about human evolution and the field of paleoanthropology, you will also encounter other primates you may consider old friends (*e.g.*, chimpanzees) and a bunch of unusual primates who are extinct. You will also explore elements of the field of archaeology, the study of material culture. This last piece is pretty unique to research into hominid evolution, as there is no archaeological record for other organisms. The closest analog we get for archaeological materials in the rest of the living world is ichnology, the study of trace fossils. Archaeology is divided into two subfields. Historical archaeology explores human culture after the development of writing. Prehistorical archaeology explores everything else. Archaeology is typically a subfield of anthropology, the study of humans.

Paleoanthropology is a confluence of paleontology and anthropology that seeks to understand our human ancestry and associated culture. The field seeks a holistic exploration. This means that paleoanthropologists are interested in human biology, social structure, cultural development, material culture (archaeology), linguistic development, and more.

There have been many rather famous paleoanthropologists. This is because the field tends to engender a good deal of interest among people outside of the scientific community. Famous names include many members of the Leakey family. Louis and Mary Leakey and their son Richard Leakey who, with his partner Meave Leakey, have contributed much of what we know about our shared African ancestry. Their work in locations like Olduvai Gorge, Tanzania and nearby environs has provided important windows into how humans evolved, migrated, and are related to our primate cousins. Ian Tattersall, a curator at the American Museum of Natural History in New York City, has worked to understand more deeply our connection to current primate cousins but also to standardize terminology in the field. Still others, like Sonia Harmand, have dedicated their research to studying human cultural development, as exemplified in her work with stone tools.

Paleoanthropology is a dynamic and very young field. New insights into our human past are being published almost monthly. To keep up with these changes, refer to two particular locations included in the section on further reading at the end, the Smithsonian Institution's human evolution program and the Institute of Human Origins at Arizona State University.

Hominid phylogeny terminology

As you dive into the study of human origins, you will encounter a bunch of confusing terms. We will begin by outlining some of these here.

Primate – Primates are a large group of social mammals that arose between 85 Ma and 55 Ma, probably from small tree-dwelling mammals. Shared characteristics of primates include relatively large brains, visual acuity, color vision, dexterous hands (prehensile with opposable thumbs for grasping), and large ranges of motion in their shoulder joints. Primates are a part of the order Primata, which contains two suborders, strepsirrhines and haplorhines. Humans are a member of this latter suborder.

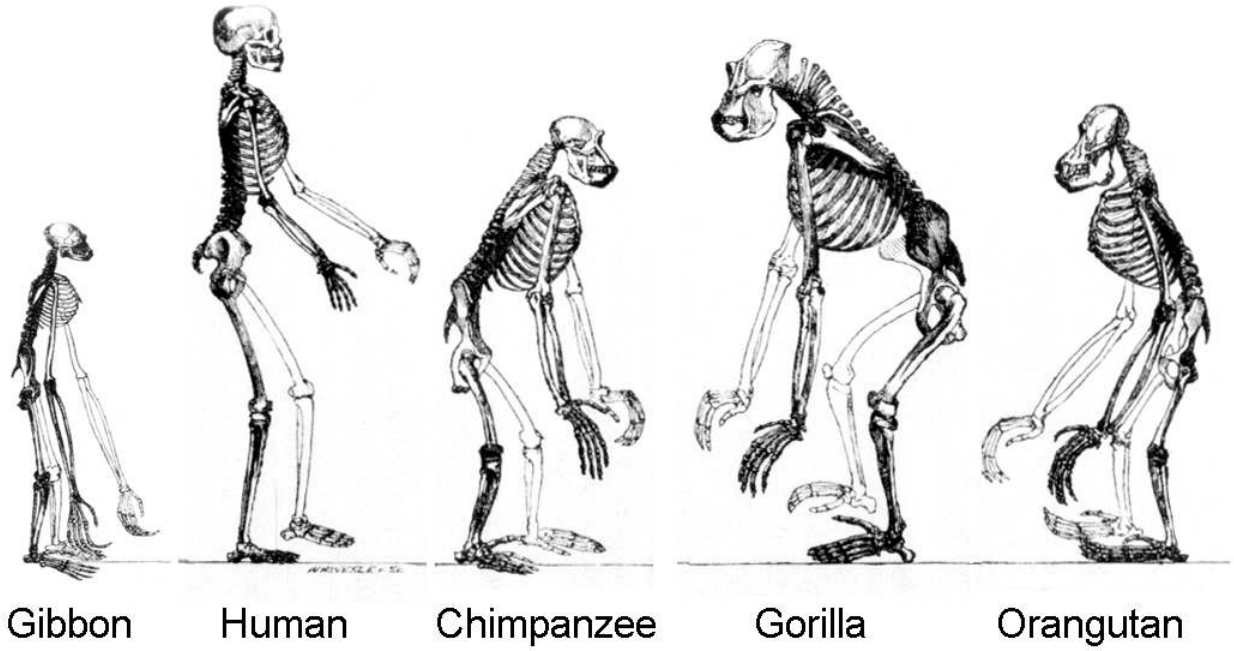


Figure 40.2.1: Skeletons of members of the ape superfamily, Hominoidea. (Image: Public Domain; Wikimedia)

Hominidae – The **hominids** are the human taxonomic family. It is more colloquially referred to as the "great apes". The term "hominid" is sometimes also used generally. Within this family, there are currently eight extant (living) species, contained within four genera. These species include us, *Homo sapiens*, Orangutans *(Pongo abelii, Pongo tapanuliensis, Pongo pygmaeus)*, Gorillas *(Gorilla gorilla, Gorilla beringei)*, and Chimpanzees *(Pan troglodytes, Pan paniscus)*.

Homininae – The **hominins** are our subfamily and refers specifically to "African Apes". This subfamily is divided into two "tribes" (the term is no connection to that which is used for indigenous human groups), Hominini and Gorillini.

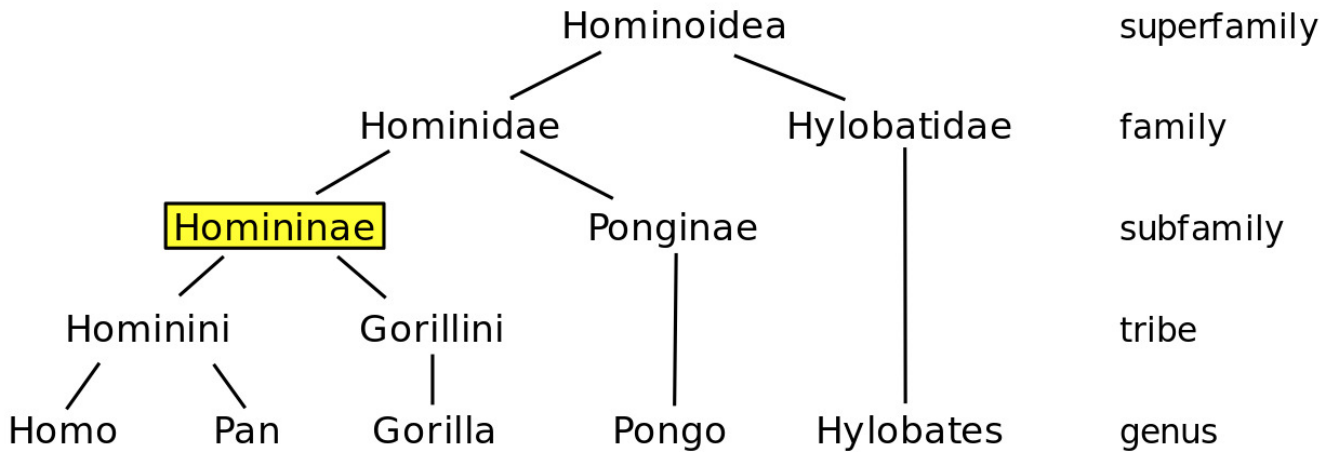


Figure 40.2.2: Evolutionary tree of the family Hominoidea. The subfamily Homininae is highlighted and Gorillini and Hominini are tribes within the subfamily Homininae. Humans are more closely related to members of the genus Pan (Chimpanzees) than to gorillas. (Image: CC BY-SA 3.0; By EoD – Own work, Wikimedia)

Hominini – The hominini is a taxonomic tribe that includes our human ancestors and chimpanzees – two genera only. Sometimes chimps (*Pan* sp.) are separated out for their own tribe. Ultimately, the search for a common human ancestor between our close cousins chimpanzees extends to about 4-8 Ma (Patterson *et al.*, 2006) but the split probably happened around 6 Ma. Once this split occurs, the phylogenetic tree is entirely focused on our ancestral lineage, all of whom are now extinct except for us. These genera include *Homo* and *Australopithecus*.

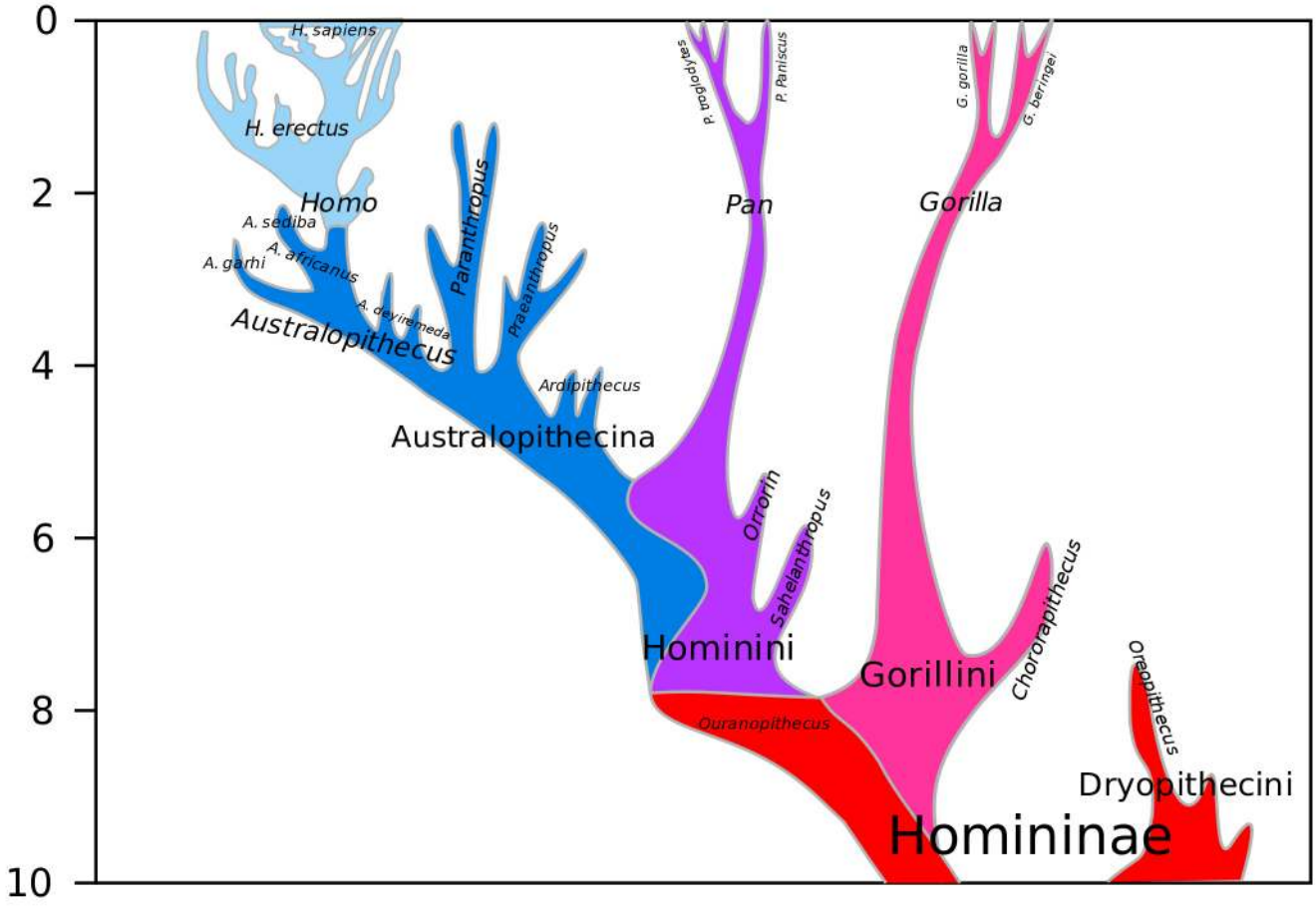


Figure 40.2.3: A Model of the phylogeny of Hominini over the past 10 million years. A pink line for Gorillini splits off around 8 Ma, a purple line for Pan splits off around 6 Ma, and Homo and Australopithecines continue on to today, represented by a lone species, *Homo sapiens*. (Image: CC BY-SA 4.0; Dbachmann – Own work, Wikimedia)

Did I Get It? Who are Hominids?

? Exercise 40.2.1

Which of the following phylogenetic terms refers to a taxonomic tribe of hominids?

a. Hominidae

b. Hominini

c. Homininae

d. Hominid

Answer

b. Hominini

Hominid Skull Terminology

Below there will be a good deal of discussion that uses hominid skull development to discuss hominid evolutionary change. Certainly, there are other anatomical features to use, when there are more complete skeletal remains. Skulls and skull fragments are

probably the most common single fossil, so they are used in this case study. Below is a summary of some of the major parts of hominid skull anatomy.

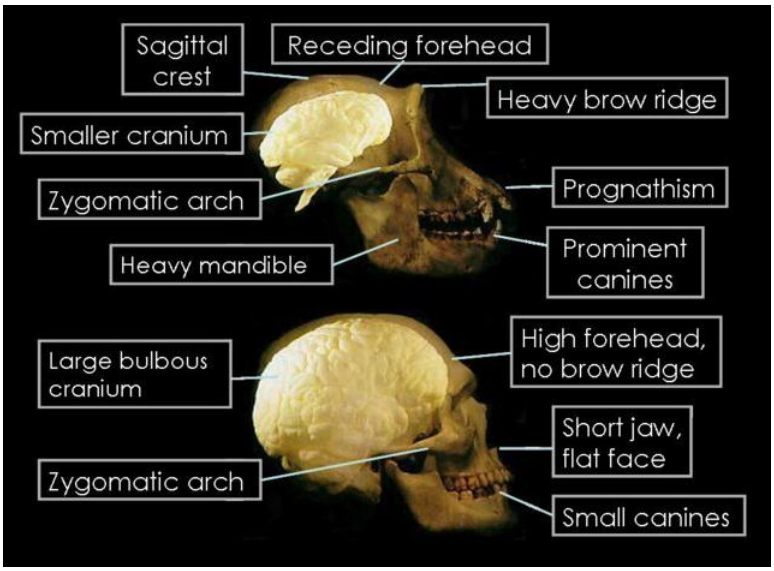


Figure 40.2.4: A summary of hominid skull features, comparing Human and Australopithecine skulls. (Image: Public Domain)

If we look in detail at parts of the skull, some of the major areas become important in the story of human development. The first of these is the mandible, or lower jaw. A key difference between apes and humans in this feature is sheer size. Apes eat a great deal more tough, fibrous material than humans do, the chewing of which requires a more robust lower jaw. Humans also developed an accentuated chin structure to help provide some additional strength to what is a much weaker area of our lower jaw.

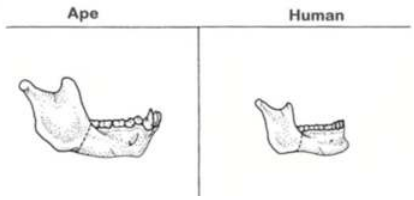


Figure 40.2.5: Mandible, or lower jaw. (Image: Public Domain)

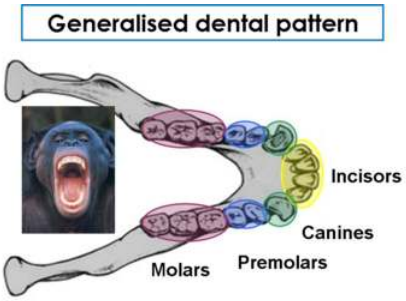


Figure 40.2.6: Dental pattern of primate jaws, including humans. (Image: Public Domain)

The arrangement of teeth in apes and hominids is very similar. The dental arrangement includes molars and premolars for chewing, incisors for cutting, and canine teeth for tearing. Teeth are also one way that apes, as opposed to humans, can display sexual dimorphism (differences between genders). Male apes tend to have much larger canine teeth than female apes. There is no real advantage to the larger canine except for sexual selection, with females giving preference over one male for this particular feature over another. While certainly humans do notice one another's teeth, mates are not selected based upon the size of their canine teeth. One other major difference between ape and human teeth is enamel thickness, with human tooth enamel being much thicker.

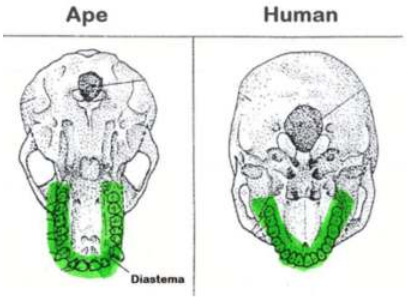


Figure 40.2.7: The dental arcade is made up of the teeth in the upper jaw of a primate. One of the key features is the diastema, a gap that exists for the large canine teeth. (Image: Public Domain)

On the upper jaw, the dental arcade provides additional areas where we can look for physiological changes in hominids. The shape of the dental arcade is an obvious starting point, where humans have a much more V-shaped arrangement and apes a more U-shaped arrangement. This is largely due to the shortening and flattening of the face through evolutionary change.

The other large difference here is the diastema, or the gap between the premolars and incisors where the large canine teeth reside. Humans have no diastema while hominid ancestors do.

Three other skull features are related to chewing, or lower jaw motion. These are the zygomaticus arch, the sagittal crest, and the supraorbital brow ridge.

The sagittal crest was lost to hominids as the demands of chewing became less strenuous. Because these crests were a primary area of attachment for the temporalis muscles that control the lower jaw, they could be quite pronounced in very early hominids.



Figure 40.2.8: The sagittal crest on a primate skull indicates the presence of very strong jaw muscles. (Image: Public Domain)

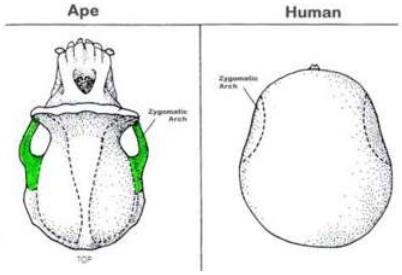


Figure 40.2.9: Zygomaticus arch. (Image: Public Domain)

The zygomaticus arch is much larger in earlier primates than it is in humans. This is due to the need for much thicker and robust temporal muscles. These muscles attached to the lower jaw through and behind the zygomaticus arch and then attach to the side of the cranium. They are the muscles that move the lower jaw and, the larger they are, the more powerful the chewing ability of the organism.

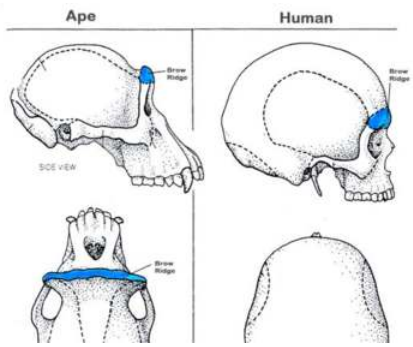


Figure 40.2.10: Supraorbital brow ridge. (Image: Public Domain)

The supraorbital brow ridge is one of the last skull features lost in hominid evolution, with humans no longer displaying it at all. This structure provided strength to the facial portion of the skull. As the powerful temporal muscles placed a good deal of strain on the face as they pulled on the cranium while food was being masticated. This ridge kept the eye sockets from collapsing under this stress.

Other features provide support for changes in locomotion, specifically the evolution from walking on all fours to mostly on the hind legs to full bipedalism, or walking upright on two feet. Today, humans are bipedal. Skull features that accompany this progressive change in stance include the nuchal area and the foramen magnum.

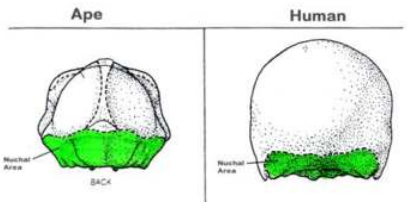


Figure 40.2.11: Nuchal area, where the neck muscles attach to the back of the skull. (Image: Public Domain)

The nuchal (pronounced “new-kul”) area is located on the base toward the rear of the skull. This is the area where neck muscles attach to the skull. Our primate ancestors have a much larger nuchal area. This is because much larger muscles are required to keep the individual looking forward when the spine is situated so far back in the skull.

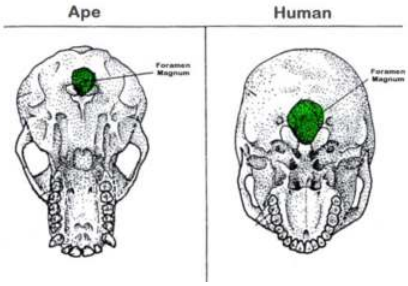


Figure 40.2.12: Foramen magnum, where the spinal cord enters the brain. (Image: Public Domain)

The foramen magnum is the hole in the base of the skull where the spinal cord enters. Depending upon the primate species, the foramen magnum is located at a fulcrum, or balance point, and so its position is highly indicative of overall locomotion. In humans, it is in a more forward position while in apes, who walk mostly on all fours, it is situated further toward the rear of the skull.

The final and perhaps most peculiar change in the hominid skull is the cranial capacity. Until the genus *Homo sp.* evolves, hominid skull capacity does not change in any substantial way. However, it certainly increases quite a bit in our most recent ancestors. In *Homo habilis*, which will be discussed below, the cranial volume is a mere 600 cm^3. In modern humans, by contrast, it is 1400 cm^3. Some Neanderthals even exceeded this, with cranial volumes approaching 1700 cm^3.



Figure 40.2.13: Changes in cranial capacity (volume) are significant through the course of hominid evolution. (Image: Public Domain)

The greater cranial volume evolved to fit a growing brain. One might be tempted to attribute this larger brain simply to greater intelligence, but this is a matter of great debate. While there does seem to be a correlation between the two variables, it is not necessarily causal, as intelligence (however you define that term) can potentially be attributed to other characteristics of the brain. Further, larger cranial capacities are also directly associated with larger body size, and adaptations to colder climates (which certainly was necessary in Neanderthals and modern humans). There is also some degree of sexual dimorphism among hominids. On average, males tend to have slightly larger cranial capacities than females of various hominid species. This is not an issue of greater intelligence, but rather cranial capacity being directly linked to larger average body size.

This is not to say that intelligence does not play a role in cranial capacity. It is more likely that brain physiology controls that. For decades, it was assumed that brain growth in humans was linked to a kind of reinforcing feedback loop between intelligence, technological development, and hunting meat. There may be some truth to this and, in the context of a discussion of *Homo habilis* below, the connection between meat and brain growth will be explored more specifically. Overall, as hominids evolved from a strictly vegetarian diet to scavenging bone marrow to directly hunting game, cranial capacity and brain growth followed. These were accentuated by technological developments that made scavenging and hunting more efficient. Simple wooden tools gave way to knapped stone scrapers. These gave way to elaborate arrow and spear points. Eventually – humans learned to smelt metals. Copper was followed by bronze and iron, each with their own "age," and now there's silicon. Are these all the result of greater cranial capacity and brain size? Perhaps some of it is, but certainly there are many other variables to consider.

Below is an excellent video showcasing hominid skull features by Dr. Daniel Lieberman, Professor of Human Evolutionary Biology at Harvard University.



Did I Get It? Hominid Skull Physiology

Exercise 40.2.1

Early hominids had powerful jaw muscles. Which skull feature protected the face from collapse under this stress and was also one of the last to go away in anatomically modern humans?

a. Mandible

b. Super Orbital Brow Ridge

c. Sagittal Crest

d. Zygomaticus Arch

Answer

b. Super Orbital Brow Ridge

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40.3: Highlights of our family tree - Australopithecina and Homo

Like all families, the human family and its ancestry are filled with interesting, odd, characters with unique traits and, in this case, a trail of important evolutionary changes and adaptations. The image below can be referred back to for context as we move through the human ancestral lineage, in this case, from right to left (note the vertical time scale). The image below does not include the first four hominids discussed below, all of which predate *Australopithecus anamensis*.

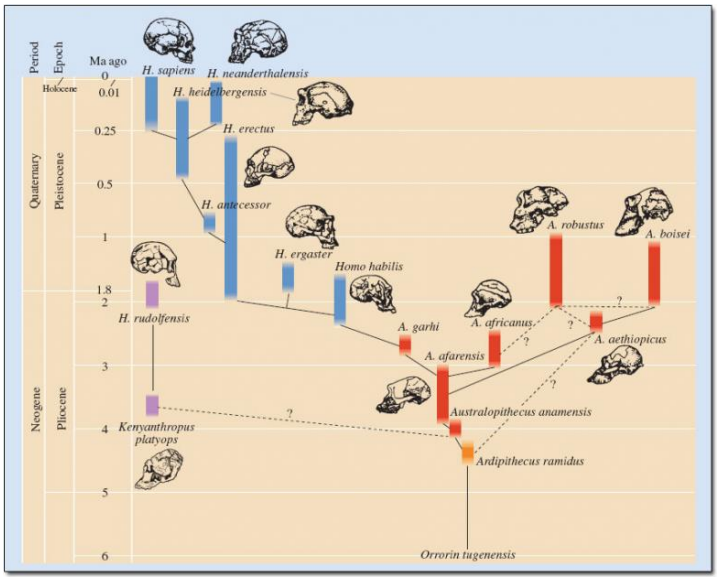


Figure 40.3.1: A current tentative arrangement of the ancestral human lineage. The Australopithecines are placed here in context with modern humans in time. (Image: Public Domain)

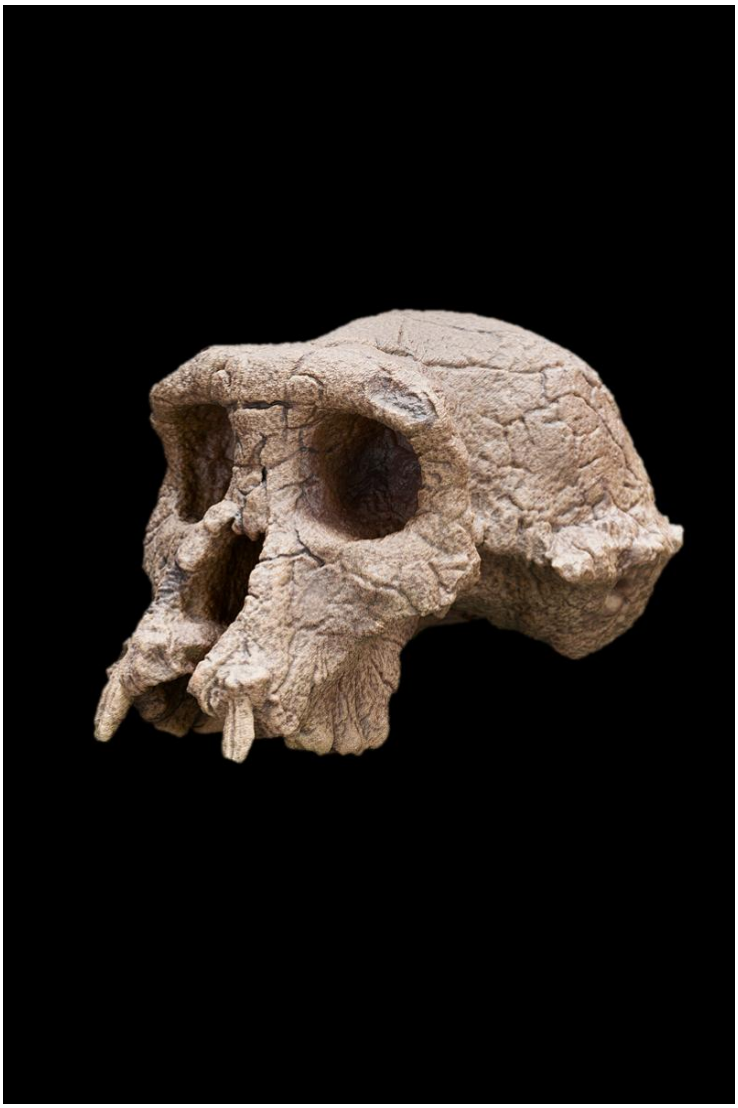


Figure 40.3.2: Sahelanthropus tchadensis, Toumai – TM 266-01-060-1, skull, 3/4 view (Image: Smithsonian Institution)

Our oldest directly connected human ancestor, *Sahelanthropus tchadensis*, is traced to the modern African nation of Chad about 6-7 Ma. Fossil evidence suggests that they walked upright. The main evidence for this is the more forward positioning of the foramen magnum, the opening for the spinal cord at the base of the skull. As this opening evolutionarily moves forward, it would indicate that the spinal column was moving to a position directly below the skull, as it is for modern humans today. *Sahelanthropous tchadensis* is known only from nine cranial specimens (the first of which was discovered by Michael Brunet in 2001), so fossil evidence for much else is very limited. We can also say that, due to its heavy tooth wear, it was likely a vegetarian.

Figure 40.3.3: Skull fragment of *Sahelanthropus tchadensis*. Note the human-like feature in the downward and forward positioned foramen magnum and the ape-like protruding mouth, large teeth, narrow and small braincase, and highly pronounced occipital ridge above the eyes. (Model: Creative Commons License; Smithsonian Institution)

Orrorin tugenensis and Ardipithecus kadabba – 5-6 Ma

Little is known about these two hominids. Both come from east Africa: Kenya and Ethiopia, respectively. From the few skull and leg fragments known, there are two characteristics that they share with modern humans. There is one upper femur specimen of *Orrorin tugenensis* that shows the same kind of bone buildup typical of a biped. Dentition, or teeth, are also important for both of these hominids. Both had thick enamel, which is a shared characteristic with modern humans. Both were about the size of a modern chimpanzee and likely subsisted on a vegetarian diet.

Ardipithecus ramidus – 4.4 Ma

Known as "Ardi" the first partial skeleton of this organism was presented in 2009, though Ardi had been known since 1994. This much more complete skeleton presents a great deal of important information. A crucial behavior advancement with Ardi appears to be an omnivorous diet. This is suggested by its tooth enamel, which was only moderately thick. The thicker the enamel, the more equipped an organism is to eat plants, as they require a good deal of crushing and grinding to begin this first part of the digestive process. Meats are much softer foods. Other skeletal characteristics indicate that both tree-climbing and bipedalism were practiced. Given that these specimens have been found alongside faunal remains that indicate a wooded environment, it seems reasonable that Ardi was not a dweller of open savannah. Rather, they lived near enough to forests where their arboreal adaptations would be beneficial. This brings up an old debate in paleoanthropology – the woodland versus savannah hypothesis for early humans. That is, what is called the "Savannah Hypothesis" posits that human bipedalism evolved as a direct result of our ancestors transitioning from forests environments to savannah environments, where an arboreal lifestyle was no longer advantageous. Ardi would seem to suggest otherwise.

A Word About Bipedalism

There is no doubt that modern humans are primarily bipedal and that our Australopithecine ancestors were not. In between, there is a clear progression from walking on fours to walking on two legs. But, why?

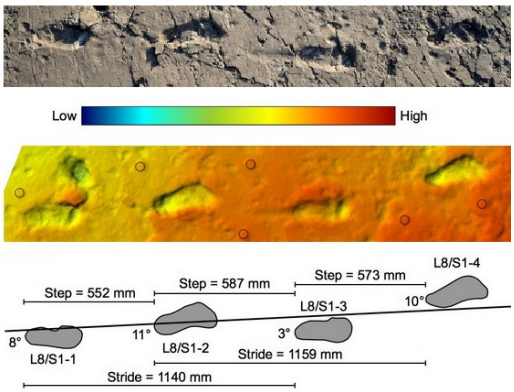
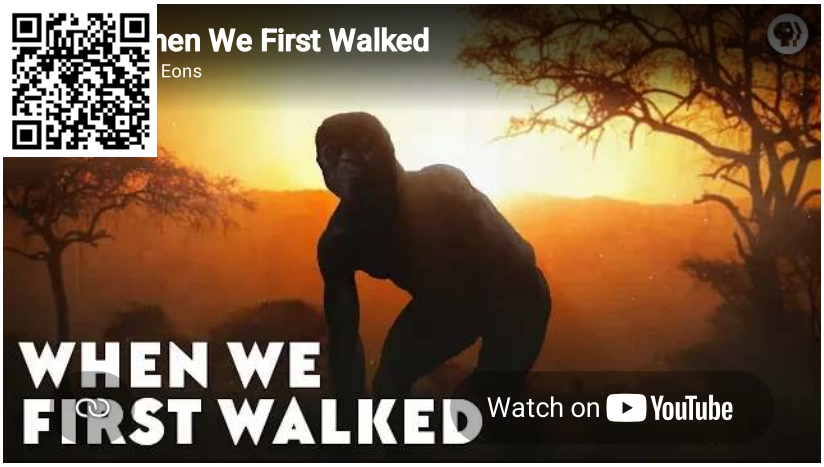


Figure 40.3.4: A portion of the Laetoli trackway, preserved in volcanic ash. (Image: Fidelis T Masao Elgidius B Ichumbaki Marco Cherin Angelo Barili Giovanni Boschian Dawid A Iurino Sofia Menconero Jacopo Moggi-Cecchi Giorgio Manzicropping/editing by Dennis Pietras – parts of 2 figures from https://elifesciences.org/content/5/e19568 licensed under CC BY 4.0)

The advantages of bipedalism are very important. Bipedalism is a critical development that allows for the free usage of the hands for things other than walking or even climbing. Suddenly, carrying young is easier. Making tools becomes more possible. It also improves our long distance locomotion, making it easier to traverse much longer distances. Closely related to the ability to travel long distances is the ability to cool off, which bipedalism makes much easier, the body not being so hunched over on itself. Finally, with the climate changing so drastically in Africa, from tropical rainforests to savannah, being bipedal becomes a survival strategy. It is now possible to see over the grasses.

There are also disadvantages to walking on two legs. The curvature of the lower spine that makes balancing on two legs possible also leads to much more lower back pain. The move toward bipedalism led to the toe positions of the back feet being changed so that the big toe rests up against the next two. This makes "pushing off" while walking much more effective, but makes climbing trees like our ancestors very clumsy. Finally – and perhaps most importantly because of its role in reproduction – bipedalism works with a narrow pelvis. This narrow pelvis makes walking upright very efficient, but it makes childbirth much more difficult, given that the size of the birth canal is part of what is sacrificed for the ability to walk upright. In fact, in most modern human births, the head is too large for the birth canal. This is accommodated in female humans through the release of a hormone, relaxin, that softens the ligaments in the pelvis, allowing the various bones to create the space necessary for the birth to occur. This is augmented by the malleability of the baby's skull, the bony plates not having yet fused. When all things don't work together as they have evolved to do, birth complications arise. Modern medicine has found ways around such situations, with procedures such as Caesarean Sections, among others.

Here is an excellent video discussion of the wider variety of anatomical changes associated with hominin bipedalism.



A WORD ABOUT BIPEDALISM FOR DISTANCE RUNNERS

Hominid bipedalism brought with it an ability to run. Running was not really as effective for our primate ancestors. It was more effective to use their arboreal, prehensile skills to climb trees. Why? Well, if you are escaping from predators or even chasing prey in a forest environment, these skills are quite useful.

But, what if the climate is changing and the environment along with it, so that tropical rainforests give way to savannah? How do you adapt and retain the ability hunt and escape – or just to travel long distances in search of food? The ability to run becomes important. And, not just the ability to run, but to run for long periods of time. Starting about 2Ma, the physiological ability to run, or really trot, become possible due to improvements in bipedal skeletal structure. These include the development of longer legs and shorter toes. Also, larger gluteal muscles, a smaller waist, and more neck flexibility played a part. The hominid foot had already also developed a larger heel bone (calcaneus) and longer Achilles tendons were in place. The ability to run also required changes in our metabolic abilities. These changes directly affected our brain, because both running and brain activity use a good deal of glucose. Changes in brain development that occur along with running include an enlarged anterior hippocampus, a change that likely led to improvements in memory (Schulkin, 2016). There are many more examples of physiological developments related to bipedalism from which we can draw in the literature. But, it is definitely worth pointing out here that while bipedalism itself brought important advantages, so did the endurance capabilities coupled with its evolutionary development for hominids.

Interestingly, sometimes evolutionary priorities conflict. In the case of the evolution of hominids, the ability to walk upright prevailed. This, despite the disadvantages that developed for childbirth, demonstrates that bipedalism must have been the more important survival strategy at the time.

Australopithecus anamensis, A. afarenesis, A. Africanus, A. garhi, and A. Sediba – 1.5-4Ma

These five species of australopithecines are all extinct, but provide important glimpses into the ancestral development of human features.

Australopithecus anamensis – 4.2-3.8Ma

Discovered near Lake Turkana, Kenya and also in Middle Awash, Ethiopia, *A. anamensis* shared traits we see in both apes and humans. Fossil evidence indicates bipedalism, particularly due to the human-like positioning of the ankle joint. But, long forearms also suggest that they climbed trees. Like Ardi, this specimen seems to argue for a still close connection to an arboreal lifestyle with some amount of bipedalism having developed. Its cranium also displays important features that connect it to humans, such as foreword projecting zygomaticus arch, or cheekbone. *A. anamensis* still retains important ape skull features, however, such as a narrow braincase and protruding face.



Figure 40.3.5: Australopithecus anamensis fossil bones, as displayed at the Royal Belgian Institute of Natural Sciences, Brussels. (Image: CC BY-SA 4.0; By Ghedoghedo – Own work)



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Figure 40.3.6: Reconstruction of the fossil skeleton of “Lucy”, Australopithecus afarensis. (Image: 120, CC BY-SA 3.0, via Wikimedia Commons)

Australopithecus afarensis is represented most famously by what is one of the most complete skeletons of an early hominid, nicknamed “Lucy”. There are over 300 individuals that comprise the fossil evidence of this species, so it is quite well studied. In many ways, this species represents a major signpost in human evolution, a place where we have enough information to confidently look both forward and backward when it comes to shared traits in our own species.

There is a fairly large degree of sexual dimorphism in *A. afarensis*, most notably in the difference in average height between adults males (151 cm) and females (105 cm). We know from skeletal evidence that, like chimpanzees today, they grew up into adulthood much faster than modern humans. Culturally, this would have left them with much less time for parental guidance and socialization during this brief childhood. Ian Tattersall (2012) has even suggested that they spent some of their times roaming savannah areas in herds like baboons are sometimes seen doing today, as opposed to living in small tribes like chimpanzees.

The species still retained many ape-like features, such as a flat nose and projecting lower jaw and a smaller braincase. It also retained the long, thin arms and long fingers that are well-adapted for tree climbing. Human characteristics come with smaller teeth and a body that stood fully on two legs and walked upright. It was an overall well-adapted species that a long-lived species, existing for nearly one million years.

Figure 40.3.7: Note some of the skull features of A. afarensis. There are still ape-like protruding mouth and small brain case, but the occipital ridge above the eyes is much less pronounced and the zygomaticus arch (cheekbone) is more forward positioned, as is the foramen magnum on the bottom of the skull. Dentition not shown. (Model: Creative Commons License; Smithsonian Institution)

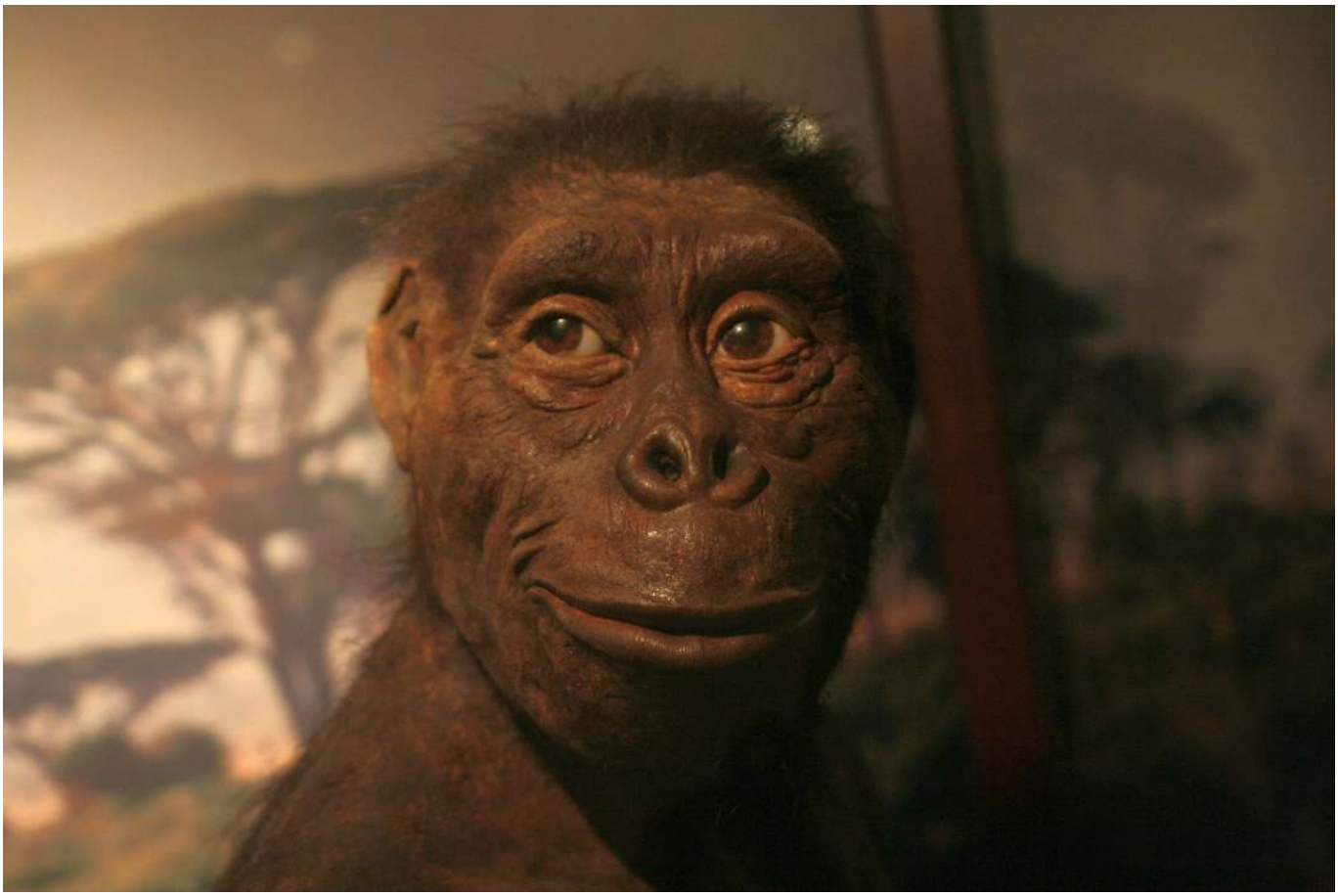


Figure 40.3.8: Reconstruction of “Lucy” in the Warsaw Museum of Natural History. (Image: CC BY-SA 3.0; By Shalom – Own work)

Did I Get It? Bipedalism

? Exercise 40.3.1

A large array of physiological changes accompanied the evolution of bipedalism. Which is these is a major change associated with the skull?

a. The foramen magnum moving forward toward the mandible.

b. Changes in hip placement.

c. The ability to grip with a thumb.

d. The changing position of the birth canal.

Answer

a. The foramen magnum moving forward toward the mandible.

Australopithecus africanus – 3.3 to 2.1 Ma

A. africanus is very similar to its immediate ancestor *A. afarensis*. Distinguishing features include a more rounded cranium with a larger brain and smaller teeth. Based on their teeth, it is thought that they were vegetarian, unlike some of their ancestors. For a long time, they were viewed as predators. This was not because any discovery of stone tools, but because some specimens had been discovered with broken animal ones nearby. Since then, it is now thought that these represent animal predation ***on*** *A. africanus*, rather than the other way around.

Still, these individuals were fully bipedal and yet were still well adapted for arboreal life. Specimens have been found primarily in South Africa, which may indicate this species was a southern cousin to *A. afarensis.*

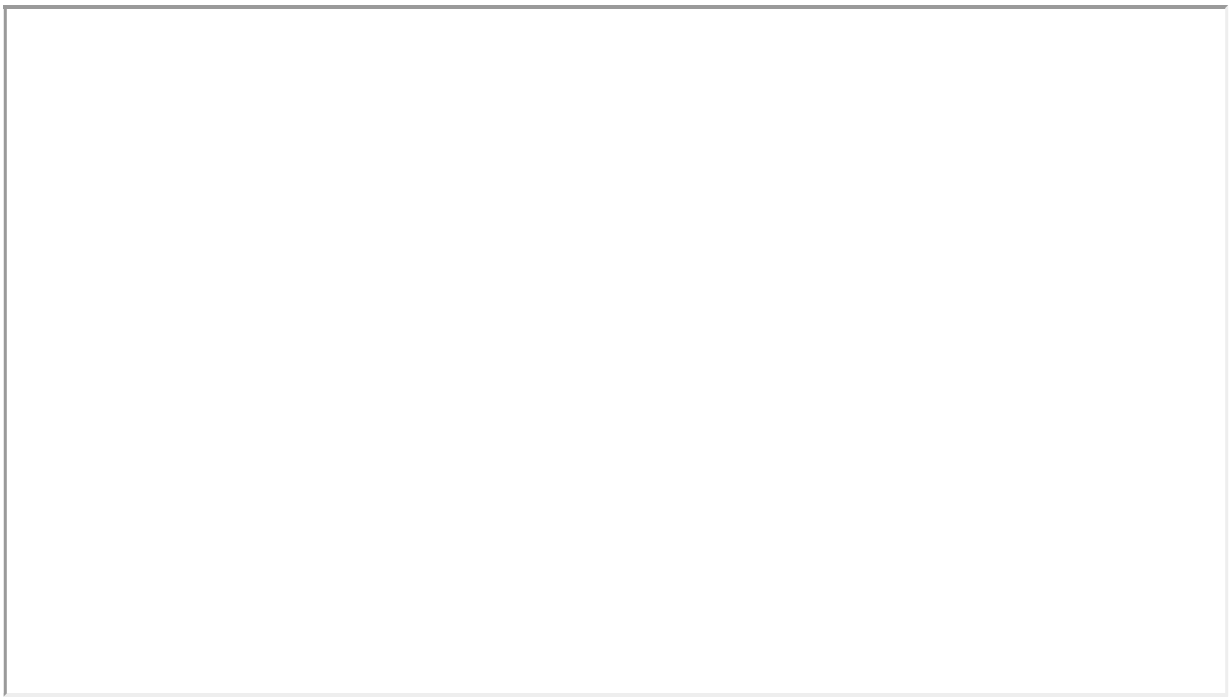


Figure 40.3.9: Above: Cranial specimen of A. africanus. There are many still-developing human features as described in above examples, but the larger braincase here is probably the most important development.

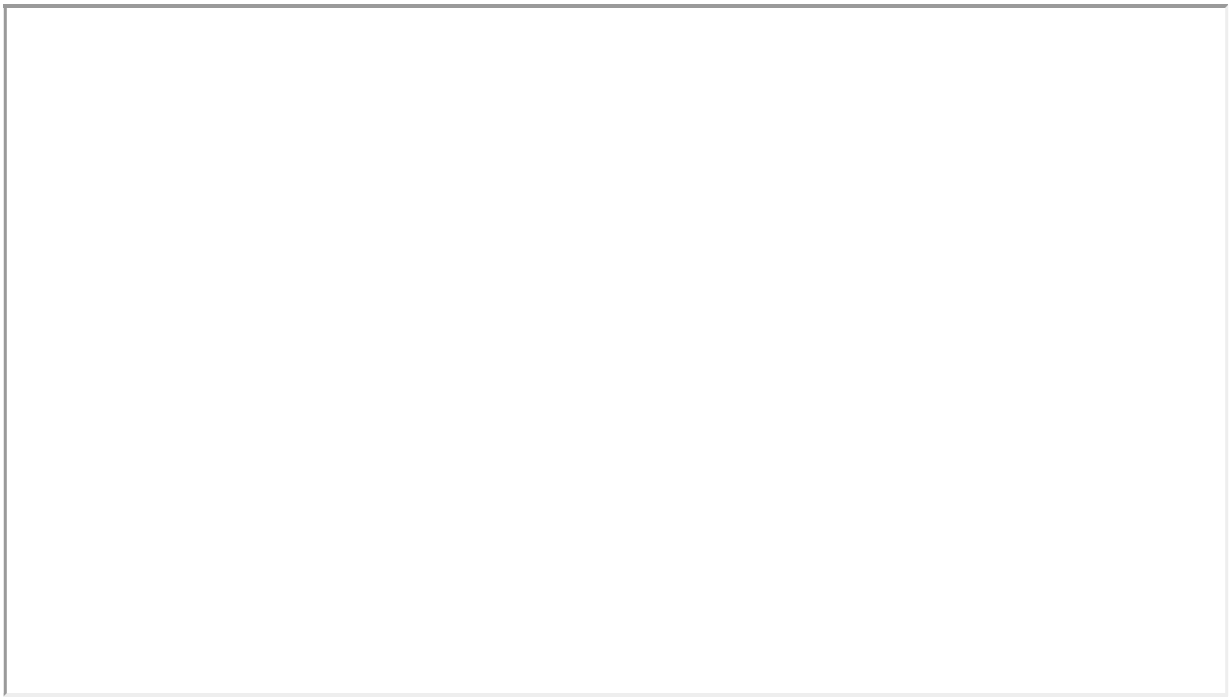


Figure 40.3.10: Above: Cranial specimen of A. africanus. This specimen allows some examination of dentition. Note the lack of a protruding, large, canine tooth, something that would be indicative of more ape-like dentition. The flat molars presented here are indicative of a vegetarian diet. (Model: Creative Commons License; Smithsonian Institution)

Australopithecus garhi and A. sediba

Of all of the Australopithecine specimens, these two latest-lived species are poorly known, as well as short-lived. *A. garhi* is only known from a few fragments. But, because it dates to a very poorly-documented time period in human evolution, 3-2 Ma, it is very important. The few fragments that do exist indicate a significant increase in bipedal stride length, and thus a more efficient gait. Another important piece of evidence are the stone tools that have been found in association with *A. garhi*. One of the first documented examples of tool usage among early humans, the fragments are solid evidence of hunting as a form of food procurement.

A. sediba is known from a very well-preserved and nearly complete skeleton found Malapa Cave, South Africa. This species lived at the same time that the genus *Homo* first appears in the fossil record. At the tail end of the Australopithecines, *A. sediba* lived alongside some of our earliest direct ancestors.

This detailed skeleton reveals some important developments. Changes in the pelvis indicate more upright walking, various tooth traits and a broad chest are more human-like, features of the femur and legs overall indicate that it spent a much greater time walking and doing so in a very similar fashion to modern humans, much more so than its other

Australopithecine ancestors. Derived features of its skull are also more human, such as small premolars and molars and facial features that are less pronounced and more human.



Figure 40.3.11: A. sebida skull. Note the more human features, such as the larger braincase, less forward protrusion in the lower jaw, and a more compact dentition. (3D Model: Creative Commons; William Mitchell)

The Genus Homo – 2.4 Ma

The genus *Homo* appears to have split from Australopithecines somewhere between *Australopithecus africanus* and *Australopithecus garhi.* The earliest direct ancestor of the genus is represented by *Homo habilis*, a species known best from specimens found in parts of eastern and southern Africa between 2.4 and 1.4 Ma.

It is very important to note that the early specimens of the genus *Homo* are all found in eastern Africa. Very rapidly after the evolution of *Homo erectus*, sometime around perhaps 1.5 Ma, human ancestors are found in other parts of the world than just Africa. At that point, the human diaspora had begun.

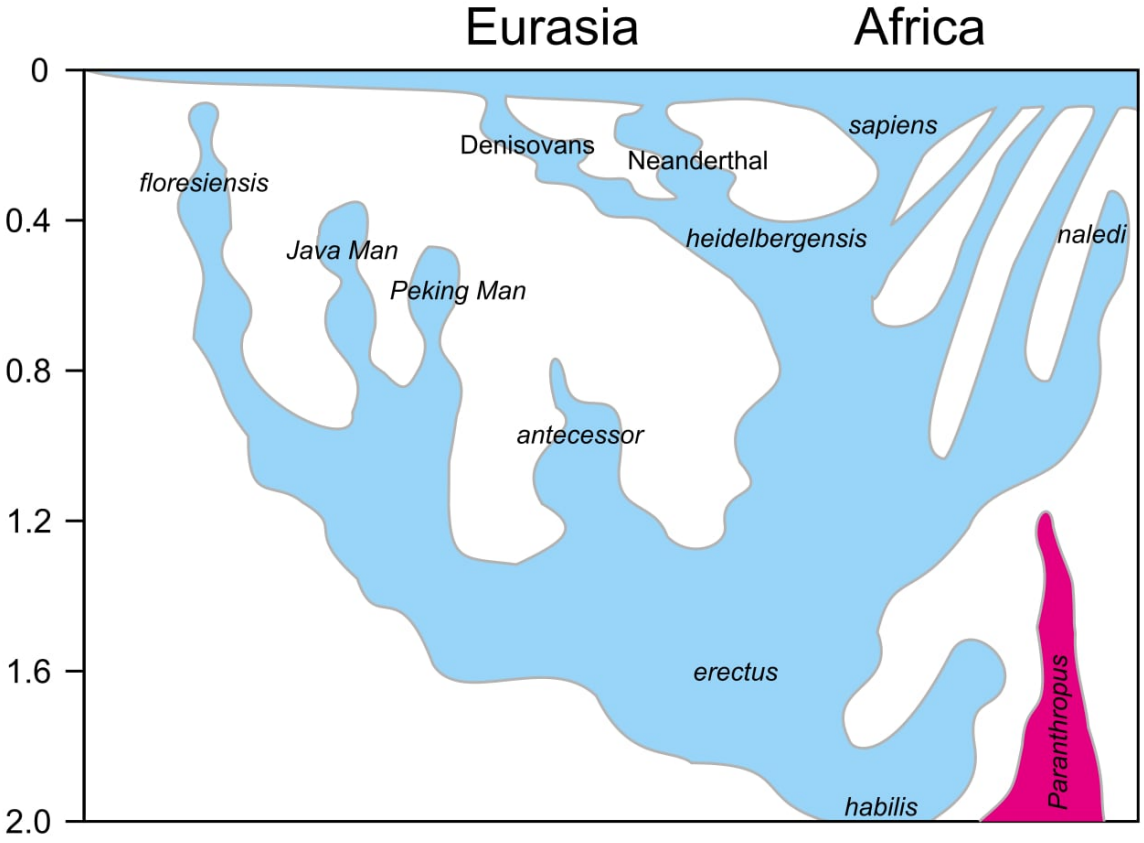


Figure 40.3.12: A model of the evolution of the genus *Homo* over the last 2 million years (vertical axis). The rapid "Out of Africa" expansion of *H. sapiens* is indicated at the top of the diagram, with admixture indicated with Neanderthals, Denisovans, and unspecified archaic African hominins. Late survival of robust australopithecines (Paranthropus) alongside Homo until 1.2 Ma is indicated in purple. (Image: CC BY-SA 4.0; Conquistador, User:Dbachmann – updated version of File:Homo-Stammbaum, Version Stringer-en.svg)

Homo habilis – 2.4-1.4 Ma

Our earliest *Homo* ancestor, *Homo habilis*, had a varied diet that appears to have been omnivorous. Originally discovered by the Leakey family — Jonathan Leakey specifically — the early specimens were found in Olduvai Gorge, Tanzania. It is the first of the genus *Homo* to be associated with the use of stone tools, due to the massive amounts of flakes and tools found in context with the skeletal remains. Later, cut and smashed animal bone would also be found in context with *H. habilis*, suggesting that meat was important to the diet on some level.

A WORD ABOUT MEAT AND BRAINS

Since the 1950s, paleoanthropologists associated increased cranial capacity to the increasing meat content in the diet of human ancestors. Evidence supporting this hypothesis include that *Homo habilis* was a known toolmaker and those tools were used in some part for meat procurement. Additionally, up until *H. habilis*, hominid cranial capacity is increasing. Discoveries such as these in ancient Ethiopia, around what was then a rather verdant lake, have been interpreted as the influx of protein that led to major

evolutionary and cultural advances in the human lineage. Yes, meat-eating made us smarter. Humans are the only primate that eats other large animals other than those of our own species (and cannibalism is well-known, too). Butchery marks on animal bones, stone tools, and the genus *Homo* all come around the same time in the fossil record.

However, new research suggests that, rather than the meat, it might have been bone marrow and associated nutritious fats that helped early human brain development and tool development (Thompson et al., 2019). Early humans may have eaten meat, but eating meat out in the open was a dangerous business for our ancestors, as other predators and scavengers lurked nearby, and the meat itself was difficult to move. And, with so many very large mammals, it is more likely that humans were some of those scavengers, opportunistically awaiting while large predators stripped hunted carcasses of meat, leaving behind bones. Bones could be then be removed and easily carried to safe locations for consumption. Further, meat quickly rots, while bone marrow stays fresh longer. The geologic epoch of this transition, the Pliocene, was also a time of major environmental change, as the climate was cooling intensely and was becoming more variable. Large swaths of what was rainforest were becoming grasslands and savannah. After 2 Ma, human ancestors were likely consuming meat in addition to bone, but it is very likely that bone scavenging and processing with stone tools came first. The conclusion is that we ate marrow before we ate meat.

Here is a nice overview of some of the ongoing discussions of the ancient dietary changes in our hominid ancestry at this point.

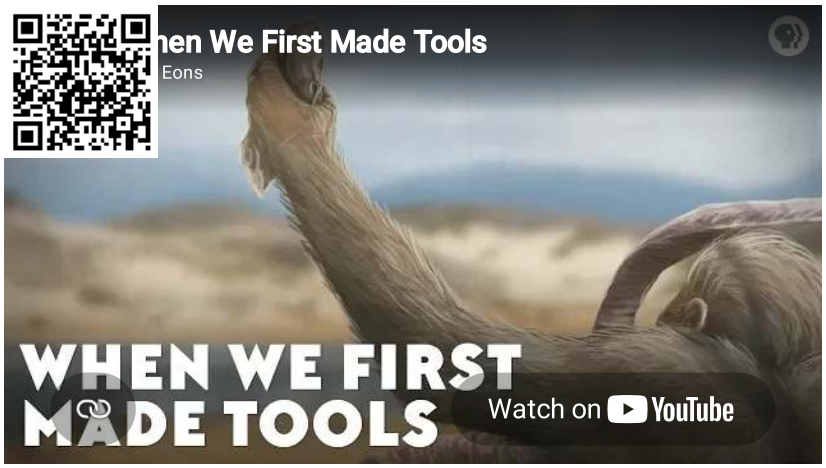


Figure 40.3.13: Homo habilis skull, where many early human, yet still very ape-like features can be seen. Human features include a larger cranium, lack of a protruding canine tooth, more compact dentition due to a less protruding mouth, and a downward/forward foramen magnum. There is still a very pronounced occipital ridge above the eyes. (3D model by **BuluLulu** on Sketchfab.)

Next, examine this chopper tool from Olduvai Gorge, Tanzania. At 1.8 Ma, this tool could have been used by *Homo habilis* to access bone marrow. Tools of this kind were often very stylized, an indication of common culture, as uniformity in stone tool manufacture indicates not only a transfer of cultural skills and knowledge, but also, potentially, specialization of skills to some degree within these early ancestral human populations. Flakes were removed form a larger core to create a sharp edge. This edge could be sharpened further until it was no longer of use.

Figure 40.3.14: Stone scraper, and early stone tool. (Model: Creative Commons License; Smithsonian Institution)

An excellent introduction to stone tool technologies and their development can be enjoyed in the video below.



TOOLS

We know that human ancestors are not the only animals to use tools. One brief example outside of our human family are chimpanzees. Jane Goodall famously documented chimpanzee tool usage while living among chimpanzee communities. Behaviors she observed included picking up leafy twigs, stripping the stems of the leaves, and using these stems to 'fish' for termites and ants. Since then, chimpanzees have been seen using tools to crack nuts as well as sharpening sticks to hunt for other mammals. In the last thirty years, gorillas and bonobos have also been observed using tools of various kinds.

Primates are not the only tool-users either. Birds, whales, sea otters, and even various predators have been observed using tools of various kinds. Given all of this, tool usage is certainly not indicative of human behavior, but something hominids likely inherited as an ancestral adaptation, but seems to be a behavior that has developed independently, evolving in multiple directions.

Did I Get It? Intelligence

? Exercise 40.3.1

The development of human intelligence is an area of intense debate, but is often discussed in the context of cranial volume. Which is a more likely measure of intelligence, rather than brain size?

a. The development of religion

b. Brain physiological features and their arrangements

c. Our ability to harness silicon and other metals

d. Speech

Answer

b. Brain physiological features and their arrangements

Homo erectus and Homo floresiensis– 1.89 Ma to 50 ka

Homo erectus has been well known since its first fossil discovery in Indonesia in 1891. This discovery actually predates all discoveries of human ancestors in Africa. The discoverer, Eugene Dubois, originally named the species *Pithecanthropus erectus*, or “erect ape-man,” but it was popularly referred to as “Java Man.” Many discoveries of this species would follow, both in China (where it was called “Peking Man”) and in east Africa.

Homo erectus was quite humanlike in its proportions. By this point, it had lost the long arms and prehensile ability of its arboreal ancestors. These traits had given way to a modern human gait and stature. *H erectus* was on average rather tall, ranging from 145-185 cm (4.7-6 ft), with an associated weight range of 40-68 kg (88-150 lb). It is very likely that not only did they spend most of their time walking, but running was likely important to their locomotion, possibly for long distances. Young appear to have grown quickly, like other apes, but the species has shown evidence that it may have cared for elders and weaker individuals. *Homo erectus* is also associated with a major stone tool development, the hand axe.

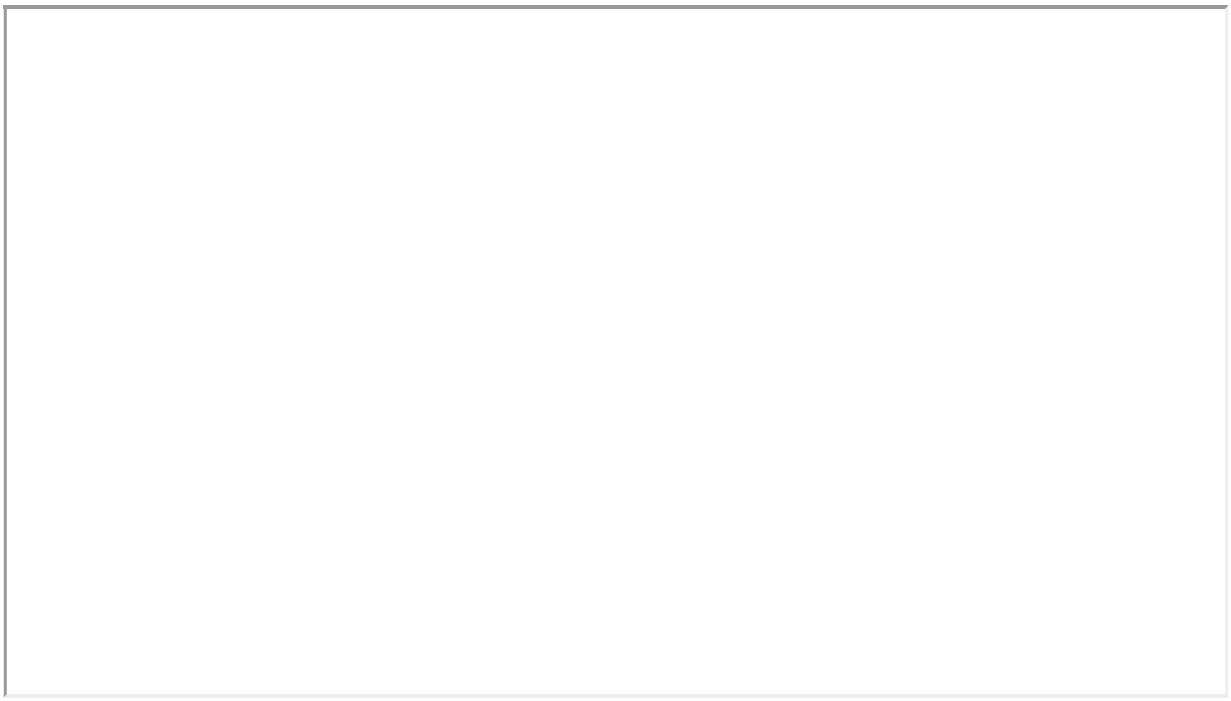


Figure 40.3.15: Homo erectus skull. Note the much more human-like characteristics, despite the still pronounced occipital ridge. The face protrusion is much less pronounced than in its ancestors and the cranial capacity is much greater. (Model: Creative Commons License; Smithsonian Institution)

Figure 40.3.16: Above: A hand axe dating back to 1.1Ma from Isampur, India, that is associated with use by Homo erectus. Stone tools like these are much more advanced in design and in the skill necessary to create them, indicating the development of some specialization in this craft. (Model: Creative Commons License; Smithsonian Institution)

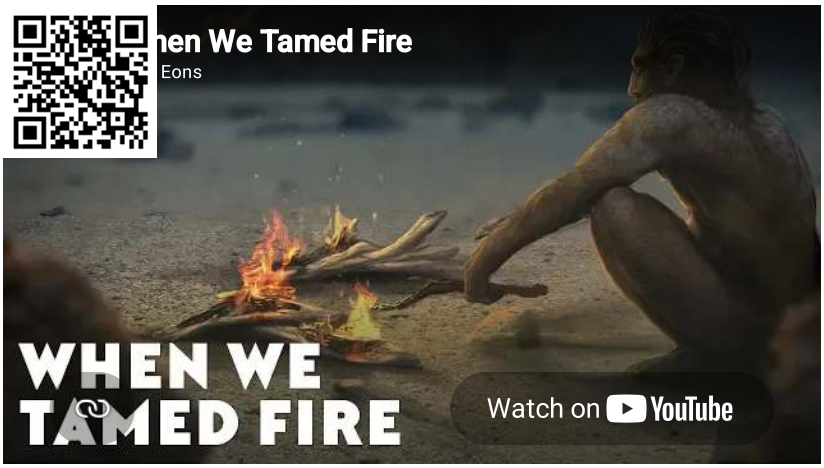
Flint-knapping, or the hand-shaping of rock tools, was not easy. It required highly specialized and learned skills, particularly as the work progressed from simple scrapers to well-hewn spearpoints. It is very difficult. The video below provides an excellent introduction to the ancient set of skills:



Video of Tyson Hughes demonstrating flint-knapping (stone tool manufacture) for Crow Canyon Archaeological Center, Cortez, Colorado.

Homo erectus is certainly the longest-lived species of *Homo.* It lived for longer than anatomically modern humans have yet managed to exist. Their tall frame and large brain required a very large amount of energy to function, making meat-eating for easily digested protein a key ingredient in their diet. Their stone tool development certainly points toward the processing of meat as being an important part of the diet. One other development that is important to point out at this point is the use of fire. Up until *Homo erectus*, there is no evidence in human ancestry of the purposeful starting and use of fires. Now, the remains of hearths are found in context. While these were almost certainly used for cooking, they were also apparently used for social interaction, warmth, and to keep predators at bay.

Here is a great video showcasing some of our developmental history with fire.



An extinct descendant of *Homo erectus* is *Homo floresiensis*. Nicknamed “the hobbit”, it has only been found very recently (2003) on the Indonesian island of Flores. *H. floresiensis* is also associated with stone tool use, but it appears to have gone extinct as late as 50 ka. Known individuals (and there are only a few) stood only 106 cm (3.5 ft) tall, had tiny brains, large teeth, and an array of unusual facial features. These include flat foreheads and less pronounced chins. Their feet were very large for their legs as well. Despite all of this, there is evidence that they subsisted in part on small elephants, large rodents, used fire, and made stone tools. Large monitor lizards (like Komodo dragons) may have preyed on them. One possibility for the distinct stature and overall form of *H. floresiensis* is **island dwarfism**.

Homo heidelbergensis-1.3 Ma-200 ka

Eventually, human ancestors would spread beyond Africa, China, and Indonesia and into Europe. The first specimen of *Homo heidelbergensis* was discovered Near Heidelberg, Germany, in 1908. This is the first species of ancestral human to live in cooler climates. Living in Europe in the midst of the Pleistocene epoch’s “Ice Ages” would not have been simple, so we can hypothesize a need for body coverings, skill with fire, and other cold-weather adaptations, including the hunting of large animals. Indeed, we have much of this. There are fire-altered tools and wood at a the site of Geshre Benot Ya-aqov in Israel, for example. *H. hiedelbergensis* also likely built shelters. Evidence of this comes from the site of Terra Amata, France.

Evidence for their hunting of big game comes from analysis of the bones of those animals where they have been found in context with *H. heidelbergensis* remains. These include butchery marks on deer, horses, elephants, hippos, and rhinos. An example of a site with such features as an early thrusting spear exists near Schöningen, Germany.

Finally, there is evidence of human ritual among *H. heidelbergensis* near Atapuerca, northern Spain. An apparently purposeful mass burial of 30 individuals dates to 400ka. Was this ritual? Certainly, there was some kind of purpose in such a burial. This pit, nicknamed *Sima de los Huesos*, or “Pit of Bones” also contained evidence of stone tool use, as a hand axe was also found in context.

Figure 40.3.17: A skull from *H. heidelbergensis*. Note how human this skull appears. The braincase is much larger, and the face much flatter. The cheekbones are less pronounced. There is, still, a rather elevated occipital ridge, which is a more ape-like feature shared with all ancestral humans up to this point. (Model: Creative Commons License; Smithsonian Institution)

Homo naledi – 335-236 ka

Remains of Homo *naledi* has so far only been found in South Africa. *H. naledi* are cousins of humans who never left Africa. At the site where this species has been found, there are 15 individuals that have been recovered. It is one of the largest single fossil hominid caches known. So, there is a fair amount of physiological information within that small population, including both adults and juveniles. Their remains have been dated to between 335,000 and 236,000 years of age.

Despite clear similarities with other species of the genus *Homo*, a number of features from *Australopithecus* are also retained. This has made the placement of *H. naledi* on the hominid phylogenetic tree challenging.

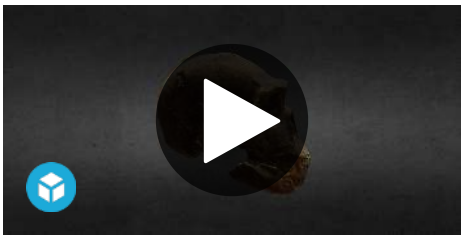


Figure 40.3.18: Homo naledi skull. While the face still retains a more ape-like protrusion and the occipital ridge is still very pronounced, the overall appearance of this South African cousin of modern humans has dentition very close to our own, a large braincase, forward cheekbones (zygomaticus arch) and seems to have certainly been bipedal. (3D model by **ArtinScienceMA**; used with permission.)

Neanderthals

Until now, the only one species of *Homo* has been known to have migrated out of Africa, *Homo erectus*. We have little information on the fate of that species, but we do know that it died out around 117,000 years ago, well before modern humans, *Homo sapiens*, would encounter some of the same environments in Asia. Thus, there is currently no evidence that these two species ever interacted.

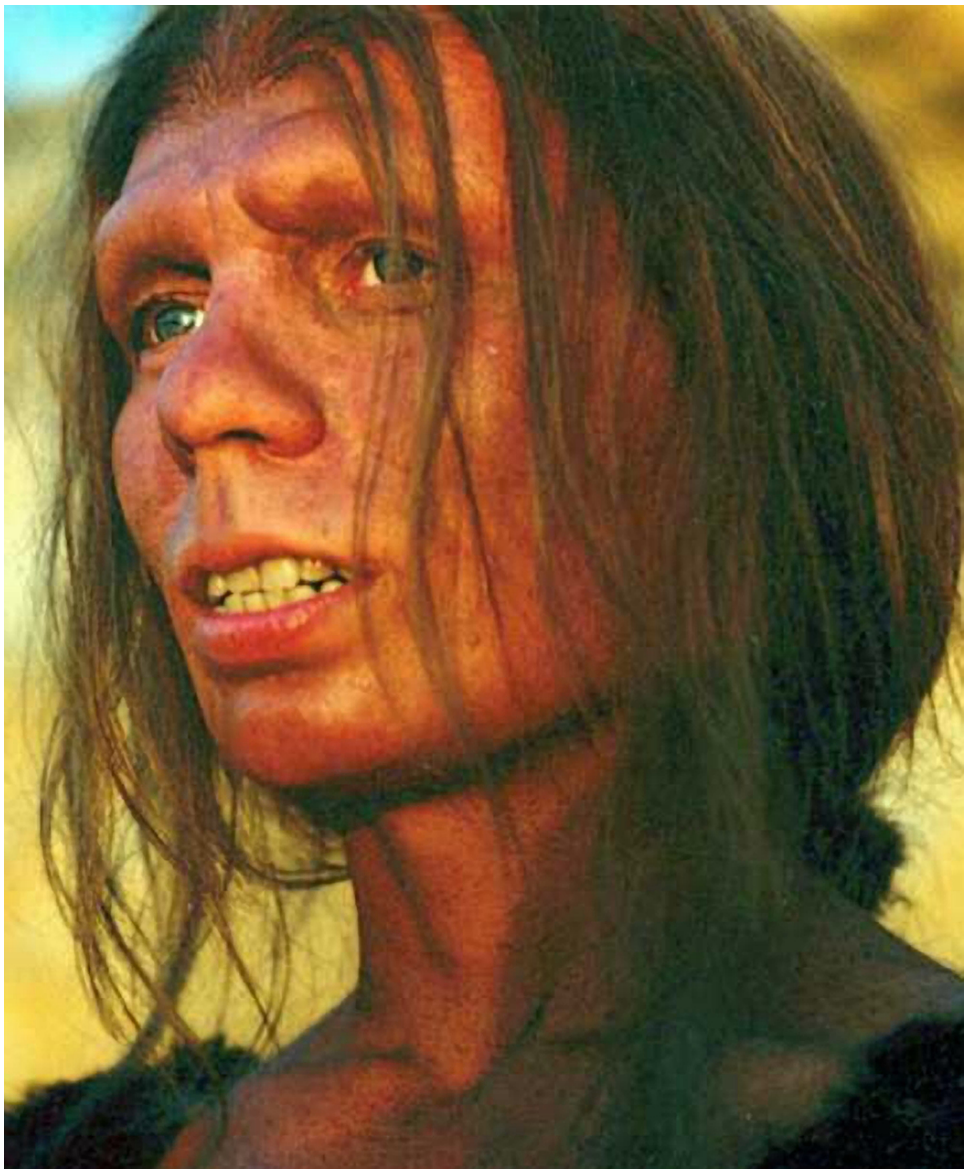


Figure 40.3.19: Reconstruction of the head and face of a Neanderthal woman. (Image: CC BY 2.5; By Bacon Cph, Wikimedia)

https://geo.libretexts.org/@go/page/22846



Figure 40.3.20: Reconstruction of an elderly Neanderthal male. (CC BY-SA 4.0; Pressebilder Neanderthal Museum, Wikimedia.)

The situation with *Homo sapiens neanderthalensis* is quite different. Neanderthals are a subspecies of our own species, and there is ample evidence that they also originated in Africa. They left the continent well before our species, *Homo sapiens sapiens* would begin its own migration (350 ka). Once they left, they migrated throughout Asia and into Europe. Given this timeline, it is possible that they interacted with *Homo erectus* individuals living in Asia at the time, though at present, there is no decisive evidence confirming this. Recent information suggests that ancient close relatives of Neanderthals, called the "Denisovans," lived alongside and interbred with our species for millennia. Known mainly by their unique DNA variants found in modern humans, they outlasted their Neanderthal relatives by at least 10,000 years, only going extinct as

recently as 15,000 years ago. The last documented evidence of a co-existence between Denisovans and modern humans, placed via DNA evidence at 30 ka, comes from New Guinea (Jacobs et al., 2019).

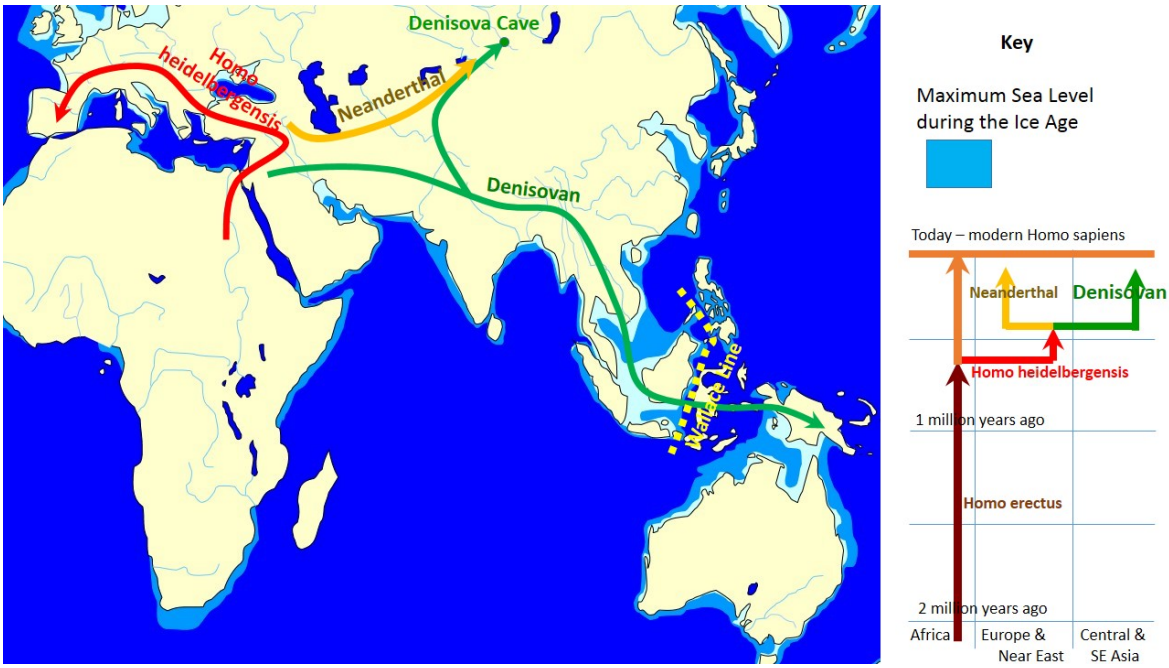


Figure 40.3.21: Spread of Denisovans and Neanderthals after 500 ka. (Image: CC BY-SA 3.0; John D. Croft, Wikipedia.)

Modern humans and Neanderthals did briefly interact in Europe before Neanderthals would also go extinct there. It is not known why Neanderthals went extinct, though a changing climate and the various natures of their interactions between modern humans have been suggested as hypotheses. Some examples of the latter include interbreeding with humans (being assimilated into the *Homo sapiens sapiens* population) and birth problems, disadvantages with humans in the same ecosystems, parasites, pathogens, and violence. One of the last known occupations of Neanderthals in Europe or Asia is found in a cave on the coast of Gibraltar, Spain. Remains there are dated to 42 ka.

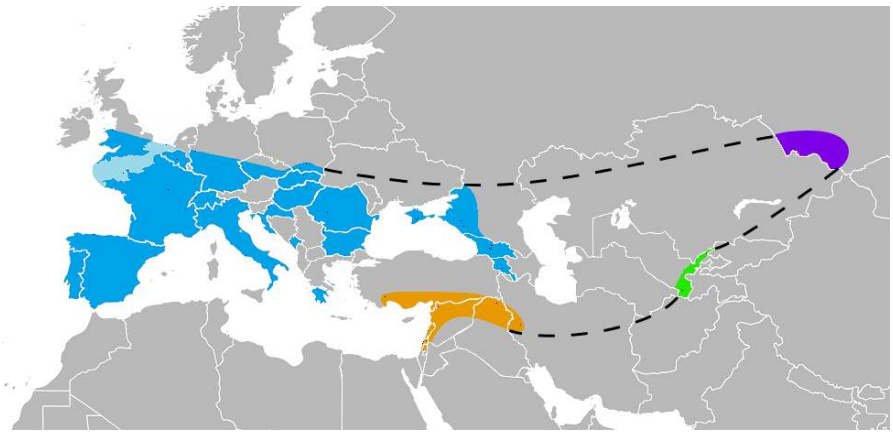


Figure 40.3.22: Known Neanderthal range in Europe (blue), Southwest Asia (orange), Uzbekistan (green), and the Altai mountains (violet), as inferred by their skeletal remains (not stone tools). (Image: CC BY-SA 3.0; Wikimedia)

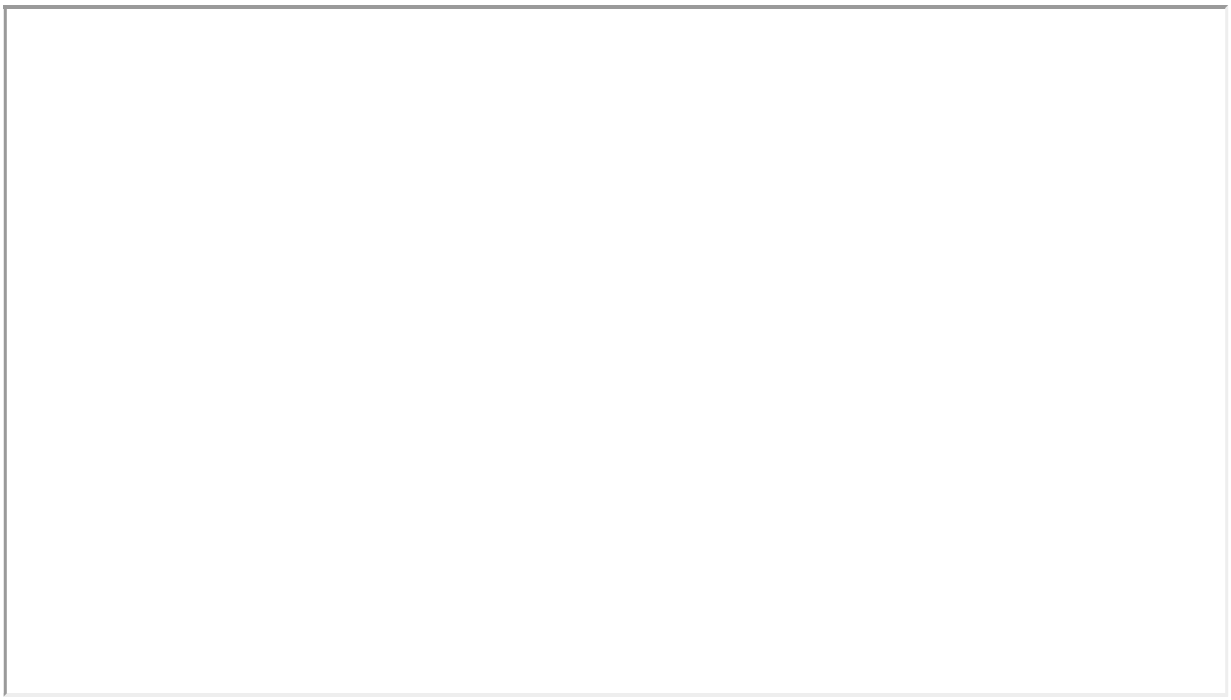


Figure 40.3.23: Upper portion of the skull of a *Homo sapiens neanderthalensis*. All of the features consistent with modern humans are present, save for a thicker occipital ridge. (Model: Creative Commons License; Smithsonian Institution)

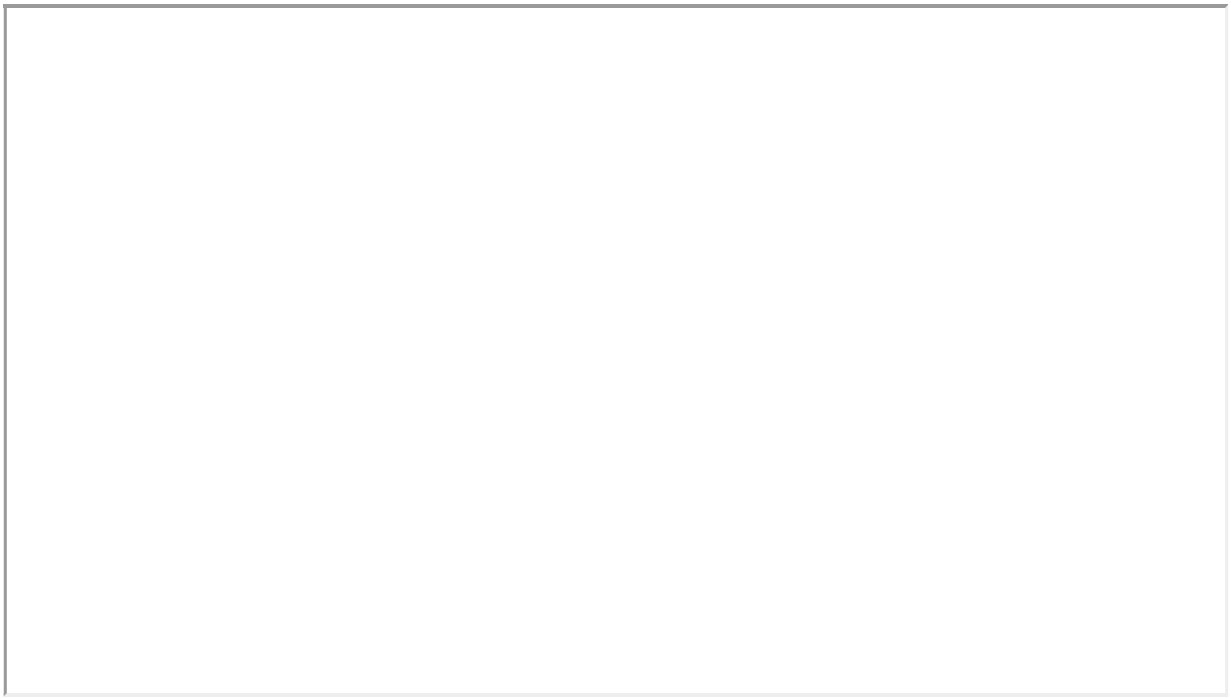


Figure 40.3.24: Neanderthal lower mandible showing dentition that is very consistent with modern humans and a mixed omnivorous diet. Preserved mandibles are more rare in fossil hominids. This mandible is included here as an example of mandible dentition in Neanderthals, as they have lost any pronounced lower jaw canine as well as the associated diastema. (Model: Creative Commons License; Smithsonian Institution)

An entire chapter could be devoted to what we know about Neanderthals. In terms of their culture, we know that they had a diverse toolkit, built shelters, harnessed fire, used clothing, hunted large game (they were apex predators), ate plants, and also participated in ritualistic behaviors. This last point is most significant. No other prior hominid shows the degree of ritualism known among Neanderthals, other than modern humans. An excellent example of this is the inclusion of what appear to be offerings such as flowers for the dead who have been buried.

In lieu of a long exposé on Neanderthals, the video below highlights some additional information that is helpful in helping us understand our own humanity, through the lens of their fossils.



Also unique to Neanderthals is the existence of actual direct DNA samples from several individuals from a wide variety of locations, including Siberia, Croatia, Spain, and Germany. By 2009, the entire known sequence of Neanderthal DNA had been sequenced by the Max Planck Institute for Evolutionary Anthropology, as a part of the Neanderthal Genome Project.

Modern Humans

This case study has been documenting your story and my story. It is a story of our hominid ancestors in deep time. The full story of modern humans is varied, complex, and beyond the scope of this case study. However, there are a few important things to point out:

Like all other hominid species, we emerged from Africa. This migration began between 100 ka and 60 ka, both along the Sinai Peninsula of Egypt and across what is now the Bab-el-Mandeb Strait between Djibouti and Yemen. At the time of migration, the last glacial period was approaching, ice sheets were advancing to the north, and sea levels were much lower. Parts of Africa that were covered with tropical rainforest were giving way to Savannah and even desert, as a part of what is sometimes referred to as a megadrought (Gibbons, 2007). Essentially, our ancestors, *Homo sapiens sapiens*, were driven to greener pastures to the east, north, and further south deeper into Africa.

While there is ample fossil evidence for humans having left Africa at this time and then having migrated throughout Asia, Europe, and into North America (across Beringia as early as 16.5 ka), it is difficult to trace this migration in detail. However, multiple studies have attempted to fill in these gaps using DNA by tracing mitochondrial DNA (on the maternal side) and via the Y-Chromosome (on the paternal side). These studies have traced all modern humans back to a "Mitochondrial Eve" who lived in east Africa about 150 ka. Based on this DNA, maternal haplogroups (individuals who share a common ancestor) can be traced across the world, including into North America. All females, even members of groups who never left Africa, share this mitochondrial signature.

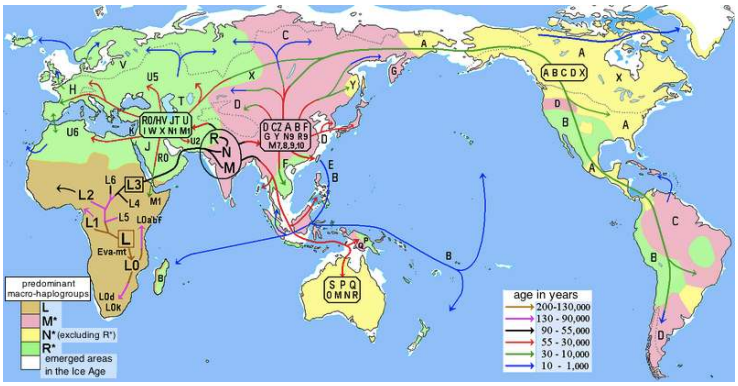


Figure 40.3.25: Hypothesized map of human migrations based on mitochondrial DNA. (Image: CCBY-SA 3.0; Maulucioni, Wikimedia.)

Likewise, paternal haplogroups can be traced (using the Y-chromosome) across the world. In this case, however, all men who share this version of the Y-chromosome are descended from a single man who appears to have migrated out of Africa about 60,000 years ago, likely taking the southern route across Yemen to do so. It's important to note that some human populations, such as members of the !Kung people of Namibia, Angola, and Botswana, do not share this Y-chromosome. This is because they are descended from other paternal line that split away prior to 60,000 years to venture deeper into western Africa.

The map below is an ArcGIS Online layer that illustrates the migration patterns of various maternal (mitochondrial) and paternal (Y-Chromosome) haplogroups among *Homo sapiens sapiens* as they migrated out of Africa. Clicking on the boxes toggles on and off the various haplogroups and you can see where members of that group migrated around the world:

Instructions for how to use this map can be accessed here. (CCBY 3.0; Haplogroup.info.)

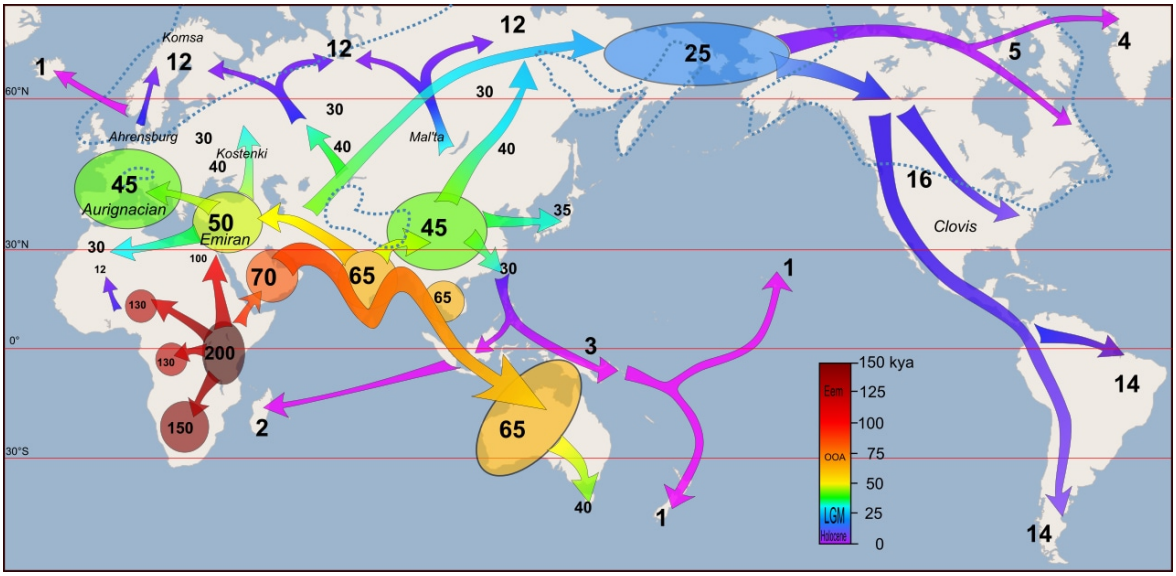


Figure 40.3.26: Peopling of the world (recent out of Africa and Upper Paleolithic). Figures are in thousands of years ago (ka). Time is color coded in a scheme of increasing "frequency", red at 100 ka to violet at 0 ka. Dotted blueish lines are meant to indicate approximate glaciation during the last glacial maximum. (Image: CC BY-SA 4.0; DBachman, Wikipedia.)

We can also trace this migration, particularly the latter portions of it, using an important evolutionary and cultural development: language. Like evolutionary change among isolated pockets of species, language development is rapid, very culturally specific, evolutionary in nature, and is thus a very useful tool for anthropological and archaeologic studies of modern human migration and culture. There are a wide variety of language groups across the world, all of which can be traced by linguists to a handful of primary language components that, combined, reveal aspects of what once may have been a "mother tongue" for that primary language family.

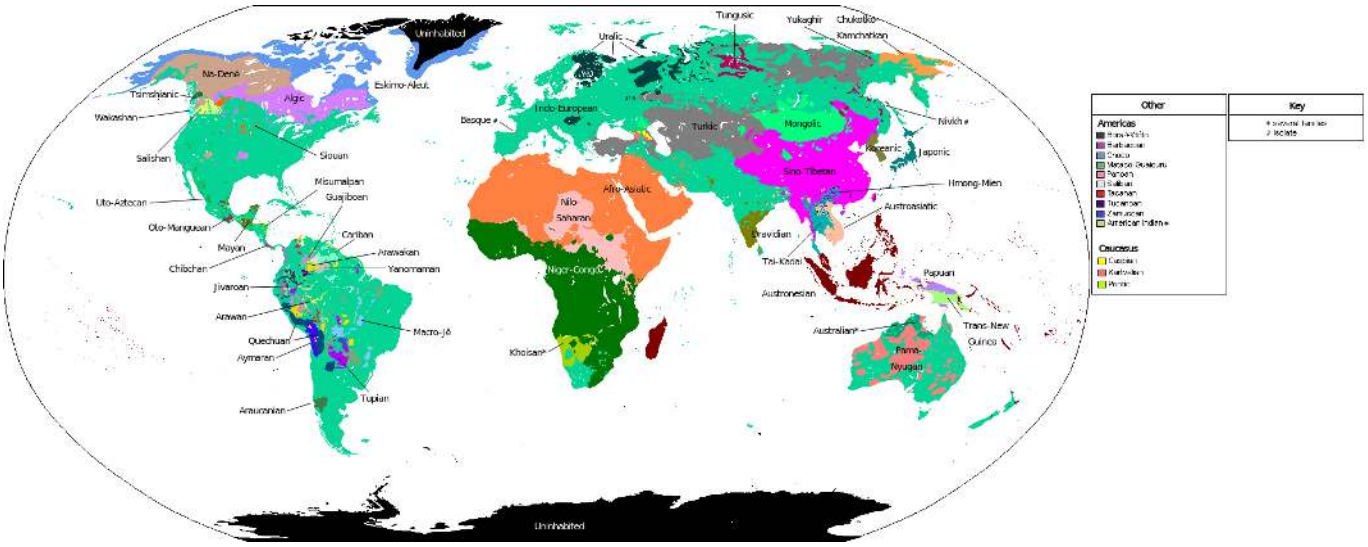


Figure 40.3.27: Contemporary distribution of the world's major language families. Only primary families are included. (Image: CC BY-SA 4.0; Alumnum, Wikipedia)

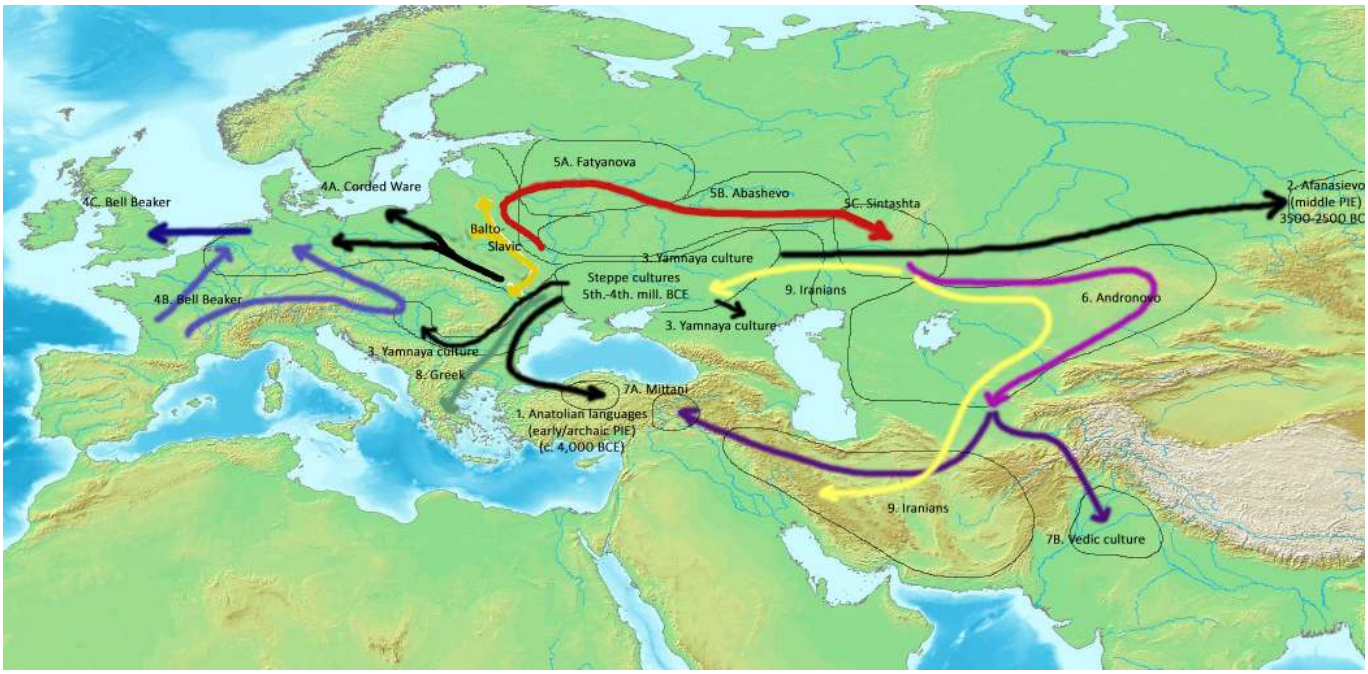
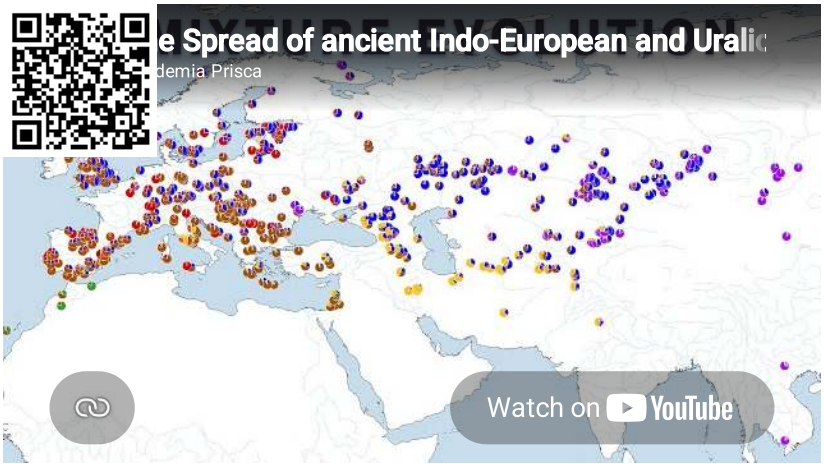


Figure 40.3.28: Scheme of Indo-European language dispersals from c. 4000 to 1000 BCE. (Image: CC BY-SA 4.0; Joshua Jonathan, Wikipedia)

You can view a video animation of the spread of Indo-European and Uralic language group speakers below:



Did I Get It? Haplogroups and Migration

? Exercise 40.3.1

Which of the following genetic markers allows scientists to trace your ancestry back to a single male who lived in Africa about 60,000ka?

a. Y-Chromosome

b. Neanderthal Genetic Variants

c. Bone Tissue DNA

d. Mitochondrial DNA

Answer

a. Y-Chromosome

Homo sapiens, Climate, and Natural Resources

What caused all of this migration? *Homo sapiens sapiens*, like all other animals, seeks to eat, procreate, and survive. Unlike many other animals, our species migrated in ways that go beyond some kind of seasonal round. We have been able to adapt through our use of tools, our diet, and in the pursuit of resources. Today, our species lives far differently (in most cases) than our ancestors did thousands of years ago and, for most of us, we live very differently than our recent ancestors just a century or even a few decades ago. What has driven these changes?

A significant part of that answer lies in climate change. Before we actively played a role in altering our climate, as we do today, our climate dictated where we called home to a large extent. Climate change contributed as a trigger for migration in many cases. Despite this, our ancestors did brave much colder climates and the adaptability challenges those climates brought with them. Technological advances have occurred irrespective of just those driven by climate, however. While certainly adapting to colder areas required innovation of our toolkits, we innovated these kits in warmer climates also. The Stone Age would give way to Neolithic technologies (agriculture, animal husbandry). To these new Neolithic ways we would eventually add our harnessing of metals such as copper, bronze, iron, and eventually silicon (today). All of this innovation requires further resource extraction. Fundamentally, this resource extraction is similar to that practiced by our microbial ancestors over a billion years ago. As with simpler organisms, all human uses of our environment have been toward the goal of providing our species with food, shelter, and a chance to reproduce.

We're animals, after all.

Did I Get It? Humans and Climate

? Exercise 40.3.1

Humans and climate change have a long history together. How has climate change significantly affected our hominid ancestors?

a. Rapid evolutionary changes

b. Forcing migrations

c. Mass extinctions

d. Major cultural shifts

Answer

b. Forcing migrations

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40.4: Further Reading

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CHAPTER OVERVIEW

41: (Case Study) Big Bang (coming soon)

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46: (Case Study) Origin of life (coming soon)

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47: (Case Study) A history of fossil plants (coming soon)

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48: (Case Study) Sauk/early Paleozoic transgression (coming soon)

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49: (Case Study) The Acadian Orogeny (coming soon)

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50: (Case Study) The Alleghanian Orogeny (coming soon)

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CHAPTER OVERVIEW

51: (Case Study) Large igneous provinces / Flood basalts (coming soon)

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52: (Case Study) Pleistocene "Ice Ages" (coming soon)

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53: (Case Study) Epicontinental seas - Pz, Mz, Cz (coming soon)

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54: (Case Study) Carboniferous coal and natural gas - cyclothems (coming soon)

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55: (Case Study) PETM (coming soon)

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CHAPTER OVERVIEW

56: (Tools of the Trade) Earth Materials - Mineral identification

56.1: Mineral Properties and Identification

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56.1: Mineral Properties and Identification

Geologists identify minerals by their physical properties. In the field, where geologists may have limited access to advanced technology and powerful machines, they can still identify minerals by testing several physical properties: luster, color, streak, hardness, crystal habit, cleavage and fracture, and some special properties. Our Big Ten Minerals make up the majority of Earth's lithosphere and are usually seen as small grains in the rock. Geologists rely on a magnifying hand lens as an essential tool for identifying most minerals in the field. Of the many properties that minerals can exhibit, we will consider only those most useful for identification as small grains surrounded by other minerals in rock.

Luster and Color

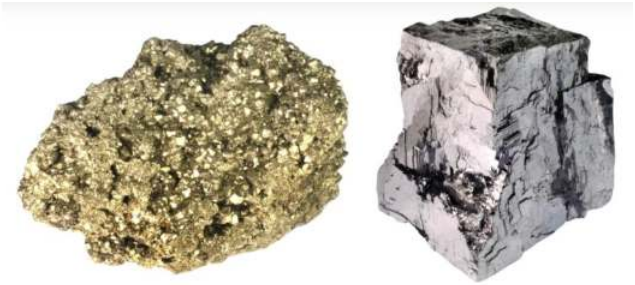


Figure 56.1.1: Metallic Luster

The first thing to notice about a mineral is its surface appearance, specifically luster and color. Luster describes how the mineral reflects light. Metallic luster looks like a shiny metal such as chrome, steel, silver, or gold. Metallic minerals are also always opaque. Non-metallic minerals may be shiny and reflect light, however, they do not look like a metal.



Figure 56.1.2: Nonmetallic Luster

Surface color may be helpful in identifying minerals, although it can be quite variable even within a single mineral family. Mineral colors are affected by the main elements present, but also by impurities, inclusions, and defects in the crystal structure. Pressure and temperature conditions can also affect mineral color. Using the mineral quartz for example, in its purest form (SiO_2), it will be clear. However, impurities by trace amounts of iron can change the color from purple (amethyst) to yellow (citrine) or pink (rose quartz); tiny inclusions of the titanium mineral rutile can give quartz a blue color, and exposure to natural radiation within the crust can provide a grey (smoky) color.

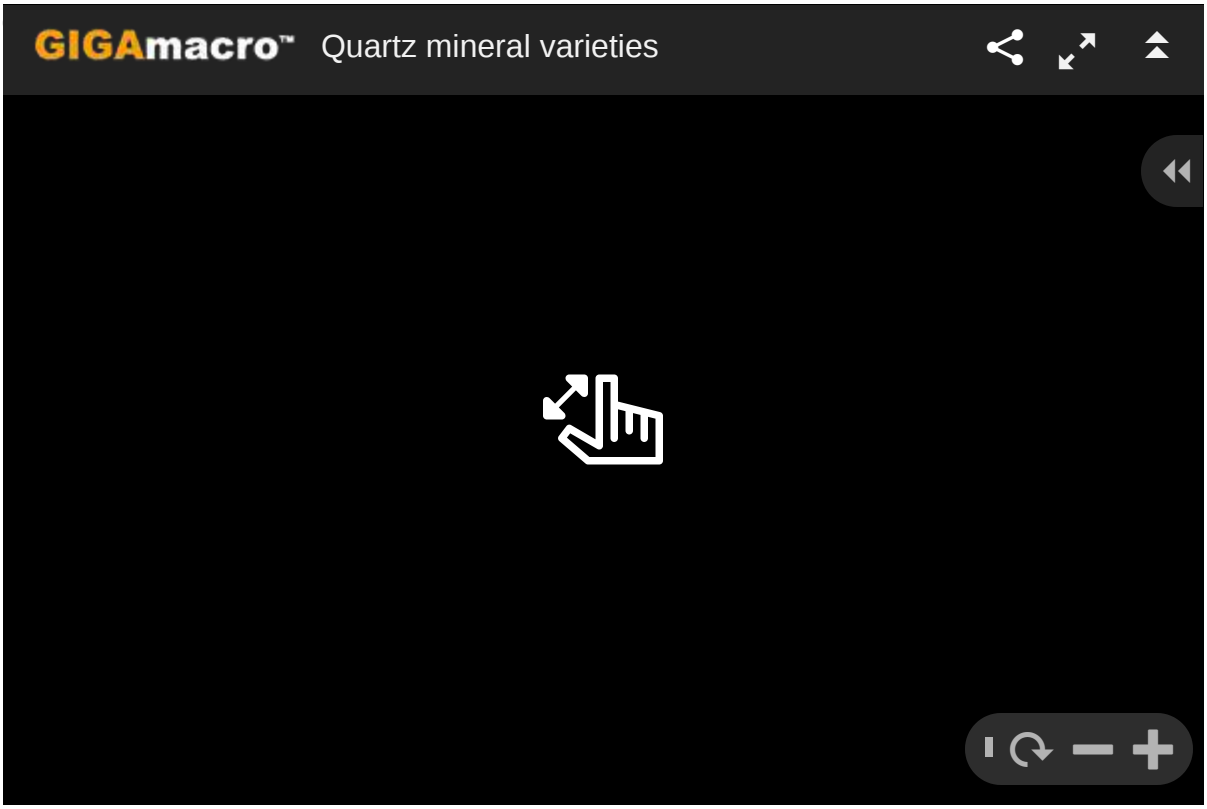


Figure 56.1.3: All of these samples are quartz. Pure quartz (SiO2) is the clear sample on the left-hand side; followed by smoky quartz, rose quartz, and amethyst quartz. Each of the colored quartz samples contains an impurity in the chemistry. (Credit: CC BY Attribution 3.0; Robin Rohrback, Mid-Atlantic Geo-Image Collection (M.A.G.I.C.) on GigaPan.)

There are some minerals that predominantly show a single color. However this is not true for our “Big Ten Minerals” or the other most common rock forming minerals. All of these minerals have a predictable range of colors due to element substitution within their crystalline structures. Feldspars, the most abundant minerals in the Earth’s crust, range in composition from calcium-rich to sodium-rich to potassium-rich varieties and many compositions in between. Feldspar colors including pink, white, green, gray and others. With notable exceptions, color is usually not a reliable diagnostic property of minerals. For identifying many minerals, a more reliable indicator is streak, which is the color of the powdered mineral.

Streak

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

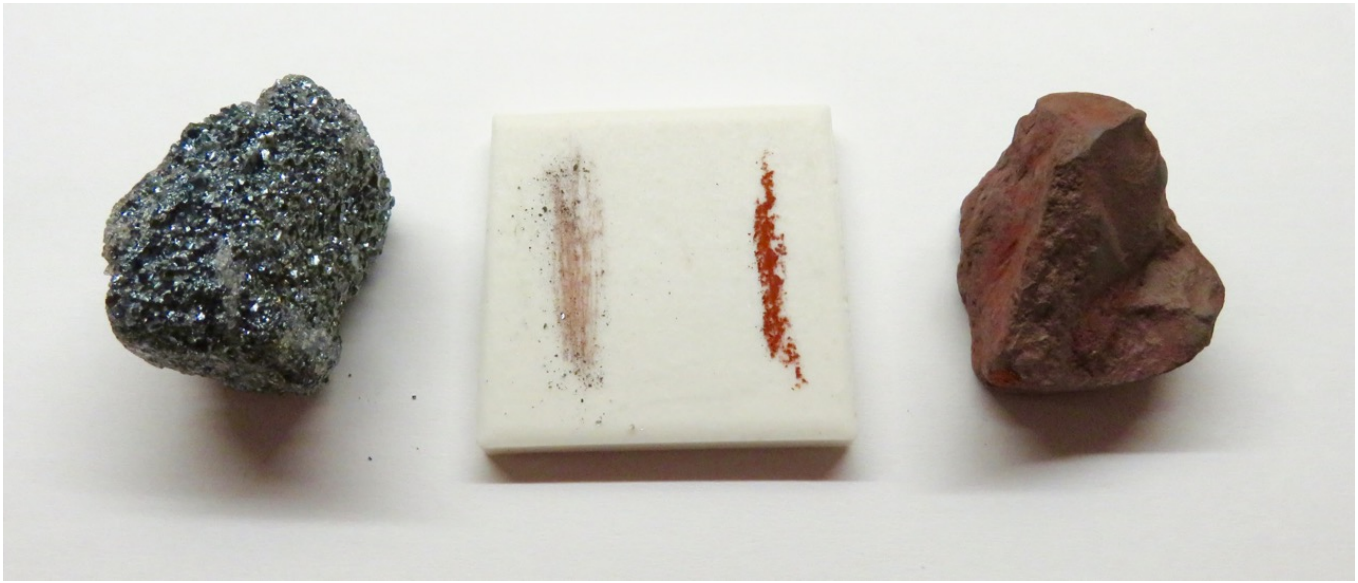


Figure 56.1.4: Hematite leaves a reddish brown streak on a ceramic plate whether the hematite has a metallic lustre or an earthy lustre. Hematite samples from the mineral collection at St. Peter's College, Muenster, SK, Canada. (licensed under the Creative Commons Attribution-Share Alike 4.0 International license; By: Karla Panchuk.)

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

Hardness



Figure 56.1.5: Mohs scale of hardness including the hardness of some common items. (Credit: Public Domain; United States National Park Service from: https://www.nps.gov/articles/mohs-hardness-scale.htm.)

Hardness measures the resistance of a mineral to abrasion. Mohs Hardness Scale assigns a number between 1 and 10 indicating the relative scratch-resistance of minerals when compared to a standardized set of minerals of increasing hardness. Mohs scale was developed by German geologist Fredrick Mohs in the early 20th century, although the idea of identifying minerals by hardness goes back thousands of years. Mohs hardness values are determined by the strength of a mineral's atomic bonds. The figure shows the minerals associated with specific hardness values, together with some common items readily available for use in field testing and mineral identification. A steel pocket knife blade, which has a hardness value of 5.5, distinguishes between hard and soft minerals on many mineral identification keys.



Figure 56.1.6: Cleavage

Minerals often show characteristic patterns of breaking along specific cleavage planes or show characteristic fracture patterns. Cleavage planes are smooth, flat, parallel planes within the crystal. The cleavage planes may show as reflective surfaces on the crystal, as parallel cracks that penetrate into the crystal, or show on the edge or side of the crystal as a series of steps like rice terraces. Cleavage arises in crystals where the atomic bonds between atomic layers are weaker along some directions than others, meaning they will break preferentially along these planes.



Figure 56.1.7: Fracture

Because they develop on atomic surfaces in the crystal, cleavage planes are optically smooth and reflect light, although the actual break on the crystal may appear jagged or uneven. In such cleavages, the cleavage surface may appear like rice terraces on a mountainside that all reflect sunlight from a particular sun angle. Some minerals have a strong cleavage, some minerals only have weak cleavage or do not typically demonstrate cleavage. But, don't be fooled: Among mineralogists, a common saying is that "cleavage in a mineral is great, when you see it. But if you don't, do not assume it is not there." Fracture is different than cleavage in that the broken surface will not be flat or even. The breakage does not occur along a distinct plane within the crystal lattice. Fracture is typically uneven however can also be very distinctive. Several common minerals, quartz and olivine, will often display a conchoidal fracture similar to that seen in obsidian or broken glass.

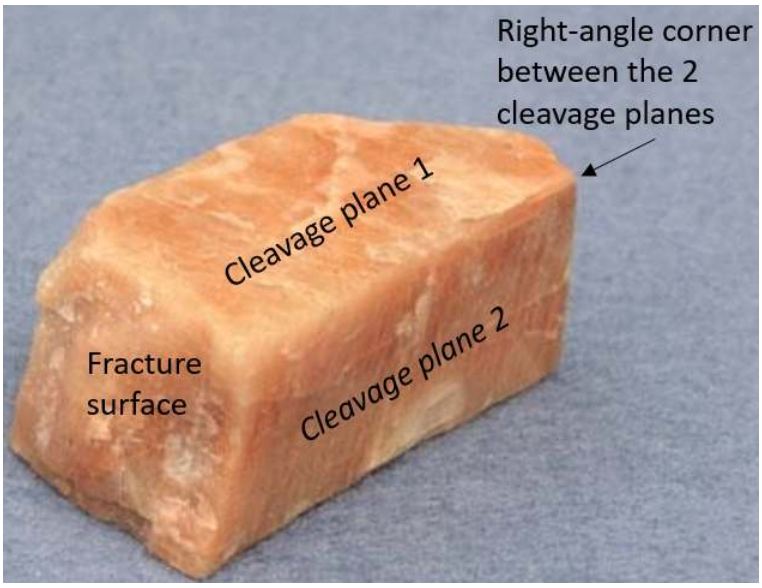


Figure 56.1.8: Potassium feldspar displaying two planes of cleavage and one irregular fracture surface. Steven Earle. (licensed under: Creative Commons Attribution 4.0 International License; https://opentextbc.ca/physicalgeology2ed/chapter/3-3-crystallization-of-magma/)



Figure 56.1.9: Conchoidal fracture displayed in smoky quartz. Width of sample 11cm. (For educational purposes by Siim Sepp from https://www.sandatlas.org/conchoidal-fracture/)

Special Properties

Special properties are unique and identifiable characteristics used to identify minerals or that allow some minerals to be used for special purposes. A simple identifying special property is taste, such as the salty flavor of halite or common table salt (NaCl).

A simple test for identifying calcite and dolomite is to drop a bit of dilute hydrochloric acid (10-15% HCl) on the specimen. If the acid drop effervesces or fizzes on the surface of the rock, the specimen is calcite. If not, and dolomite is suspected, the specimen can be scratched to make a small amount of powder. If the acid drop fizzes slowly on the powdered mineral, the specimen is dolomite. The video below compares these two minerals and their effervescence reaction. Geologists often carry a small dropper bottle of dilute HCl in their field kit.



Some iron-oxide minerals are magnetic and are attracted to magnets. Common magnetic minerals include magnetite (Fe_3O_4) and ilmenite ($FeTiO_3$). Magnetite is strongly attracted to magnets. Ilmenite and some types of hematite are weakly magnetic. Geologists often carry a free-swinging magnet in their field kit to test for magnetism.

Review the following video to learn more about the basic properties of the rock forming minerals.



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CHAPTER OVERVIEW

57: (Tools of the Trade) Evolution Part II - Paleontological applications



Figure 57.1: An exquisite speciment of Cryptolithus bellulus from the Alexandria Submember of the Kope Formation, Cincinnati, OH (Kohrs, 2003).

Key Concepts

At end end of this chapter, students should be able to:

- Describe the development of the use of fossils as evolutionary tools.
- Identify major patterns seen in the fossil record as a result of evolutionary change.
- Describe the basics of cladistic analysis using fossils.
- Understand the connections between speciation and extinction.
- Identify several important mass extinction events in the Earth's past.
- Discuss the difference between background extinction and mass extinction.

> "The theory of punctuated equilibrium, proposed by Niles Eldredge and myself, is not, as so often misunderstood, a radical claim for truly sudden change, but a recognition that ordinary processes of speciation, properly conceived as glacially slow by the standard of our own life-span, do not resolve into geological time as long sequences of insensibly graded intermediates (the traditional, or gradualistic, view), but as geologically "sudden" origins at single bedding planes."
>
> — *Stephen J. Gould, "Time's Arrow, Time's Cycle", 1987*

57.1: Changing Perspectives on Fossils - Evolution's Silent Sentinels
57.2: Mass Extinctions of The Past and Possible Causes
57.3: The Fossil Record - Making Sense of Evolution's Past
57.4: Further Reading

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57.1: Changing Perspectives on Fossils - Evolution's Silent Sentinels

Figured Stones

Fossils are evidence of ancient life. However, they were not always seen as such. The evolution of our understanding of these figures in stone is a classic example of how science progresses with new knowledge. Since the time of the ancient Greeks, people have noticed seashells in stones far from shore and well above current sea level. Greek writers including Xenophanes, Herodotus, Strabo, and Eratosthenes interpreted them being left behind by the ebb and flow of the sea. They also viewed the Earth as eternal. Perhaps because later Christian thinkers saw this view of the Earth as too Pagan, in conflict with the Biblical account in Genesis, these views eventually fell away from acceptance in Western Europe. Still, people wondered about the odd shapes in local rocks, referred to at the time as "figured stones". Today, we know them as fossils.

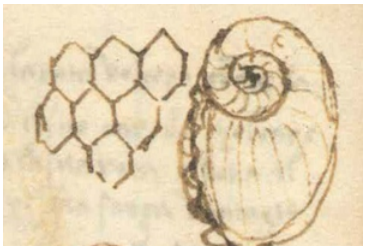


Figure 57.1.1: Drawing of Palaeodytion specimen, Leonardo da Vinci, Paris Manuscript I

A drawing of fossil structures, *Palaeodyction*, from a "figured stone" by Leonardo Da Vinci, from his unpublished *Paris Manuscript I*.

They saw them contained within the building blocks of cathedrals, along pathways, and in the mountains. Fossils of marine shellfish, Long pulled from the limestones of the Alps, were not seen as evidence of former life despite their similar appearance with living shellfish of the time. Or if they were originally from the sea, the preferred interpretation was that they were a remnant of the global deluge, a result of the deposition of organisms along a globally flooded, and eventually inundated land surface. Others interpreted the figures in the rocks not as evidence of flood deposits but as having been placed there during creation itself. Divinely created as is, they were considered to be miraculous. Still the ultimate nature of fossils remained elusive to most, until Nicholas Steno (1638-1686), a Danish scientist, had another idea.

Steno was trained as an anatomist. He also had many other interests. Cutler (2003) recounts Steno's life, beginning with his famed skill at dissection. These skills led him to many discoveries, including the salivary glands in mammals. This fame led him to the Medici court in Florence, Italy, where he not only found financial support, but also was given the opportunity to study anything he wanted. This freedom of curiosity and creativity would lead eventually to his geological explorations. It would be these geological discoveries, rather than anatomical insights, that would make Steno famous as the "father" of geology.

Steno was curious about the "figured stones". Some contained what looked like seashells while still others contained structures that the local people called "tongue stones" (Figure 57.1.2). After studying in detail the head of a shark specimen reeled in by two fishermen in 1666, Steno would make the connection between the similar shape of that shark's teeth and the "tongue stones" found encased in rocks nearby. He was the first to publish accurate interpretations regarding how such a "solid within a solid" could be possible, suggesting that these tongue stones represented evidence of ancient life: tongue stones were the teeth of long-dead sharks! He argued that the encased solid (the shark tooth in this case) had to first exist on its own, prior to being encased. This line of reasoning is an early form of his "principle of relative dating by inclusions", that is still utilized today. His recognition of these "figured stones" as fossils would lay the groundwork for a rebirth of thought about Earth's past that would carry into the minds of enlightenment thinkers just decades later. As scientific understandings of observation and experimentation would advance, so too would our appreciation of fossils as evidence of the history of life itself, including our species' own backstory.

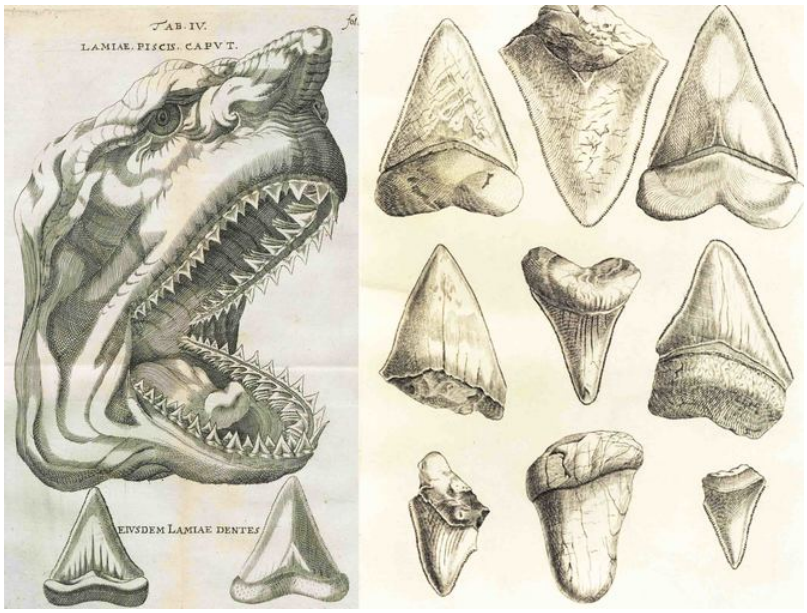


Figure 57.1.2: Woodcut of figured stones from Nicholas Steno's 1667 monograph on the dissection of a Great White Shark. Similar stones, found embedded in rocks in the northern Italian Alps, Steno identified as sharks' teeth. From "Canis carchariae dissectum caput" (Steno, 1669)

Figured Stones to Fossils: Evolutionary Foundations

Other scientists of Steno's time would fully accept fossils as being evidence of ancient life. One such contemporary was Robert Hooke (1635-1703), who studied fossils under the microscope. While conducting this work, he noticed similarities between features in certain fossils and extant organisms. He noted that mineralization could occur between wood fibers, producing permineralized wood. He also determined that shells could be turned into fossils after being filled with mud or clay. After Steno, two key individuals would emerge whose work and ideas would lay the groundwork for the use of fossils toward understanding past life.

The first was Carl von Linné (a.k.a. Carolus Linnaeus; 1707-1778). Linneaus was interested in fossils, and while he did not recognize them yet as evidence of prior life, he did acknowledge they were distinctively different than other kinds of rocks and minerals. Linnaeus' original classification system included three kingdoms. These were *Regnum Animale* (Animal), *Regnum Vegetabile* (Vegetable), *and Regnum Lapideum* (Mineral). While these labels are still used today, our current classification schemes for animals, plants (vegetable), and geological materials (rocks and minerals) are very different. In Linnaeus' Mineral Kingdom, he did take pains to do what others before him struggled with. He separated fossils out as being different from rocks. He did not call them evidence of ancient life, but he recognized that they were different in some way.

The second important paleontologist of this time was George Cuvier (1769-1832). He fully accepted fossils as being evidence of ancient life. Because Cuvier saw fossils as ancient life, he was the first to include them in the Linnaean system along with living forms. During his life he collected a massive amount of fossil material from the New World. Using this collection, he would eventually argue for the existence of extinction through his study of megafaunal fossils, in particular. The extinction of an entire species was not thought to be possible, or at least common, in nature. Cuvier's revolutionary ideas surrounding extinction, supported by the fossil record, represent a paradigm shift in how fossils were viewed. Not only were they evidence of ancient life, but they were also evidence of wide-scale ancient death. And, extinctions make way for evolutionary change.

Evolutionary Patterns in the Fossils

PATTERN 1: INHERENTLY BIASED DATA

Evolution is well supported in the fossil record. Even so, the limitations of this record create significant challenges for understanding the history of extinct life. For instance, using cladistics to organize known fossils into clades based upon shared and derived characteristics is very useful. However, only a small percentage of life is fortunate enough to come down to us as a fossil, allowing its story to be told. Significant gaps, or missing data, occur in the record. This bias in the record tends to be against taxa

with no hard skeletal parts, those that are generally fragile, and those that live in habitats where rapid burial is unlikely or other circumstances are not ideal for fossilization. Within the time range of a species, there are great uncertainties about its origination and its extinction, when it appears and when it dies off. In fact, these exact moments in the life of a species are entirely impossible to determine. The cause of this effect is a sampling bias, related to a variety of causes, but most critically to the overall problem that not everything gets fossilized and, what has been fossilized is not all that has been found. Philip Signor and Jere Lipps (1982) described this sampling bias, now referred to as the Signor-Lipps Effect, because the oldest appearances of a given taxon appear to come on gradually just as the youngest appear to die away gradually, even if the extinction (or origination) itself was actually quite sudden.

The Cretaceous/Paleogene extinction event that ended the reign of reptiles is an excellent case study for this challenge (Alvarez et al., 1980; Signor and Lipps, 1982). Until Walter and Luis Alvarez identified an asteroid as a cause of this geologically sudden extinction, mass extinctions were thought to come on quite gradually because of the apparently slow disappearance of fauna in uppermost Cretaceous strata. The Signor-Lipps Effect signal in the rock record below the extinction boundary suggested that it was gradual – a matter of relative appearance. Their work showed otherwise and so proved that such events can happen suddenly, despite what the record of fossils might appear to be showing.

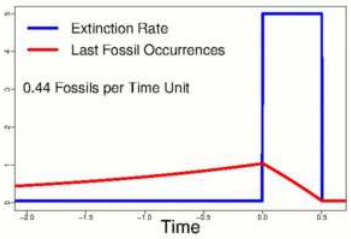


Figure 57.1.3: An animation of the Signor-Lipps effect, demonstrating how the rate of last fossil occurrences approximate the extinction rate as the number of fossils per time unit increases (Wikipedia: NikelsenK)

PATTERN 2: FOSSIL SUCCESSION

Because the fossil record shows evolution occurring through time, we can reconstruct the story of life. Cladistic analysis, the discovery of transitional fossils, and the abundance and variety of organisms that have been discovered all support the evolutionary story of descent with modification. Because of this, when we find a rock with a particular fossil in it, we can assign a relative age to it because we have a good understanding of the succession, or sequence, of organisms in the record. Biologist J.B.S. Haldane (purportedly in a conversation with Paul Harvey in 1970), once asked how his confidence in the theory of evolution could be challenged, replied, “fossil rabbits in the Precambrian”. There is little concern that such an outlandish discovery would, however, negate evolutionary theory. Rather, such a radical change in our understanding of fossil succession would merely cause scientists to re-evaluate timelines, originations of species, and other associated evidence toward a recasting of the theory in the light of such new knowledge.

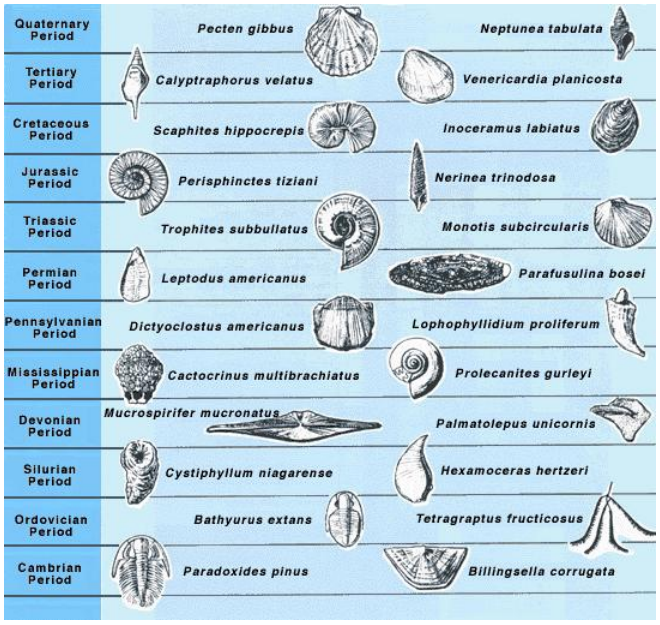


Figure 57.1.4: Faunal Succession through geologic time. Certain fossils are widespread in certain time intervals and can come to define them. Over time, they represent evolutionary change. (USGS)

One excellent example of how succession works is with trilobites. These classic and important fossils among invertebrates were a key part of the "Cambrian Fauna" (Sepkoski, 1984). Originating most abundantly in the Cambrian and truly flourishing through the Ordovician Period, they went extinct at the end of the Permian Period, victims of the largest mass extinction in Earth's history. Because of this, we find no examples of trilobites prior to their origination during the latest Precambrian and none after the start of the Triassic. Knowing the time of their origination and the time of their extinction and the ability to reconstruct the clades within this fossil group's phylogeny provides a variety of key index fossils that can be used to identify rocks of particular ages, all because scientists have been able to reconstruct their evolutionary history. For instance, *Paradoxides pinus* is a trilobite typical of the middle Cambrian while the *Isotelus maximus* is characteristic of the middle to late Ordovician. The presence of these specimens in a rock is key evidence that the host rocks were deposited during these periods of time.

PATTERN 3: TRANSITIONAL FOSSILS

Transitional fossils are fossilized remains that share ancestral characteristics ***and*** those of the derived group of descendants. By this definition, and considering how rapid cladogenesis can occur, there can be a wide interpretation of exactly what constitutes a transitional fossil. Some workers would say that all fossils are transitional fossils! Generally the idea is invoked only between major groups. Table 1 illustrates a wide variety of these, but the classics that come to mind most rapidly for most are *Archaeopteryx* sp., an organism that is transitional between non-avian dinosaurs and birds (Dengler, 2018), *Pakicetus* sp., a transitional organism between land mammals and whales (Gingerich et al., 2018), and *Tiktaalik*, an organism that shares the characteristics of lobe-finned fish while also displaying the derived characters seen in later terrestrial vertebrates (Carroll, 2016).

Fossil Group or Clade	Organism	Features	Image	Between:	Period
Avialae	*Archaeopteryx lithographica*	Jaws with sharp teeth, Long bony tail, Feathers		Non-Avian Dinosaurs and Birds	Late Jurassic (150 Ma)
Stegocephalia	*Tiktaalik roseae*	Scales, Gills, Flattened head, Sturdy interior fin bones, Ears capable of hearing in both air and water		Fish and Amphibians	Late Devonian (375 Ma)
Hominin	*Ardipithecus ramidus*	Grasping big toe, Reduced canine tooth size, bipedal features in the pelvis and base of skull		Common ancestor of chimpanzees and humans	Late Miocene (4.4 Ma)
Epicynodontia	*Thrinaxodon liorhinus*	Possibly had hair, secondary palate, Nasals pitted with foramina, inner ear bone		Synapsids to Mammals	Early Triassic (250 Ma)
Cetacea	*Basilosaurus cetoides*	First fully aquatic mammal, nearly complete loss of hindlimbs		Land Mammals to Whales	Late Eocene (40 ma)

PATTERN 4: PALEOBIOGEOGRAPHY

One of the major factors in fossilization that is important to consider when contemplating the record is that not all organisms lived in every environment and everywhere around the world. As is true of most organisms today, even humans, geographic distribution of organisms is determined by the environments to which they have adapted. There have been fossils, such as the *Glossopteris* sp. ferns



Figure 57.1.5: Glossopteris sp. leaves from the Permian of Australia.

and *Paradoxides* sp. trilobites that have been found in myriad locations, apparently far apart from one another. These far flung fossil sites have helped reconstruct paleogeography (positions of continents) because of the abundance of specimens and similarity of environment.



Figure 57.1.6: The trilobite Paradoxides davidus.

Because of our understanding of modern ecology and evolutionary principles, we can go beyond just describing fossils and organizing them by appearance, we can also apply biogeographic understandings to place them within ecosystems and, sometimes, even describe communities within those ecosystems. At times, there are very interesting stories that emerge when examining the paleobiogeography of an organism or group, when epiboles or outages (sudden appearances and disappearances) occur, giving us insights into the stability and resilience of long-extinct organisms and environments.

For further exploration of paleobiogeographic material, the Paleobiobology Database Navigator is an amazing tool.

PATTERN 5: HOMOLOGY, ANALOGY, AND VESTIGIALITY

Humans are tetrapods, a trait we share with all four-legged organisms going back to the earliest amphibians and fish. Such homology, or similarity in structures, supports cladistic analysis. Organisms with similar developmental biology, or embryology, will tend to develop the same structures. However, through evolutionary change, or descent with modification, these same structures can be adapted for different uses.Structures within a lineage, such as the arm bone sequence within tetrapods (one upper arm bone, two lower arm bones, many hand bones), are considered to be homologous structures. These shared characteristics, derived from a much earlier ancestor, are key identifiers within a lineage. For our own species, the arm bone sequence is a classic case that connects us to the earliest tetrapod, *Tiktaalik rosaceae*, the earliest freshwater organism to carry this trait onto land that we know of from the fossil record. Homology provides critical insights into how heredity works. Thus, as Pierre Belon noticed in 1555, humans and birds have similar structures in their arms. Birds being adapted to flight, their wings are otherwise useless for the manipulation of tools, climbing trees, or some of the other uses for which a primate descendent might use them. Homology implies divergent evolution, or the splitting off of lineages through the processes of speciation, because advantageous traits develop in different ways for different organisms.

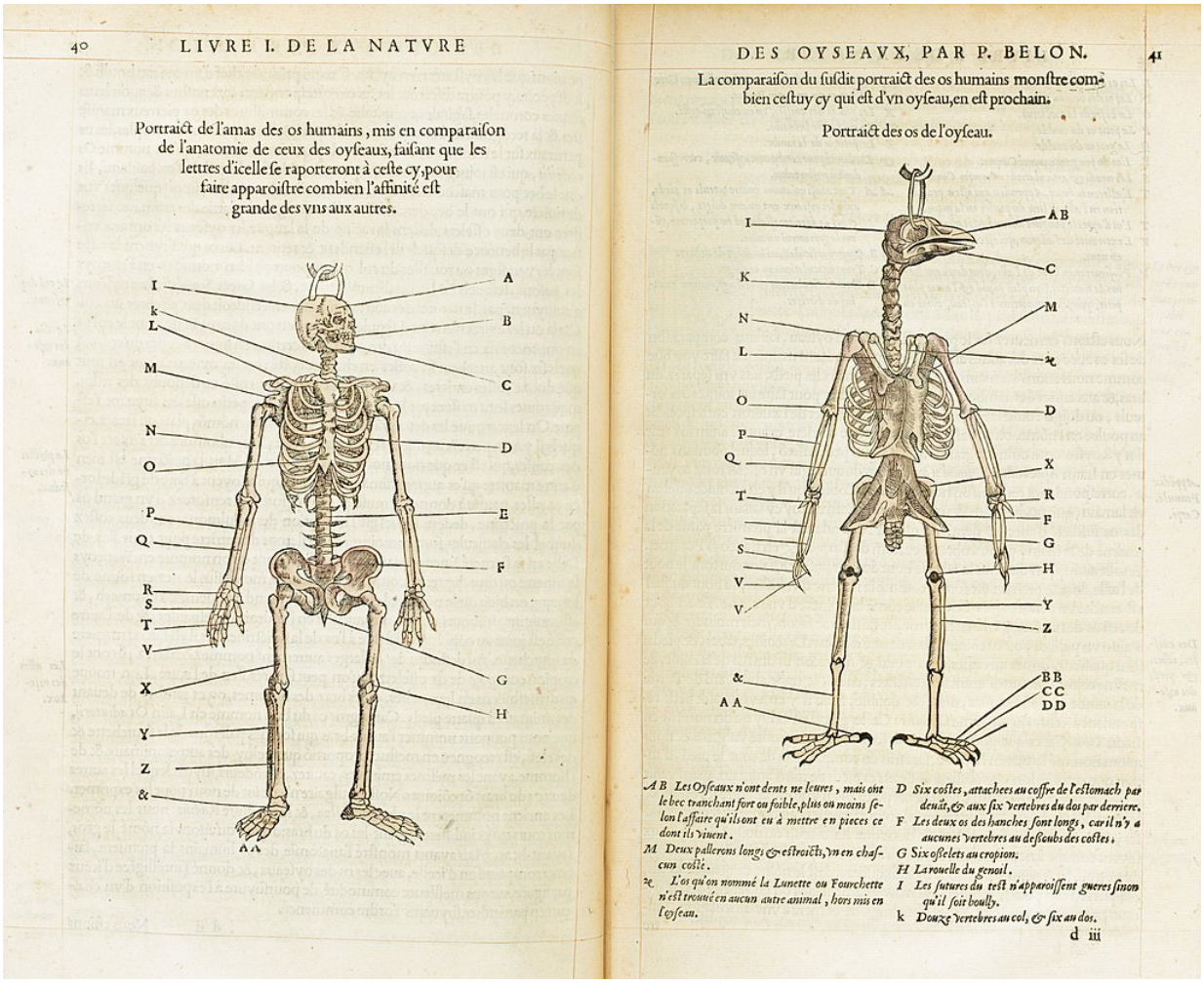


Figure 57.1.7: Pierre Belon's early attempt at an analysis of homologous traits in his book, Natural History of Birds, 1555.

Not all similar structures are homologous. Some structures are analogous. Wings are an excellent example, for while birds have wings as mentioned above, so do butterflies. The reason bird and butterfly weeks are analogous is that these structures evolved independently of one another. These analogous structures are examples of convergent evolution, or the independent evolution of similar structures among different taxa that, can also share a similar adaptational purpose, such as wings assisting with flight in the case or birds and butterflies. There are many other examples of convergent evolution and analogy. Consider the similar torpedo-shaped swimming forms of tuna, icthyosaurs, dolphins, and penguins. Some of the best examples occur among plants, such as the over 60 independent instances of C4 photosynthesis, one of several photosynthetic pathways for producing sugars, that evolved among 19 different families of plants over time (Osborne and Sack, 2012), wind-blown seed dispersal, and carnivory.

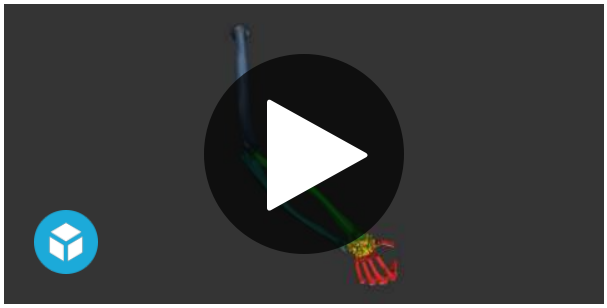
Vestigial Structures

Finally, it is worth considering vestigial structures. These are anatomical features, cladistically shared with ancestral taxa, that serve little to no function purpose or can even be maladaptive. Such structures can be hard to identify but are not uncommon. They are the result of a lack of positive selective pressure, meaning that the structure no longer provides a function that leads to an evolutionary advantage. These structures, sometimes also referred to as vestigial organs, can be found in an atrophied, or reduced form, in many instances. A classic example is the loss of sight in blind cave fish. Living a life in total darkness, these fish no longer have need to expend energy maintaining a sensory organ evolved for light detection. While they retain eyes, they are no longer functional. Other examples of organisms retaining structures that are mere vestiges of their past usefulness are wings in some island-dwelling birds, such as penguins or Galapagos cormorants, or hindlimbs that are retained in whales and snakes.

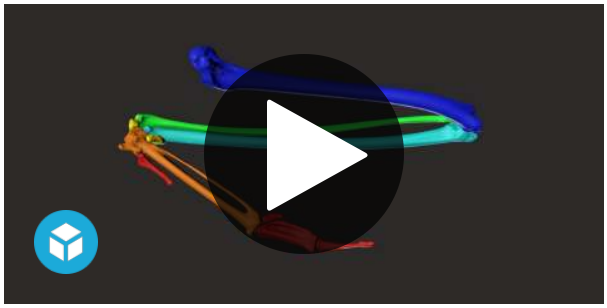
Figure 57.1.8: Example of whale evolution. Ambulocetus with four legs evolved into Basilosaurus, who retained a vestigial hind limb. The modern descendant, the sperm whale, retains a rudimentary hind limb.

Care must be taken in labeling structures definitively as vestigial for the reason of exaptation (Gould, 1982), a kind of re-adaptation of an evolved structure for a different purpose. Feathers are an example of a feature that could have become vestigial but instead was the subject of exaptation. Having originally been evolved to aid in thermal management, feathers eventually became just as important for purposes of display.

Chimpanzee right forelimb



Snowy egret right forelimb



Pan troglodytes right forelimb by Blackburn Lab on SketchfabSnowy Egret Wing Skeleton by Blackburn Lab on Sketchfab

Figure 57.1.9: Homologous structures in the limbs of tetrapods. The bone sequence depicted here, one upper arm bone, two lower arm bones, and many hand bones, is a set of shared characteristics of all tetrapods today going all the way back to the lobe-finned fish of the Devonian Period.

PATTERN 6: ORIGINATION AS AN EVOLUTIONARY MECHANISM

Origination is speciation. It is the process by which new species evolve from ancestral forms. As this evolution occurs, phylogenies grow. Though often visually depicting a sense of gradual change, phylogenies are anything but gradual in the way they develop. They branch out as they develop over time, but off-shooting branches tend to occur pretty suddenly in the fossil record. Speciation, which ultimately creates new groups, does not occur at a constant rate over time. Rather, it is most typically quite choppy. Much like a real tree, the tree of life has branches leading out from a trunk and smaller branches popping out quite suddenly. To analyze evolutionary relationships among groups, Huxley (1959) introduced the concept of a clade, or group of related organisms, including a common ancestor and all its related descendants. Cladistic analysis provides insights into relationships between organisms through character analysis–studying the observable features of organisms in order to establish whether they are shared or derived from an ancestral taxon. Shared characters are traits that evolved early in a clade and are visible among the entire group. All primates, humans included, have opposing thumbs, for example. These are also called synapomorphies. Derived characters are traits that distinguish a taxon within its clade. They are also called apomorphies. While shared traits define an entire clade, derived traits represent evolutionary innovation that leads to the formation of new clades through speciation. Distinguishing these traits among a population of organisms or fossils is often done using a character matrix, where traits of taxa are listed one one axis and the tax themselves on the other.

Figure 57.1.10: A character matrix can be used to analyze evoutionary relationships among organisms. Shared derived characters, rather than shared ancestral traits, are used to parsimoniously differentiate between a group of organisms. (Image: University of California Museum of Palentology, https://evolution.berkeley.edu/admin/media)

Figure 57.1.11: From a character table, a cladogram can be created to visually explore shared derived characters, their development through speciation (origination), and their development over time. In this case, connections between sharks, primates and other mammals, and dinosaurs are explored in a larger scale. Red bars indicate the development of a shared derived trait and subsequent speciation events that led to the development of additional clades over time. (Image: University of California Museum of Palentology, https://evolution.berkeley.edu/admin/media)

Such character analysis makes it possible to create cladograms (Figure 57.1.11) by organising taxa first by shared characters and then through the identification of derived characters over time.The cladograms used to represent evolutionary relationships very intentionally imply a sense that most of the time, life is enjoying a kind of stasis but that, every once in a while, is punctuated by a major event, called species origination, that creates a new branch off of the tree. This more modern view of speciation, or cladogenesis, as the generation of a new clade through speciation is called punctuated equilibrium (Gould and Niles Eldredge, 1972). This set of ideas relies on the idea that environments retain a sense of stability unless changed, such as through major

environmental upheavals. Punctuated equilibrium predicts that groups of organisms will otherwise remain in periods of stasis with little evolutionary change through most of their existence. Cladogenesis is, then, a rather geologically sudden process.

Within a cladogram, groups of organisms are ordered according to time across the horizontal axis, and evolutionary change along the vertical axis. There are three ways organisms can be grouped in a cladogram, based upon how they are related to one another. A monophyletic group is a clade, or a group of organisms that all share common traits derived from a common ancestor and all of its descendants. Such groups include not only the ancestral organism from which all shared characters are derived, every organism included across time and evolutionary change beyond that. Paraphyletic groups are not defined as clades. In contrast with monophyletic groups, these groupings of organisms retain common characteristics among related organisms, but not all of the descendent groups of organisms. The lemurs, lorises, and tarsiers below are examples, but so are reptiles, with regard to other amniota. While they contain a great many of amniotic organisms within their paraphyletic group, they do not contain the mammals or birds, which are also amniotes. The final situation is that which results from convergent evolution, or homoplasy. Polyphyletic groups are situations where organisms may have evolved similar characteristics but these characters are not derived from a common ancestor. In the case of lorises and tarsiers, both groups are nocturnal, a behavioral trait that is not shared with or derived from any other primates and must have evolved independently of one another. Another example would be mammals and birds, who are both haemothermic, or warm-blooded, a characteristic not derived from a common ancestor but also evolved through homoplasy.

Cladogenesis and Anagenesis

Cladistics is the exploration of speciation, defined as the splitting off of a clade (Figure 57.1.11). Speciation in this context is called cladogenesis. During cladogenesis, a parent species branches off, forming one or more new species. All of these organisms may also still live among one another, though they can no longer interbreed. Such episodes represent the visible norm of evolutionary trajectories and are consistent with punctuated equilibrium, as such events are preceded and followed by periods of stasis.

Another form of speciation occurs within a clade and is called anagenesis. Anagenetic change occurs more gradually, eventually leading to a new species, as heritable changes accumulate within a population through small episodes of natural selection. At this point, the originating species no longer exists, giving way to the new, derived species. One example of potential anagesis still discussed occurs within our own genetic lineage, *Homo*, where one earlier species may have evolved directly into subsequent species (Clee and Gonder, 2012).

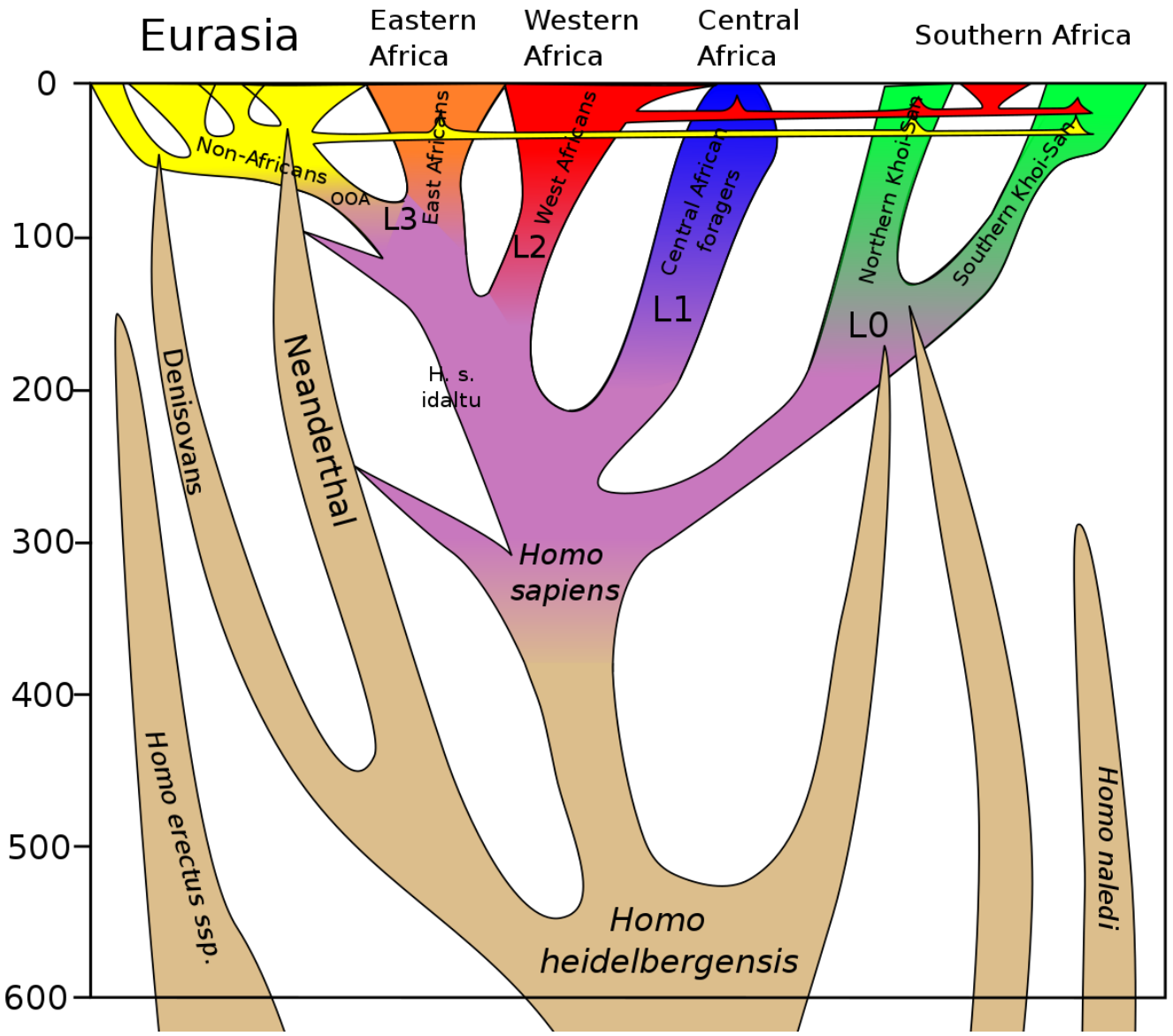


Figure 57.1.12: Phylogeny of Homo sapiens across time and place through anagenesis.

This may have been the case as *Homo heidelbergenesis* anagentically evolved into *Homo neanderthalensis* and then into modern humans, *Homo sapiens*. One possible support for this thinking is that, in the case of the last two species, we have ample evidence that there was at one time interbreeding populations but that this ended with the extinction of Neaderthals.

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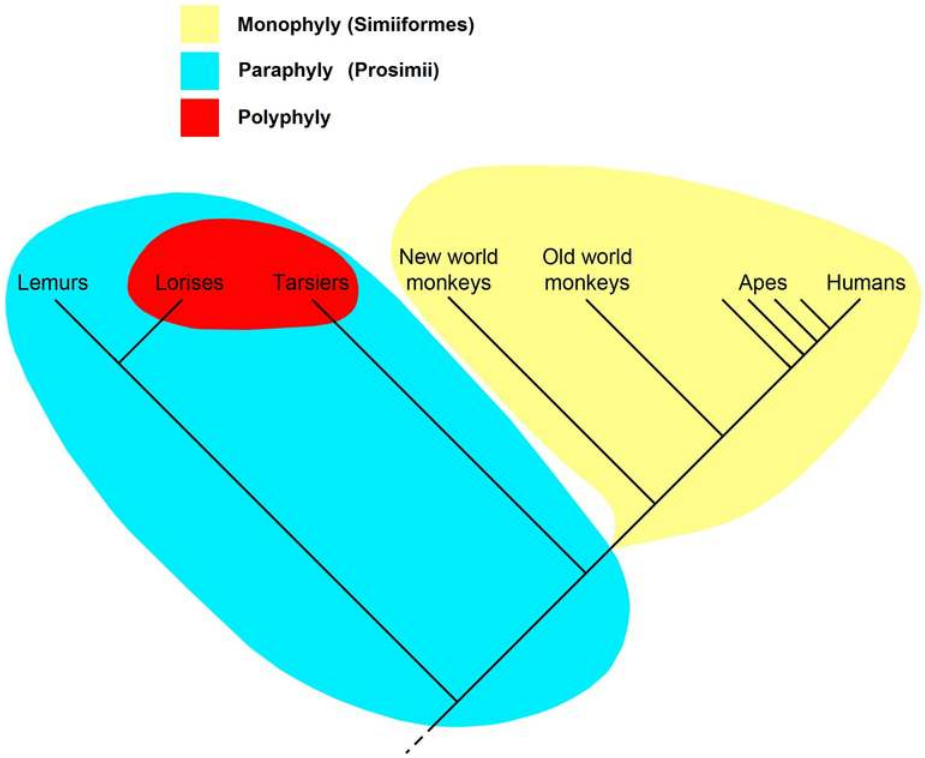


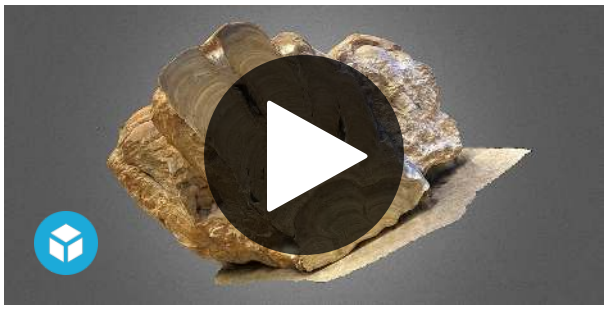
Figure 57.1.13: A primate cladogram showing the differences, or groups, of different kinds of taxon. Monophyletic taxon are in yellow. Paraphyletic taxon are outlined in blue. Polyphyletic taxon are outlined in red.

Just a few “Highlights” in the Story of Origination

Abiogenesis

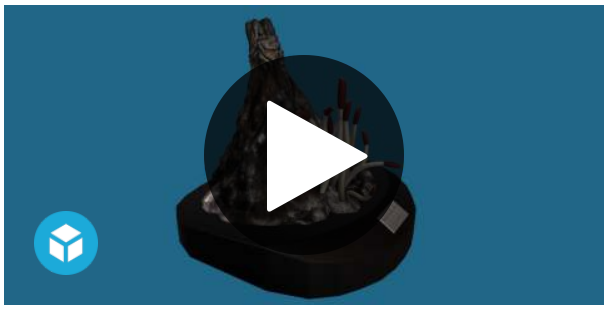
Abiogenesis is the development of life from non-living matter. At one point in Earth’s past, there was no life. Because of the development of a series of processes that eventually led to the eventual emergence of cell membranes, life began. How this occurred is still a subject of some uncertainty. The mechanisms of abiogenesis are poorly understood. In 1952, the Miller-Urey Experiment attempted to address the question of abiogenesis. Through this classic experiment, combining inorganic compounds with a source of energy, the experiment demonstrated that it was possible to synthesize proteins important for organic life. This result would only be the beginning of a chain of processes that would need to include the ability to self-replicate, self-assemble, leads to a rudimentary cell membrane, all through a series of self-perpetuating reactions. However this occurred on Earth, the earliest known such life dates back at least 3.5ba. These early life forms provide the foundation of all subsequent life and are our most distant ancestors.

An excellent example of prokaryotic life billions of years ago come in the form of stromatolites. Still around today, these algal mounds are most commonly made up of single-celled cyanobacteria. Cyanobacteria are photosynthetic. Photosynthesis is a later evolutionary pathway that will be discussed below, but the need to produce food using sunlight means that cyanobacteria would need to be living within ready access to sunlight. Stromatolites are still around today.



Evolution of Photosynthesis

Photosynthesis is one of the most important energy production pathways that primary producers use in the biosphere. The ability to change sunlight and water into glucose allowed for major evolutionary advances. Before photosynthesis, a form of chemosynthesis was most prevalent. At mid-ocean ridge hydrothermal vents, at a depth where sunlight does not penetrate, organisms still use this pathway today. Tube worms use symbiotic bacteria to help them fix carbon, using hydrogen sulfide (energy source), to produce the sugars and amino acids they need.



Organisms that use photosynthesis for energy production have evolved to do so in a variety of ways. Ultimately, the basic chemistry is the same. Cells take in carbon dioxide, water, and light energy to produce carbohydrates, such as glucose, which they can store to be used right away or at a later time. Photosynthesis may have evolved as early as 3.4 Ga.

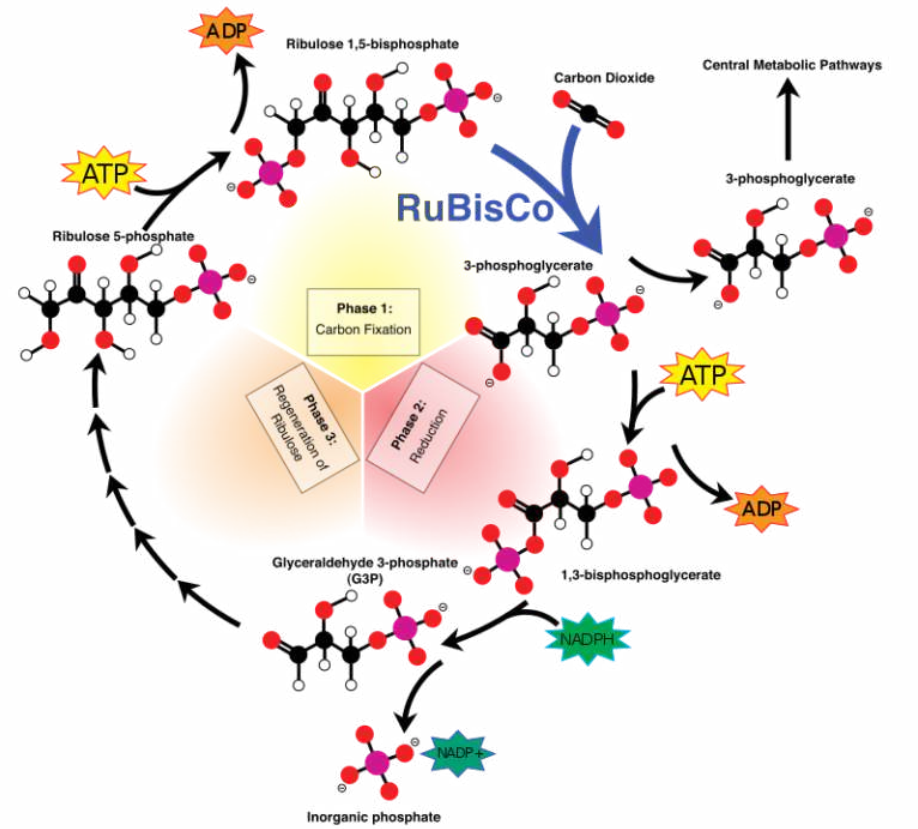


Figure 57.1.14: The Calvin Cycle in Photosynthesis. Occurring in 3 phases, this process fixes carbon from the atmosphere (carbon dioxide), reduces it, and then produces ATP and ADP, organic chemicals that provide the energy that drives many of the processes occurring in cells.

Endosymbiosis

For most of life's history, it existed in single-celled prokaryotic forms. Of the three domains of life today, two of them are made up of these kinds of organisms, the archaebacteria and eubacteria. Prokaryotic cells contain no organelles, but simply DNA, ribosomes, and the containment structure consisting of a cell wall, plasma membrane, and capsule. Scientists think that the process of endosymbiosis between some of these prokaryotes led to the first eukaryotic cells. Endosymbionts are organisms that live within the body or cells of another, sometimes mutualistically, sometimes parasitically. Within eukaryotic cells, several of the organelles are thought to have a direct bacterial origin. These include the mitochondria, plastids, and chloroplasts. Endosymbiotic bacterial forms may have evolved to live directly within other bacteria, took on specific roles as a part of that tiny community, eventually leading to fully eukaryotic cells. This evolution took place nearly 2 billion years after life began, around 1.6-2.1 Ga. Since then, eukaryotic cells have evolved into many forms, most notably from two main structures, plant cells and animal cells.

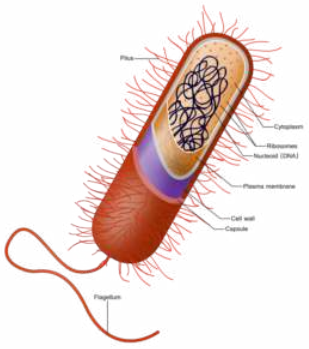


Figure 57.1.15: A model of a prokaryotic cell.

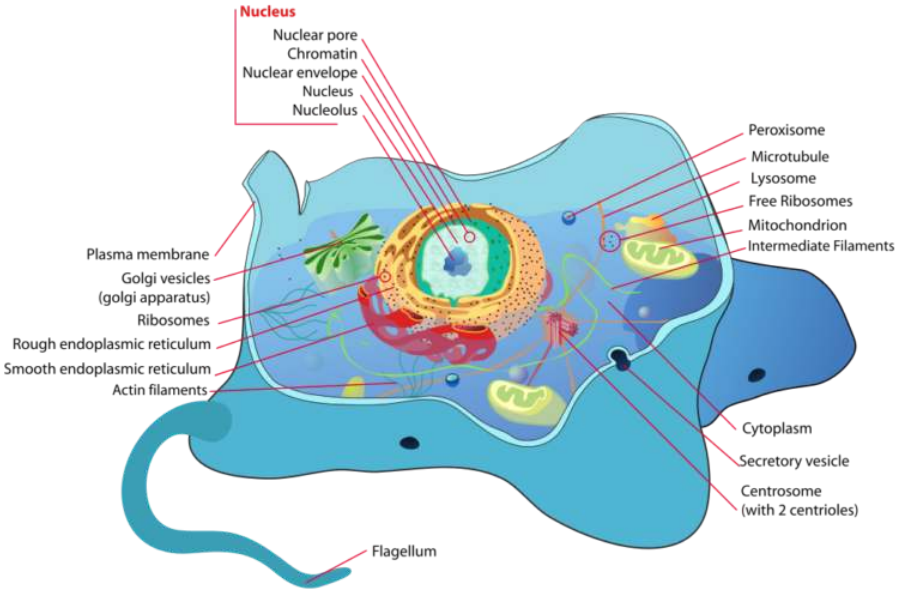


Figure 57.1.16: A model of a typical eukaryotic animal cell.

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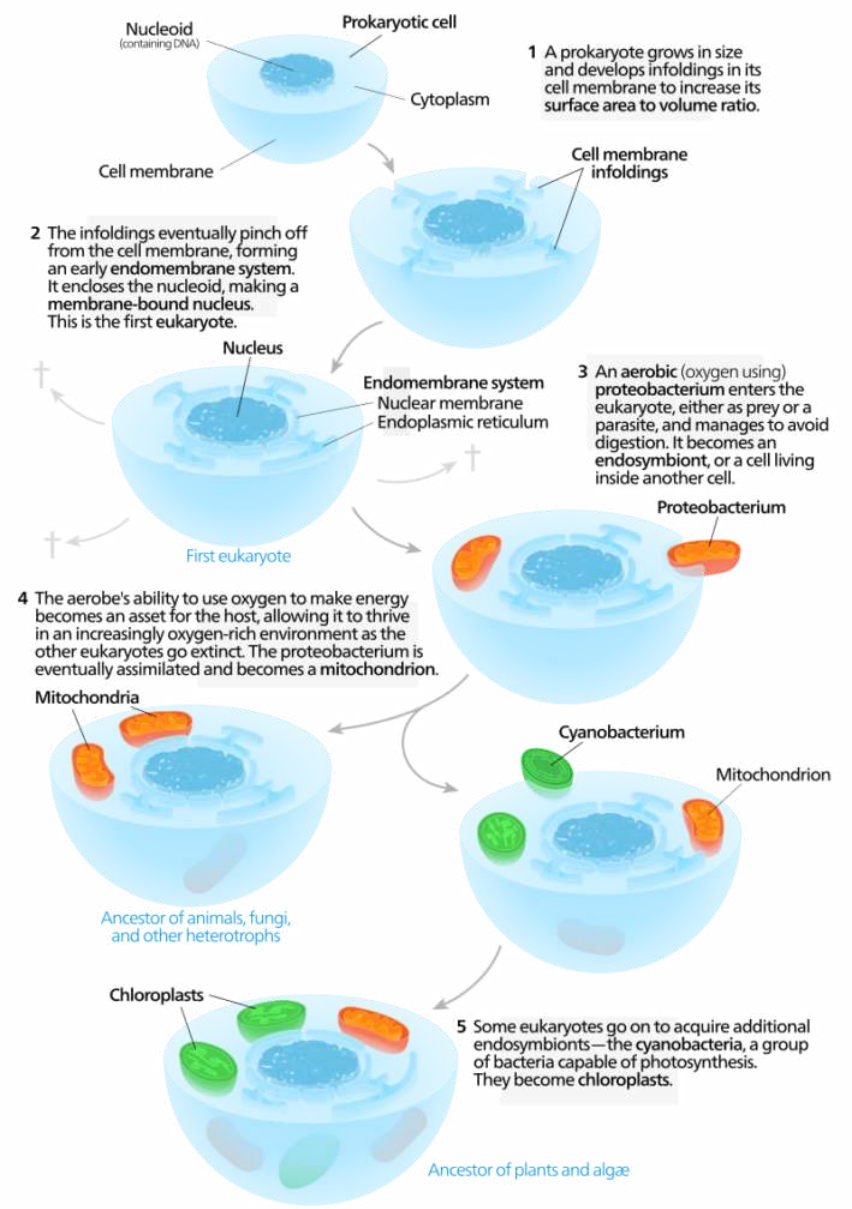
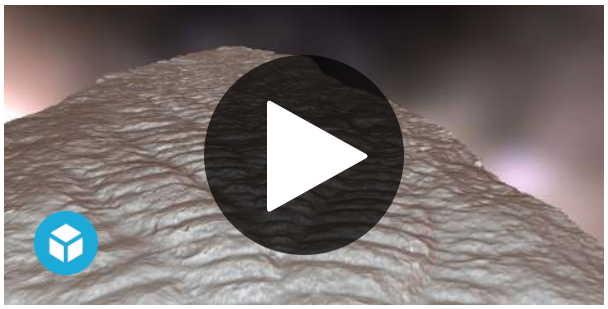


Figure 57.1.17: Endosymbiosis. This example of the process is illustrated through the inclusion of an aerobic bacteria into a host. The aerobes ability to use oxygen to make energy becomes an asset, and this trait and its DNA are then incorporated into the prokaryotic cell, serving as mitochondria, one of the key organelles that exist within eukaryotic cells.

Vendian Radiation

About 560ma, some of the earliest soft-bodied organisms that would come down to use as body fossils existed. Variously referred to as “Vendian”, after the Period in which they evolved, or “Ediacaran”, for the hills in Australia where they were first discovered, their forms are enigmatic and like nothing we see today. They have been described in many ways, but some only appear to be blobs of matter, leaving great challenges to interpretation. Some of them look like worms, others like sponges, others trace fossils. It is very likely that they are ancestral to arthropods. In any case, they are important because they are the first animals in the fossil record!





Origination is Critical for Evolution

More of the story of “Wonderful Life” will be explored in another chapter of this book. These few examples from the Precambrian time highlight the importance of origination. While we know a great deal about how some organisms originated because of the fossil remains that have come to us in the rocks, we also have many gaps yet to fill. Nevertheless, the fossil record has helped inform our understanding of how species originate today.

PATTERN 7: EXTINCTION AS AN EVOLUTIONARY MECHANISM

Extinction is the opposite of origination. It is the permanent termination of an organism or a group of organisms. At this moment in time, 99% of all species that have ever existed have been subject to extinction. In actual numbers, about 5 billion species are thought to have succumbed to such a demise. The lifespan of some species is certainly longer than others, and geographic distributions also vary. Because of these two factors, it can be very challenging to determine the actual moment of extinction. Nevertheless, extinction is very common and perfectly normal. At any given moment in the geologic past, species are going extinct. It is the rate of extinction that varies. Background extinction rates are much different from the rate of extinction during mass extinction events.

Background Extinction

Normal background extinction rates vary because species life spans vary. Estimates for the life spans of various taxa range from a high of 11 million years to as short as 0.5 million years for all fossil groups (Lawton and May, 1995). This means that some species, not individual organisms of course, could be around for as long as 11 million years. Our own species, *Homo sapiens*, has been around for about the last 300,000 years (0.3 million years). Comparing this span of time to the low end average estimate, it could be said that we have a long way to go before our species becomes extinct. But, such a numerical comparison does not necessitate such a reality!

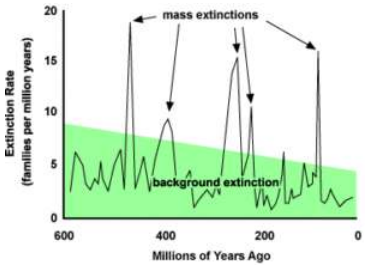


Figure 57.1.18: Mass extinctions versus background extinctions. The green shaded area represents extinction that is occurring all the time. Species turnover at low rates, or even when rates are a bit elevated above background levels, are not unusual (Permission: University of California Museum of Paleontology's Understanding Evolution (http://evolution.berkeley.edu).

In terms of background extinction rates, estimates hover between 0.1 genera per million genera year for fossil taxon and 1 extinction per million species per year for living taxa. This translates to about 10-100 species per year going extinct, globally. The fossil record estimates are coarse, due to preservation bias (favorite marine organisms with mineralized skeletons) and inherent destruction in the record over time (the near-present is better represented than the ancient past). Modern estimates are difficult also due to the fact that so many species are not yet known and because the challenge of taking species census data is quite large. The modern numbers, when parsed more simply, suggest that for birds we would expect one extinction every 400 years or one mammal every 200 years. These estimates take into account the lifespan of species. Mammals, as a group, are thought to have an average species lifespan of about 1 million years.

Mass Extinctions and their causes

When extinction rates increase sharply above background rates for a geologically short period of time, the event is referred to as a mass extinction. Mass extinctions are rare, but when they happen they make huge adjustments to the patterns of life on the planet. Traditionally, there five really large mass extinctions that are set aside as special. In order, they mark the end of the Ordovician Period, the end of the Devonian Period, the end of the Permian Period, the end of the Triassic Period, and the end of the Cretaceous Period. Very recently, a sixth ancient mass extinction is being discussed that would directly precede the End-Permian event. Called the end-Guadalupian extinction, it would have occurred just 8 million years prior to the traditional end-Permian event. Ultimately, Raup and Sepkoski (1982) first identified the "Big Five".

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57.2: Mass Extinctions of The Past and Possible Causes

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Age (Ma)	Percentage of Genera Lost	Hypothesized Trigger or Cause	Affected Groups
Ordovician/Silurian - 439 Ma	86%	The shift in continents closer to poles caused glaciation, low sea levels; too many plants removed too much CO_2 from the air which reduced temperature causing an ice age; falling sea levels due to huge ices sheets in the southern hemisphere; climate changes; ocean chemistries changed; the Appalachian Mountains forming	Marine species (most organisms were marine at this time) such as some bryozoans, reef-building brachiopods, trilobites, graptolites, and conodonts
Devonian - 364 Ma	75%	Flourishing of land plants released nutrients in the oceans causing algal blooms that depleted oceans of O_2. Bacteria flourished, Volcanic ash cooled the Earth's temperatures. Sea levels changes; asteroid impact; climate change;	Life is shallow seas such as corals, other shallow marine organisms including brachiopods, and single-celled foraminiferans; Terrestrial species such as spiders, scorpion-type creatures; Elipistostegalians; helped with vertebrate evolution.
Permian - 251 Ma	96%	"The Great Dying" Volcanic activity in Siberian Traps released lava flows the size of the United States and over 1,000 feet thick. Emitted large amounts of CH_4. Global warming and atmospheric changes; oceans warmed and became acidic. Pangaea also taking shape, changing ocean currents; meteorite may have hit too. Climate change	96% of all marine species 70% of all land vertebrates The only mass extinction event that significantly impacted insects; Earth's largest extinction event, decimating most marine species such as all trilobites, plus insects and terrestrial animals; life descended from the 4% of surviving species;
Triassic - 200 Ma	48%	Volcanism leading to flooding basalts and a tremendous amount of sulfur dioxide and carbon dioxide in the atmosphere, causing climate change; an asteroid impact	Many types of animals died, including marine reptiles, large amphibians, reef-building creatures, cephalopod mollusks; plants were not affected. This extinction allowed for the evolution of the reptiles
Cretaceous - 65 Ma	76%	Most famous event; a.k.a. "K-T Extinction"; Astroid imact hypothesis - asteroid impact in Yucatan Peninsula; Deccan Trap flood basalts were also erupting in India, but while they may have been a	Extinction of many species in both marine and terrestrial habitats; ammonites, many flowering plants including pterosaurs, mosasaurs, many other dinosaurs, and other marine reptiles, many insects, and all non-Avian dinosaurs.

Age (Ma)	Percentage of Genera Lost	Hypothesized Trigger or Cause	Affected Groups
		contributor, were likely not the ultimate cause	This extinction allowed for the evolution of mammals.

The "Big Five" mass extinctions in the Earth's past and hypothesized causal factors. There are very likely more mass extinctions than listed here. And, we may also be in the midst of one today. How will it compare?

When extinction rates increase sharply above background rates for a geologically short period of time, the event is referred to as a mass extinction. Mass extinctions are rare, but when they happen they make huge adjustments to the patterns of life on the planet. Traditionally, there five really large mass extinctions that are set aside as special. In order, they mark the end of the Ordovician Period, the end of the Devonian Period, the end of the Permian Period, the end of the Triassic Period, and the end of the Cretaceous Period.

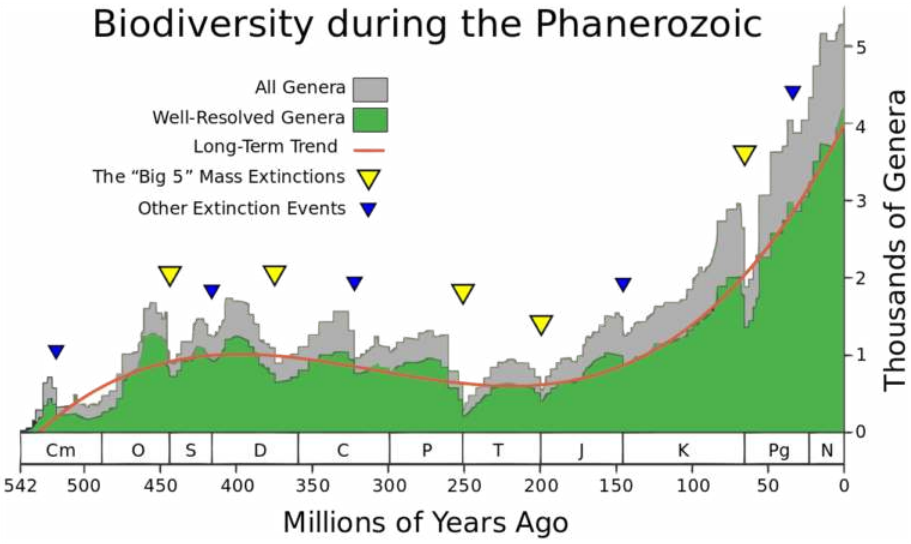


Figure 57.2.1: Biodiversity during the Phanerozoic Eon according to the fossil record. There are a wide variety of extinction events that have occurred between the Cambrian Period (Cm) and the Neogene Period (N). The "Big 5" are marked with the yellow triangles and include the end-Ordovician, the three-stage end-Devonian, the end-Permian, the end-Triassic, and the end-Cretaceous (K). The Permian end-Guadalupian Stage extinction is not marked, but can be seen in the rapid biodiversity drop just before the end-Permian event. Despite the extinctions, note the increase in biodiversity since the Trirassic (T)

Our understanding of these events changes as time and research themselves evolve. Compared to one another, the "Big Five" certainly vary in intensity, with the end-Permian being the most intense of them, often being referred to as "The Great Dying" because of the loss of an estimated 95% of all life on Earth. Many scientists feel that today we could be in the midst of a sixth mass extinction, caused by human activity. There are challenges in comparing today's biodiversity loss to what we see in the fossil record. Whether or not today's extinctions make the cut and join the other five is still an open question. Eventually, as things in the science progress, so will our conclusions about these events.

Mass extinctions are caused by powerful natural events. Every mass extinction contains similarities in their story lines. There is usually a significant element of climate change, whether as the main driver or a secondary result of the primary cause. There is massive loss of life, though how this plays out on land and in the sea can vary by event.

Ordovician Extinction

The end-Ordovician event is described as being caused by global cooling. More extreme cooling events during the Proterozoic Eon, referred to as "Snowball Earth" events, certainly occurred. But, there is no record of mass extinction. Most life was single-celled and it did not fossilize well. The end-Devonian, accompanied by massive black shale deposits, is unusual in that it actually happened over millions of years. The evolution of vascular plants on land may have been a factor, as the increase biomass to waterways could have caused mass eutrophication across large sections of the shallow epeiric (inland) seas that existed.

Cretaceous/Paleogene Extinction

The 3D model below illustrates the famous K/T boundary extinction event in the rock record. This event, caused by the Chicxulub asteroid impact, caused the extinction of the dinosaurs. Located near Gubbio, Italy, this is the very outcrop where Walter Alvarez collected samples to analyze for microfossils across the boundary. He noted that below the boundary, they were abundant, diverse, and some were very large. Above the boundary, they were sparse and tiny. Enlisting his Nobel-prize winning physicist father, Luis Alvarez, they determined that the boundary clay (the inset area in the model that looks like a gap) was unusually rich in iridium. Many such similar boundary clays rich in iridium have since been identified globally.



Are we currently in the midst of a sixth (or seventh!) mass extinction? In order to see the rate of species loss we see in the record over the last century, between 2,000 and 10,000 years should have been necessary. Given a conservative estimate of the background, pre-human extinction rate of about 0.1 extinction per million species years, the current observable rate is about 1,000 to as much as 10,000 times higher, at a bout 100 extinctions per million species years or more. These rates are quite alarming. Causes of current biodiversity loss include a wide variety of factors. Climate change, overhunting, habitat destruction, and habitat fragmentation are some of the largest. These are all directly related to human activity. At this point, this would be be the only mass extinction in geologic history caused by a single species.

Extinction as Opportunity

Extinction leads to origination. While it is certainly a way organisms fade into the memory of the rock record, these events also create opportunity. With extinction comes new niches, into which species can move. These new niches provide new environments, food sources, and habitats. They also provide new opportunities for evolutionary change. Organisms that survive extinctions and that move into environments that are changed or even marginal for their current life mode may adapt well enough to pass on their genes to another generation who may be better adapted to the new environment. While extinctions on various scales are certainly tragic for those groups or organisms that experience them, they can also provide the foundation for success stories. As humans, we have inherited the successes of our evolutionary ancestors.

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57.3: The Fossil Record - Making Sense of Evolution’s Past

Like any scientific theory, evolutionary theory is robust, supported by multiple lines of evidence, explains observations make, and allows us to predict and test. As a theory, it does an excellent job of helping us make sense of the patterns we see around us in ecosystems, within populations of organisms, among natural systems, and within our own bodies. This cohesive framework is still being adjusted with new discoveries in different ways. It is also providing scientists with directions for research that shines light on the world around us but research that also leads to critical new medical treatments.

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57.4: Further Reading

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Tree of Life Navigators

University of California Museum of Paleontology Taxon Lift Tree of Life

Wikispecies Taxon Navigator

Tree of Life Web Project

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CHAPTER OVERVIEW

58: (Tools of the Trade) Sequence stratigraphy

Stratigraphic Tools: Basic Sequence Stratigraphy



Figure 58.1: Conformable Ordovician limestone layers near Carthage, Tennessee

Key Concepts

After reading this chapter, students should be able to:

- Define a sequence in the context of stratigraphy.
- Describe basic sequence hierarchy and architecture.
- Describe a basic understanding of accommodation space and the variables that affect that space.

58.1: An Introduction to Sequence Stratigraphy

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58.1: An Introduction to Sequence Stratigraphy

A sequence is a pattern of observable phenomena that occur in a predictable order. In sequence stratigraphy, these are layers of sediment deposited over time in a predictable manner due to the effects of environmental changes that affect tectonics, sediment supply, and sea level.

Over time, environments change due to a wide variety of long-term and short-term events within the Earth system. Earth system events can be extrinsic, coming from the space environment, or intrinsic, resulting from an event within one of the Earth's spheres. Examples of an intrinsic event might be a large flood-basalt eruption. Such a geosphere event will directly impact the hydrosphere, atmosphere, and biosphere. An example of an extrinsic event would be the gradual changes in Earth's orbital or tilt parameters (Milankovitch Cycles). These affect the amount of incident solar radiation at a number of levels, which impacts all of the Earth's spheres.

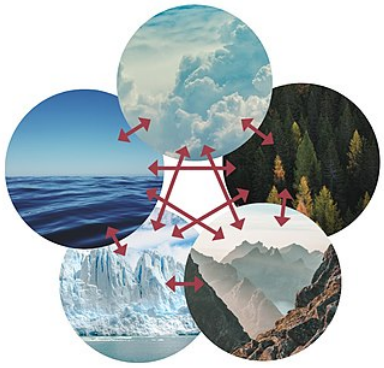


Figure 58.1.1: A representation of five "spheres" that are a part of the Earth system. From top around clockwise, the atmosphere, biosphere, geosphere, cryosphere, and hydrosphere. Events that occur within the system have direct and indirect effects that are transferred through interactions between systems. Such interactions lead to changes in sedimentary deposition that is recorded in basins.

Events like these that occurred in Earth's deep past are often hard to understand as a whole, let alone when attempting to examine their genesis, progress, or ultimate result. Geoscience historians care about these events because they have direct and indirect effects on the Earth's individual spheres. And, understanding how systems like the hydrosphere, geosphere, and atmosphere have interacted in the past provides a great deal of insight into similar processes occurring today. If we can understand how Earth system events lead to change, we can create predictive models for what that change looks like in the rock record. In deep time, those records are recorded in sedimentary basins. Sequence stratigraphy is a field that provides important tools for understanding the course and cause of these events in Earth's past. And, we can apply these lessons to understanding them today.

The ultimate goal of sequence stratigraphy is to understand the Earth's past as recorded in the stratigraphic history of a basin. Sometimes, the impetus for this work comes through the search for natural resources, like coal and petroleum. Sometimes, the impetus is pure research and curiosity. We can then use the stratigraphic tools like lithostratigraphy and biostratigraphy to measure and analyze stacks of rock layers. Then, correlating units across a region, it is possible to reconstruct this history. At this point, it becomes possible to quantify changes in water depth, sedimentation rates, rates of tectonic change, and rates of global (eustatic) sea level change. Over time, all of these factors will produce packages of sediment that are related to one another and bounded on the top and bottom by unconformities, or sequence boundaries, that can be correlated over distances. These related and predictable packages of rock are referred to as sequences. Sequences of strata are genetically-related packages of rock bounded by unconformities. Within sequences, Walther's Principle of facies migration apply. Across these sequence boundaries, it does not.

SEDIMENT NEEDS ITS SPACE: ACCOMMODATION, SEA LEVEL, AND SEDIMENT SUPPLY

Basins are not static features. Basins are influenced by the physical processes around them as they record the history of the surrounding land and of their own past. They also come in a variety of shapes and sizes. Some basins, like the Atlantic or Pacific Ocean basins today, are massive. As sediments erode,transport downhill via wind and streams, and accumulate, they fill in the "accommodation space" that is available in that basin. Like a bowl of cereal, a basin only holds so much sediment, and fills in over

time. Sedimentary systems strive toward an equilibrium between between sediment infill and available accommodation space, but various Earth events will upset this balance.

In order to calculate the accommodation space within a given current or past basin, the following equation is used:

T + E = S + W

Where:

T = Rate of tectonic subsidence, or sinking, in the basin

E = Rate of eustatic, or global, sea level change, relative to a reference ellipsoid

S = Rate of sedimentation/accumulation in the basin

W = Rate of change in water depth within basin

This equation represents a simple balance. If the left side of the equation is positive, then there is increasing accommodation space in the basin and, if negative, decreasing space. On the right, if the number is positive then there is increasing sedimentation and negative if there is not. Thus, if the numbers on both sides of the equation are both positive, then both space and sedimentation (S) are increasing. If the left is negative and the right side positive, then we can predict that the basin will be infilling over time until no more sedimentation is possible. While it seems simple enough, this equation is very difficult to actually use in practice and better serves as a model for thinking about and considering stratigraphic packages of sedimentary rocks (Holland, 2018).

Absolute sea level (ASL), also known as eustatic sea level, is an attempt at calculating a global average. It is the level that, under the force of gravity and if all waters were perfectly calm. Today, when we think of sea level rise due to global warming, this is the kind of change to which we refer. If so much ice melts in Antarctica or Greenland, sea level globally will rise to some amount (10cm to multiple meters, perhaps) over the next century. Due to global warming, the ocean waters are undergoing thermal expansion, causing the rises we see along our shorelines today. The rate of this change (E) can be difficult to measure. Absolute sea level is measured against a reference ellipsoid, an imaginary sphere within the Earth that can be used as a datum. From this, a geoid can be produced for sea level, which is the level at which it would be under the force of gravity and the Earth's rotation alone, taking out tidal and other short-term variations. Changes in ASL happen slowly, but occur in reaction to global events, such as changes in climate.

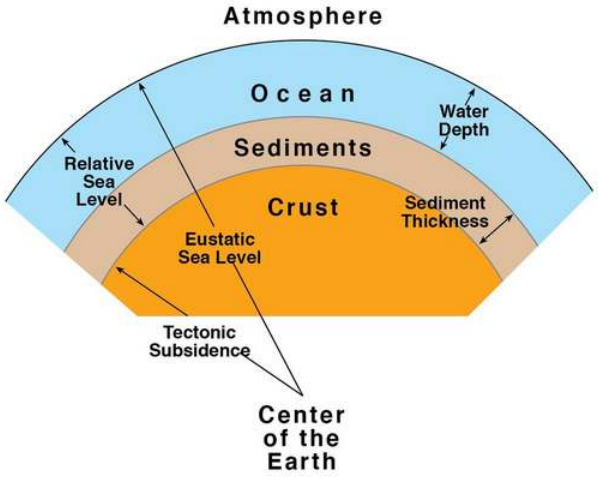


Figure 58.1.2: Cross-section of the Earth showing important sequence stratigraphic definitions (from Holland, 2018).

Tectonic changes include variables like the uplift of land, isostatic rebound or the sinking of land, subsidence. Plate tectonic changes are the primary driver of all of the changes in these variables. Recall that oceanic lithosphere is primarily made up of basalt and that continental lithosphere is primarily made up of granite. Lithospheric plates are like ice cubes. These different rock types have different bulk densities, basalt being much more dense. This causes oceanic lithosphere to sit lower on the Asthenosphere than continental lithosphere. At subduction zones, new mountain ranges are thrust up as lithosphere buckles and

thickens. This concentrates mass and leads to the subsidence of the land as foreland basins are created through downward flexure. These kinds of processes tend to deepen basins.

Other situations, such as ice melt on land, tend to shallow basins. During ice ages, times when the continents bear the burden of massive ice sheets, the extra mass of ice weighs down the plates. When the ice melts, local basins will rise as continents isostatically adjust their elevation. The rate of subsidence or uplift (T) is very important to know when calculating the accommodation space of a basin. For the purposes of accommodation space, the rate of subsidence and the rate of eustatic sea level change are the biggest and longest term factors. If eustatic sea level drops at the same time that subsidence increases, no new sedimentation space is created. If eustatic sea level drops at a slower rate than subsidence, new space is made, and available for deposition.

Seas are dynamic. It is a bit misleading to refer to “sea level”, in the sense that there is no absolute level at which the water rests, when measured against the floor of a basin. Measured relative sea level is an average depth, relative to this basin floor in a particular region. Relative sea level (RSL) is the water depth (W) of the basin. It is affected by many factors that can cause it to rise or fall. Some of these factors include increases in water mass, thermal expansion of water with a rise in temperature, changes in ocean currents or seawater salinity, and isostatic rebound adjustments of continents. When looking at NOAA tidal gage data today, it is the case that the eastern shore of Virginia is experiencing rising seas while the Gulf coast of Alaska is experiencing a net drop in sea level.

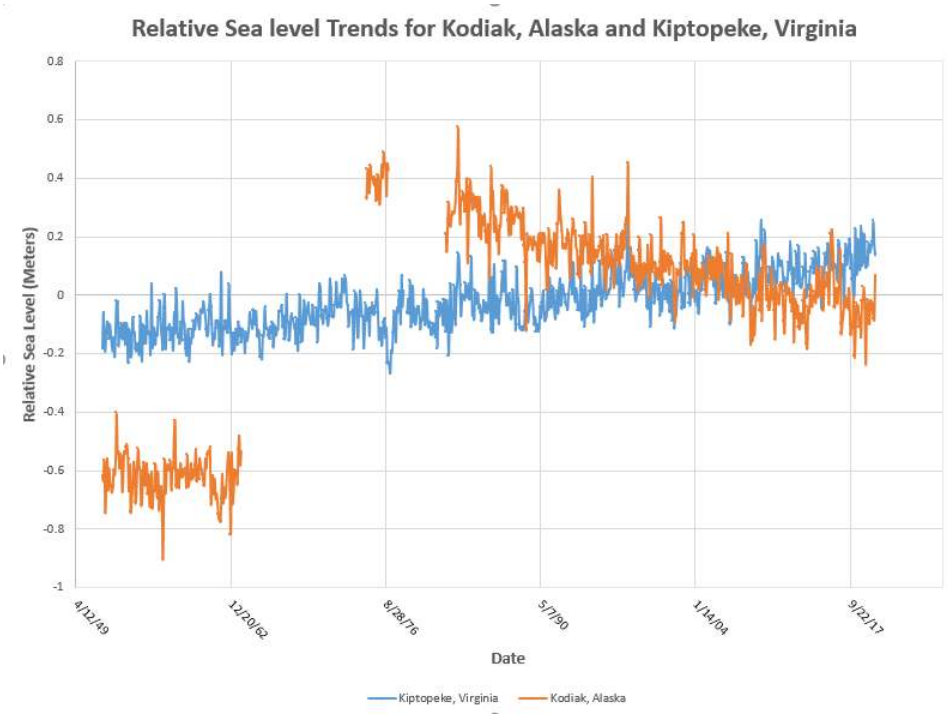


Figure 58.1.3: Relative sea level (RSL) trends for two sites on opposite sides of the North American craton. Kiptopeke, Virginia, is experiencing a relative rise in sea level of 3.97 mm/yr where Kodiak, Alaska (southern coast) is experiencing a drop in relative sea level of -9.92 mm/yr. While global (eustatic) sea levels are rising, local records will vary. Forecast trends for both sites, according to NOAA, indicate positive sea level trends into the next century. Note: The Kodiak, Alaska data has a large gap around the time of the 1964 Good Friday earthquake. Prior to the quake, the sea level trend was flat. After the quake, the sea level trend is negative. This negative trend is temporary, however, and may be a tectonic remnant of this event.

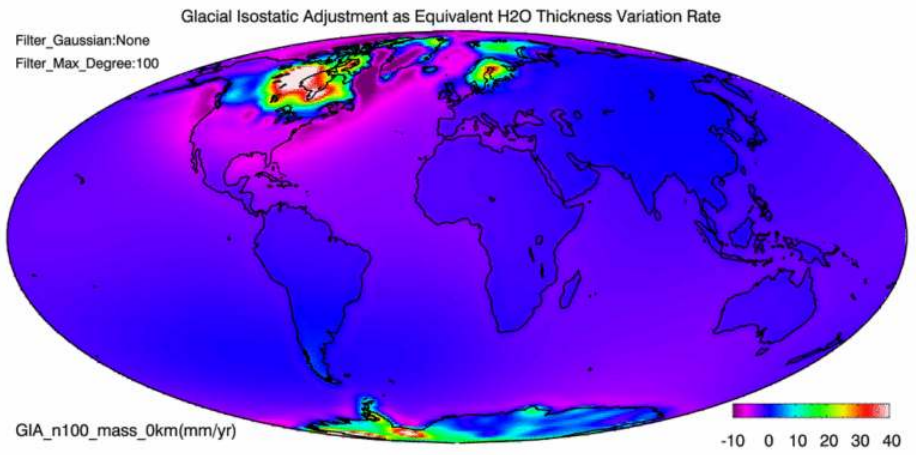


Figure 58.1.4: Glacial Isostatic Adjustment as equivalent H_2O thickeness variation rate. In other words, gravitational anomalies (changes) as a result of tectonic rebound from the last glacial advance of the Pleistocene. Note the high levels of adjustment over areas of northern North America, Europe, and also Antarctica. Gravitation, as measured by NASA's GRACE mission, is balanced against the gradual recovery of ocean basins as they slowly spring upward, causing relative sea level around these noted areas to drop. Gravitational anomalies are due to mass loss, which can be mantle material moving to accommodate areas of the crust that are rebounding. Rates of mass change have been estimated to be as high as 104 gigatons under Greenland (Wu et al., 2010). Image: NASA

The graph of tidal gauge data for Kodiak, Alaska, illustrates nicely the effect tectonics can have on relative sea level. Prior to the great 1964 "Good Friday" quake, sea level was stable. After the quake, not only did sea level start out much higher than the original datum prior to 1964, but it has been dropping ever since. This indicates that the land surface is undergoing uplift due to tectonic changes since the earthquake. Long term trends (not shown) according to NOAA do see sea level eventually rising here also, but overcoming tectonic effects takes time. Another important measure of relative sea level can be analyzed using satellite gravity measurements. In areas of the Earth's lithosphere where there is more mass (thicker lithosphere), gravity is higher than normal, and vice versa. As in the GRACE data in the image above, areas subject to uplift now due to the retreat of glaciers since the last ice age are experiencing intense changes in gravity, indicating increasing mass in those areas, most likely due to mantle material moving into place beneath uplifting lithosphere. such tectonic change is important to consider also throughout Earth's history, as a component of both relative and global changes in sea level. It is not always the water that changes, but sometimes it is the land.

When water levels are rising and a basin deepens, we call this change a **transgression**. When water levels are falling and the basin gets more shallow, we call it a **regression**. Whether it happens due to rising or lowering water or rising or lowering land, the effect is the same.

Sedimentation rates (S) vary a great deal. As already seen, they are most closely linked with the rate of change in water level (W) in the equation above. How do sedimentation rates change? There are a wide variety of factors. As mountains rise, sedimentation rates will also rise, as the land moves upward into the atmosphere and is exposed to ever more chemical and physical weathering. Likewise, the eventual erosion of such mountains will lead to drop offs in sedimentation into the basin. Sedimentation is also affected by the climate and latitude. Humid environments will have more runoff. This will lead to higher sedimentation rates. The type of geologic setting also matters. Siliciclastic environments source their sediment from outside the basin and so in these situations sediment supply is a much more significant factor in basin analysis. In carbonate settings, the sediment is produced within the basin by organisms precipitating $CaCO_3$ directly from seawater.

Carbonate settings are very good at keeping up with sea level rise and this leads to the remarkable diversity in depositional settings. Today, carbonate platforms are more rare, existing in the Caribbean, Persian Gulf, and portions of the Pacific basin. In the past, particularly in the early to middle Paleozoic Era, such environments were much more common. The Great American Carbonate Bank, represented by the thousands of feet of variously fossiliferous Sauk Sequence limestones throughout the Mid-Atlantic region and the Cincinnati, Ohio area, are excellent evidence of this past. When it comes to sediment supply, we can also describe how relative sea level behaves with respect to sedimentation rates. When sediment supplies are increasing more rapidly than water levels, it is said that the coastal land is prograding. When sediment supplies are decreasing, it is said that the coastal land is retrograding. Thus, the rate of sedimentation (S) is critical to grasp when trying to discern whether sedimentation or water level is

controlling changes in accommodation space. If progradation of a coastline occurs and sea level regresses, then accommodation space decreases. If coastlines are in retrograde and sea levels are transgress, then accommodation space in the basin will increase.

SEQUENCE STRATIGRAPHIC: A MODEL FOR DISCERNING BASINAL AND ENVIRONMENTAL CHANGE

In the stratigraphic record of a basin, there can be packages of beds that repeat in a cyclic fashion. Sometimes, these are bounded by clear unconformities and mark major changes in tectonic regimes. In other situations, they are a remnant of changes in sea level. An entire sub-field of stratigraphy, sequence stratigraphy, is dedicated to the study of such cyclic phenomena. It will only be treated very lightly here. These cycles, however, can occur on a variety of scales. Sometimes, they represent tens of millions of years and sometimes only days. Their causes are open to interpretation, depending upon the scale of time involved, and the evidence contained in the record. One thing that is very clear is the important story these cyclic beds tell about the nature of repetitive phenomenon in nature. Climates change, continents move, and sea levels rise and fall. Such cycles have been occurring during the entirety of Earth's history. However, stratigraphy can give us important insights into how one cycle differs from another or, in our modern situation, the sediments we leave today will tell the story of human impact on the environment. Packages of cyclic sediment being deposited today represent different inputs created by our activities.

Seismic data collected in the search for new petroleum reserves in the 1960s and 1970s led to the identification of packages of **genetically-related strata**. These strata, bound by unconformity surfaces, came in various thicknesses. Later on, the field of sequence stratigraphy was developed that sought to make sense of these packages of strata. The longest of these cycles, called 1st Order Megasequences, have periods of several hundred million years. The shortest of these, what are referred to as 5th order or higher, could have a cyclicity of just days, a remarkable degree of resolution for the geologic record. This stratigraphic revolution has led not only to new successes in economic geology, but arguably more important understandings of how natural cyclicity works on a wide variety of scales, from global to a very specific location.

SEQUENCE HIERARCHY

Stratigraphic sequences are superimposed on one another. This is because environmental phenomena that create cycles of various orders are also happening concurrently. One of the jobs of a sequence stratigrapher is to separate these storylines. Nested within the time span of a 1st order megasequence, a number of 2nd order supersequences will occur. Within a supersequence, a number of smaller sequences occur, and so on. Each of these different packages of genetically-related strata are bounded by unconformities, created by erosional flooding surfaces (transgressions), that can be traced over distances. Strata are genetically-related if all facies within the sequence were deposited in lateral continuity with one another, following Walther's Principle. Flooding surfaces are sharp contacts that separate underlying shallower-water facies from overlying deeper water facies. Sequences of any order can be related across time and studied for what they have to teach about the Earth's past, a region's past, and a particular location's past.

The causes of sequences of various hierarchical levels is subject to debate. 1st Order Megasequences are known to reflect global phenomena, such as the formation and breakup of supercontinents. The ancient supercontinents of Rodinia and Pangaea would be examples of sequence boundaries. An unconformity often called "The Great Unconformity" is a good example of a bounding surface that marks the lower boundary of the megasequence that would eventually lead to the formation of Pangaea, starting with the flooded cratons that make up today's continents. 2nd order supersequences are thought to be more regional in scope and again driven by tectonics, but this time within an ocean basin. During the Paleozoic Era, a global megasequence developed, within which four 2nd Order Supersequences would progress (The Sauk, Tippecanoe, Kaskaskia, and Absaroka). The Sauk Sequence represents the earliest of these, bounded by the Great Unconformity and the Knox Unconformity. The Tippecanoe Sequence came next, bounded by the Knox Unconformity on the bottom and the Wallbridge Unconformity on top. Each of these supersequences carry with them signals for shorter, 3rd order sequences. Within those, 4th order, and within those, 5th order. By the level of a 5th order sequence, the data is entirely within a single stratigraphic formation. These shortest sequences, 4th and 5th Order, can be driven by a number of factors, but Milankovitch orbital variations are certainly likely drivers.

Sequence Order	Duration	Title	Average thickness of strata	Rise/Fall rate (cm/1000 yr)	Areal extent of deposition (Sq. mi.)
First	50-100 million years	Sloss Megasequence	1000+	<1	Global
Second	5-50 million years	Supersequence	500-5000+	1-3	Regional
Third	0.5-5 million years	Sequence	500-1500	1-10	500-50,000
Fourth	0.1-0.5 million	Parasequence Set	20-800	40-500	20-2,000
Fifth	0.01-0.1 million years	Parasequence	10-200	600-700	20-2,000

SYSTEMS TRACTS

A sequence can be divided up into four phases, sandwiched on top and bottom by sequence boundaries. In sequence stratigraphy, these phases are referred to as **systems tracts**. These systems tracts are directly tied to behavior of relative water level within the basin. The placement of a particular set of strata within a specific systems tract is done using chronostratigraphic and lithostratigraphic tools. The systems tract phases of a sea level cycle include, **Falling Stage Systems Tract and Sequence Boundary Formation (FST), Lowstand systems Tract (LST)**, **Transgressive Systems Tract (TST)**, **Highstand Systems Tract (HST)**, and back to **Falling Stage Systems Tract and Sequence Boundary Formation (FST)**. Systems tracts are made up of all sediments and material that accumulate during one sea level cycle and, as much as possible, are bounded by chronostratigraphic markers, or global stratotype sections and points (GSSP).

A sequence begins with the lowstand systems tract. At this point in the sea level cycle, the main process is sedimentation. The thickest strata deposited during this tract will be located in the center of the basin. Deeper water deposits, such as turbidites, are often common at this phase of the cycle and progradation of the coastline is the norm. Sedimentation rates are high. Accommodation space decreases. Sea level is at its lowest ebb in the basin. This phase is bounded at its bottom by the sequence boundary, always an erosional unconformity following the prior falling stage systems tract.

Once sea level begins to rise, the transgressive systems tract begins. As sea levels rise, a depositional pattern of strata displaying a retrograding (receding) shoreline and rising sea level appear. Sedimentation rates during this phase are greatest around the basin margin and lower in the center of the basin. Eventually, the basin center will see a significant drop off in sedimentation and coastal areas will show evidence of tidal influence. Such evidence might include stromatolite growth or the successive deposits of mud atop evaporative salt casts. In either situation, this evidence will largely be microstratigraphic.

Once sea level reaches its peak for the cycle, the highstand systems tract has been reached. This is the peak of the cycle. At this point, sediment supply is much greater and a progradational pattern will begin to emerge once more, as accommodation space decreases on the back front side of the sea level curve. Coastal deposits during such times will be dominated by waves and fluvial inputs and will tend to be thin, but with great areal extent. Once this highstand systems tract gives way to sea level fall, erosion will begin again during the regression of sea level and a new sequence boundary will form during the falling stage systems tract. The end of a sea level cycle is then marked by the onset once again of the lowstand systems tract.

Ultimately, there are two very important observations to be made about the sea level cycle and the progression through these systems tracts. Within one sea level cycle, there may be many, much smaller sea level cycles occurring, forming their own unique depositional systems tracts, sequence boundaries, and leaving important evidence of environmental changes in the stratigraphic record. Second, the depositional portion of a systems tract always begins with lowstand and ends with the highstand. The formation of sequence boundaries, which are unconformities, is by definition non-depositional or erosional. As such, in sequence stratigraphy, all sediment deposited in a basin is deposited during an overall sea level transgression. Within a sequence, there may be brief periods of sedimentation interruption, called diastems, but otherwise there are no substantial periods of erosion or nondeposition.

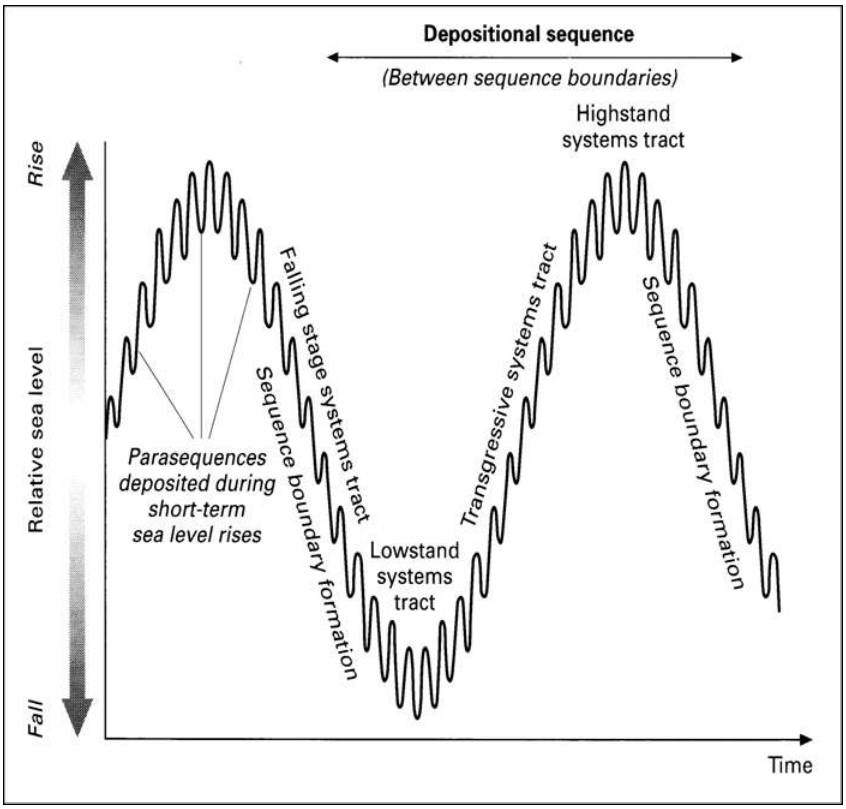


Figure 58.1.5: Model of a typical sequence stratigraphic progression. Relative sea level will transgress or regress based upon actual sea level rise or the progradation or retrogradation of the shoreline with changes in sediment supply. Beginning at highstand, where a flooding surface is formed, sea level enters the falling stage systems tract. This relative regression leaves an erosional sequence boundary behind. As lowstand approaches, relative sea level is at its lowest. The trasgressive systems tract follows the lowstand and leads to the bulk of deposition as increased sediment supply begins a relative rise in sea level, eventually leading to a new highstand.

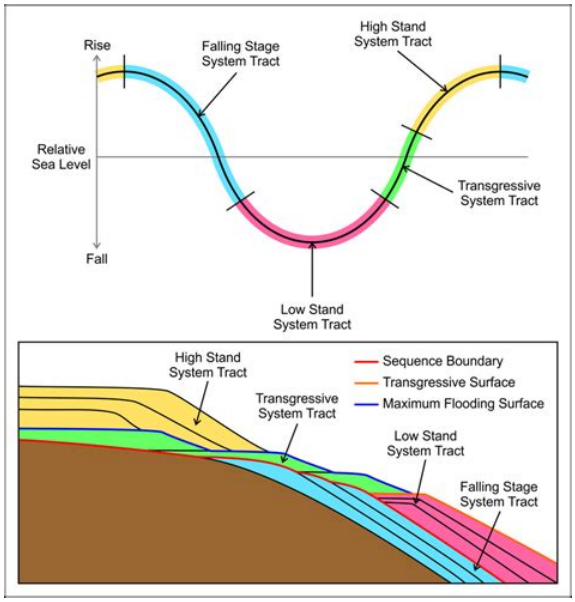


Figure 58.1.6: Sequence stratigraphy systems tracts. From sequence boundary to sequence boundary (Falling Stage Systems Tracts), a sequence repeats. This model provides a lens through which packages of rock strata, deposited along the margins of a basin, can be interpreted (Wright, 2013).

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THE PARASEQUENCE

The 5th Order Sequence, the parasequence, provides an accessible pattern for understanding all of this. These sequences represent some of the shortest depositional periods, the smallest areal extent, and the most rapid change in sea level, providing a nice laboratory for exploring how these systems tracts work. Two examples of parasequences, or parasequence sets (as in the second figure below) are provided. In both cases, parasequences represent a single episode of shoreline progradation, or seaward movement of the shoreline. This creates a typical shallowing-upward succession of strata typical of a more rapid filling of accommodation space than may have been created at the time. The first represents a set of facies that progressively become coarser in grain size as deposition occurs, which is typical in a deltaic environment along a coast. The sequence boundary at the bottom is an unconformity that represents the end of the falling stage systems tract from the prior sea level cycle. The fine-grained muds that begin the sequence represent the deeper waters of the offshore facies residing in deeper waters. As the sequence develops, a typically shallowing-upward prograding progression will develop. This is a predictable pattern for this kind of environment. Sea level is still transgressing, but sedimentation rates are very high.

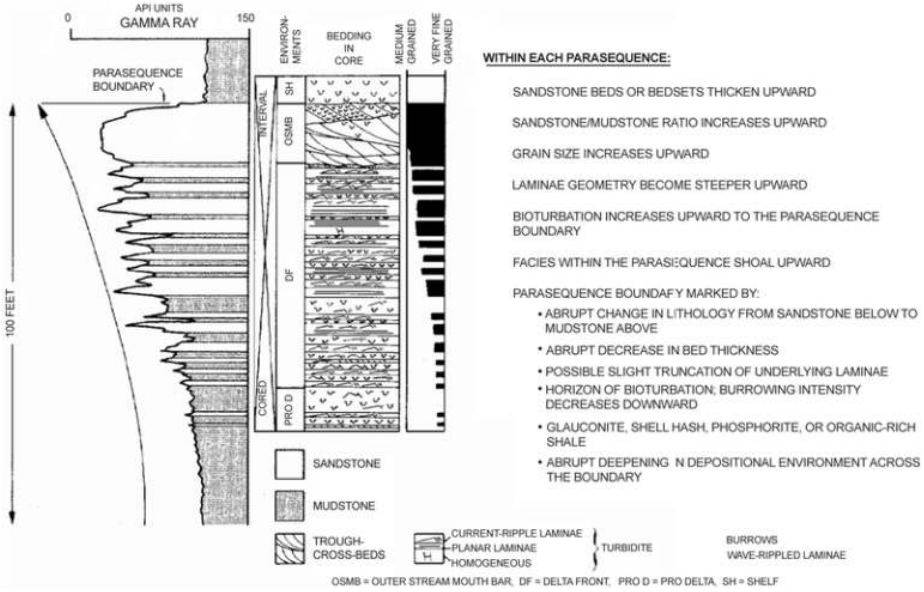


Figure 58.1.7: A Model Parasequence. Muds at the base (gray) represent the sea level highstand systems tract. As the parasequence deposits over time (bottom to top), sea level is rising as sediments are prograding seaward. Accommodation space in the basin is being filled faster than space is created. The sequence is typically shallowing-upward, or coarsening-upward.

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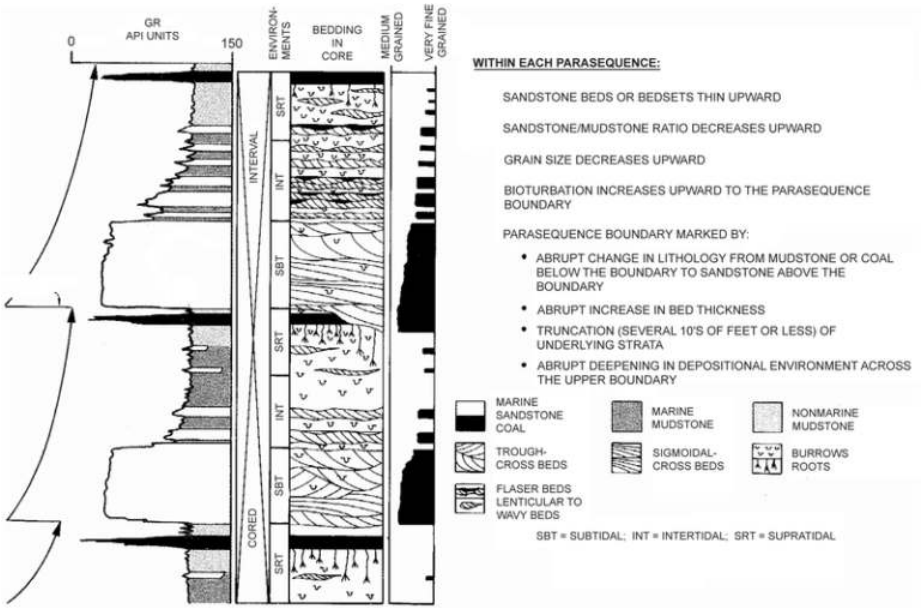


Figure 58.1.8: Characteristics of two upward-fining parasequences formed in a tidal flat to subtidal environment. (Van Wagoner et al., 1995).

In the field, tracing out sequence boundaries is challenging. Given the spotty nature of the rock record and the variation that occurs as beds pinch out or new beds emerge over distances, it is impossible to ever have all of the data necessary to reconstruct ancient environments or the entire history of a basin. These sequence stratigraphic tools have provided valuable help in identifying important deposits of economic importance and for better understanding how a wide variety of environmental processes work. Such insights are important as we strive to understand the processes we are experiencing today along our shorelines and beyond.

A deeper exploration of sequence stratigraphy is beyond the scope of this work. Introducing the concept of sequences in this chapter serves as a way to tie together all of the various lines of discussion surrounding stratigraphy. The use of basic stratigraphic principles to understand how beds are correlated across distances is foundational and descriptive. Sequence stratigraphy provides a set of models that allow for deeper interpretations of basin history. They also provide opportunities for further research. Testable questions abound through the application of these models of change within basins.

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CHAPTER OVERVIEW

59: (Tools of the Trade) Geologic structures

59.1: Tectonic Movement

59.2: Types of Stress

59.3: Deformation - folding

59.4: Deformation - Faults

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59.1: Tectonic Movement

The Earth's lithosphere is in constant motion at tectonic plate boundaries. The velocity of this movement ***on average*** is often compared with the growth rate of your fingernails. That does not sound like much but this movement has a dramatic effect on Earth's landscape. What we view as craggy mountain peaks and valleys have all been sculpted by the movement of tectonic plates and the work of weathering and erosion. The tectonic map of the Earth found below, provides directional arrows that indicate plate movement along the major boundaries.

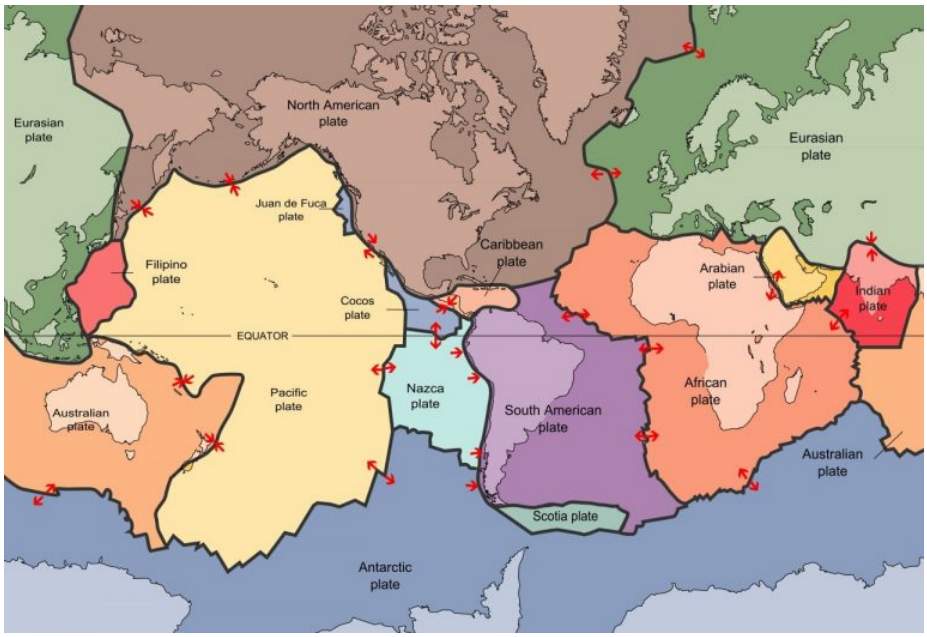


Figure 59.1.1: Tectonic Plate Boundary Map. Arrows at the plate boundaries indicate the direction of movement along that particular boundary. (Public domain; from Wikipedia.)

Most of the movement along the boundaries is imperceptible to humans in our lifespan. However, this tectonic movement over many thousands to millions of years does make major changes that account for features as dramatic as the Himalayan Mountains. Some of the movement can take place in virtually the blink of an eye. This happens when stress builds up as these solid plates attempt to re-equilibrate. The dramatic reaction to sudden release of stress results in earthquakes which can make visible changes to Earth's surface in the form of faults, or breaks, within the solid crust. Look at the amazing before and after Google Earth images of the displacement caused by the June 2019, 7.1 magnitude earthquake in Ridgecrest, CA. This GIF was created by using satellite imagery; in these photos, you are looking down on the land surface as if you are flying overhead in a plane.



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59.2: Types of Stress

The arrows that appear on the tectonic map above display three types of movement. These are the main types of stress that affect the rock of the lithosphere at plate boundaries. ***Compressional stress*** takes place where the arrows point toward each other and plates collide. This type of plate boundary is referred to as ***convergent*** because the plates are colliding. ***Tensional stress*** occurs where the arrows point away from each other and separation occurs between these plates. This type of plate boundary is referred to as ***divergent*** as the plates are moving away from each other. ***Shear stress*** results where plates attempt to slide past each other in opposite directions. This type of plate boundary is referred to as ***transform***.

Stress is the ***force*** being exerted on the rock at each of these boundaries. ***Strain*** is the ***physical change*** that results in response to that force. The visible strain that we see in the rock is called ***deformation***.

Let's explain this in visual terms relative to the before/after GIF from the Ridgecrest earthquake above. The Ridgecrest earthquake is related to movement along the transform plate boundary which slices through southwestern California. Movement along this plate boundary also created the infamous San Andreas fault. This transform plate boundary has created a ***system of faults*** as it tries to slide a solid slice of southwestern California, which sits on the Pacific Plate, past the rest of California which resides on the North American Plate. This movement produces shear stress. The force of the shear stress caused a fracture to form in the crust which is a fault. The fault is the strain that occured in response to the stress produced by the shearing force. This type of physical fracturing of Earth's crust is referred to as ***brittle deformation***.

When rock of the Earth's crust is subjected to increasing stress it passes through 3 successive stages of deformation:

- Elastic deformation where the strain is reversible.
- Ductile deformation (also referred to as plastic deformation); the strain is irreversible.
- Brittle deformation where the strain results in a permanent fracture of the rock.

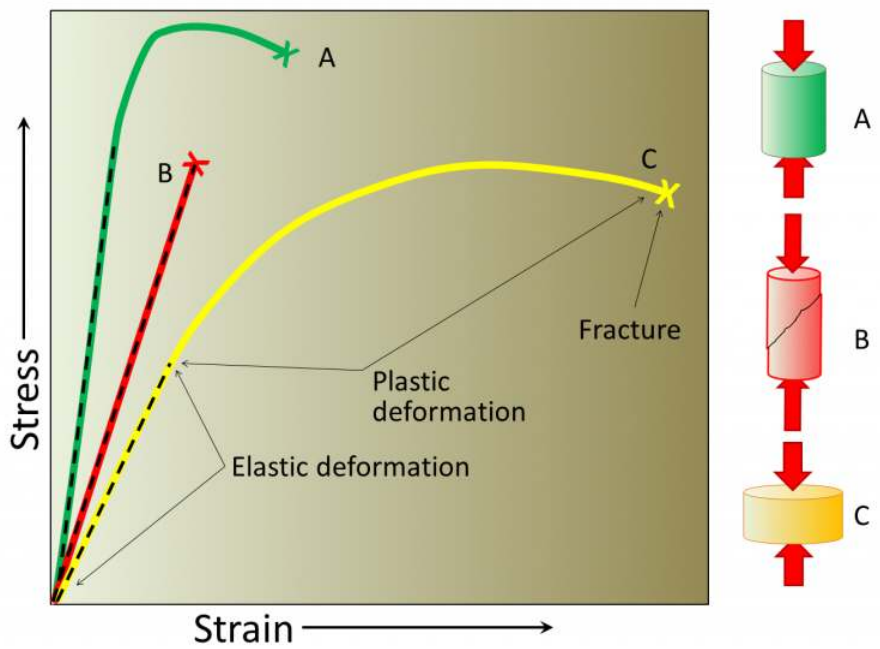


Figure 59.2.1: Different materials deform differently when stress is applied. Material "A" has relatively little deformation when undergoing large amounts of stress, before undergoing plastic deformation, and finally brittlely failing. The common metamorphic rock gneiss would fall into this category. Material "B" only elastically deforms before brittlely failing. The common igneous rock gabbro would fit here. Material "C" undergoes significant plastic deformation before finally failing brittlely. Most sedimentary rock types fit into this category. (CC BY 4.0; By Steven Earle from Physical Geology.)

Elastic deformation of the crust would not be noticed. Ductile deformation is where things become interesting in terms of visual effects displayed in solid rock (see photo below). When sufficient stress is applied, the response will be a change in shape. This type of deformation is permanent. Brittle deformation, as described above with respect to the Ridgecrest earthquake, occurs when the physical strength of the rock is surpassed and the rock will break resulting in a permanent fracture, or fault.



Figure 59.2.2: Compressive stress has produced folding in the layers limestone, Mt. Kidd, Alberta. The “upfolds” are geologic structures called anticlines and the “downfolds” are synclines. (With permission for educational purposes from Marli Miller Photography.)

Many factors contribute to how a rock will respond to applied stress including the composition and mineral structure of the rock and the temperature of the surrounding environment. Sedimentary rock often contain water in mineral structures like clay and in pore spaces between grains. This allows sedimentary rock to respond to applied stress by deforming ductilely. Most igneous rock, however, is composed of interlocking mineral crystals and contains very little water. The strong fabric of igneous rock will result in rupture to the same applied stress.

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59.3: Deformation - folding

Tectonic compressional stress is the dominant force in mountain building. As plates collide, compressional stress will cause the crust to buckle and warp to produce folds. A compressional fold belt can be found in all mountain ranges around the world which have been caused by tectonic plate collision. View the Google Earth image below of the Ridge and Valley region in the Appalachian Mountains of Pennsylvania. The rock in this region includes mostly ductile sedimentary rock layers which were folded during the collision of Africa with North America as the Pangean supercontinent was being pieced together. We're viewing this section of the Appalachians from an altitude of about 200 km. The folding pattern displayed is quite a spectacular feature seen from space.



Figure 59.3.1: Google Earth Image of the Appalachian Mountain fold belt in western Pennsylvania at 200 km eye altitude.

Fold Classification

Folds are classified based on their geometry and fall into 3 general categories: monoclines, anticlines and synclines.

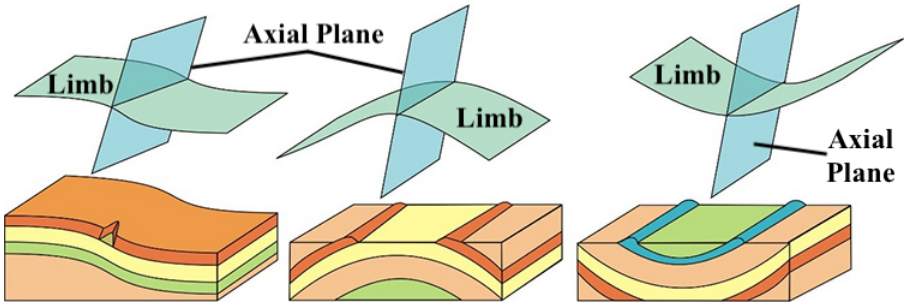


Figure 59.3.2: Monoclines, anticlines, and synclines are the three basic types of folds. Here they are shown in block diagrams, as well as in diagrams showing their essential geometry: the relationship of the strata to the axial surface.

Anticlines and Synclines

Anticlines (upfolds) and synclines (downfolds) are very common geologic structures that form in pairs in response to compression. Anticlines and synclines will share a limb of the fold. Click on the starred placemarks in the gigapixel image from eastern West Virginia, below, to learn more about specific characteristics of anticlines and synclines.



Monoclines

Monoclines (see image below) are not as common and usually form as an adjustment in the surface layers to some type of tectonic activity happening deep within the Earth's surface. An example of this would be an uplift of a block of deeply buried "basement" rock of the continental crust.



Figure 59.3.3: West Bighorn monocline in the Bighorn Range, Wyoming. (By Callan Bentley.)

Overturned and recumbent folds

The stress of tectonic compression is commonly applied more forcefully in one direction as one tectonic plate collides with another. The directionally applied stress may result in folding that appears to be "pushed" from one direction. When this happens, folds may

begin to turn over on themselves. A fold will become increasingly asymmetric as the axial plane tips toward 90° from vertical at which time it will be classified as recumbent.

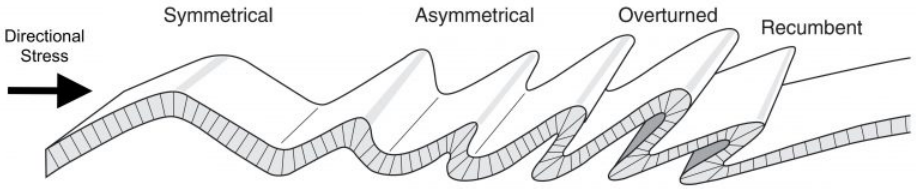


Figure 59.3.4: Directionally applied stress will result in folds becoming more asymmetrical as the axial plane dips away from the vertical. (Modified after unknown source.)

Overturned Fold



Figure 59.3.5: Overturned anticline-syncline pair in marble and quartzite. (By: Marli Miller Photography.)

Recumbent Fold



Figure 59.3.6: Recumbent Fold, Port au Port Peninsula, Newfoundland. (CC BY-NC 4.0; from: https://openeducationalberta.ca/introductorystructuralgeology/chapter/e-folded-surfaces/)

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59.4: Deformation - Faults

All rock types will ultimately reach a point of failure where they will break permanently and form a fault. Faults are classified based on the type of stress that produced them as well as the movement associated with the two separate blocks of rock. Deformational folding, as discussed above, is largely caused by compressional stress of tectonic collision. Faulting is also the result of tectonic movement however, different fault types are associated with the different types of plate boundaries. The three types of stress, ***compressional***, ***tensional***, and ***shear***, are associated with movement along the three different types of plate boundaries, ***convergent***, ***divergent*** and ***transform***.

Reverse and Thrust Faults

Geologists use block diagrams in an attempt to demonstrate three dimensional movement. The terms ***hanging wall*** and ***footwall*** refer to the relative position of the blocks after movement. Economic minerals, such as gold, often form along faults planes, and these terms come from where a miner would stand, and where they would hang their lantern as they were mining the minerals along the fault plane.

The block diagram displays horizontal sedimentary strata (layers, or beds, of sedimentary rock). When geologists look at these structures as they occur at a rock exposure, they trace ***marker beds*** to determine which block has moved and in what direction. Study the block diagram below of a reverse/thrust fault. Trace the rock layers from one side of the fault to the other.

A reverse fault is caused by compressional stress at convergent plate boundaries.

In a reverse or thrust fault, the hanging wall has moved up relative to the footwall.

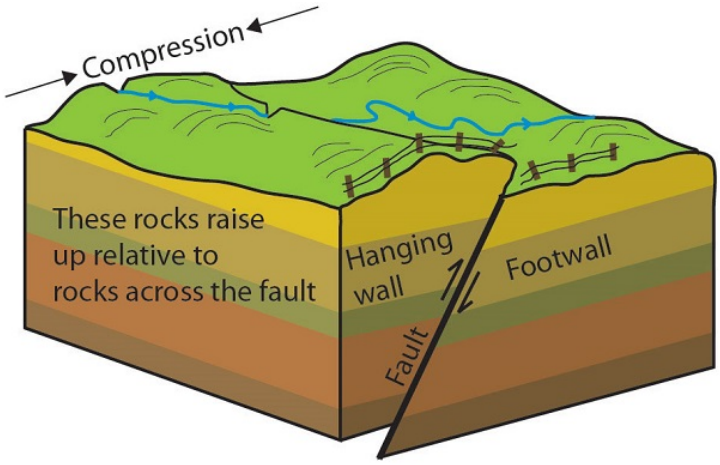


Figure 59.4.1: A reverse fault formed by the tectonic force of compression along a convergent plate boundary. (Public Domain; By: Trista L. Thornberry-Ehrlich, Colorado State University, From: National Park Service.)

The difference between a reverse and thrust fault is the angle of the fault plane. A reverse has a high angle fault plane (the surface separating the two blocks) while a thrust fault has a low angle fault plane. Take a look at the 3D models below to investigate. Once the models load, you can grab them to rotate in 3D. One common misconception by students in identifying faults is that the arrangement of the blocks is static as in the diagram above. This may lead a student to believe that the hanging wall is always on one side in a given type fault (such as on the left in the diagram above). Being able to look at these blocks in 3D allows the student to see that the hanging wall and footwall can be oriented in any direction as they occur in the Earth's crust.



Figure 59.4.2: Reverse Fault

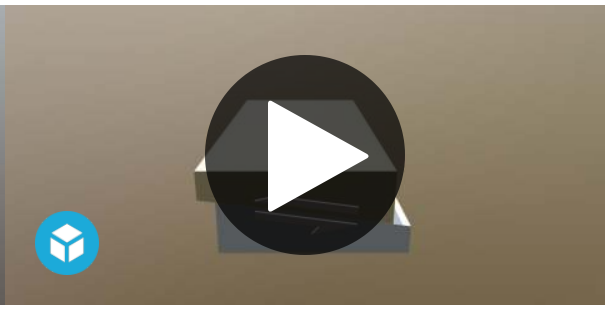


Figure 59.4.3: Thrust Fault

Examples of a Reverse and Thrust Fault, below, to compare with the 3D images above.

Reverse Fault



Figure 59.4.4: Ketobe Knob in the San Rafael Swell of Utah displays an example of a thrust fault. (CC BY-NC 4.0 International; Ketobe Knob photo by Ron Schott, From: Flickr)

Thrust Fault



Figure 59.4.5: Annotated Thrust Fault from southwest Montana. (With permission for educational purposes by: Marli Miller Photography.)

Normal Faults

A normal fault is caused by extensional (tensional) stress at divergent plate boundaries as two plates are moving in opposite directions, away from each other.

In a normal fault, the hanging wall has moved down relative to the footwall.

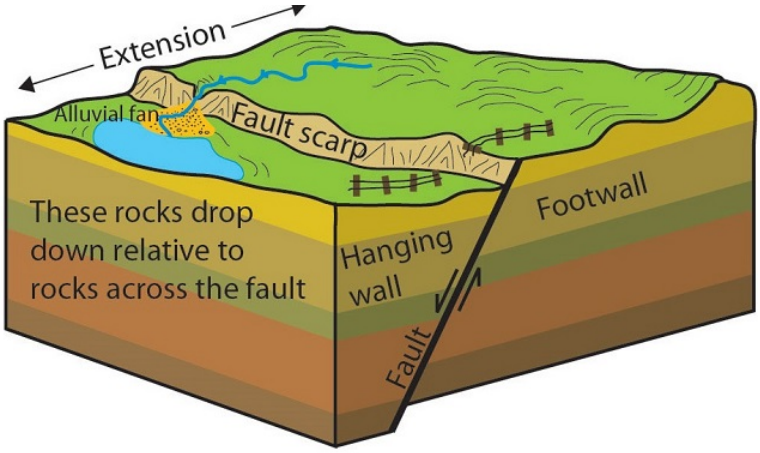


Figure 59.4.6: A normal fault formed by the tectonic force of extension (tensional stress) at a divergent plate boundary. (Public Domain; By: Trista L. Thornberry-Ehrlich, Colorado State University. From: National Park Service.)

Explore this 3D block structure of a normal fault. Once the models load, you can grab them to rotate in 3D. One common misconception by students in identifying faults is that the arrangement of the blocks is static as in the diagram above. This may lead a student to believe that the hanging wall is always on one side in a given fault (such as on the left in the diagram above). Being able to look at these blocks in 3D allows the student to see that the hanging wall and footwall can be oriented in any direction as they occur in Earth's crust.



Figure 59.4.7: Normal Fault



Figure 59.4.8: A normal fault near Tblisi in Georgia. Note marker beds on the right (the hanging wall block) have moved down along along the fault plane relative to the left (footwall). (CC BY 3.0; Credit: Luka Adikashvili from imaggeo.egu.eu)



Figure 59.4.9: Normal fault from Lima Peru. Notice that the marker bed on the right (the hanging wall block) has moved down along along the fault plane relative to the left (footwall). (CC BY 2.0; By: Miguel Vera León on Flickr.)

Strike-Slip Faults

Shear stress at a transform plate boundary will produce a strike-slip fault. The fault is named strike-slip because the fracture is along the Earth's surface and the movement between the blocks is not up or down but "along strike" – a line with a specific direction along Earth's surface – the line of strike – which is also known as the "fault trace." The fault plane is typically vertical into the crust and the "slip," or movement, occurs along the plane. The strike slip fault in the diagram is labeled a "right-lateral" strike-slip fault. If you are standing on one side of the fault trace and looking across to the other side, the movement would appear to be laterally to the right. Note how the stream jogs to the right along the fault trace. In a left-lateral strike-slip fault, the blocks would be moving in the opposite direction.

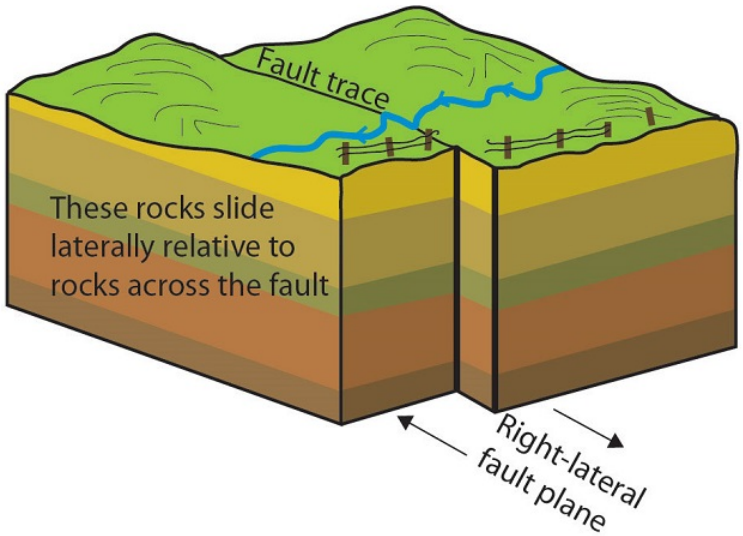


Figure 59.4.10: A strike-slip fault formed by shear tectonic force along a transform plate boundary. (Public Domain; By: Trista L. Thornberry-Ehrlich, Colorado State University. From: National Park Service.)

Let's take a look at another one of these fantastic before/after Google Earth image GIFs from the Ridgecrest, CA earthquake. This GIF was created by using satellite imagery; in these photos, you are looking down on the land surface as if you are flying overhead

in a plane. As you analyze the movement, put yourself on one side of the fault and look across to the the other side. Is this a right-lateral or left-lateral strike-slip fault? Find out below.



Did I Get It? - Quiz

? Exercise 59.4.1

Is the Google Earth image GIF in the text a left-lateral or right lateral strike-slip fault?

a. Left-lateral strike-slip fault

b. Right-lateral strike-slip fault

c. It's nobody's fault!!! :-)

Answer

b. Right-lateral strike-slip fault

The photo in the aerial photo below is of a left-lateral strike-slip fault. Note: if you were standing where the word fault is written and looking across the fault trace to the other side, the motion along fault plane would be to the left.



Figure 59.4.11: Satellite image of the Piqiang Fault, a northwest trending left-lateral strike-slip fault in the Taklamakan Desert south of the Tian Shan Mountains, China. (Public Domain; NASA Earth Observatory images by Robert Simmon and Jesse Allen, using Landsat data from the USGS Earth Explorer.)

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CHAPTER OVERVIEW

60: (Tools of the Trade) Geologic maps

60.1: Introduction
60.2: History of geologic maps
60.3: Elements of a geologic map
60.4: Orientation of structures
60.5: Folds and faults
60.6: The map's "explanation"
60.7: Examples to help train your eye
60.8: Practice
60.9: Conclusion

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60.1: Introduction

Geologic maps are maps that depict the rock units that crop out at Earth's surface. Typically, they use different colors (or different fill patterns) to distinguish between different geologic units (or **formations**). Units (members, formations, groups, supergroups, etc.) meet at **contacts**, which can be of several varieties. To make relationships between rock units more clear, many geologic maps include cross-sections, which show a conceptual "slice" through the Earth along a straight line or multi-segment "polyline" on the map. The patterns that rock units (formations) make can convey key information about the geologic history of a region.

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60.2: History of geologic maps

The oldest known use of mapping to depict the distribution of rock types on Earth's surface was the Turin Papyrus Map, made in 1150 BCE in central eastern Egypt. It shows multiple rock types by virtue of what color their mountainous outcrops appear. Landmarks such as gold mines, shrines, and roads are also shown:

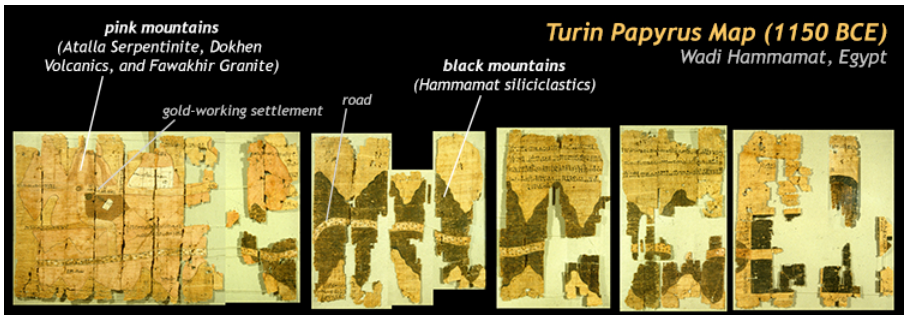


Figure 60.2.1: The Turin Papyrus Map is sometimes cited as the oldest example of a geological map.

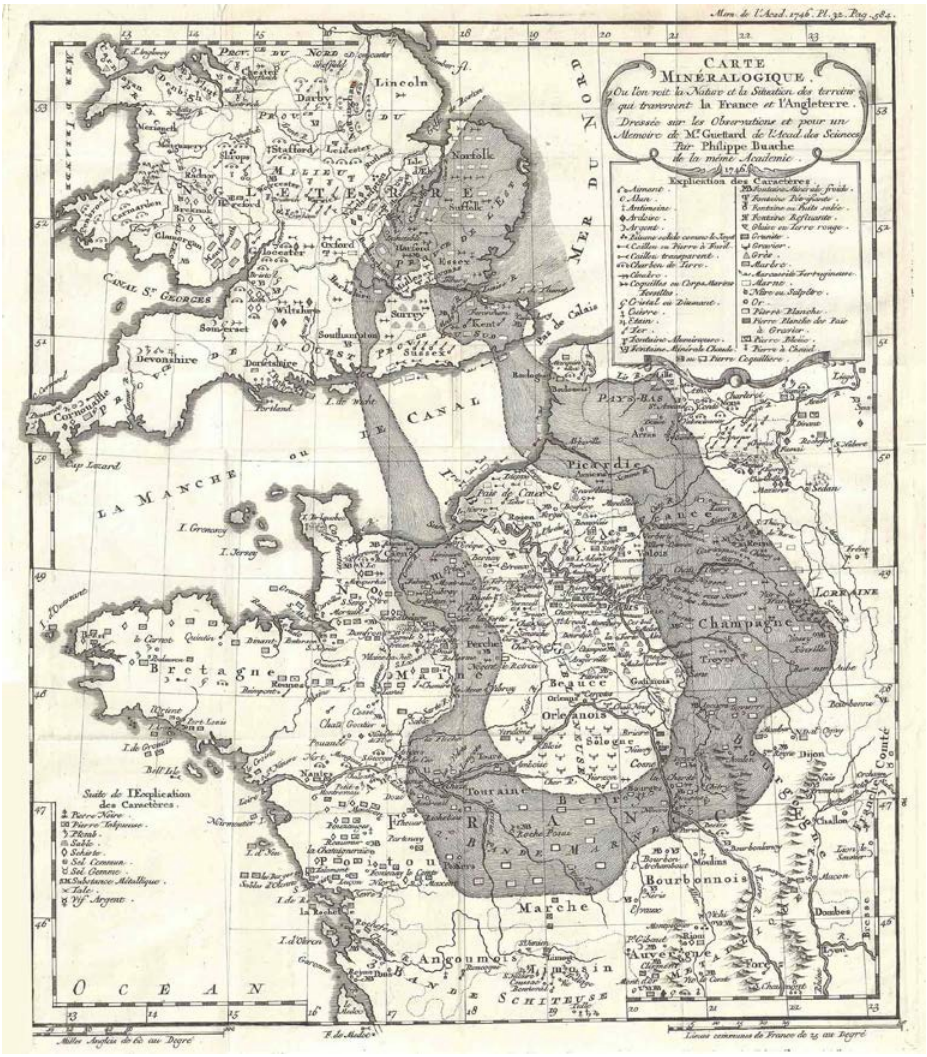


Figure 60.2.2: Jean-Étienne Guettard and Philippe Buache's map showing the distribution of chalk around the Paris Basin (and extending across the English Channel into southern Britain), 1746.

This innovative use of landscape color was forgotten by the time French geologists Jean-Étienne Guettard and Philippe Buache published their map of chalk deposits in 1746. However, it was still an important milestone. This map summarizes in a single image a large amount of geological information about a region. As a matter of historical significance, it shows for the first time the

distribution of a single geological unit across space; indeed across multiple countries. In this case, the lumpy gray “doughnut” shape shows the area where surface outcrops of “the chalk” (namesake of the Cretaceous) may be found in France and England. It implies that strata in the “doughnut hole” are younger than the chalk, and strata outside the ring’s periphery are therefore older than the chalk. This map is an important milestone, but it’s not what most geologists would instantly recognize as a geologic map because it emphasizes just one unit, excluding all others, and because it uses only one color.



Figure 60.2.3: William Smith’s 1815 geologic map of Britain, a major milestone in geologic mapping.

The next step forward was taken in England, where geologist William Smith is renowned among geologists for his achievement of depicting more than one formation in a single map. In fact, Smith depicted a whole country’s worth of formations, and rendered the first ‘modern’ geologic map. He made the decision to represent the surface occurrence of different rock units by depicting them in different colors, a practice that is still used today.

Smith gathered his data through his day job as a canal engineer. His career occurred during the heyday of British canals, and Smith was on the front lines as new trenches were being dug. As excavation revealed rock, he noted whether the strata were sandstone, shale, limestone, conglomerate, or coal. In particular, he paid attention to fossil content such as ammonites, trilobites, and brachiopods. Correlating the strata on the basis of these fossil “timestamps,” he was able to match up outcrops across an entire nation.

His map elegantly summarizes years of field work, and thousands of outcrops into a single picture. It’s a marvelously data-rich image. The dark gray units are the Carboniferous “coal measures.” The luminescent yellow stripe down the middle is a belt of honey-colored Jurassic limestones. Likewise, each other color corresponds to a particular period of geologic time, marked by a single type or rock or group of related strata.

The other important innovation that Smith achieved was to depict a simplified geologic cross section, showing how the various strata related to one another, from the perspective of a “side view” looking north at an east-west cross-section through the entire island.

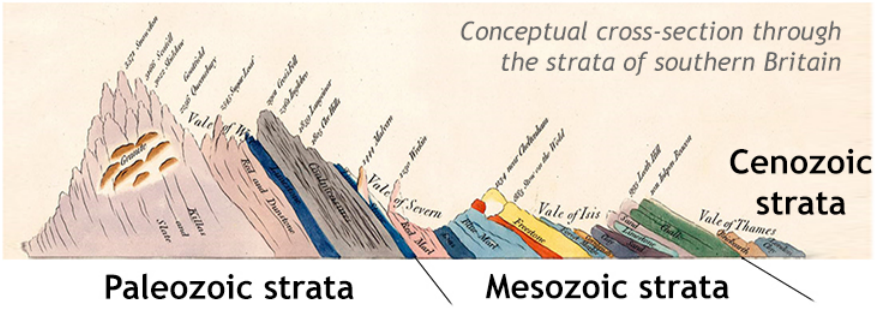


Figure 60.2.4: William Smith's "Sketch of the succession of strata and their relative altitudes," an inset on his geological map of England and Wales (with era names added).

Though this perspective is impossible to physically achieve in real life, to visualize the cross-section in our imaginations is illuminating; it shows at one glance the overall structure of the island: a series of strata dipping off to the east. It is rather similar to the "layer cake" of the Grand Canyon, but tipped over to one side.

Since Smith's innovations, geologists have continued to draw cross-sections, and continued the practice of using color (or patterns) to distinguish different geologic units or formations. In the modern day, the National Geologic Map Database provides an elegant visual organization to U.S. geologic maps (although the current map display is Flash-based, so will be incompatible with some browsers). Back in William Smith's home, the British Geological Survey has a similar browser-map-based database search with their Geology of Britain viewer. Globally, OneGeology offers a global compilation of maps, again with a user-friendly, map-based interface that they call the OneGeology Portal.

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60.3: Elements of a geologic map

Given that all these maps of all these different places depict a great variety of different rocks, what do they all have in common? In general, geologic maps consist of several key elements, or components. They include:

- A **base map**, usually a topographic map of the region. Topographic maps depict the shape of the land using contour lines of a set elevation.
- The usual stuff that goes along with a topographic base map, such as a scale, a reference map, latitude and longitude, and magnetic declination (the angle of difference between magnetic north and geographic north).
- Distinct geologic **units**, differentiated from one another by some distinction, like different colors or different fill patterns, as well as an abbreviated identifier code. For instance, the Cambrian-aged Weverton Formation might be tan in color, and would be abbreviated $\mathbb{C}_{\mathrm{W}}$.
- **Contacts** between different geologic formations. These contacts might be comformable stratigraphic contacts, unconformities, faults, or intrusions.
- **Symbols** showing the *measured orientation* of rock structures, such as bedding, foliation, and faults.
- Other **symbols** showing the *interpreted structure* of the region, such as anticlines, synclines, normal faults, thrust faults, and transform faults.

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60.4: Orientation of structures

Knowing the position of geologic structures is important for geologic mapping. These structures include sedimentary features such as beds, metamorphic features such as foliation, and igneous features such as dikes. Their positions in space can be measured in the field and recorded accurately on the map using symbols and numbers.

There are several approaches to expressing this orientation, but the basics are that we want to (a) describe the position of the structure relative to the compass (north, south, east, west, etc.) — we call this 'declination' — and (b) we want to be able to describe how inclined that structure is relative to horizontal (*i.e.*, in reference to the Earth's surface) — this is its 'inclination' or 'dip.' We measure the declination using either 'dip-direction' or 'strike.'

Most students find it easier to think about measuring bed orientation in "dip and dip-direction," but many professors have been trained with "strike and dip." Let's briefly examine both.

Dip & dip-direction (vs. strike & dip)

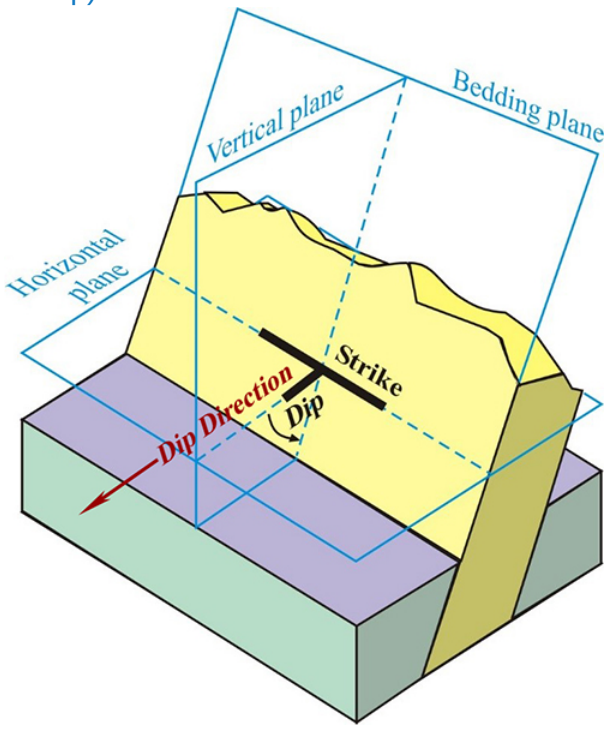


Figure 60.4.1: A diagram to show the relationship between different planes, resulting in the "T" shape of the strike & dip symbol.

This block diagram shows the relationship of a plane-shaped feature of interest (in this case, a bed of yellowish rock, perhaps a sandstone) with two planes that intersect it: one is a horizontal plane and one is vertical. The line of **strike** is the intersection between the bedding plane and the horizontal plane. The **dip-direction** is perpendicular to the strike: it is a line at the intersection of a vertical plane and the horizontal plane. It's the direction in which the rock dips: if you were to tilt a bottle of water over the outcrop surface, the little stream of water would run downhill in the direction of dip. But which way is that?

Both strike and dip-direction are compass directions, measured with reference to north. They are most efficiently expressed as a three-digit number that varies between 000° and 360°. This is "azimuth notation." Some geologists prefer "quadrant notation;" to measure east or west of north or south by the appropriate number of degrees and then express that as something like "N40°W," which would mean "40° west (counterclockwise) from north." That same orientation could be described in azimuth as "320°" (that is the 360° of true north minus 40°.)

If that sounds confusing, don't worry: On geologic maps none of those numbers appear. Instead, the orientation of the line of strike is simply shown graphically. Usually "north" is at the top of the map page, so a line running from the top of the page toward the bottom would be a northerly strike. From a north-south strike, the beds could dip either to the east, or to the west. Those are the

only two directions that are perpendicular to north. The dip-direction is always perpendicular to the strike. The little dip-tick line (the "stem" of the T shape, jutting out from the strike line at a right angle) shows dip-direction.

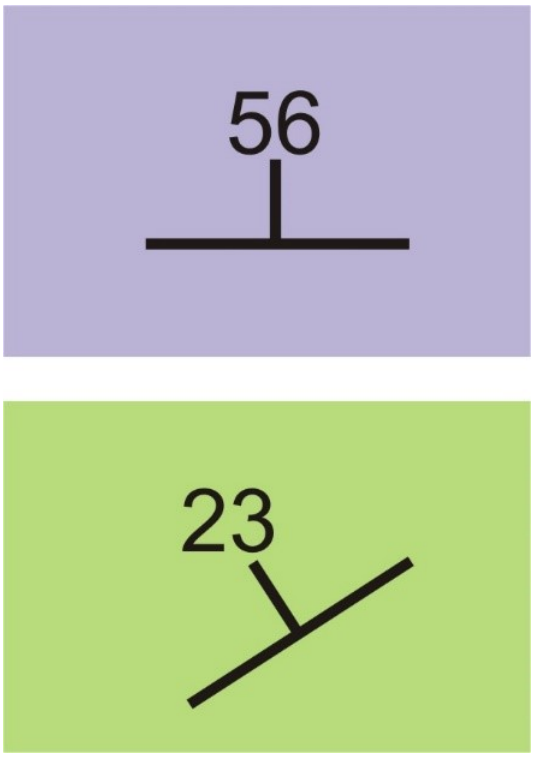


Figure 60.4.2: Two examples of strike and dip (dip and dip-direction) symbols: In the upper example, the beds dip 56° to the north. In the second example, strata dip to the northwest (toward ~330°) at a shallower angle, only 23°. In the first example, the beds strike east-west (090°). In the second example, the strike is northeast-southwest (~240°).

The **dip** itself is the *angle* of inclination, measured from horizontal down to the bedding plane. It is most efficiently expressed as a two-digit number that varies between horizontal (00°) and vertical (90°). This number is written at the end of the dip-tick line, as shown here in the two examples of strike and dip symbols as they appear on geologic maps. Note that nowhere is the dip-direction or the strike direction written down as numbers. Instead, they are depicted graphically with the orientation of the T shape of the symbol. The long line (the top of the "T") is the line of strike. The short line that comes out of it (the stem of the "T") is the dip-direction. The number at its end is the dip angle in degrees from horizontal.

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60.5: Folds and faults

Folds and faults are deformational structures, also called tectonic structures. They form as a consequence of tectonic deformation of pre-existing rocks. Where we have tectonic compression, we get folds and thrust faults. Where we have tectonic extension, we get normal faults.

Folds

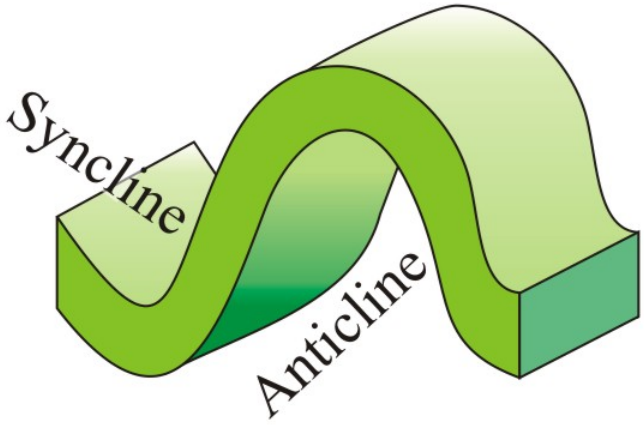


Figure 60.5.1: Anticlines and synclines are different aspects of the same crumpling of a sedimentary layer.

Rock strata can be folded under compressional stresses applied over a long period of time. Generally, we recognize two major categories of folds, and one minor category: **anticlines, synclines,** and **monoclines**. Anticlines and synclines occur together, and share a "limb" in common. Where you find anticlines, you often find synclines, too.

Monoclines are features associated with the adjustment of sedimentary layers over top of basement uplifts. They are folds where the layers go from horizontal (on top of the uplift) to tilted, to horizontal again (next to the uplift). They are reasonably common in the Rocky Mountain west and Colorado Plateau region of North America, but not really anywhere else. Anticlines and synclines, in contrast, have a global distribution anywhere there is a mountain belt.

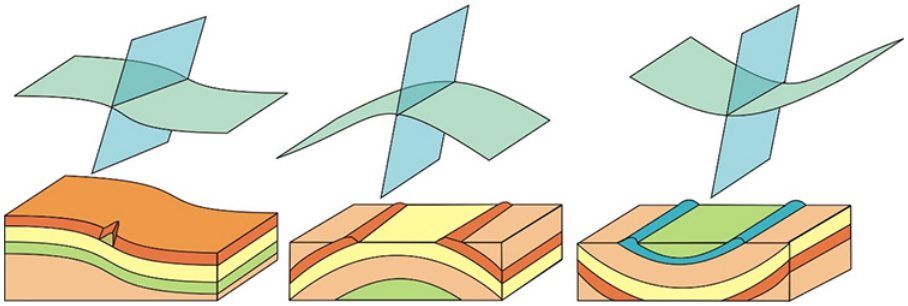
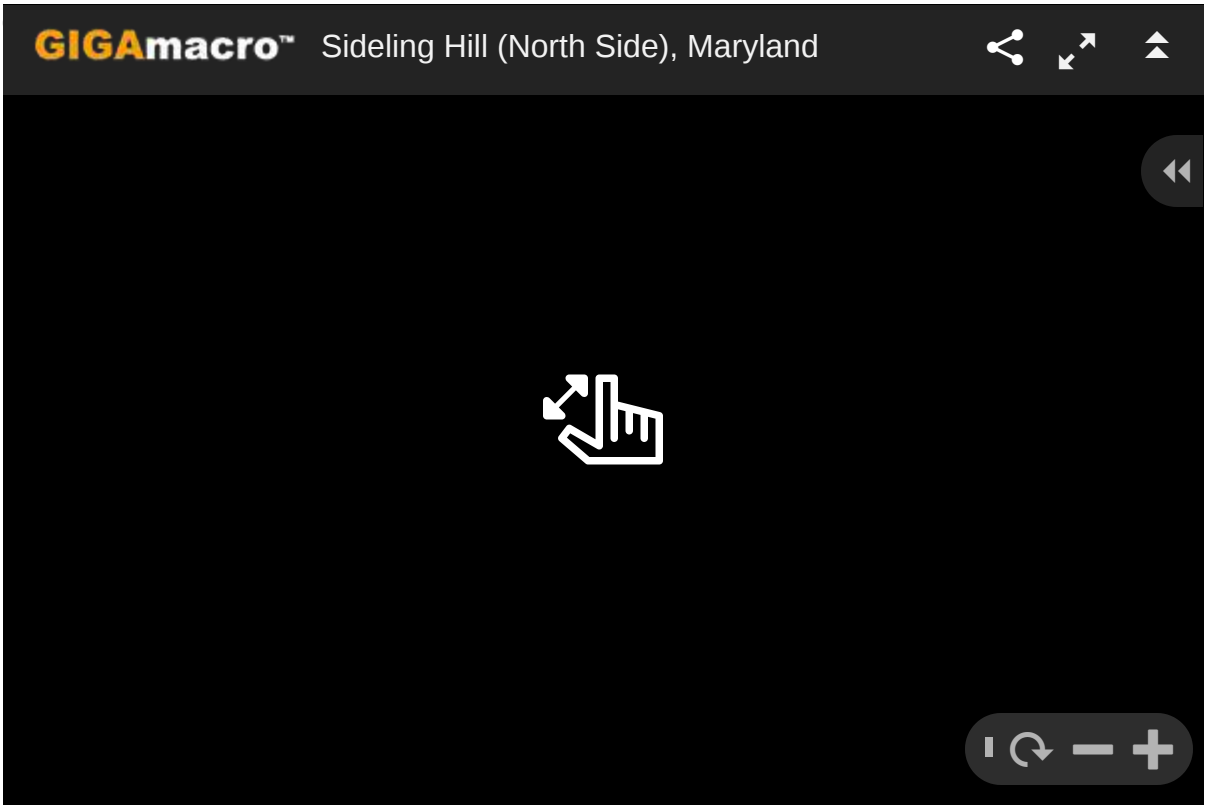


Figure 60.5.2: Monoclines, anticlines, and synclines are the three basic types of folds. Here they are shown in block diagrams, as well as in diagrams showing their essential geometry: the relationship of the strata to the axial surface.

It's important to note that just because a fold goes down in the middle (*i.e.*, a syncline), that doesn't mean the landscape will also go down (*i.e.*, a valley). The land surface's shape depends on many variables, including the very important role played by the different rates at which different rock units weather.

For instance, Sideling Hill in Maryland is a mountain ridge (a "hill," a topographic high) but a syncline (a structural low):



It's only a mountain ridge because the strata at the center of the syncline are quartz-rich and therefore weather more slowly than the silty strata beneath them. The opposite situation is just as common: anticlines with easily-weathered material at the center are etched out so the structural high is weathered out to form a topographic low. Bottom line: There is no guaranteed relationship between the structure of a fold and the shape of the overlying landscape.

Faults

Faults are breaks in the crust along which movement has occurred. There are three main sorts of movement we observe on faults:

(a) the fault plane is dipping at some non-vertical angle, and the upper block of rock has slid down the fault surface under the influence of gravity. The movement is downward along the dip line of the fault, and is called **normal faulting**. It occurs in situations of tectonic extension, such as at divergent plate boundaries. Low-angle (close to horizontal) normal faulting is called "detachment faulting."

(b) the fault plane is dipping at some non-vertical angle, and the upper block of rock has moved up the fault surface, getting shoved into position by tectonic compression. The movement is upward along the dip line of the fault, and is called **reverse faulting**. It occurs in situations of tectonic convergence, such as mountain belts and subduction zones. Low-angle (close to horizontal) reverse faulting is called "**thrust faulting**."

(c) the fault plane is vertical, and the two blocks of rock move laterally past one another in opposite directions. These are "strike-slip" faults, and they occur at transform plate boundaries. All the slip is parallel to the strike of the fault. **Strike-slip faults** can be either right-lateral or left-lateral. If you imagine standing on one side, and looking across the trace of the fault at the opposite side, and the rock over there looks like it's moving to the left, then the fault is left-lateral. To an observer on the opposite side, it will look like you are moving to *their* left!

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60.6: The map's "explanation"

It’s not a “legend.” It’s not a “key,” though those are common terms for similar features. On a geologic map, we call the area where the various units are named, described, and their colors and symbols defined by the name “explanation.”

Consider this example, which is a close-up look at one of the different units depicted on the geologic map of Pennsylvania: Ss is the symbol for the Shawangunk Formation, depicted on the map in a dark brown color. You can read a description for the formation, including its four sub-units, called members, as well as the other geologic formations with which they have been correlated (and scroll around to see a few more examples):



Figure \(\PageIndex{1}\): Berg, T.M., Edmunds, W.E., Geyer, A.R., Glover, A.D., Hoskins, D.M., MacLachlan, D.B., Root, S.I., Sevon, W.D., and Socolow, A.A., 1980, Geologic map of Pennsylvania (2nd ed.): Pennsylvania Geological Survey, Map 1, scale 1:250,000. Explore more here and here.

Various abbreviations are used for the various named portions of geologic time:

Table \(\PageIndex{1}\)

Geologic time unit (era or period)	**Abbreviation**	**Geologic time unit** (era or period)	**Abbreviation**
Archean	A	Pennsylvanian	PP
Paleoproterozoic	X	Permian	P
Mesoproterozoic	Y	Triassic*	T\(_R\)
Neoproterozoic	Z	Jurassic	J
Cambrian*	Ȼ	Cretaceous*	K
Ordovician	O	Tertiary* (archaic)	T
Silurian	S	Paleogene	Pg
Devonian	D	Neogene	Ng
Carboniferous*	C	Quaternary	Q
Mississippian	M		

*A few of these are a little confusing: Carboniferous got the “C” first, so Cambrian got a C with a slash through it, “Ȼ,” and poor Cretaceous was stuck with “K.” As far as Ts go, Tertiary got the T first, which left Triassic to be symbolized with a T that had a subsidiary R coming off its stem. However, when “Tertiary” was officially stricken from the ICS geologic timescale in 2003, to be partially replaced by the Paleogene, that freed up the solo T, which is now officially designated as symbolizing Triassic. However, a lot of old maps use “T” to mean Tertiary, not Triassic. So that’s confusing! There were already a bunch of Ms and Ps by the time it came to subdivide the Proterozoic, so that’s the reason for the random X, Y, and Z there.

Various symbols are used to show all these features on the geologic map:

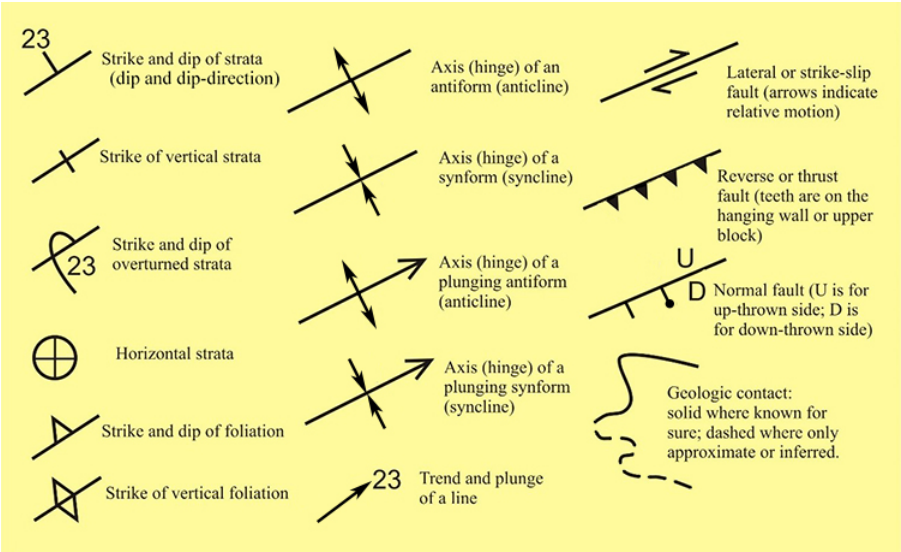


Figure \(\PageIndex{2}\): A summary of some different symbols used on geologic maps.

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60.7: Examples to help train your eye

Example 60.7.1: Horizontal strata

When strata are horizontal, you only see the top layer unless erosion has cut down through it (in stream channels, say), exposing deeper layers. The overall pattern is that the geologic contacts follow the contour lines, and don't vary much in elevation. In areas of dendritic drainage, such as the Atlantic Coastal Plain, the Great Plains, or the Colorado Plateau, this produces a distinctive branching pattern to the geologic map, as in these two examples:



Figure 60.7.1: Bick, K.F., and Coch, N.K., 1969, Geology of the Williamsburg, Hog Island, and Bacons Castle quadrangles, Virginia: Virginia Division of Mineral Resources, Report of Investigations 18, scale 1:24,000. Explore more here.

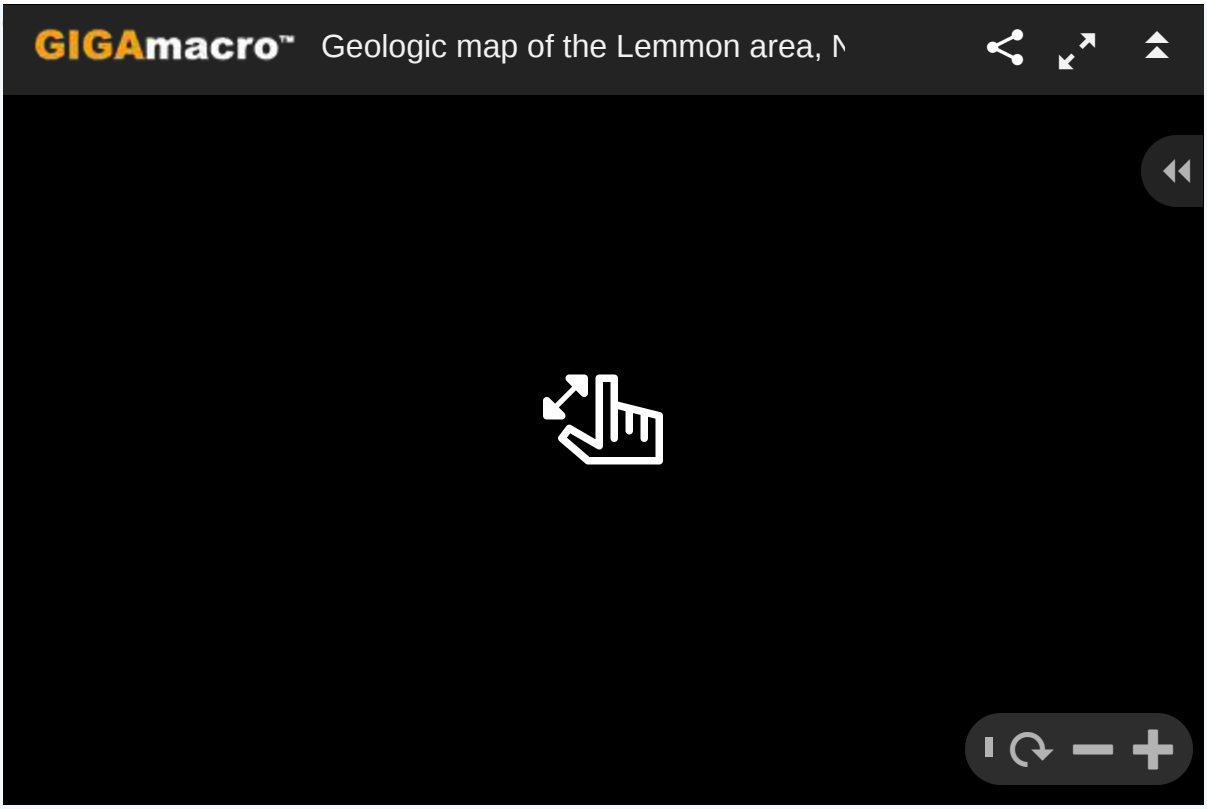


Figure 60.7.2: Sawyer, J.F., and Fahrenbach, M.D., 2011, Geologic map of the Lemmon 1-degree x 2-degree quadrangle, South Dakota and North Dakota: South Dakota Geological Survey, Geologic Quadrangle Map GQ250K-1, scale 1:250,000. Explore more here.

Example 60.7.2: Angular unconformities

Angular unconformities are as distinctive on geologic maps as they are in cross-section: two sets of formations meet each other at an abrupt angle. The lower, older formations will typically be folded or tilted, while the upper, younger formations will typically be close to horizontal.



Figure 60.7.3: Haley, B.R., Glick, E.E., Bush, W.V., Clardy, B.F., Stone, C.G., Woodward, M.B., and Zachry, D.L., 1993, Geologic map of Arkansas: U.S. Geological Survey, scale 1:500,000. Explore more here.

Example 60.7.3: Folds

So how do you recognize anticlines vs. synclines on geologic maps? There are two main techniques: (1) look for strike and dip symbols, and (2) look for age patterns. In an anticline, the strata dip outward away from the axis (hinge) of the fold. The anticline's oldest strata will appear in the middle, and younger strata will appear on the limbs. In a syncline, the strata dip inward toward the axis (hinge) of the fold. The syncline's youngest strata will appear in the middle, and older strata will appear on the limbs.

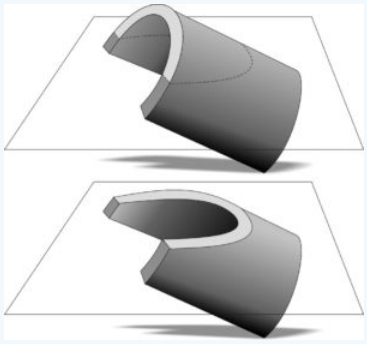


Figure 60.7.4: Plunging folds, like this anticline, produce V-shaped outcrop patterns.

When a fold plunges into the Earth at some angle, its outcrop pattern changes from a series of parallel stripes to a series of nested "V" shapes. As with non-plunging folds (those that have horizontal hinges), in a plunging anticline, the oldest strata are in the middle, and the strata dip outward on the two limbs. And as with a with non-plunging synclines, plunging synclines also have the oldest strata on the limbs, and the strata dip inward toward the youngest strata in the middle.

In this example, you can tell it's an anticline, because it has older strata in the middle (Devonian, D) and younger strata on the limbs (Mississippian and Pennsylvanian, M and PP):



Figure 60.7.5: Berg, T.M., Edmunds, W.E., Geyer, A.R., Glover, A.D., Hoskins, D.M., MacLachlan, D.B., Root, S.I., Sevon, W.D., and Socolow, A.A., 1980, Geologic map of Pennsylvania (2nd ed.): Pennsylvania Geological Survey, Map 1, scale 1:250,000. Explore more here and here.

In contrast, here (same map, different fold), you can see the pattern is reversed. This allows you to deduce that it's a syncline, because Mississippian strata are the middle (younger) and Devonian and Silurian strata occur on the limbs of the fold (older).



Figure 60.7.6: Berg, T.M., Edmunds, W.E., Geyer, A.R., Glover, A.D., Hoskins, D.M., MacLachlan, D.B., Root, S.I., Sevon, W.D., and Socolow, A.A., 1980, Geologic map of Pennsylvania (2nd ed.): Pennsylvania Geological Survey, Map 1, scale 1:250,000. Explore more here and here.

In this example, you can see the strike and dip symbols more or less parallel to the contacts between the beds, and dipping westward on the west limb, and eastward on the east limb. This tells you it must be an anticline:



Figure 60.7.7: Reynolds, M.W., and Brandt, T.R., 2006, Preliminary geologic map of the Townsend 30′ x 60′ quadrangle, Montana: U.S. Geological Survey, Open-File Report OF-2006-1138, scale 1:100,000. Explore more here.

Example 60.7.4: Faults

Here are two examples showing normal faults...

The first one is marked with D/U notation showing which side went down and which went up:



Figure 60.7.8: Wilson, E.D., Moore, R.T., and Cooper, J.R., 1969, Geologic map of Arizona: Arizona Bureau of Mines, scale 1:500,000. Explore more here.

...And this one shows the alternate style, with a little dip-tick coming off the fault trace, on the down-dropped side. Sometimes, as with this example, the mapmaker puts a little “ball” at the end of the tick-mark, making it look like a lollypop:



Figure 60.7.9: Wolfe, E.W., and Morris, Jean, 1996, Geologic map of the Island of Hawaii: U.S. Geological Survey, Miscellaneous Geologic Investigations Map I-2524-A, scale 1:100,000. Explore more here.

With thrust faults, a saw-toothed pattern is used. The little triangular “teeth” are on the upper block of rock. There are a lot of examples in this map!



Figure 60.7.10: Mudge, M.R., 1972, Pre-Quaternary rocks in the Sun River Canyon area, northwestern Montana: U.S. Geological Survey, Professional Paper PP-663-A, scale 1:48,000. Explore more here.

Example 60.7.5: Structural domes & basins

Unlike anticlines and synclines, structural domes and basins are areas where the strata are flexed around a central point, rather than a line (fold axis or hinge). In map view, they appear like "bull's eyes" with a series of concentric rings of strata. If the strata dip outward away from that central point, then it is a structural dome. Structural domes expose their oldest strata in their center, while strata get younger toward the outer edges of the structure. The reverse is true for structural basins: they are structurally lowest in the middle, so all the strata dip inward toward that "deepest" part of the structure. This means that structural basins expose their ***youngest*** strata in their center, while strata get ***older*** toward the outer edges of the structure.

Here is an example from Utah: the San Rafael Swell is a structural dome. If you zoom in, you'll see the pale blue strata exposed in the middle are Permian, with strata getting younger (Triassic, then Jurassic, and then Cretaceous) as you work your way outward toward the edges.



Figure 60.7.11: Hintze, L.F., Willis, G.C., Laes, D.Y.M., Sprinkel, D.A., and Brown, K.D., 2000, Digital Geologic Map of Utah: Utah Geological Survey, Map 179DM, scale 1:500,000. Explore more here.

Another example, of the opposite situation: here, in northern New Mexico, is a structural basin, with the youngest strata (Tsj) is surrounded by progressively older and older strata (Tn, Toa), into the Cretaceous series (they start with K):



Figure 60.7.12: Anderson, O.J., and Jones, G.E., 1994, Geologic map of New Mexico: New Mexico Bureau of Mines and Mineral Resources, Open-File Report 408, scale 1:500,000. Explore more here.

As a friendly reminder, the 1746 Guettard / Buache map of the chalk outcrops around the Paris Basin was another example of a structural basin.

Now that we've covered these basics, it's time to test yourself...

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60.8: Practice

To hone your skills with geologic maps consider these three problems, take this self-quiz (5 questions):

Geologic map self-quiz

Test your understanding of basic geologic map patterns with a handful of questions.

? Exercise 60.8.1

What major feature(s) is/are depicted in this portion of the North American geologic map?



a. Lots of normal faults

b. Folding

c. Mostly horizontal strata

d. An angular unconformity

e. A structural dome

f. A structural basin

g. Lots of thrust faults

h. Strike slip faulting

Answer

g. Lots of thrust faults

Exercise 60.8.2

What major feature(s) is/are depicted in this portion of the North American geologic map?



a. Strike-slip faulting

b. A structural basin

c. Angular unconformity

d. Mostly horizontal strata

e. Lots of normal faults

f. Lots of thrust faults

g. A structural dome

h. Folding

Answer

d. Mostly horizontal strata

Exercise 60.8.3

What major feature(s) is/are depicted in this portion of the North American geologic map?



a. A structural dome

b. A structural basin

c. Strike-slip faulting

d. Mostly horizontal strata

e. Lots of normal faults

f. Folding

g. Angular unconformity

Answer

g. Angular unconformity

? Exercise 60.8.4

What major feature(s) is/are depicted in this portion of the North American geologic map?



a. A structural basin

b. Lots of thrust faults

c. Lots of normal faults

d. An angular unconformity

e. Strike-slip faulting

f. A structural dome

g. Mostly horizontal strata

Answer

c. Lots of normal faults

? Exercise 60.8.5

What major feature(s) is/are depicted in this portion of the North American geologic map?



a. Lots of thrust faults

b. A structural basin

c. An angular unconformity

d. A structural dome

e. Lots of normal faults

f. Mostly horizontal strata

g. Strike-slip faulting

Answer

b. A structural basin

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60.9: Conclusion

Geologic maps are vital representations of the surface outcrop patterns of geologic formations. To those who master the skill of reading them, they quickly and efficiently communicate a wealth of information about the types and ages of rocks found in an area, whether they've been subjected to tectonic stresses, and how they control the landscape.

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CHAPTER OVERVIEW

61: (Tools of the Trade) Way-up structures

61.1: Introduction

61.2: Sedimentary way-up structures

61.3: Igneous way-up structures

61.4: Conclusion

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61.1: Introduction

“Which way’s up?”

With rocks, the answer is not always clear.

If layered rocks have experienced mountain building, they may be rotated from their original horizontal positions into vertical orientations, potentially confusing geologists, since the principle of superposition no longer applies.





Figure 61.1.1: An exciting outcrop: but in order to interpret it correctly, we need to know which strata are older, and which are younger.

Consider the hypothetical example illustrated in the image at right: you discover a vertical sequence of strata. In one layer, you find evidence of a mass extinction event. In another layer, you find evidence of glaciation. Did an ice age cause the extinction? Or did the extinction somehow trigger the ice age? In order to pose intelligent questions about causality, you need to know which one is older. The older event can influence the more recent event, but not the other way around.

Even more extreme is when compressive stresses associated with convergent plate tectonic settings manage to fold beds into up-side-down positions. If the beds are up-side-down in such a tectonic inversion, it could really throw off the interpretations of the Historical Geologist: they would be reading Earth history backward!



Figure 61.1.2: A tight, recumbent fold in the Sierra de Juarez, Mexico, shows both upright and up-side-down bedding. Way-up (or "geopetal") structures allow us to tell the two apart.

In such situations, we need reliable tools in order to accurately interpret which direction was "up" when the rock originally formed. We call these patterns that look different right-side-up compared to up-side-down by the general term "**way-up structures**." Some geologists also call them "geopetal structures."

Way-up indicators are critical for figuring out the correct sequence of geologic units. The help us determine "younging direction," the direction in which strata get younger. (This is the same as paleo-"up" or "facing direction.")

The welter of terminology shouldn't turn us off: it's an indication that geologists put a strong emphasis on finding and relying on way-up structures. Consider this example:

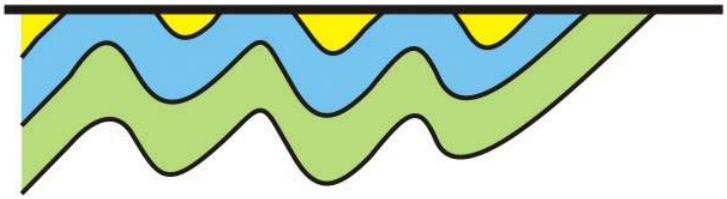


Figure 61.1.3: Geopetal structures that point in the paleo-“up” direction (red arrows) are of critical value in deciphering the story told by a sequence of strata.

Three layers are shown in outcrop at Earth’s surface, color-coded green, blue, and yellow. They are folded into what appears to be a series of anticlines and synclines. Given superposition alone, we would assume the lowest one is oldest, and the uppermost one is youngest. However, way-up structures in these three units point **downward** as the “up” or younging direction. Therefore, the exposed layers are part of a larger-scale fold, and the upper (upright) portion of that fold has been removed by the forces of erosion. Without the way-up indicators, we would have been fooled. So they are very important.

So what are these geopetal tools, exactly?

The key thing is that a way-up structure must be display some difference between its top and its bottom. **They always look different up-side-down compared to right-side-up.** In sedimentary rocks, the following way-up structures can aid the historical geologist in figuring out the paleo-“up” direction:

- cavity fills
- crossbeds
- ripple marks
- mudcracks
- graded beds
- loading structures
- sole structures
- burrows
- stromatolites

In igneous rocks, there are fewer options, but a few that are handy include:

- vesicle concentration
- apophyses
- pillow structures
- in a few rare intrusions, primary structures including grading and cross-bedding(!)

Metamorphic rocks develop no way-up structures as a consequence of metamorphism, but sometimes primary structures in a sedimentary or volcanic protolith can potentially survive as discernible patterns in lower-grade metamorphic rocks. Higher-grade metamorphic rocks are so thoroughly recrystallized that any original geopetal structures would be destroyed.

A note of caution

Be cautious about leaping to big, important conclusions from a single way-up structure. The way-up indicators shown on this page are hopefully straightforward, but nature is vast and varied. Examples seen in the field can frequently be ambiguous. The careful historical geologist will search for as many examples as possible; to see if they agree with one another. Careful geologists seek a preponderance of evidence before settling on an interpretation.

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61.2: Sedimentary way-up structures

We'll begin with an examination of way-up structures in sedimentary deposits, whether lithified into sedimentary rock or in an unconsolidated state (but after deposition).

Cavity fills

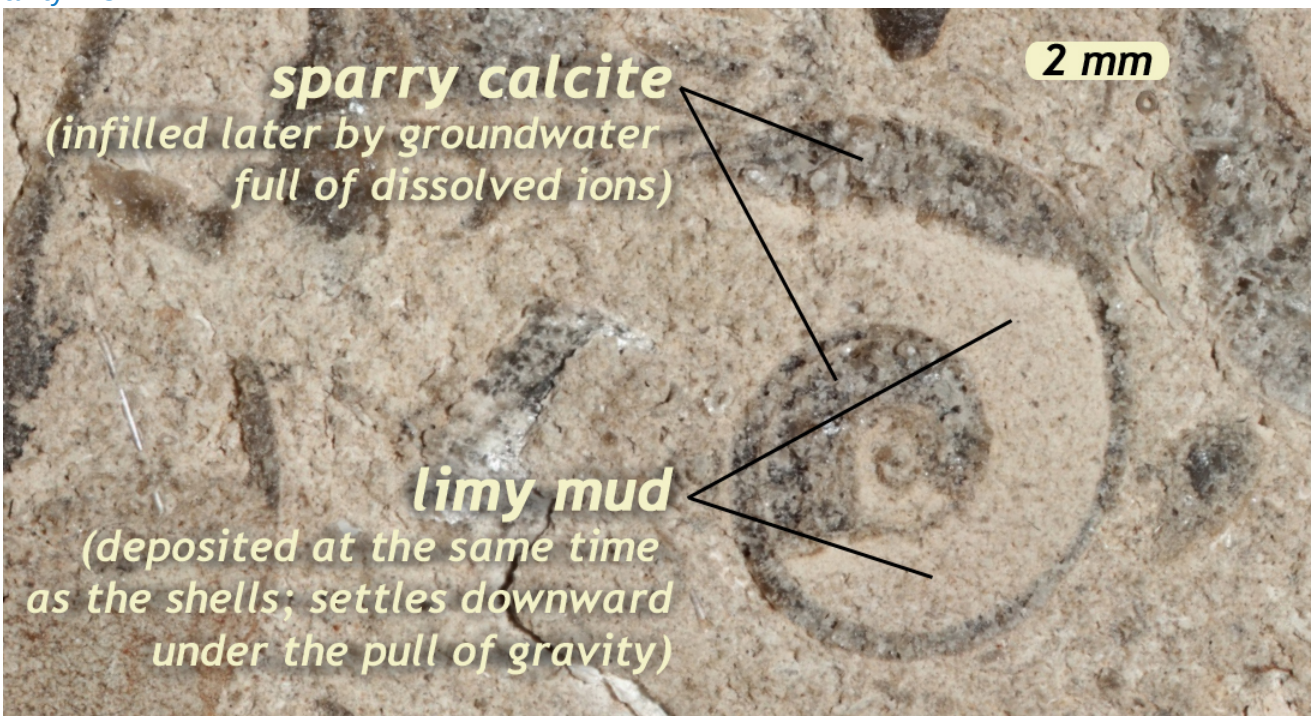


Figure 61.2.1: In cavity fills, mud fills a portion of the empty space at the bottom and later diagenetic deposition of calcite fills in the remaining empty space (at the top) with spar.

When an empty space exists in a sedimentary deposit, such as the protected little hollow under a shell, it can be partially filled by sediment. We call this little pocket of space a "void" or a "cavity." If the cavity is ***entirely*** filled with sediment, it's useless as a geopetal structure. It only has geopetal value if it is ***partially*** filled in with mud. In fact, this is **geopetal** in the strictest sense of the term: it refers to these cavity fillings. Here is how it works: the mud settles under the influence of gravity, lining the bottom of the space (but not the top of the space). Later, as groundwater moves through the sediment, it carries with it dissolved ions, which may bond together, filling the available space with crystals. In the example at right, the crystals formed are of the mineral calcite, which occurs in a coarse form geologists call "spar."

Use this principle to examine the following image, showing a cross-section through a bed of limestone collected in West Virginia. We have intentionally placed it in a vertical position for you to examine. As you will see, the limestone includes many snail shells (gastropod fossils). A lot of them show this key pattern of infilling with two substrates: gray limy mud and white calcite spar. Note that because the shells are made of calcite also, both the original shell and the subsequent spar appear the same coarse crystalline white:



Which way is depositional “up” in this sample? You can use the rotation tool at lower right to test various orientations, until you get the mud at the (original) bottom and the spar above it in the (original) top of the available space.

Hopefully you determined that the side of the sample next to the pencil was originally “down.”

The sample principle applies with minerals other than calcite. For instance, consider this slab of “*Turitella* agate,” a rock made of whole snail shells filled in with the tiny, clear, bean-shaped shells of small arthropods called ostracodes.

Compare and contrast these two shells from the slab, with their different way-up implications:



Figure 61.2.2: A gastropod shell serves as a cavity, partly filled with a sedimentary deposit of ostracode shells, leaving an upper chamber open for quartz crystals to attempt to fill. This one is right-side-up.



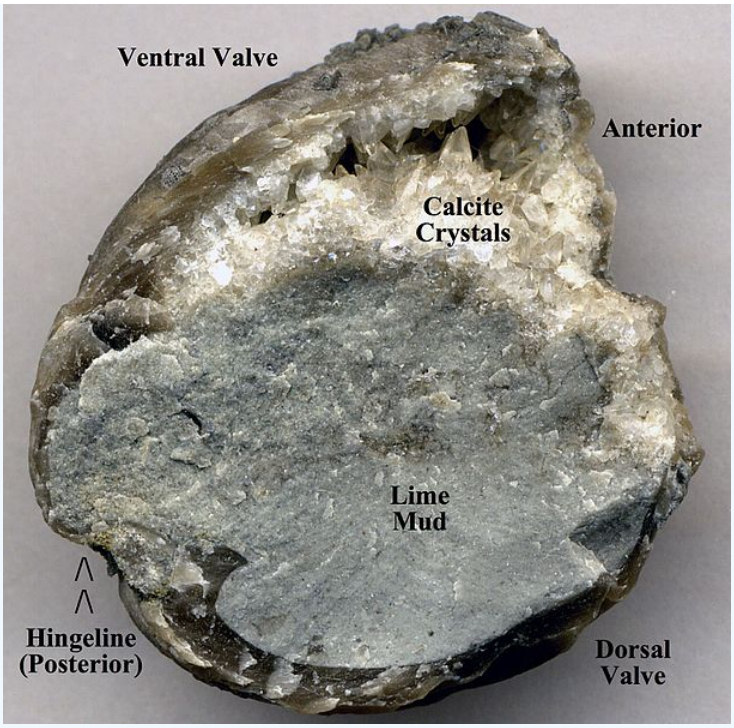
Figure 61.2.3: A gastropod shell serves as a cavity, partly filled with a sedimentary deposit of ostracode shells, leaving an upper chamber open for quartz crystals to attempt to fill. But this one is up-side-down!

Explore the slab on your own if you want to search for additional examples. Or if you are ready to test yourself on a similar sample, from Texas, take the quiz below:

Did I Get It? - Quiz

? Exercise 61.2.1

Is this sample right-side-up or up-side-down? Explain.



a. Right-side-up, since the bottom is filled with mud, and the empty cavity at the top is partially filled with sparry calcite crystals.

b. Up-side-down, since the bottom is partially filled with sparry calcite crystals, and the top is filled with mud.

Answer

a. Right-side-up, since the bottom is filled with mud, and the empty cavity at the top is partially filled with sparry calcite crystals.

? Exercise 61.2.2

Examine this sample for cavity fills.

https://geo.libretexts.org/@go/page/22870



Which way is depositional "up?"

a. The top of the screen (where the scale is) is the original depositional "up" direction.

b. The left of the screen (where the biggest shells are) is the original depositional "up" direction.

c. The bottom of the screen (opposite the side where the scale is) is the original depositional "up" direction.

Answer

a. The top of the screen (where the scale is) is the original depositional "up" direction.

Crossbedding

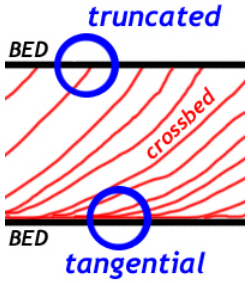


Figure 61.2.4: A cartoon showing the different crossbed/bed relationships at the upper and lower portion of the bed.

Cross-bedding is a series of laminations included within a larger sedimentary bed. A directional current allows the laminations to build up on the leeward (downstream) side of a migrating bedform called a ripple. (Bigger versions of ripples are called dunes, and they create cross-beds as they migrate, too.) Ripples and dunes form in a directional current of either water or air. They accumulate at the angle of repose for that size of sedimentary particle, and within that medium (saltwater, freshwater, or air). In general, it is a gentle angle of around 30° of dip.

In terms of shape, cross bedding is most useful when the cross beds show a pronounced concavity, with the scoop-shape curving upward, like a smiley face. This concave-up shape is a reflection of the cross-beds curving into a gentler orientation, approaching

parallel to the base of the main bed. We call this “tangential,” from the geometric term describing a line intersecting a circle at one point. At the top of the bed, in contrast, they are often truncated (“stopped short” or “cut off”), because the original upper tangential portion of the leeward side of the ripple has evidently been “planed off” by post-depositional erosion.

This leads to important insights, as shown in the case study of the Mixed-Up Quartzites of Cape Agulhas.

Examine this gigapixel panorama of a specimen of cross-bedded red sandstone. There are two halves to the sample shown. One is right-side-up. The other is up-side-down. Explore the two samples to see if you can figure out which one is which, applying the criteria explained above.



Really large cross-beds form via aeolian (wind) deposition in dune fields. The shape and geopetal implications are the same, but the cross-beds are much larger. Here is an example (right-way-up) from coastal exposures in the western Orkney Islands:



Exceptions to the rule

Cross-bedding can be tricky. Sometimes cross-bed laminae accumulate without the tangential, concave-up portion, and sometimes they preserve the upper convex-up portion at the ripple crest. Even crazier, sometimes the accumuate not on the leeward side of a dune, but the upstream (stoss) side! Let us take a moment to see what these complications look like...

CLIMBING RIPPLES



Figure 61.2.5: Climbing ripples in cross-sectional view, Morrison Formation, Greybull, Wyoming.

Climbing ripples form when downstream migration of a ripple or dune is accompanied by rapid vertical aggradation of sediment. This tends to occur when the sedimentary load is higher than the capacity of the current that's carrying it. Climbing ripples are distinct structures indicating critical to supercritical flow, but they often look about the same right-side-up as they do up-side-down. The concave-up portion of the laminae in the lee of the ripple is matched by the convex-up crest of the ripple itself, buried as soon as it forms. As a result, they can mislead the historical geologist.

Consider the example below; if you flipped the photo up-side-down, it would be difficult to tell:



Figure 61.2.6: Climbing ripples. (Photo by Brian Romans; reproduced with permission.)

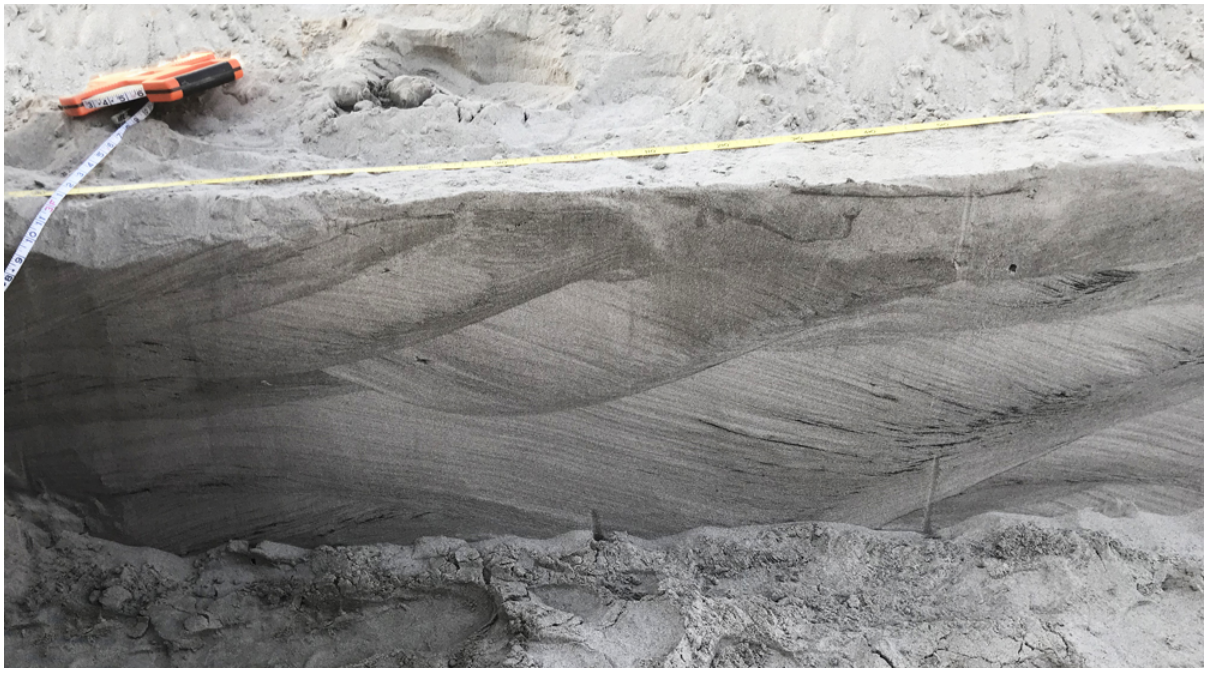


Figure 61.2.7: Climbing ripples in modern delta sediments, Louisiana. (Photograph by Robert Mahon; reproduced with permission.)

The image at right (click to enlarge) shows a cross-section through three sets of climbing ripples from modern (unlithified) sand deposits in the Mississippi Delta region of Louisiana. Note how the largest, oldest, deepest set of cross beds shows an undulating set of internal laminations, and these "climb" up and to the right, with tangential bases. This is likely indicative of supercritical flow. The second set has a trough-like bottom that cuts into the older set of cross-beds, and is truncated by the most recent, uppermost, thinnest set of cross-beds.

Finally, here is a gigapixel panorama of a specimen showing climbing ripples in cross-section:



ANTIDUNES



Figure 61.2.8: Standing surface waves will have corresponding antidunes on the sediment bed, in this small stream draining a beach at low tide.

If water is flowing fast enough to be in the upper flow regime, antidunes can form. They are mostly found in shallow channels (*e.g.*, fluvial and tidal channels). If you are lucky enough to see them forming, you can recognize the situation when you see standing surface waves – watch closely and you will see the waves migrate **upstream over time**, meaning that the are bedforms developing immediately below the standing waves. If high flow is maintained, the antidunes will also migrate upstream. They do this by adding new sediment on their upstream (stoss) side rather than the usual situation for migrating ripples, which deposit sediment on the downstream (lee) side. So antidune cross-beds dip in the **opposite** direction from normal ripple or dune cross-beds. The good news is that the preservation potential of antidunes is low, because as soon as flow slackens they tend to wash out and be eroded.

Did I Get It? - Quiz

? Exercise 61.2.1

Is this outcrop of sandstone right-side-up or tectonically inverted? Why?



a. It must be tectonically inverted, since the cross-beds are up-side-down: they are truncated at the bottom, and tangential at the top.

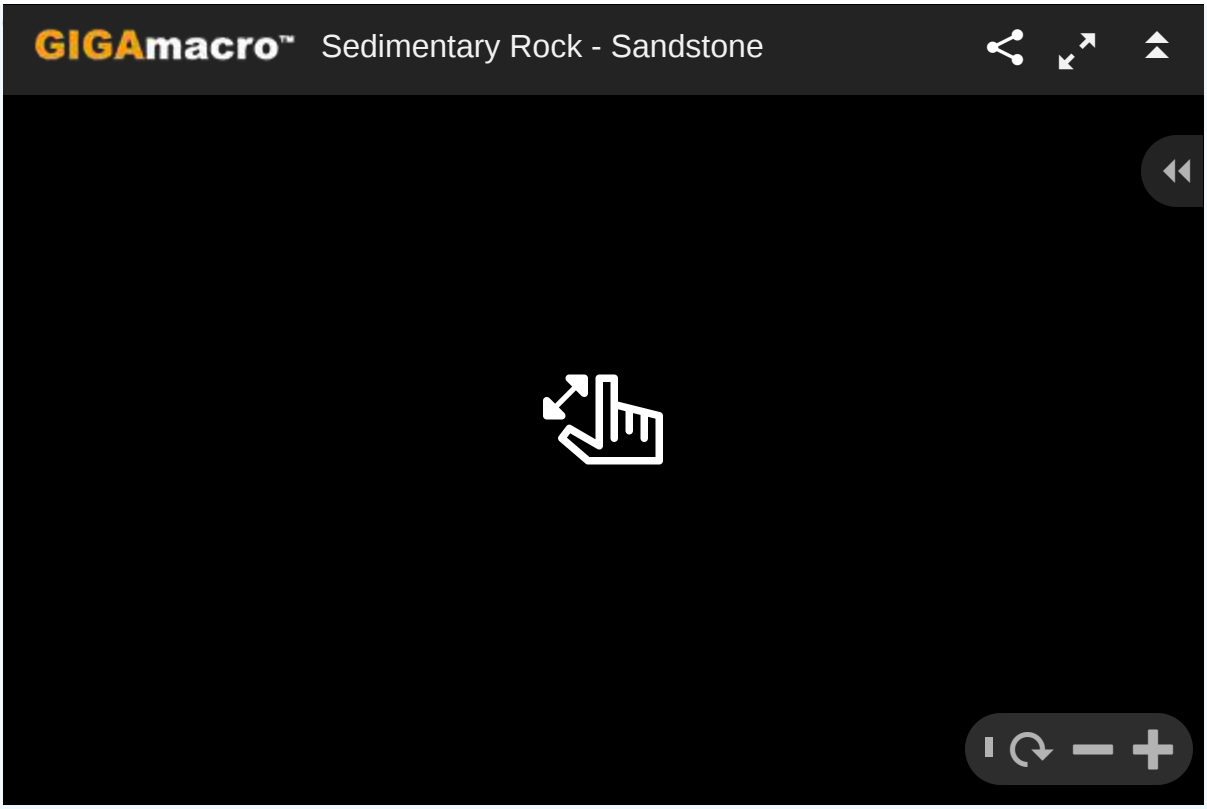
b. It appears to be right-side-up, since the cross-beds in the middle of the outcrop show tangential relationships to the main bed at the bed's bottom, and are truncated at the top.

Answer

b. It appears to be right-side-up, since the cross-beds in the middle of the outcrop show tangential relationships to the main bed at the bed's bottom, and are truncated at the top.

? Exercise 61.2.2

Is this sample of sandstone in its upright depositional position, or up-side-down?



a. It is right-side-up.

b. It is up-side-down.

Answer

b. It is up-side-down.

? Exercise 61.2.3

Are these cross-beds likely to be fluvial (river-deposited) or aeolian (wind-deposited)? Why?



a. Fluvial, because the outcrop is broken by a couple of weathered-out joints (vertical fractures).

b. Aeolian, since they are so large.

c. Aeolian, because the outcrop has some snow on it.

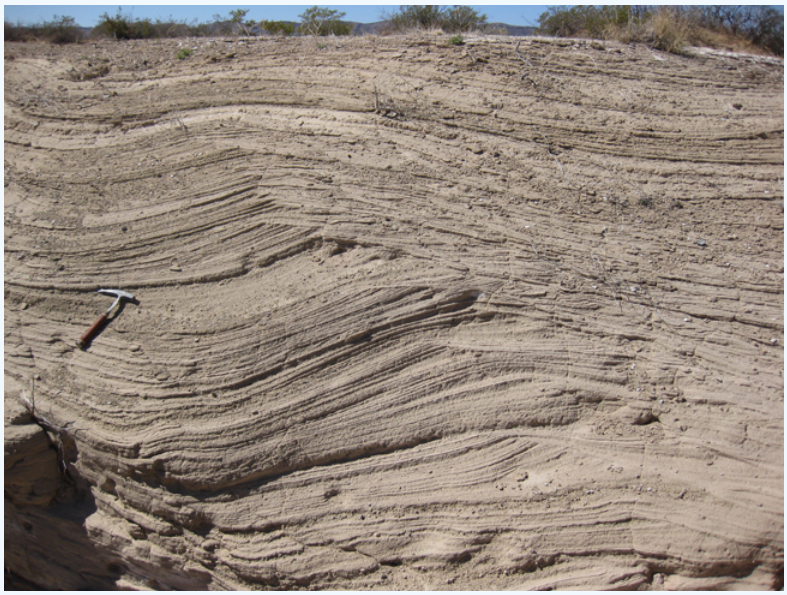
d. Fluvial, since they are right-side-up.

Answer

b. Aeolian, since they are so large.

Exercise 61.2.4

Which way-up structure is shown here? Are the strata right-side-up, or up-side-down?



a. Graded bedding; the strata are up-side-down.

b. Cavity fills; the strata are up-side-down.

c. Climbing ripples; the strata are right-side-up.

d. Climbing ripples; the strata are up-side-down.

e. Cavity fills; the strata are right-side-up.

f. Graded bedding; the strata are right-side-up.

Answer

c. Climbing ripples; the strata are right-side-up.

Exercise 61.2.5

Which way is paleo-"up" in this sample of quartzite and red argillite? Why?



a. Paleo-"up" is to the right, on the opposite side from where the scale bar is. This is because the central set of cross beds are tangential to the main bed at the right, but truncated by white sand at the left.

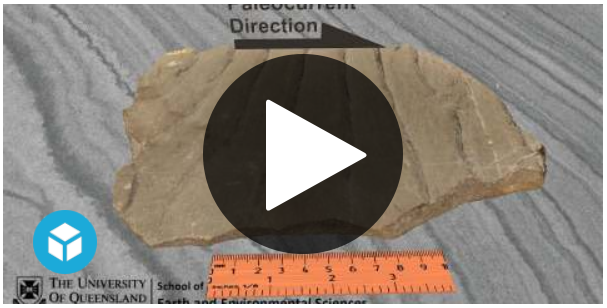
b. Paleo-"up" is to the left, where the scale bar is. This is because the central set of cross beds are tangential to the main bed at the right, but truncated by white sand at the left.

Answer

b. Paleo-"up" is to the left, where the scale bar is. This is because the central set of cross beds are tangential to the main bed at the right, but truncated by white sand at the left.

Ripple marks

Ripple marks are the 3D expression of the same phenomenon as cross-bedding: the actual bedform's shape exposed in lithified form. They can be either symmetrical or asymmetrical. The 3D model below shows a great example of aysmmetric ripple marks, the kind that form in a unidirectional current. Rotate the model so you are looking down the crest of the central ripples so you can see their shape in cross-section: the left side is gently sloped compared to the right side. The left side is therefore the upstream (stoss) side, and the right side is therefore the downstream (leeward) side.



Here is a 3D model showing some ripple marks exposed on multiple bedding planes within the shallow-water Foreknobs Formation (Devonian):



One bedding surface shows fairly linear ridge crests, but an older (deeper) layer to the right shows interference ripples between two perpendicular wave orientations. These create little “point” like features that protrude outward from the bed in the paleo-“up” direction.

Did I Get It? - Quiz

Exercise 61.2.1

What primary sedimentary structure is shown in this 3D model?

Which way was the current flowing when this sediment was deposited?



a. Ripple marks. Current flowed from location "5" toward location "3."

b. Ripple marks. Current flowed from location "3" toward location "5."

c. Ripple marks. Current flowed from location "4" toward location "2."

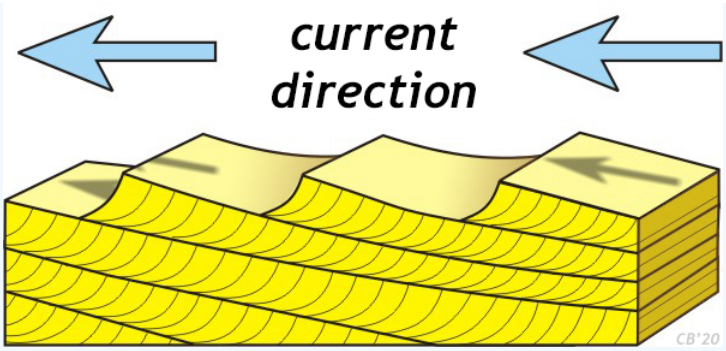
d. Ripple marks. Current flowed from location "2" toward location "4."

Answer

b. Ripple marks. Current flowed from location "3" toward location "5."

Exercise 61.2.2

Does this diagram show symmetrical or asymmetrical ripples?



a. Symmetrical

b. Asymmetrical

Answer

b. Asymmetrical

Mud cracks



Figure 61.2.9: Pleistocene mud cracks between sand dunes at Mesquite Dunes, Death Valley, California

The desiccation of mud produces shrinkage when stresses induced by volume loss through evaporation are greater than the cohesive strength of the mud. Cracks initiate and grow, dividing the mud up into a series of irregular polygons. The photograph here shows the sort of thing we think about when we think of mudcracks: a vast dry field of polygons, sometimes with air between them, and sometimes filled with later sediment.

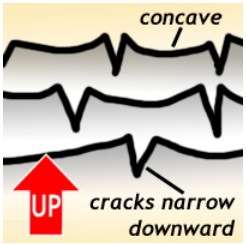
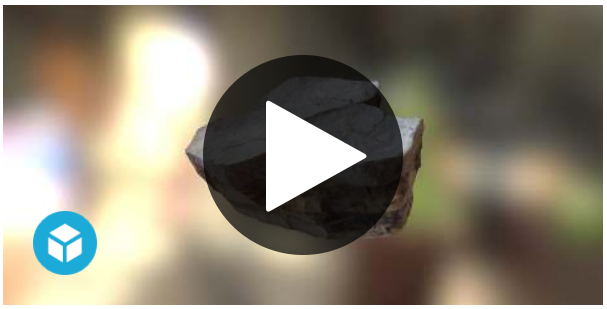


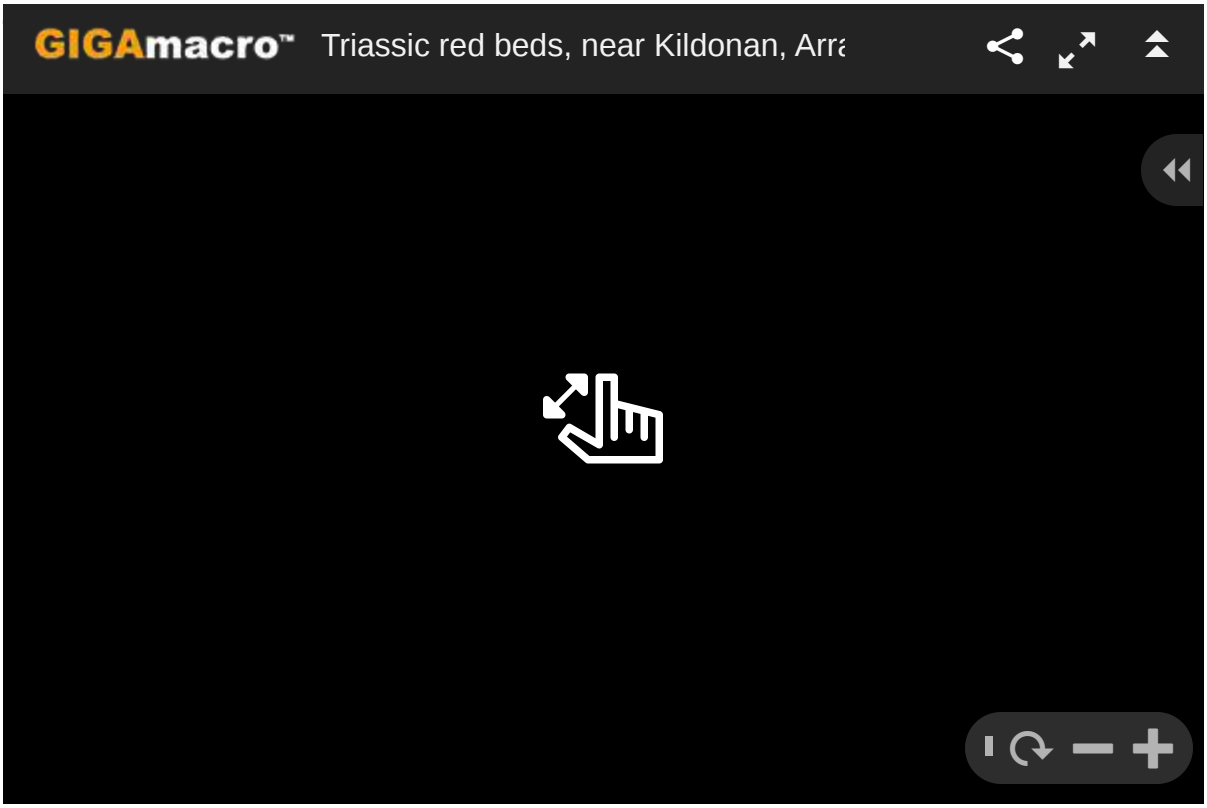
Figure 61.2.10: Key aspects of mud crack morphology that can aid in interpreting way-up direction.

But the geopetal value of mud cracks comes from their cross-sectional view, which is perhaps not as intuitive but is worth considering, because it is very useful. The cracks propagate downward through the mud deposit in an ever-widening gap. This gap is widest at the top and most narrow at its deepest propagating tip, giving it an overall V shape in cross-section. When a deposit of new sediment gets dumped atop this open gap, it can fill it in. In cross-section, this creates a V-shaped projection down through the desiccated muddy layer.

This 3D model shows these aspects of the shape of well-developed desiccation cracks well:

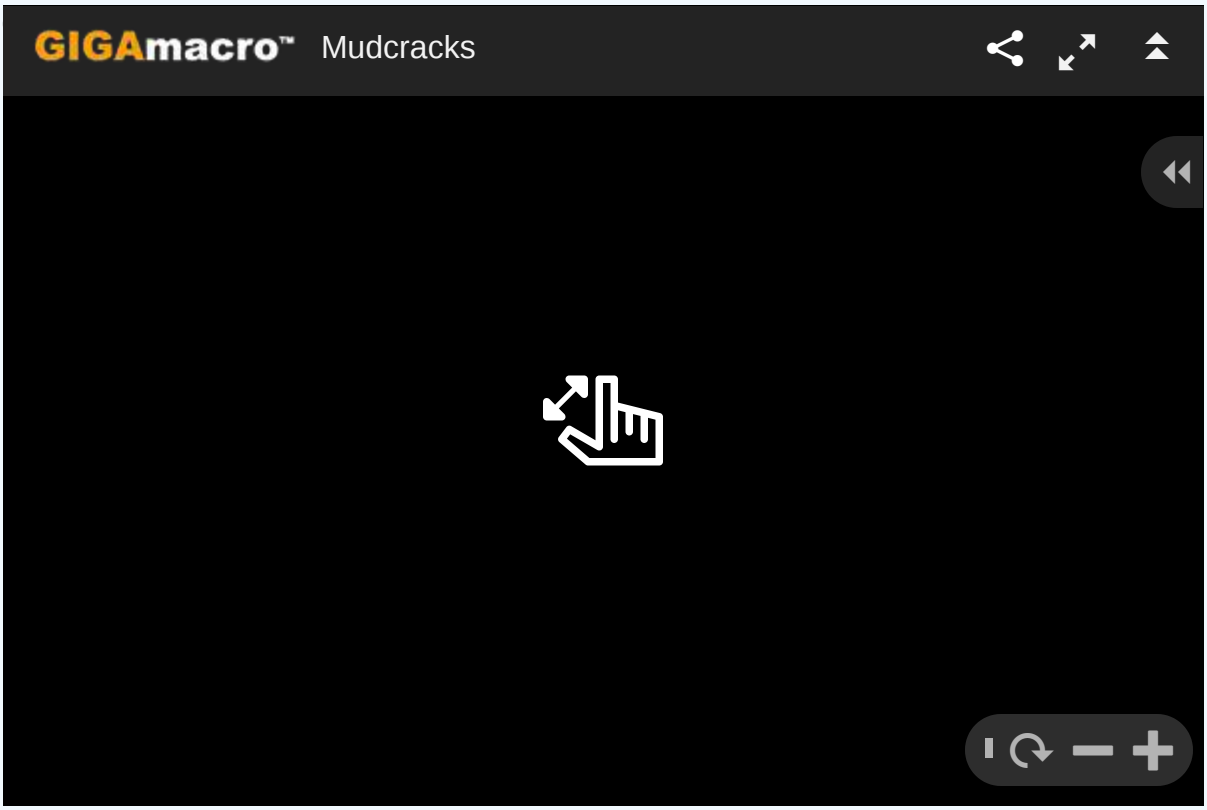


In this gigapixel panorama, you can see a cross-sectional view of some red shale (former mud) and white sandstone (former sand), from the Triassic of Scotland. Note how the mud cracks appear in cross-section: almost like spiky white teeth in a red mouth!



Exercise 61.2.1

What primary sedimentary structure is shown here?



a. Cavity fill

b. Mud cracks (desiccation cracks)

c. Flame structure

d. Cross-bedding

e. Graded bedding

Answer

b. Mud cracks (desiccation cracks)

Exercise 61.2.2

What primary sedimentary structure is shown here?



a. Climbing ripples

b. Ball & pillow

c. Mud cracks (desiccation cracks)

d. Graded bedding

e. Cross-bedding

Answer

c. Mud cracks (desiccation cracks)

? Exercise 61.2.3

Identify the primary sedimentary structure shown here. Are we looking at the top or the bottom of the bed? Explain.



a. Cross-bedding is shown. The current must have been flowing from right to left, and paleo-"up" is toward the bottom of the photo.

b. Cross-bedding is shown. The current must have been flowing from left to right, and paleo-"up" is toward the top of the photo.

c. Mud cracks are shown. Because the sandstone filling in the gaps between the mud cracks pokes outward from the face of the slab, this must be the underside of a sandstone bed that filled in the mudcracks. We are looking "upward" at the bottom of this post-mud-crack sand bed.

d. Mud cracks are shown. Because the sandstone filling in the gaps between the mud cracks pokes downward into the face of the slab, this must be the upper surface of the mud layer that dried out. We are looking "downward" at the top of this muddy layer.

Answer

c. Mud cracks are shown. Because the sandstone filling in the gaps between the mud cracks pokes outward from the face of the slab, this must be the underside of a sandstone bed that filled in the mudcracks. We are looking "upward" at the bottom of this post-mud-crack sand bed.

? Exercise 61.2.4

What primary sedimentary structure is shown here? What are the implications?



a. Mud cracks: So mud must have been deposited by water, then dried out in the open air.

b. Graded bedding: sand and mud were deposited by a turbidity current, probably in a deep ocean setting.

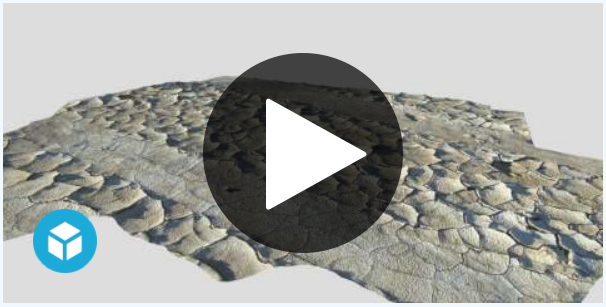
c. Cross-bedding: sand was deposited by a flowing directional current of water, running from left to right.

Answer

a. Mud cracks: So mud must have been deposited by water, then dried out in the open air.

Exercise 61.2.5

What primary sedimentary structure(s) is/are shown? What are the implications?



a. Mud cracks are shown, as well as raindrop impressions and what appear to be automobile tire tracks. The sequence of events implied is (1) a vehicle driving over the mus, (2) a spattering of rain drops, (3) drying of mud, making mud cracks, and finally (4) deposition of mud.

b. Mud cracks are shown, as well as raindrop impressions and what appear to be automobile tire tracks. The sequence of events implied is (1) deposition of mud, (2) drying of mud, making mud cracks, (3) a vehicle driving over the mudcracks, and finally (4) a spattering of rain drops.

Answer

b. Mud cracks are shown, as well as raindrop impressions and what appear to be automobile tire tracks. The sequence of events implied is (1) deposition of mud, (2) drying of mud, making mud cracks, (3) a vehicle driving over the mudcracks, and finally (4) a spattering of rain drops.

Graded bedding

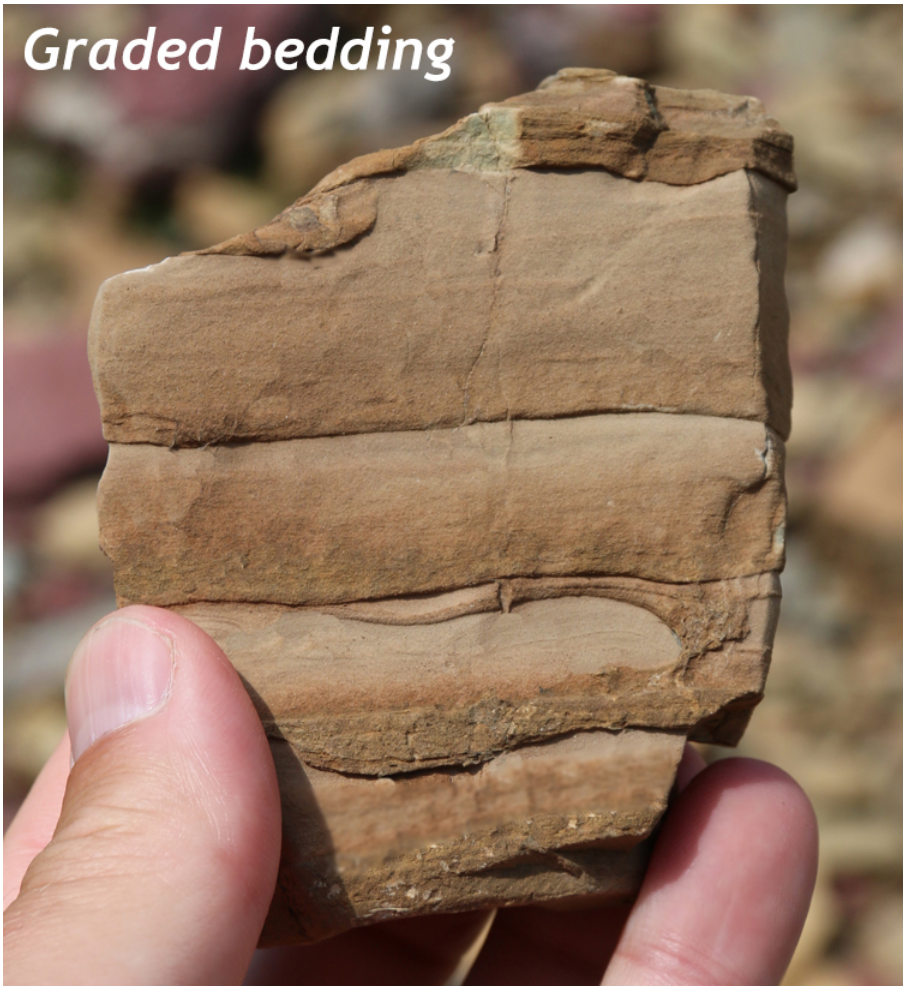


Figure 61.2.11: Five graded beds in one sample of sandy limestone from Glacier National Park, Montana. The sample is in its original stratigraphic orientation, with relatively coarse dark sand beginning abruptly at the bottom of each bed, and gradually fining to very fine-grained light-colored limestone at the top.

Graded bedding is a distinctive and widespread primary sedimentary structure where the grain size changes systematically from the bottom to the top of a bed. Because graded bedding forms as suspended grains settle from a turbidity current, the heaviest grains tend to settle out first, and those tend to be the biggest grains. A graded bed therefore has the largest grains at the bottom, and then they gradually get finer toward the top of the bed. This change in grain size within the bed is transitional and gradational, but the contact between graded beds will typically be quite crisp – a sudden switch from the fine grains at the top of one graded bed and coarser grains at the bottom of the overlying bed.

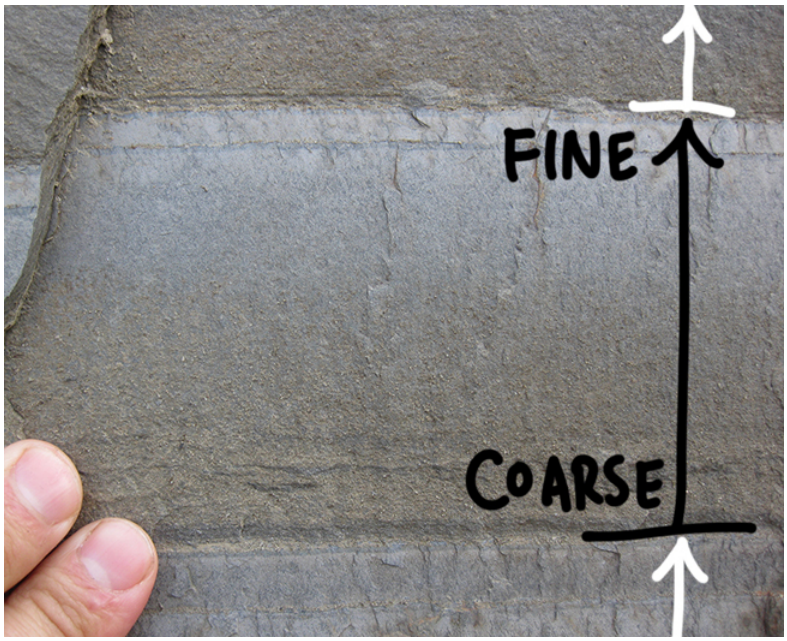


Figure 61.2.12: Graded bedding between find sand and clay-rich mud in meta-turbidites of the Miette Group, Banff National Park, Alberta, Canada.

Graded bedding is most commonly found in submarine turbidity current deposits (*e.g.*, on an abyssal fan), but it can occur in other settings too, such as lakes, floodplains, and tidal flats.

It also is not limited to siliciclastic rocks, and may be found in carbonate sedimentary deposits as well. In the GIGAmacro image below, you can examine an exquisite sample from the Silurian-aged tidal flat carbonates of the Tonoloway Formation (West Virginia), showing a series of graded beds and corroborating sedimentary structures (annotated in yellow circles) with ambiguous evidence (annotated with pink circles) and misleading evidence (annotated with the lone gray circle). Please click on the various annotations to learn the lessons this sample has to teach about interpreting which way is “up.”



Exercise 61.2.1

Examine this piece of limestone. Which way is "up?"



a. "Up" is up (in other words, where the scale bar is at the top of the screen is the original depositional "up" direction).

b. "Down" is up (in other words, the bottom of the screen, opposite the scale bar, is the original depositional "up" direction).

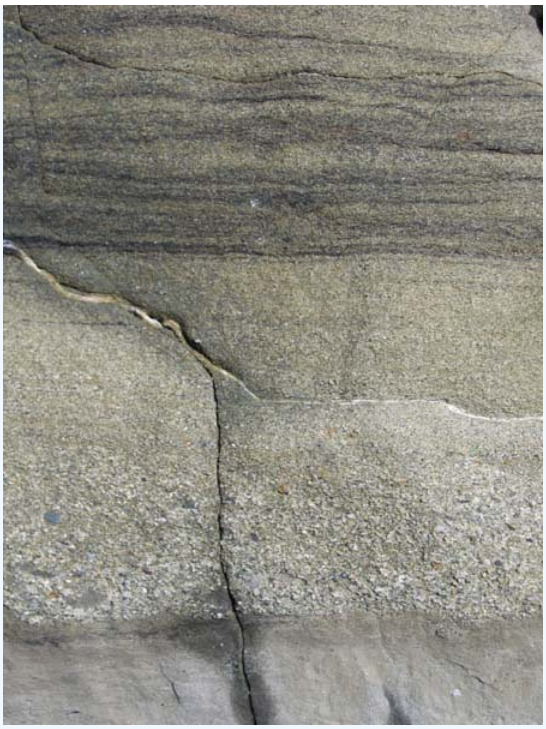
c. "Left" is up (in other words, counterclockwise from the scale bar, along the left edge of the screen, is the original depositional "up" direction).

Answer

a. "Up" is up (in other words, where the scale bar is at the top of the screen is the original depositional "up" direction).

Exercise 61.2.2

Examine this outcrop. What way-up structure is shown? Which way is up?



a. Flame structures are shown. The top of the photograph is the paleo-"up" direction.

b. Flame structures are shown. The bottom of the photograph is the paleo-"up" direction.

c. Graded bedding is shown. The top of the photograph is the paleo-"up" direction.

d. Graded bedding is shown. The bottom of the photograph is the paleo-"up" direction.

Answer

c. Graded bedding is shown. The top of the photograph is the paleo-"up" direction.

Exercise 61.2.3

Examine this outcrop. Which way is up?

(Note that there are some offsets of the bedding along small faults; don't let that throw you off.)



a. "Down" is up. The youngest part of this deposit is at the bottom, on the opposite side from where the pencil is.

b. "Up" is up. The youngest part of this deposit is at the top where the pencil is.

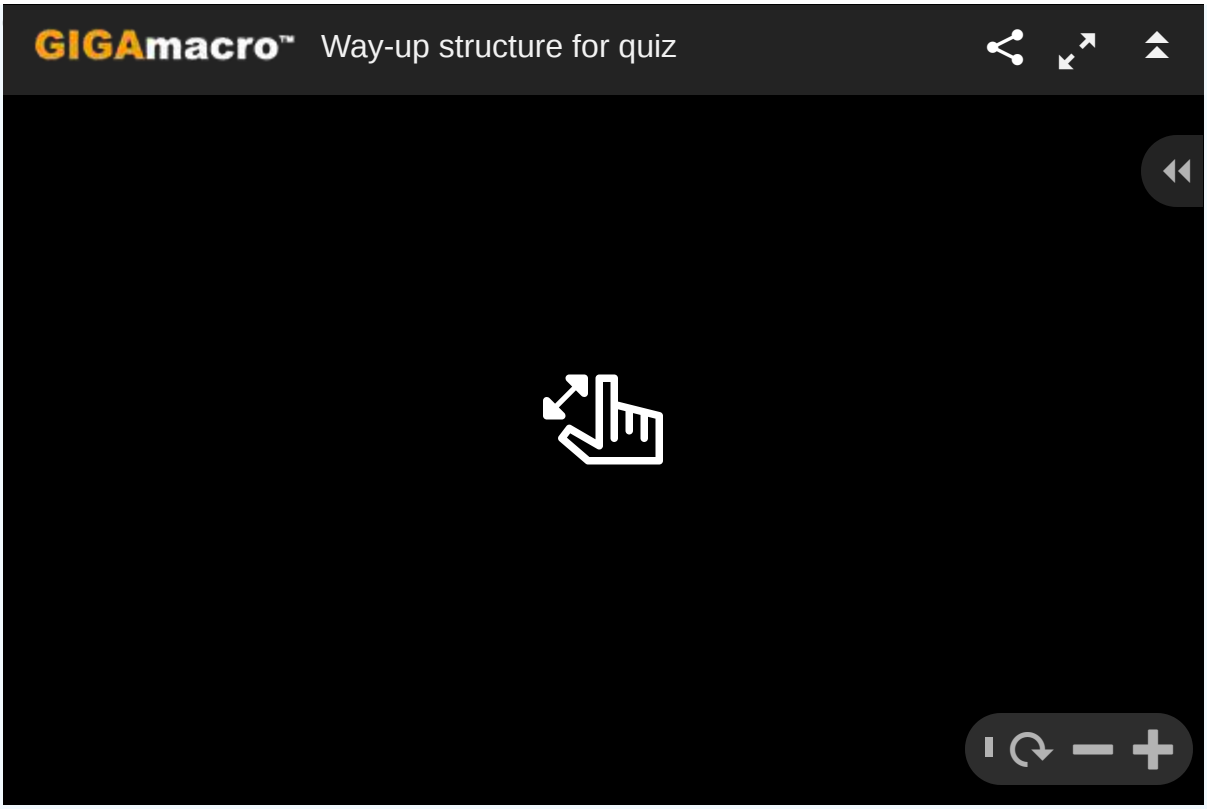
c. "Right" is up. The youngest part of this deposit is on the right side of the photo, underneath the shady overhanging chunk of rock.

Answer

a. "Down" is up. The youngest part of this deposit is at the bottom, on the opposite side from where the pencil is.

Exercise 61.2.4

Examine this outcrop. Which way is up?



a. The graded bed that starts in the middle of the outcrop gets coarser going downward, so these strata are right-side-up.

b. The graded bed that starts in the middle of the outcrop gets finer going downward, so these strata are up-side-down.

Answer

b. The graded bed that starts in the middle of the outcrop gets finer going downward, so these strata are up-side-down.

Exceptions to the rule

In **normal** graded bedding, we see coarse at the bottom and fine at the top, with a gradational transition in between. Therefore, in most cases, we can find graded bedding in an outcrop of sedimentary rock or volcanic ash, and find the coarsest bit and head in the direction of the finest bit, and be confident the strata get younger in that direction. However, *caveat emptor*: nature is vast and varied, and she has a few more tricks up her sleeve.



Figure 61.2.13: A reverse-graded bed within the Purisima Formation.

For instance, there are natural examples of what is called "**reverse** graded bedding," with exactly the opposite pattern. (It's also called "inverse graded bedding.") In particular, debris flows sort grains in the direction that to normal graded bedding would be considered "up-side-down." In the example at left, we see an example from coastal California: an upright submarine debris flow recorded by a stack of muddy sand transitioning gradationally upward into coarse sand and pebbles, and finally to a surface scum of subangular cobbles. The biggest particles rise to the top due to a phenomenon called "the Brazil Nut Effect."



Figure 61.2.14: A reverse graded bed in lahar deposits of the Absaroka Volcanic Field. (Callan Bentley photo.)



Figure 61.2.15: An example of metamorphically-reversed graded bedding from Mount Washington, New Hampshire.

Not only that, but metamorphism can make it even more complicated. Consider this example from the Littleton Formation, in the Presidential Range of New Hampshire:

It shows a Devonian-aged turbidite that has undergone serious metamorphism. But not all the minerals in the rock reacted the same way to metamorphic conditions. The clay in the finer, upper part of the graded bed was more susceptible to metamorphic reactions than the quartz making up the sand in the coarser, lower part of the graded bed. As a result, not much happened with the quartz at high temperature and pressure, while the clay reacted profusely to generate huge porphyroblasts of andalusite — crystals many times larger than the grains of sand which were *originally* the coarsest components of the rock. So the sense of grading was reversed through metamorphic reactions. The key thing to recognizing this phenomena is to understand that the fine portion of a siliciclastic graded bed is likely made of different minerals (clay) than the coarse portion (quartz and perhaps feldspar) and so will show a different susceptibility to metamorphism under high pressure / temperature conditions.

This isn't a fatal flaw, since we understand the metamorphic reactions that convert clay into metamorphic minerals such as garnet, staurolite, sillimanite, kyanite, and andalusite, and we know that quartz is chemically stable over a wide range of temperature and pressure conditions. In fact, we could consider this a blessing in disguise – because we get information both about the metamorphic conditions the rock experienced (andalusite only forms at low pressures and moderate temperatures) but we also still preserve the crisp/gradational/crisp character of the original graded beds. The primary sedimentary structure is altered, but not beyond recognition. It is not destroyed.

Did I Get It? - Quiz

? Exercise 61.2.1

What variety of graded bedding is shown in this upright layer of Absaroka volcanic breccia?



a. Recrystallized (metamorphically inverse-coarsened) graded bedding

b. Normal graded bedding

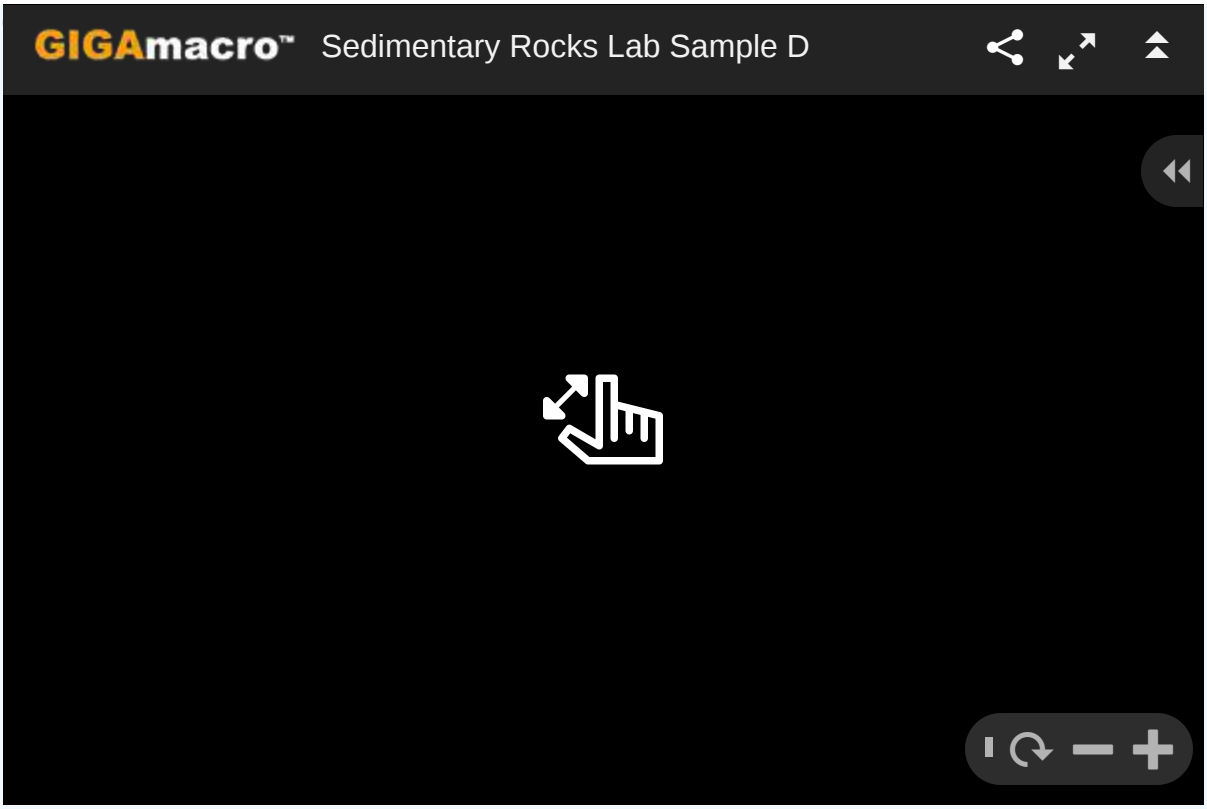
c. Reverse graded bedding

Answer

c. Reverse graded bedding

? Exercise 61.2.2

Here is a normally graded turbidite sequence. Which way is paleo-"up?"



a. Down is paleo-"up," since the central sand body has a crisp edge near the black shale at the top, and the sand gets finer toward the bottom.

b. Up is paleo-"up," since the central sand body has a crisp edge near the black shale at the top, and the sand gets finer toward the bottom.

c. Left is paleo-"up," since the central sand body has a crisp edge near the black shale at the top, and the sand gets finer toward the bottom.

Answer

a. Down is paleo-"up," since the central sand body has a crisp edge near the black shale at the top, and the sand gets finer toward the bottom.

? Exercise 61.2.3

Examine this sample. Which way is the paleo-"up" direction, and why?



a. Up is paleo-"up" because the beds are upright.

b. Right is paleo-"up" because the is a lot of garnet there, indicating the rich former mud at top of the graded bed.

c. Left is paleo-"up" because the beds show a gradation from quartz rich former sand at the base of a graded bed to garnet rich former mud at top of the graded bed.

Answer

c. Left is paleo-"up" because the beds show a gradation from quartz rich former sand at the base of a graded bed to garnet rich former mud at top of the graded bed.

? Exercise 61.2.4

Does this outcrop show normal grading or reverse grading?



a. Normal grading; this looks like a turbidity current deposit.

b. Reverse grading; it must be a debris flow deposit.

Answer

b. Reverse grading; it must be a debris flow deposit.

Exercise 61.2.5

This outcrop is on the slopes of a volcano in Ecuador. Examine the layer behind the geologist's jacket. Identify the way-up structure there and interpret it.



a. Normal graded bedding. This is a submarine turbidity current deposit.

b. Reverse graded bedding. This is a lahar (volcanic mudflow) deposit.

Answer

b. Reverse graded bedding. This is a lahar (volcanic mudflow) deposit.

Recap graded beds as way-up indicators with this slideshow:



Figure 61.2.16

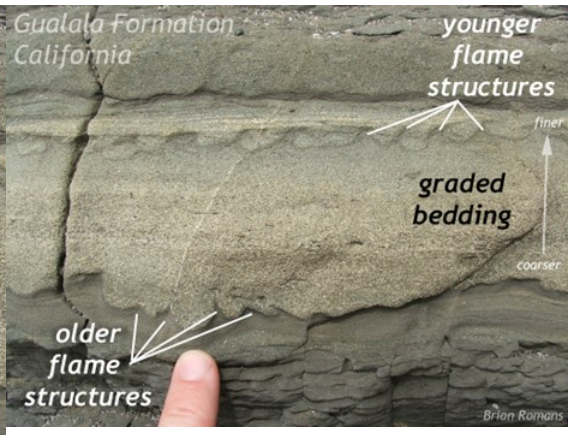


Figure 61.2.17



Figure 61.2.18: Graded bed in turbidite, Usal Beach, California. Bedding is vertical, and fines (gets younger) to the right. Note the crisp left edge to the central sand body.

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Figure 61.2.19: Graded bedding in Lake Vermilion Formation graywacke, Minnesota; younging direction is to the left.

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Figure 61.2.20: Relict graded bed in Mather Gorge Formation metagraywacke, near Potomac, Maryland. Fining direction (and thus geopetal "up") is to the top of the photo. In this case, "up" is up!



Figure 61.2.21: A stack of graded beds in the Buchan block of Scotland. (Original photo by Tim Johnson; digitally cleaned up by CB.)

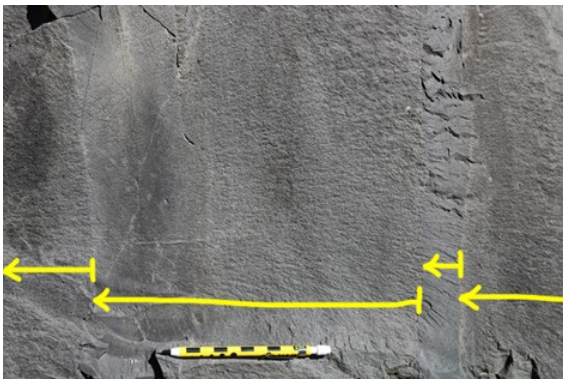


Figure 61.2.22: Vertical graded beds in turbidites of the late Ordovician Martinsburg Formation, Virginia.

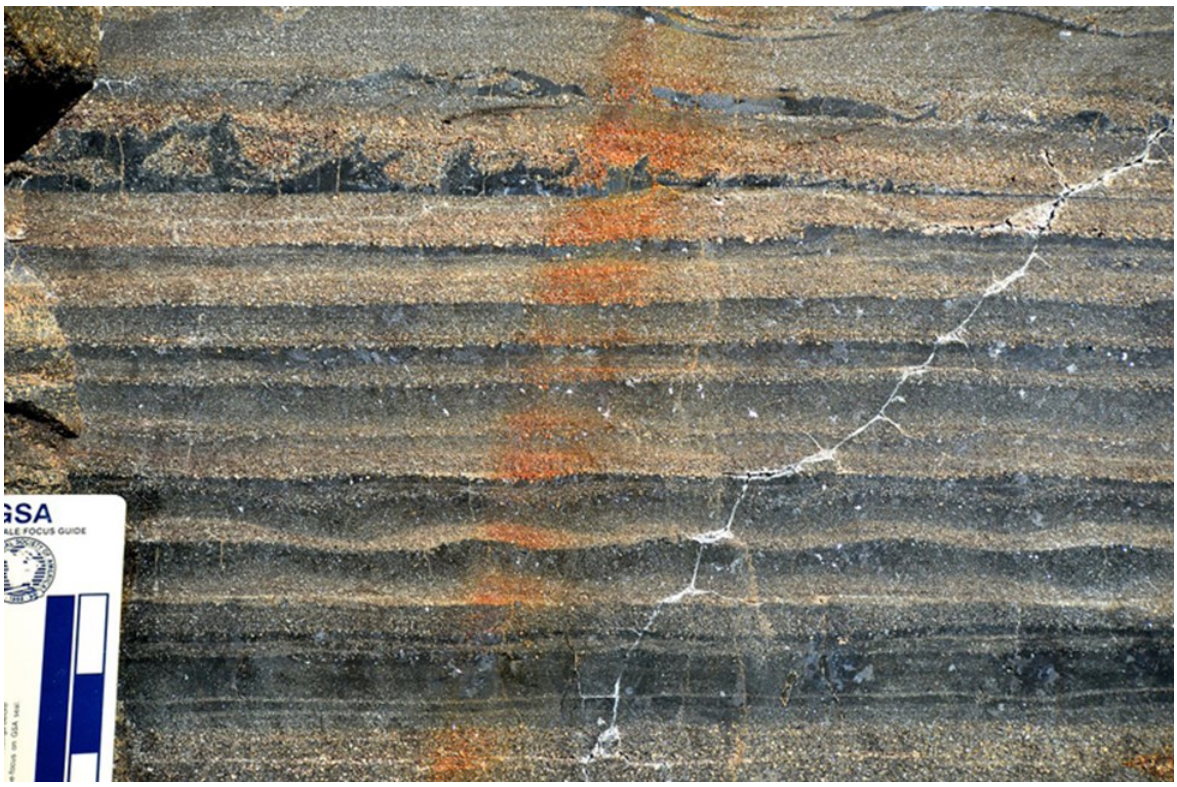


Figure 61.2.23: Normal graded beds in Patagonian turbidites; also note the flame structures near the top. (Photograph by Zoltán Sylvester; reproduced with permission.)

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Figure 61.2.24: Normal grading in a turbidite from the Talara Basin, Peru. (Photograph by Zoltán Sylvester; reproduced with permission; digitally altered by CB.)

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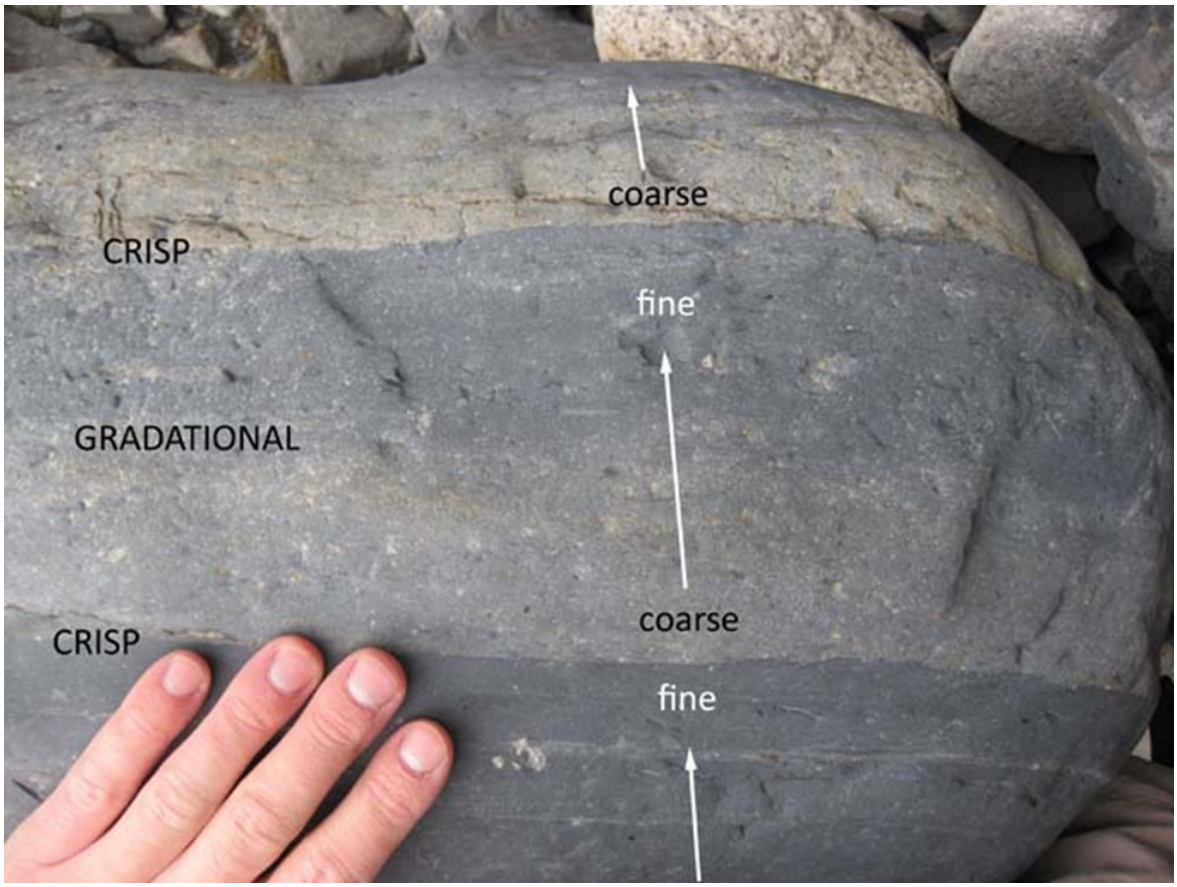


Figure 61.2.25: Cretaceous turbidites in Patagonia, Chile, showing portions of three beds in cross-section.



Figure 61.2.26: A cross-sectional view through Bouma layers A and C, with prominent graded bedding shown in the lower 2/3 of the sample, and cross-bedding in the upper 1/3. This example from Proterozoic Malmesbury Group turbidites, near Rooi-Els, South Africa.

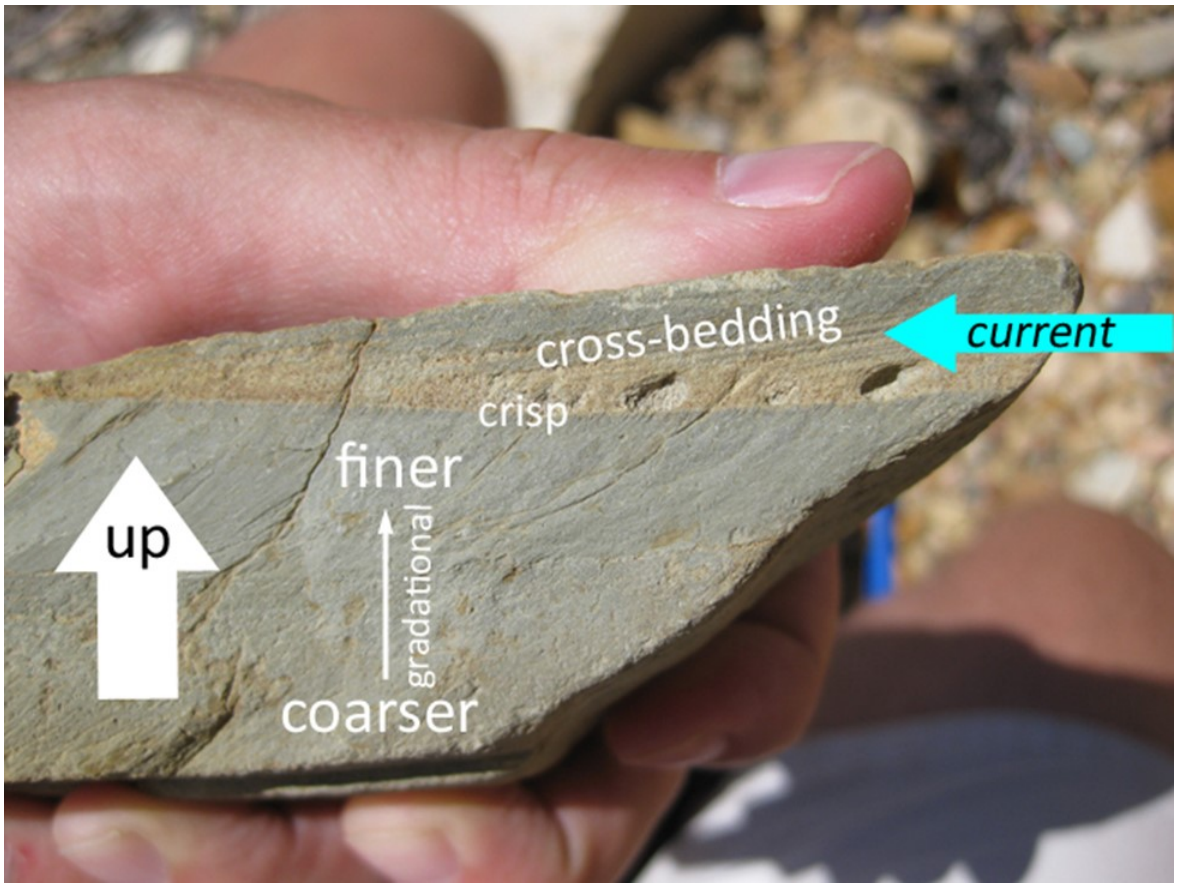


Figure 61.2.27: Annotated cross-sectional view through Bouma layers A and C, with prominent graded bedding shown in the lower 2/3 of the sample, and cross-bedding in the upper 1/3. This example from Proterozoic Malmesbury Group turbidites, near Rooi-Els, South Africa.

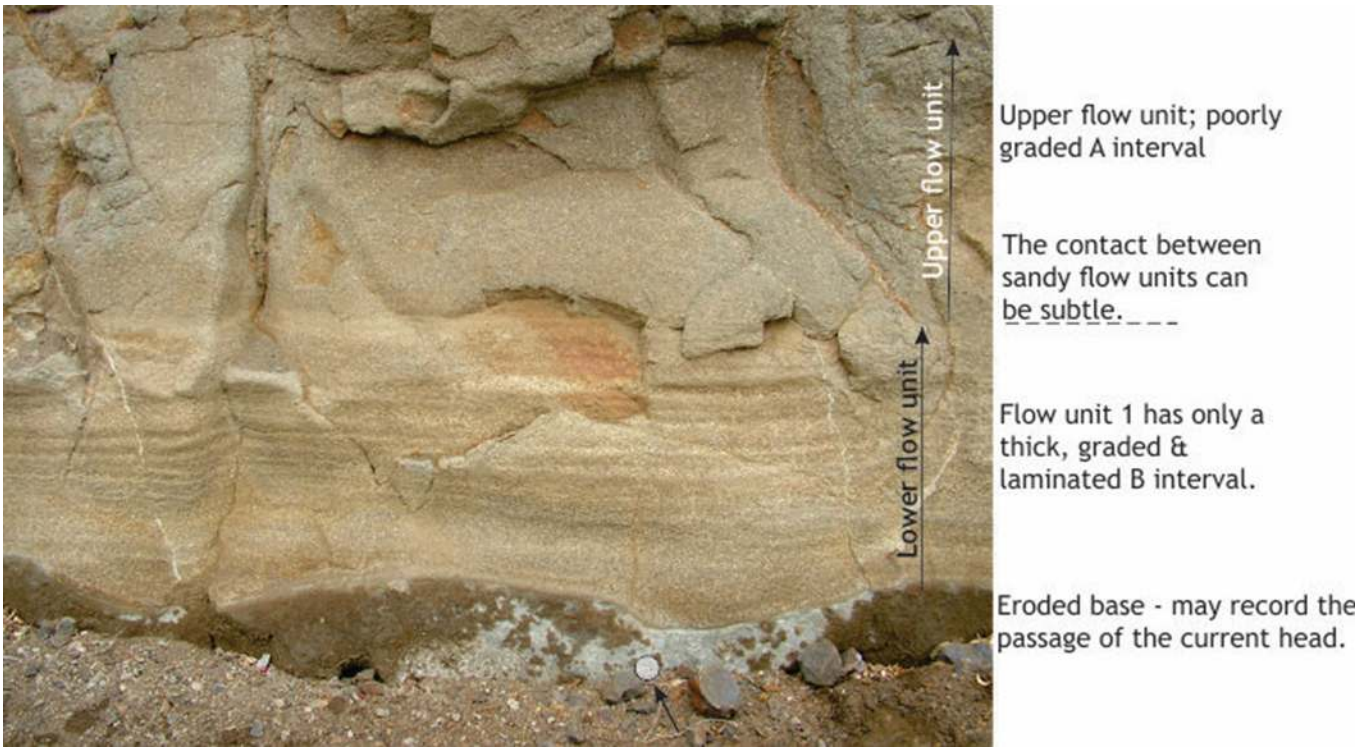


Figure 61.2.28

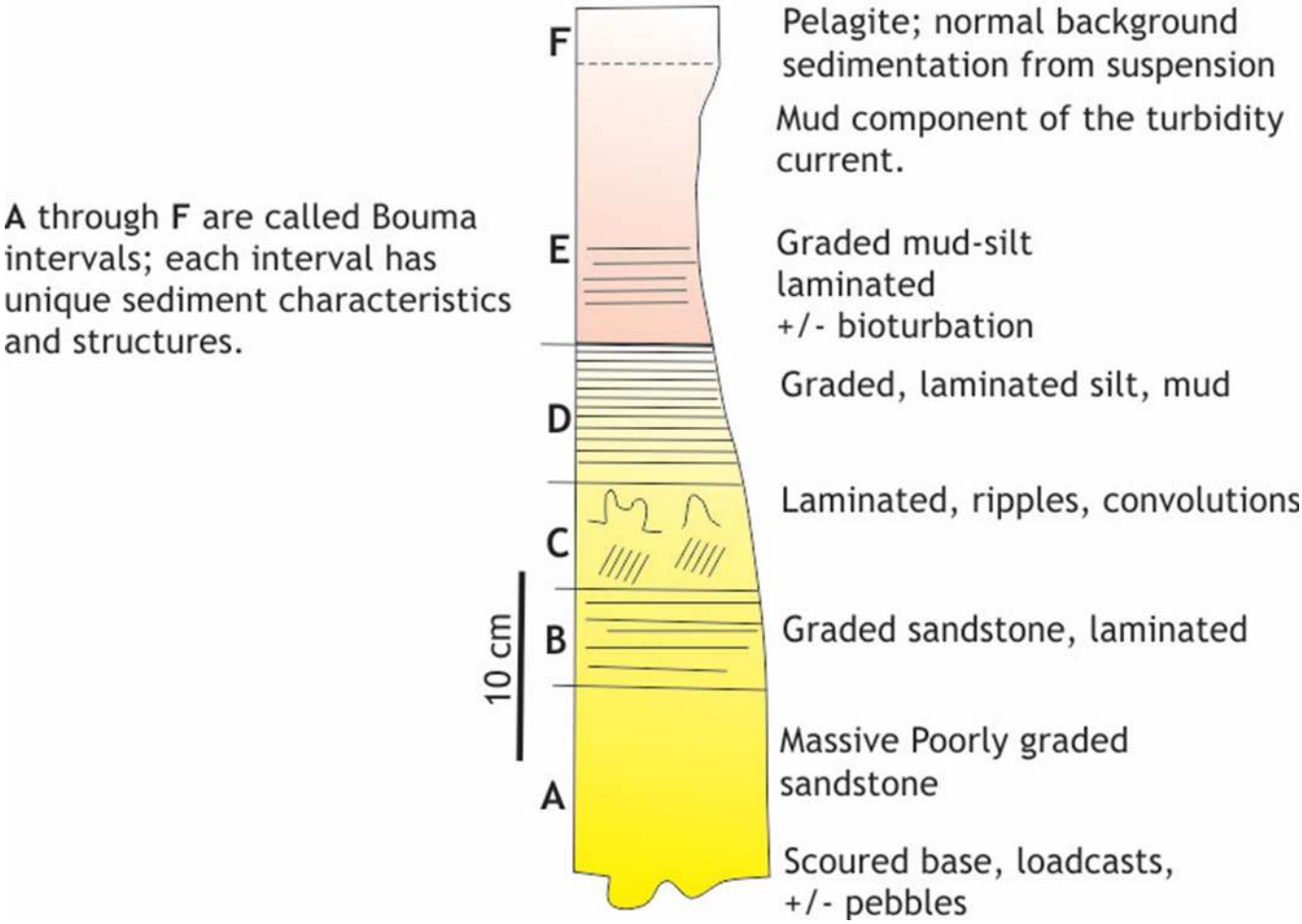


Figure 61.2.29: Classic Bouma model for sand-silt-mud turbidites



Figure 61.2.30: Graded laminae and beds in turbidite deposits of the Late Triassic Huntington Formation, Oregon. Because the grain size fines upward, these beds are upright. (Photo by Todd LaMaskin; reproduced with permission.)

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Figure 61.2.31: Turbidite metaconglomerate atop tuffaceous metasandstone in the Candelaria Formation in the High Sierra of California. Grain size decreases to the left; hence that is the younging direction (paleo-"up"). (Photograph by Ryan Hollister; reproduced with permission.)



Figure 61.2.32: Graded bed in Miette Group metasedimentary strata, Banff National Park, Alberta. (Callan Bentley photo.)

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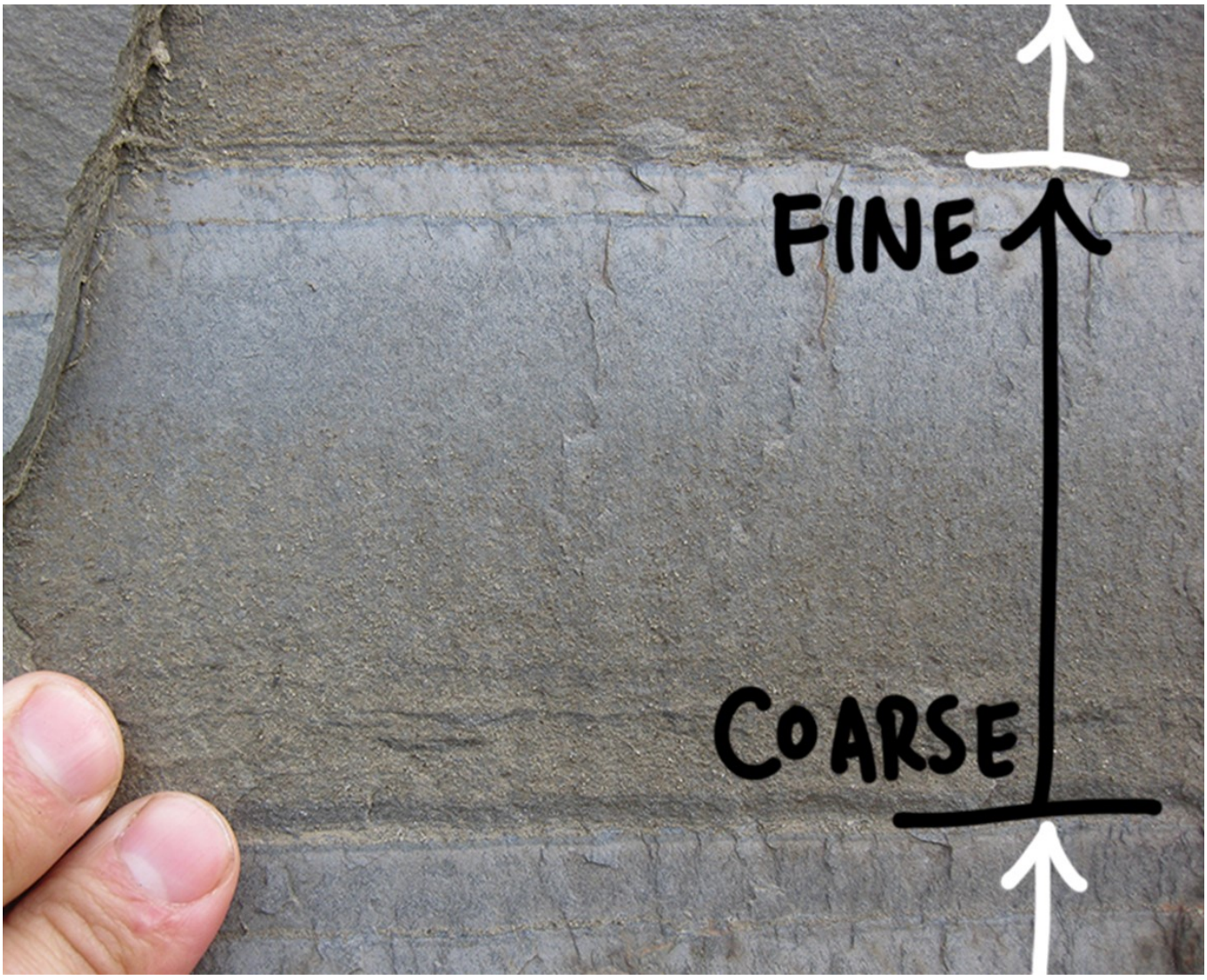


Figure 61.2.33: Graded bed in Miette Group metasedimentary strata, Banff National Park, Alberta. (Callan Bentley photo.)

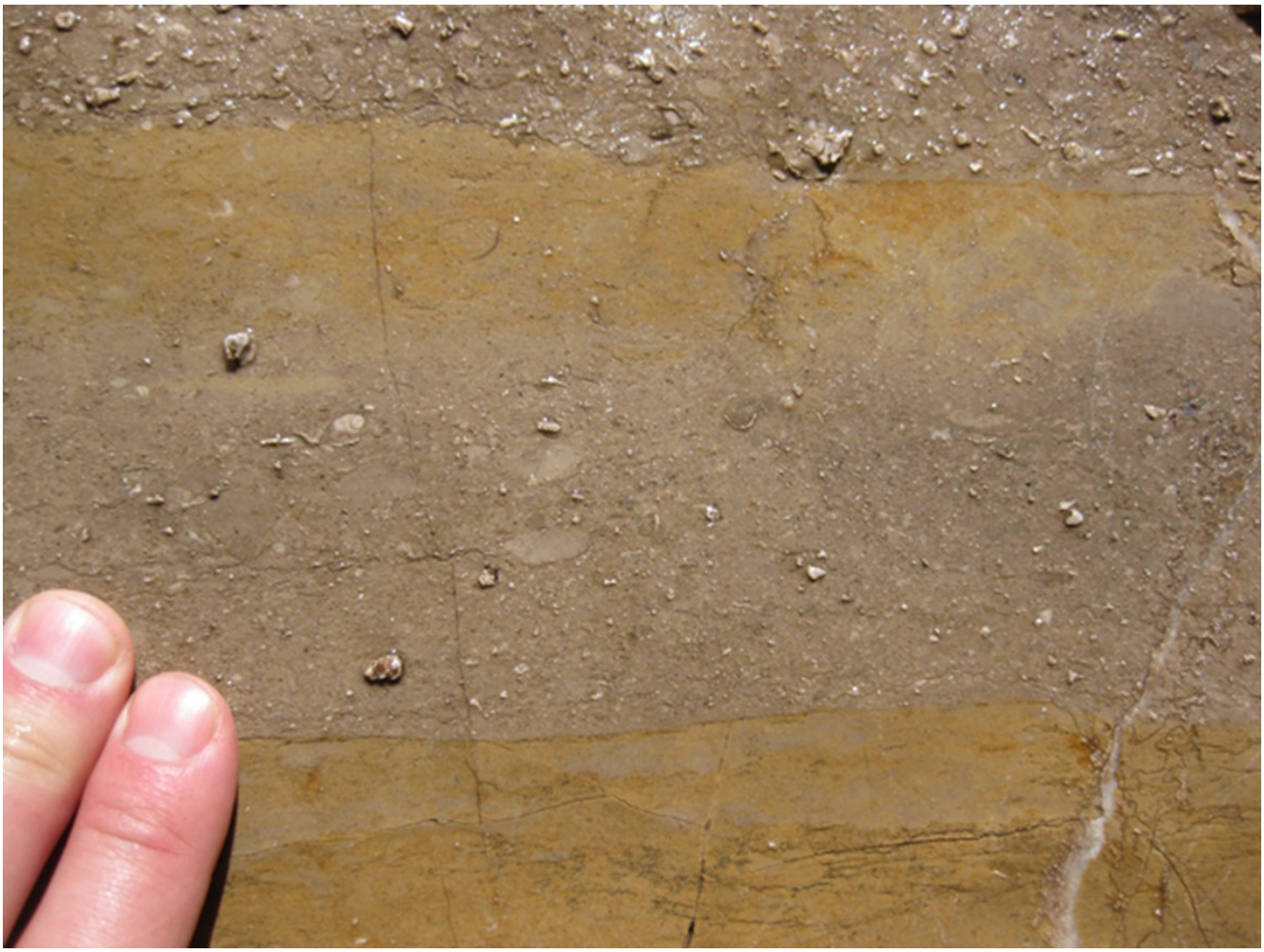


Figure 61.2.34: Graded beds aren't just for siliciclastic sediments! Here is a forereef carbonate turbidite in limestone of the Bell Canyon Formation, Guadalupe National Park, Texas. (Callan Bentley photo.)

Loading structures



Figure 61.2.35: Ball & pillow structures often form along the same sand/mud interface that produces flame structures. White arrows show downward motion; red arrows show upward motion.

Loading structures form due to density inversions, when heavy wet sand is deposited suddenly atop squishy waterlogged mud. The sand sags downward in broad lobes, and the mud squirts out of the way between those lobes. Once lithified, these bulging sandy lobes and pointy shale fingers serve as excellent way-up indicators. They are best viewed in cross-section, as seen here.

Ball & pillow



Figure 61.2.36: Ball & pillow loading structures made when sand was deposited on red mud, Graafwater Formation, Table Mountain, South Africa.

We call the sandy lobes that are broad and curved "ball & pillow." These lumps make blob-like shapes that bulge downward, pointing toward gravity's paleo-pull direction. They are lobate in form: convex downward, and concave upward. Any internal structures within the sand body, such as laminations or crossbeds, are likewise distorted downward most in the middle and least at the edges, resulting in a "smiley face" appearance in cross-section.

Here is a 3D model of some ball & pillow that formed in Devonian-aged redbeds of the Hampshire Formation in the Valley & Ridge province of West Virginia, when a thick layer of sand (today sandstone) was laid down atop a squishy layer of waterlogged mud (today red shale):



Flame structures



Figure 61.2.37: Flame structures in the Gualala Formation of California. (Photo by Brian Romans; reproduced with permission.)

Flame structures are what results when the mud squirts out of the way of the intrusion of the sagging sand lobes from above. It forms cuspate features that somewhat resemble flames, poking upward between the ball & pillow structures.

Did I Get It? - Quiz

Exercise 61.2.1

What structure(s) can be found along the contact between the upper conglomerate and the muddy sandstone layer below it? (i.e. just above the head of the older geologist) What are the implications?



a. This contact shows ball & pillow structures (on the base of the conglomerate), and flame structures (on the top edge of the muddy sandstone). The implication is that there was sedimentary loading here, creating a density inversion, and the heavy gravel sagged down into waterlogged muddy sand, which squirted up around the lobes of the "ball & pillow" to form large flame structures.

b. The contact shows cross-beds, which imply deposition in a directional current. Based on the orientation of the cross-beds, the current must have been flowing from the left toward the right. The size of these cross-beds implies a migrating submarine dune, rather than a ripple. (Because of the size of the cobbles in the gravel, it could not be aeolian deposition.)

Answer

> a. This contact shows ball & pillow structures (on the base of the conglomerate), and flame structures (on the top edge of the muddy sandstone). The implication is that there was sedimentary loading here, creating a density inversion, and the heavy gravel sagged down into waterlogged muddy sand, which squirted up around the lobes of the "ball & pillow" to form large flame structures.

Exercise 61.2.2

Which way is up in this outcrop? Why?

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a. Down is paleo-"up," since the beds get coarser upward rather than downward, and also because of the one layer that shows up-side-down flame structures in the mud, in between upward-bulging ball & pillow structures made of sand. The beds have therefore been tectonically inverted.

b. Up is paleo-"up," since the beds get finer upward, and also because of the one layer that shows flame structures in the mud, pointing upward in between downward-bulging ball & pillow structures made of sand. The beds are therefore in their original depositional orientation.

Answer

a. Down is paleo-"up," since the beds get coarser upward rather than downward, and also because of the one layer that shows up-side-down flame structures in the mud, in between upward-bulging ball & pillow structures made of sand. The beds have therefore been tectonically inverted.

Exercise 61.2.3

Which way-up structure is shown in this 3D model?



a. Graded bedding

b. Ball & pillow

c. Cross-bedding

d. Mud cracks

Answer

b. Ball & pillow

? Exercise 61.2.4

Identify the way-up structure in this outcrop, and offer an interpretation.



a. The central sandstone layer shows "ball & pillow" load structures on its lower surface. Because they are convex-downward, the strata appear to be in their original upright position. Paleo-"up" is toward the top of the photo.

b. The central sandstone layer shows cross-bedding. Because the crossbeds are tangential at the bottom and truncated at the top, the strata appear to be in their original upright position. Paleo-"up" is toward the top of the photo.

Answer

a. The central sandstone layer shows "ball & pillow" load structures on its lower surface. Because they are convex-downward, the strata appear to be in their original upright position. Paleo-"up" is toward the top of the photo.

Sole structures

Sole structures are erosional features formed along the sedimentary interface by powerful currents that then deposit a bed atop their erosional divots and carvings. They include flutes, tool marks, scours, and gutters. All share in common their mode of formation: (1) there must be a pre-existing sedimentary substrate, which (2) gets eroded, making a hole or depression, and then (3) the hole gets filled in with an overlying deposit of sediment. Most all sole structures are preserved not as their original negative shape (a hole) but as the positive shape that results when that hole is filled (a cast). Let’s examine this relationship in the context of flute marks first.



Figure 61.2.38: Flute casts on the underside of a sandstone bed. Devonian of Pennsylvania.

Often the erosion takes place on/in a substrate of mud (a calm-water deposit), when current energy later increases. The increased current cuts into the soft mud, but also brings in sand (a coarser sized particle). As current energy wanes, the sand is deposited atop the mud, including in the new holes and grooves, making "casts" of them. Much later, when the sediment has all been lithified and the rock has been uplifted to Earth's surface, the two different rock types react differently to weathering. The mud has turned to shale, which rapidly degrades and falls apart. But the sand has turned to sandstone, which typically stands up much more sturdily to the forces of weathering. It resists degradation. The result is that we typically find the bulging ***cast*** rather than the original concavity that was eroded out as a ***hole***.

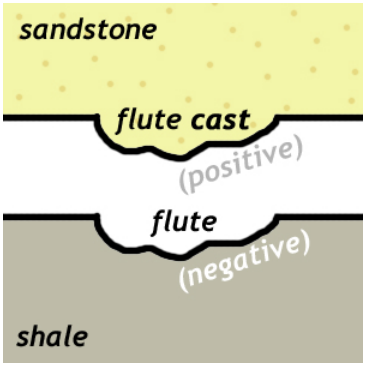
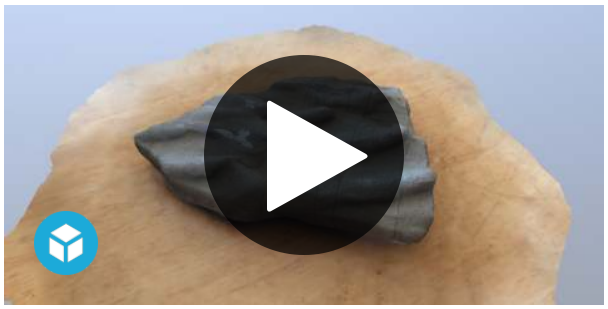


Figure 61.2.39: The relationship between a "negative" flute cut into mud (later lithified to shale) and the "positive" flute cast at the base of an overlying sand deposit (later lithified to sandstone).

One way of describing the situation is like this: the scour is a "negative" space (a hole), while the cast is a "positive" feature, poking out as a bulge or bump. Whether we're lucky enough to see the original flute, or only the secondary cast of that flute, the relationship is plain. Negative spaces are cut into the paleo-"down" direction, and their infilling (the cast) mimics this same direction.

Because both the concave-up flutes and their convex-down casts bulge in the direction of paleo-"down," they are useful as way-up indicators.

Here is a 3D model of flute casts, showing the underside of the sand bed that filled in the original flute:



Here are a few more examples to scroll through:



Figure 61.2.40: Close-spaced flute casts in an Eocene shelf sandstone; flow direction indicated with blue arrow.



Figure 61.2.41: Large flute casts from Malaysia. Flow direction was from the top of the screen toward the bottom. (The beds have been rotated into a vertical orientation since deposition and lithification.) (Photograph by Zoltán Sylvester; reproduced with permission.)



Figure 61.2.42: Sole structures at the base of a submarine fan channel conglomerate. Flow was from left to right. (Photograph by Zoltán Sylvester; reproduced with permission.)

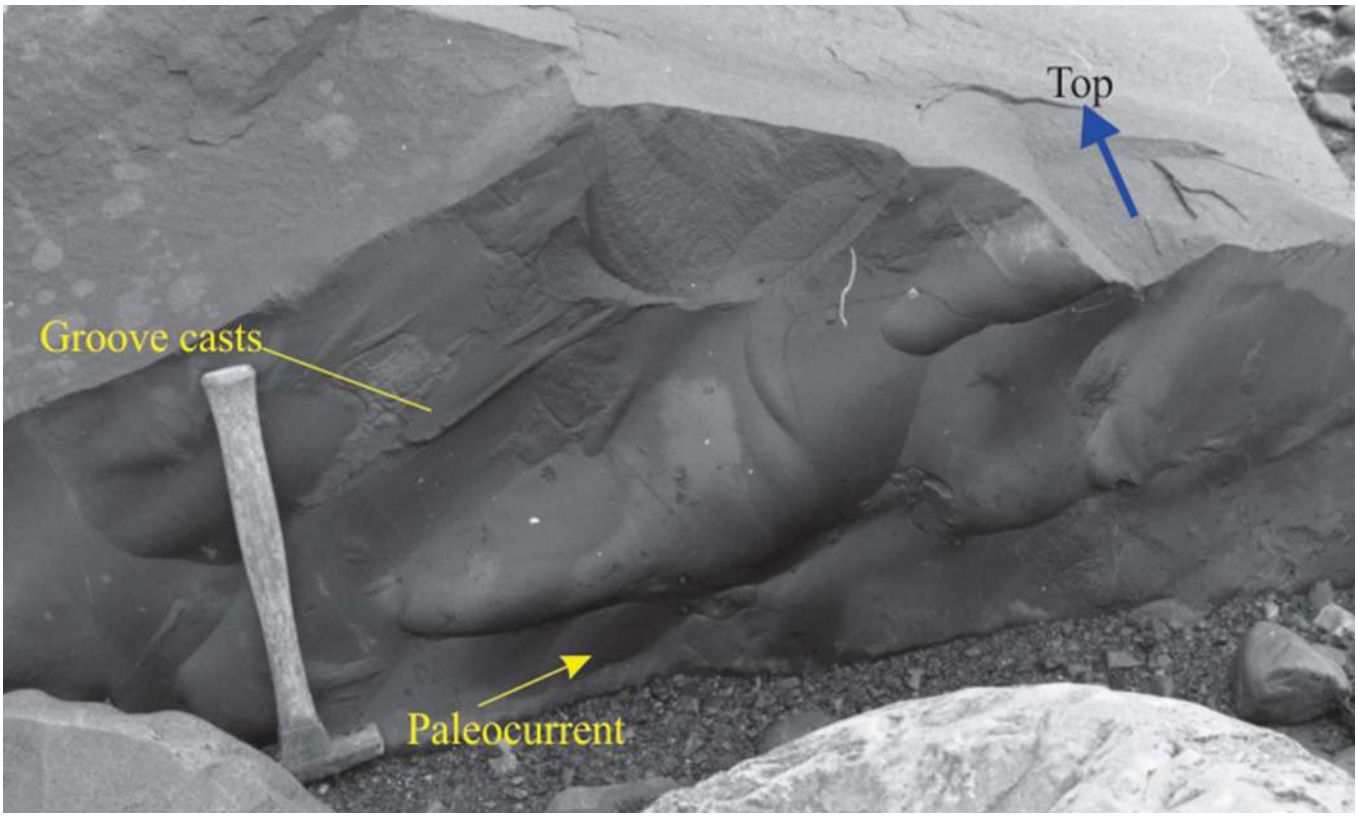


Figure 61.2.43: Large flute casts at the base of a turbidite bed. Current flow (indicated with yellow arrow) determined from flutes is generally unique. The groove casts parallel the flutes but inferred flow is ambiguous.



Figure 61.2.44: Flute casts in late Ordovician Martinsburg Formation, Page County, Virginia. Flow direction was from the upper right toward the lower left.



Figure 61.2.45: Flute casts on the base of a bed of sandstone from the Inverness Formation (Pennsylvanian), western Cape Breton, Nova Scotia. Flute casts are bulbous on the upstream side and taper on the downstream side; in this case recording paleoflow from right to left. (CC-BY; Photograph by M.C. Rygel.)



Figure 61.2.46: Flute casts on the lower bedding surface of rocks of Neoproterozoic (Brioverian) age near Keric Bihan in Brittany, France. The view is about 50 centimetres across. (CC-BY; Photograph by Sophie Coat.)

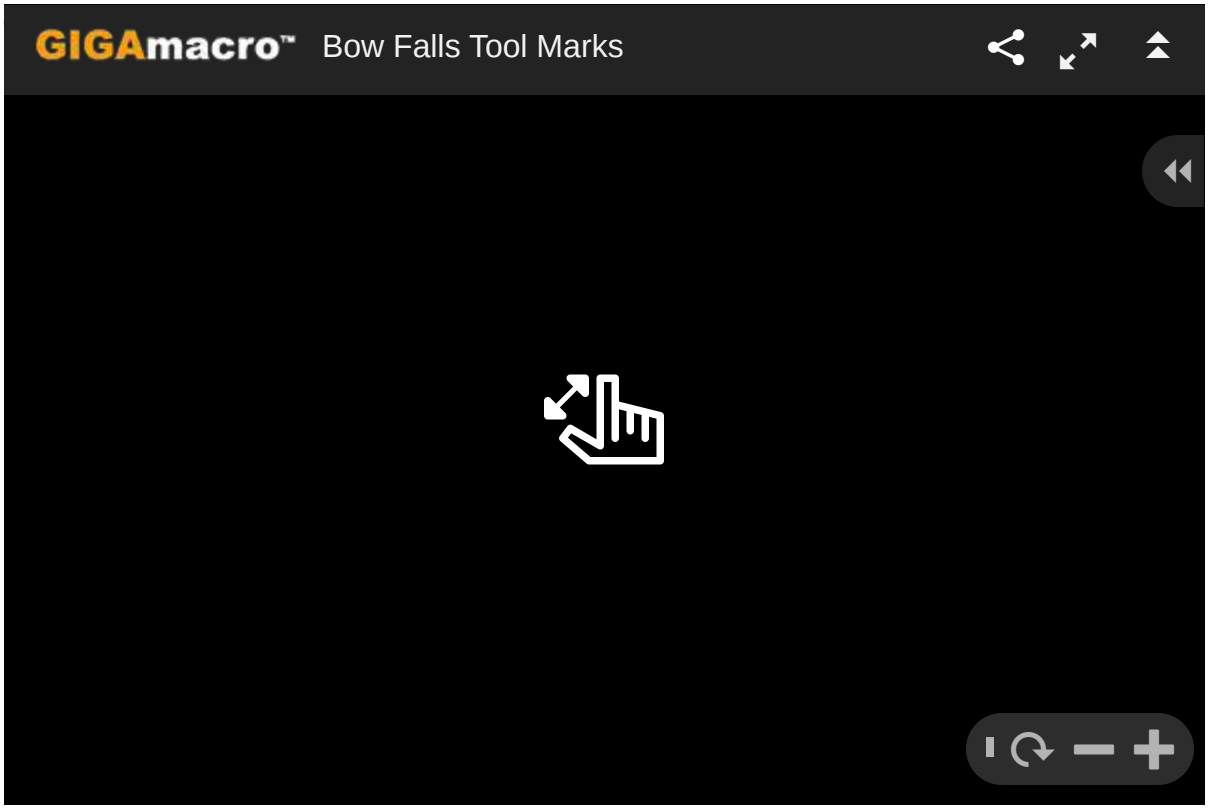


Figure 61.2.47: Flute cast formed by the action of turbidity currents on sediments deposited in shallow seas. This layer of sediment appears on the cliffs of Punta Carnero in the Spanish town of Algeciras. Because the flute casts project *upward* toward the sky, this bed has been tectonically inverted. (CC-BY; Photo by Falconaumanni.)

Tool marks

Tool marks form when hard objects such as cobbles, bones, wood, or shells get bounced along the bottom of a body of water, carried along by a forceful current. As the object bangs, crashes, drags, and bounces, it leaves behind a series of scratches and divots in the pre-existing sedimentary substrate. Later, these holes can be filled in with subsequent deposits, lithified, and preserved in the geologic record.

Tool marks (and casts thereof) are similar to flutes in their way-up implications: they project downward into the pre-existing sedimentary substrate, and when they are filled in by subsequent deposition, their shape is preserved as a cast on the bottom of the overlying bed. Here is an example, looking “up” at vertical Spray River Group siliciclastic strata from “below.” As such, the tool marks project off the bottom of the bed, toward our viewing perspective:



Did I Get It? - Quiz

? Exercise 61.2.1

Inspect this 3D model: Which way-up structure is shown? What are the implications?



a. This sample shows mud cracks on its upper surface; therefore this is the underside of a bed which filled in mud cracks that cut into a previously existing muddy substrate (now removed by erosion).

b. This sample shows flame structures on its upper surface; therefore this is the top side of a bed which was squished by heavier sediment being dumped on top of it (a layer now removed by erosion).

c. This sample shows flute casts on its upper surface; therefore this is the underside of a bed which filled in flutes that were cut into a previously existing substrate (now removed by erosion).

Answer

c. This sample shows flute casts on its upper surface; therefore this is the underside of a bed which filled in flutes that were cut into a previously existing substrate (now removed by erosion).

Exercise 61.2.2

Which way-up structure is this? Which way is up?



a. This sample shows casts of tool marks. We are therefore looking at the bottom of the bed.

b. This sample shows mud cracks. We are therefore looking at the bottom of the bed.

c. This sample shows casts of mud cracks. We are therefore looking at the top of the bed.

d. This sample shows tool marks. We are therefore looking at the top of the bed.

Answer

a. This sample shows casts of tool marks. We are therefore looking at the bottom of the bed.

Exercise 61.2.3

Identify the way-up structure shown in circled region of this sample (presented as a gigapixel panorama). What are the implications?



a. Three flutes are shown. The implication is that the current which deposited this bed was traveling from the right to the left, and this is the top of the resulting bed. So paleo-"up" is toward our perspective; we are looking "stratigraphically downward."

b. Two flute casts are shown. The implication is that the current which deposited this bed was traveling from the left to the right, and this is the bottom of the resulting bed. So paleo-"up" is away from our perspective; we are looking "stratigraphically upward."

c. A ball & pillow loading structure is shown. The implication is that the current which deposited this bed deposited a lot of heavy sediment quickly, and this is the bottom of the resulting bed. So paleo-"up" is away from our perspective; we are looking "stratigraphically upward."

d. Fifteen mud cracks are shown. The implication is that the sediment dried out in the open air, and this is the top of the resulting bed. So paleo-"up" is toward our perspective; we are looking "stratigraphically downward."

Answer

b. Two flute casts are shown. The implication is that the current which deposited this bed was traveling from the left to the right, and this is the bottom of the resulting bed. So paleo-"up" is away from our perspective; we are looking "stratigraphically upward."



Figure 61.2.48: A burrow cannot be cut into sand unless that sand is already there. So the burrow points to “paleo-down.”

A similar logic applies to burrows, the holes dug by organisms into sedimentary substrates. An animal can only dig a hole in the mud if the mud is already there to dig in. The hole cannot precede the material into which the hole is dug. So open burrows that are later filled in with contrasting sediment can also serve as a geopetal indicator. Consider the example at right.

Similar, plants cannot take root in soil that isn’t yet there, so root traces also must penetrated in the paleo-“down” direction, at least in the most broad view. Once roots get deep enough, they can of course grow laterally or even upward in addition to growing downward.

The same idea applies to footprints, which can only press downward into extant sediment, not push upward into sedimentary deposits that haven’t yet been deposited. There are thousands of examples, but let’s consider two here: Consider the images below.



Figure 61.2.49: Cross-sectional view of a sauropod dinosaur footprint, Jurassic aged Morrison Formation. Dinosaur Ridge, Colorado. (Photograph by James St. John.)



Figure 61.2.50: Trace fossil track in Turkish siltstone.

One image shows a side-view of a sauropod dinosaur track in disrupting sandstone and shale layers. The other photo shows the sinuous trackway left by some kind of invertebrate in silt, looking down on the bedding plane. We have two different perspectives here, and two different organisms, but what is in common to these bedding-surface traces is that they push/erode down into the pre-existing sedimentary substrate.

Here is a 3D model of the enigmatic Cambrian trace fossil *Climactichnites*, on display on a wall of a museum at the University of Wisconsin:



Note that we must be looking down on this trace, since it cuts across the ripples (and not the other way around). So this is the top of the bed of sandstone including the trace fossil, and the trace projects downward into the sand, disrupting its previous surface feature (the ripples).

The opposite is true here:

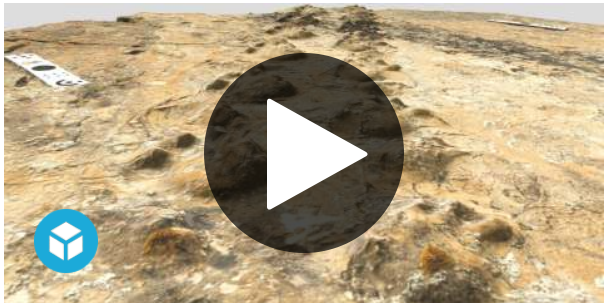




Figure 61.2.51: Permian reptile footprints on the underside of a dune cross-bed within the Coconino Sandstone, Grand Canyon, Arizona. The photo is lit from the left: these are positive-weathering features (they poke out of the bed surface). Field of view is about 1 meter wide by 2 meters tall.

Here, in a Permian-aged reptile trackway from the Grand Canyon, the footprints poke upward from the surface of the slab of rock, implying that this is a **cast** of the reptile's footprints, not the original divots pushed downward into the sediment.

Whether a monster ancient slug or a more familiar reptile, all organisms leave tracks that push **downward** into the sedimentary substrate over which they move. As with sole structures, the concave 'negative space' pokes downward into the older sedimentary layer, and the infilling creates a convex 'positive space' feature that projects downward off the bottom of the younger sedimentary layer.

Did I Get It? - Quiz

? Exercise 61.2.1

This 3D model shows the tracks of a giant eurypterid (sea scorpion) on a vertical sandstone bed in Scotland. What is our perspective on the orientation of the bed? Explain.



a. We are looking at the BOTTOM of the bed, since the tracks poke out toward our perspective.

b. We are looking at the TOP of the bed, since the tracks poke downward into the sandstone, away from our perspective.

c. We are looking at a CROSS-SECTIONAL view of the bed, since we can see how the tracks poke down into the sandstone toward the bottom of the screen.

Answer

a. We are looking at the BOTTOM of the bed, since the tracks poke out toward our perspective.

? Exercise 61.2.2

Are we looking at the top or the bottom of this bed of Jurassic sandstone in Colorado?



a. Bottom, since the dinosaur footprints project outward from the bed's surface as casts of the original footprints.

b. Top, since the dinosaur footprints are pressed downward into the bed's surface, making "negative space" hollows.

Answer

b. Top, since the dinosaur footprints are pressed downward into the bed's surface, making "negative space" hollows.

Exercise 61.2.3

This photograph shows a bedding surface in a dune sandstone. It is lit from the upper left. What surface of the bed is this, and why?



a. It is the top of the bed, since the fossil footprints are depressed into the surface, making hollow holes.

b. It is a cross-section of the bed, since we can see a sideways-view of the footprint shape.

c. It is the underside of the bed, since casts of the fossil footprints poke out of its surface.

Answer

c. It is the underside of the bed, since casts of the fossil footprints poke out of its surface.

Exercise 61.2.4

Are we looking at the top or the bottom of this bed of late Cretaceous limestone in Bolivia?



a. Bottom, since the dinosaur footprints project outward from the bed's surface as casts of the original footprints.

b. Top, since the dinosaur footprints are pressed downward into the bed's surface, making "negative space" hollows.

Answer

b. Top, since the dinosaur footprints are pressed downward into the bed's surface, making "negative space" hollows.

Exercise 61.2.5

Are we looking at the top or the bottom of this bed of Cambrian sandstone in Arizona?



a. Bottom, since the trace fossil pokes out at us as a positive-weathering feature. This is a cast of a trilobite crawling trace.

b. Top, since the trace fossil is depressed as a hollow (or "negative space") feature on the surface of the bed.

Answer

a. Bottom, since the trace fossil pokes out at us as a positive-weathering feature. This is a cast of a trilobite crawling trace.

Exercise 61.2.6

Which way is the geologist pointing? Why?



a. The geologist's hand is pointing downward, showing the direction a dinosaur's foot squished into sedimentary layers. Later, the footprint was filled in with a massive sand layer, preserving the trace fossil as a cast.

b. The geologist is pointing in the paleo-"up" direction, showing an up-side-down stromatolite. It bulges downward (convex down), which means these strata must have been tectonically inverted.

Answer

a. The geologist's hand is pointing downward, showing the direction a dinosaur's foot squished into sedimentary layers. Later, the footprint was filled in with a massive sand layer, preserving the trace fossil as a cast.

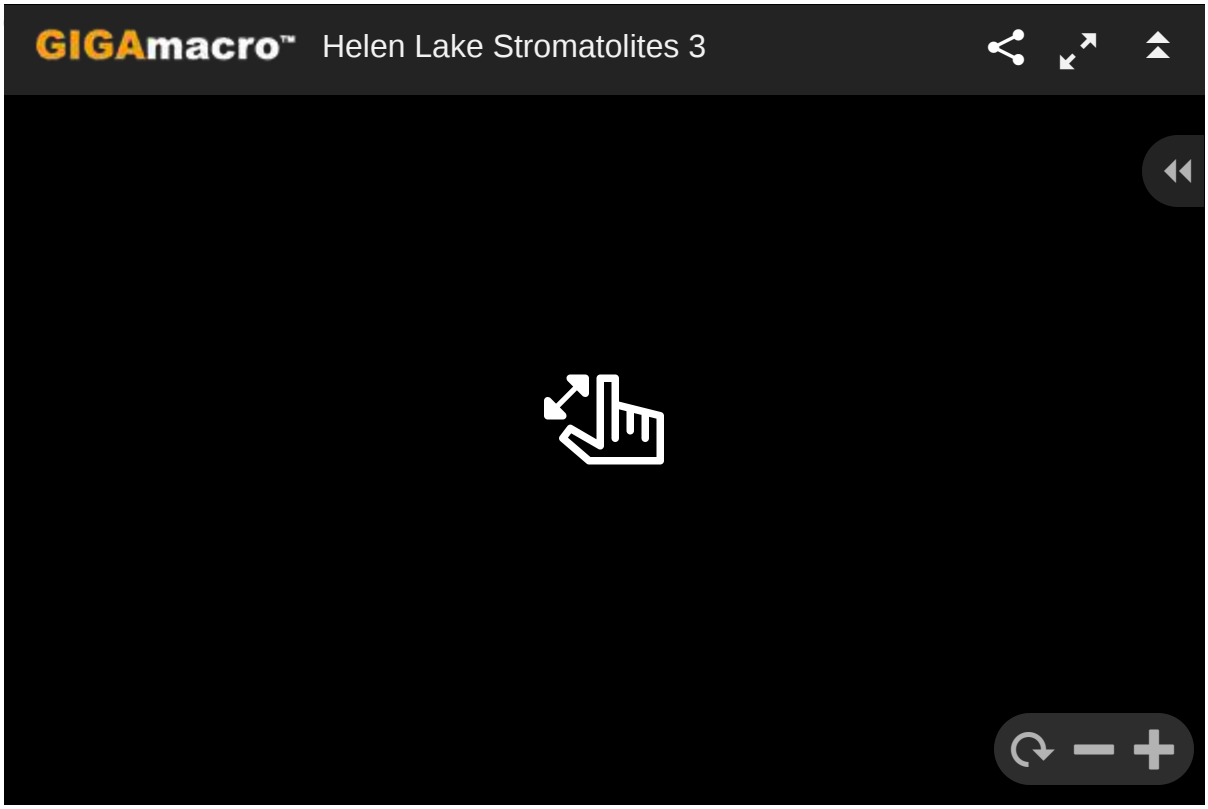
Stromatolites



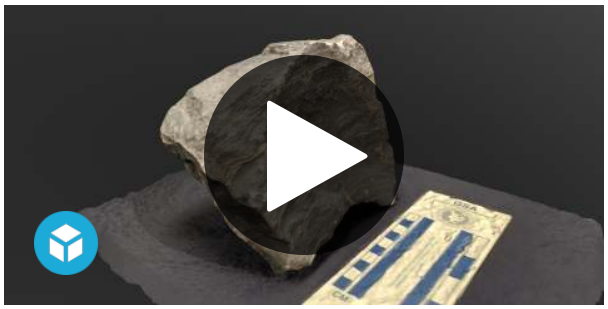
Figure 61.2.52: Stromatolite-bearing layers in the Crystal Springs Formation, a carbonate unit in the strata of the Death Valley region, California.

Stromatolites are layered microbial mats, common in Precambrian shallow-water sediments, particularly carbonates, but also sometimes siliciclastic sediments or even banded iron formations. Their photosynthetic life habit causes them to grow upward toward the sunlight, giving them convex-upward (dome-like) shapes. If concave-up instead, they can be interpreted to be up-side-down.

Here is a gigapixel panorama of an outcrop of stromatolites in Banff National Park, Alberta. Explore it to get a sense of their shape when right-side-up:



Here is a 3D model of a stromatolite, from the Mesoproterozoic Belt Supergroup of Montana. Rotate it around to get a sense of how the cross-sectional views on each side relate to the 3D dome-like structure of the fossil (visible on top):



Microbial mats can take on many shapes and textures, but this upward-doming shape of stromatolites is a reliable way-up indicator, especially when there are many individual stromatolites all pointing the same direction.

Test your identification of stromatolite younging direction with this self-quiz:

Did I Get It? - Quiz

? Exercise 61.2.1

Identify which way-up structure is shown in this vertical outcrop face. Are the strata right-side-up, or in some other orientation?



a. There are numerous cross-beds shown here. Because the cross-beds show tangential bottoms at the bottom of the photo and truncated upper surfaces toward the top of the photo (i.e. they are concave-up), they must be in their original depositional orientation.

b. There are numerous stromatolites shown here. Because the stromatolites dome downward toward the bottom of the photo (i.e. they are concave-up), they must be tectonically inverted.

c. There are numerous stromatolites shown here. Because the stromatolites dome upward toward the top of the photo (i.e. they are convex-up), they must be in their original depositional orientation.

d. There are numerous cross-beds shown here. Because the cross-beds show tangential contacts with the main bed at the top of the photo and truncated surfaces toward the bottom of the photo (i.e. they are concave-down), they must be tectonically inverted.

Answer

c. There are numerous stromatolites shown here. Because the stromatolites dome upward toward the top of the photo (i.e. they are convex-up), they must be in their original depositional orientation.

? Exercise 61.2.2

Examine this gigapixel panorama of a boulder of limestone containing stromatolites. Which way is paleo-"up," and why?

https://geo.libretexts.org/@go/page/22870



a. Left is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

b. Down is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

c. Up is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

Answer

b. Down is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

? Exercise 61.2.3

Examine this 3D model of a sample showing a series of small stromatolites. Which way is the paleo-"up" direction, and why?



a. As the model first loaded up, LEFT is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

b. As the model first loaded up, DOWN is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

c. As the model first loaded up, UP is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

Answer

a. As the model first loaded up, LEFT is the paleo-"up" direction, since the stromatolites' convex shapes dome in that direction.

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61.3: Igneous way-up structures

Next up, we will consider the handful of igneous way-up structures. There are not as many as there are in sedimentary rocks, but though few in number, they can come in handy in deformed terranes with igneous strata or intrusions.

Vesicle concentration



Figure 61.3.1: Vesicles are less dense than fluid lava, so they rise within a flow. Paleo-“up” is to the left in this sample.

In lava flows that have significant gas content, bubbles (vesicles) form as the lava degases upon eruption. These bubbles have low density, and so want to rise to the top of the fluid flow. Given a sufficiently runny lava, they can make their way upward, accumulating beneath the upper cooling crust of the flow. This creates a concentration of vesicles at the top of of flow, and more massive fine-grained textures at the base of the flow.

Caveat: Lava flows are dynamic systems, and pulsing currents of lava and differential cooling can mix up the vesicle-rich portions and vesicle-free portions. Search for other clues which corroborate the vesicles’ way-up implications.

Apophyses



Figure 61.3.2: In Utah's House Range, a Jurassic monzonite (an intrusive igneous rock like a granite) intruded an apophysis into Cambrian sedimentary strata.

Apophyses are small injection features, like small dikes, that protrude out from a lava flow or sill into surrounding pre-existing rocks. In a lava flow, they can only project downward, but in a sill they can project both upward and downward. Therefore, to use apophyses as a way-up indicator, a necessary first step is to verify that the apophyses are coming off a lava flow, not an intrusive sill. The

Post-igneous sedimentary strata laid down atop a lava flow may show their own "apophyses" as cracks in the top of the lava flow are filled in with sediment. Below is a cartoon summary of the nature of apophyses for telling sills from lava flows, and telling if lava flows are right-side-up or not:

Pillow lavas



Figure 61.3.3: Pillow structures in Columbia River flood basalt, eastern Washington. Yellow field notebook for scale.

When lava erupts underwater, lava pillows form. On the outside, they have a shape like toothpaste extruded from a toothpaste tube. In cross-section, they show a glassy outer rind and fine-grained (aphanitic) interior, often with radial fractures. As a freshly formed pillow, with its still-molten interior, flows and settles into the submarine topography of older pillows, it often sags a bit in the

middle, producing a little downward “pooch.” It looks something like an up-side-down teardrop shape, or a “word bubble” in a comic strip.

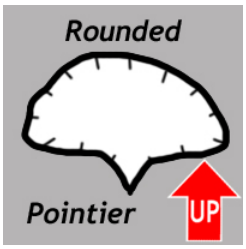


Figure 61.3.4: Pillow basalts frequently show a distinctive shape that is useful as a way-up indicator.

Bear in mind that the eruption of lava underwater is a messy process, and a variety of 3D pillow shapes can be produced. Further complications arise when we view these 3D shapes in 2D outcrops, as cross-sectional views. Many pillows are just round or cylindrical, but enough pillows show this distinctive shape that in a decent outcrop, there should be enough that you can figure out the younging direction. Sometimes, they will show more than one downward projection.

In this video, geologist Christie Rowe shows an outcrop of pillows near San Francisco, California:



Plutonic grading and cross-bedding



Figure 61.3.5: Graded “bedding” in the Skaergaard intrusion, Greenland. Each “bed” shows a concentration of heavier mafic minerals at left, and lower-density plagioclase feldspar (white colored) at right. (Photo by Kurt Hollacher; reproduced with permission.)

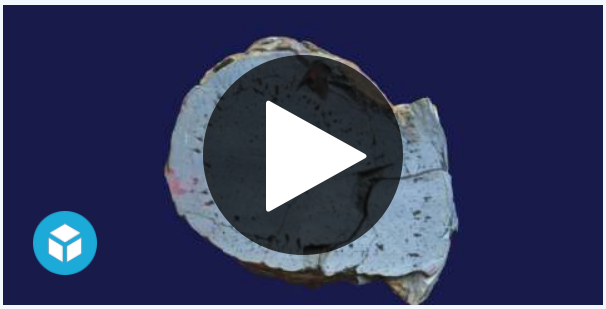
Crazy as it may sound, the dynamics inside a cooling pluton can mimic some of the same situations we observe at the surface with sediments deposited in flowing water. These include cross-bedding formed by magma flowing through a crystal mush, and graded bedding that results as crystals settle out from a cooling magma. The key here is that in a partially crystallized magma chamber,

some minerals are more dense than the melt, and others are less dense. Ferromagnesian (mafic) minerals such as olivine and pyroxene are denser than their surroundings, and will sink if given the chance. Plagioclase feldspar is less dense, and so will “float” upward. Because the ferromagnesian minerals are dark, and the plagioclase is almost white in color, this can create visually striking patterns that are as informative as they are beautiful.

Did I Get It? - Quiz

Exercise 61.3.1

What primary igneous structure is shown here?



a. Lava pillow

b. Vesicle grading

c. Graded layering

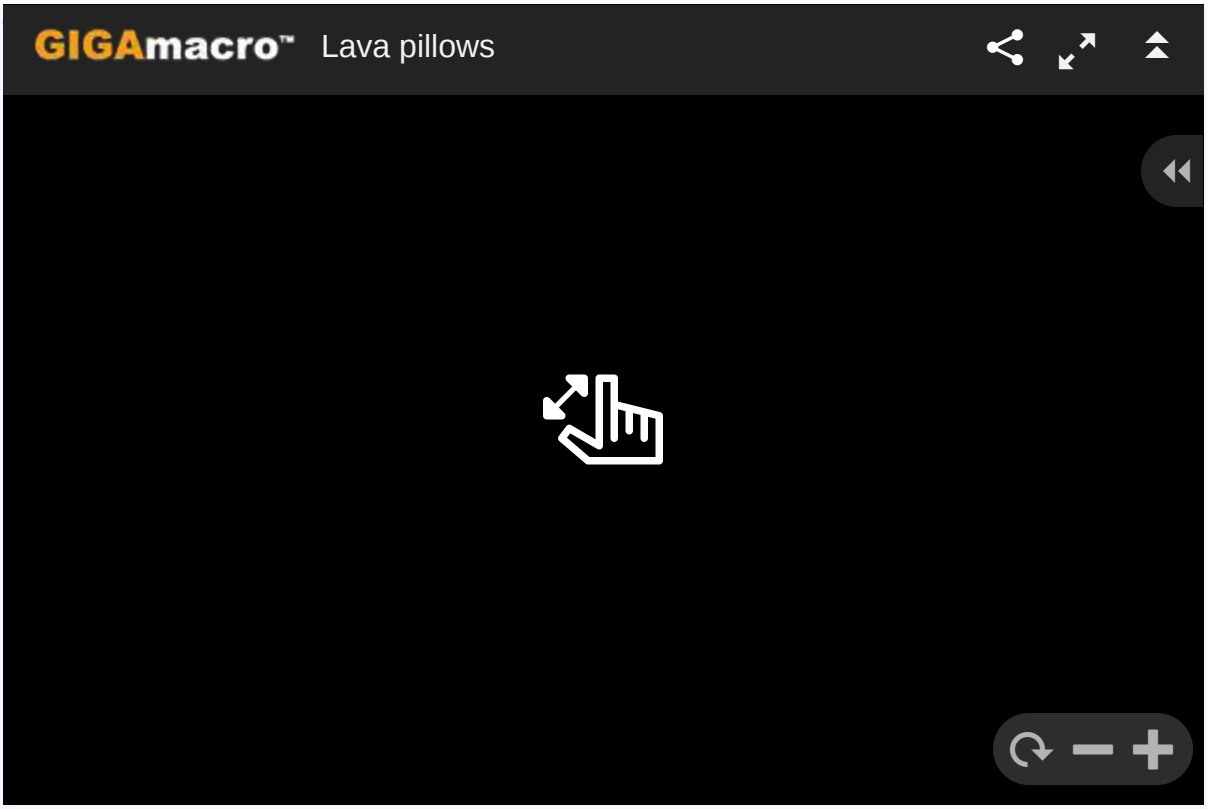
c. Apophysis

Answer

a. Lava pillow

Exercise 61.3.2

Observe this scene. What conclusions can you draw?



a. This is a site where lava once erupted underwater. Since then, these rocks have been rotated out of their original position. Paleo-"up" it to the upper left, and paleo-"down" is to the lower right.

b. This is a site where lava once erupted underwater. Since then, these rocks have been rotated out of their original position. Paleo-"up" it to the lower left, and paleo-"down" is to the upper right.

c. This is a site where magma once intruded into sediments. Since then, these rocks have been rotated out of their original position. Paleo-"up" it to the lower left, and paleo-"down" is to the lower right.

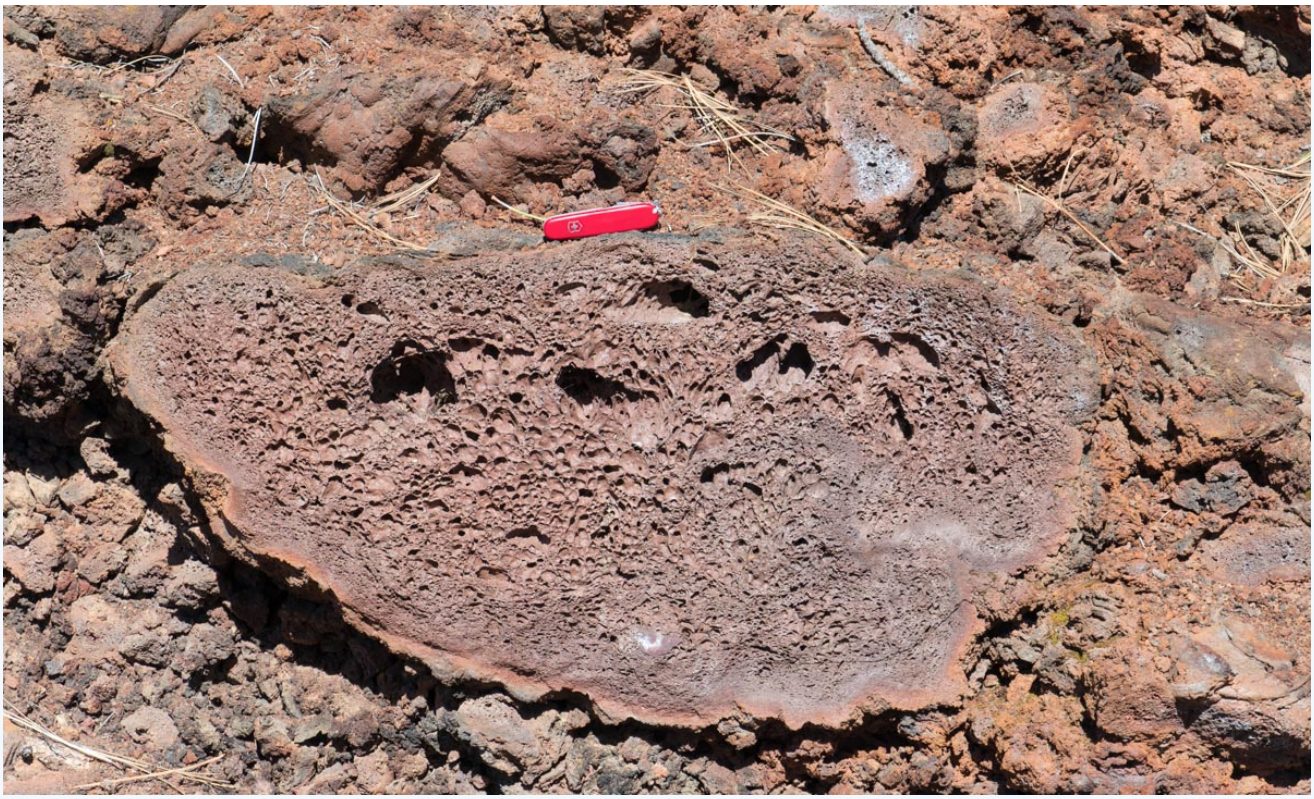
d. This is a site where magma once intruded into sediments. Since then, these rocks have been rotated out of their original position. Paleo-"up" it to the upper left, and paleo-"down" is to the lower right.

Answer

a. This is a site where lava once erupted underwater. Since then, these rocks have been rotated out of their original position. Paleo-"up" it to the upper left, and paleo-"down" is to the lower right.

Exercise 61.3.3

This lava bomb contains vesicles. Is it right-side-up or up-side-down? Explain.



a. It is right-side-up, since both the number and the size of the vesicles is larger at the top of the bomb's interior.

b. It is up-side-down, since both the number and the size of the vesicles is larger at the bottom of the bomb's interior.

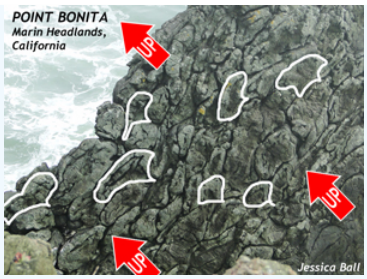
Answer

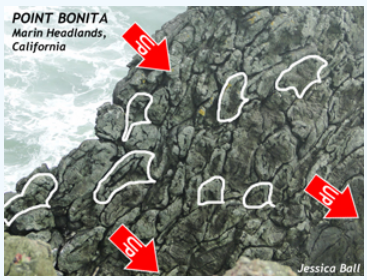
a. It is right-side-up, since both the number and the size of the vesicles is larger at the top of the bomb's interior.

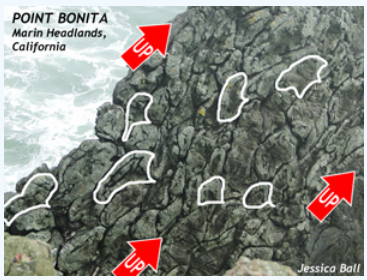
? Exercise 61.3.4

Which way is up in this image? Several way-up indicators have been traced out to aid your interpretation.

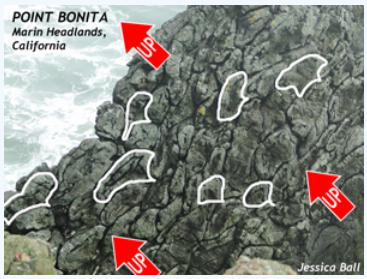


a. 

b. 

c. 

Answer

a. 

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61.4: Conclusion

Many of the structures produced in rocks through geological processes do not record information about paleo-“up” direction. But there are a dozen or so “way-up indicators” that occur in sedimentary strata and some igneous rocks that can help the historical geologist infer younging direction with confidence. Knowing these way-up structures is a critical piece of knowledge in the successful interpretation of stratigraphic sequences in deformed terranes.

Further reading

“Way-up structures” on Wikipedia.

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CHAPTER OVERVIEW

62: Virtual Field Experiences (VFEs)

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62.2: Fossil identification and use in correlation, part 2
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62.4: Glacier National Park, Montana
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62.1: Fossil correlation

VFE: Fossil identification and use in correlation

Part 1: learn the relevant fossil taxa

This webpage is your identification guide so you can learn to recognize each fossil species. Each fossil's name (usually genus or ichnogenus, sometimes genus and species) is given, plus any relevant taxonomic or anatomical information, followed by its geologic range (the time during which it is found in the geologic record, *i.e.*, the time between its first instance and its last instance).

In reviewing these images, your goal is to **get familiar with the different fossils**, so you can then recognize them in Part 2 of this VFE, and thereby use their geologic ranges to figure out how old a sequence of sedimentary strata are.

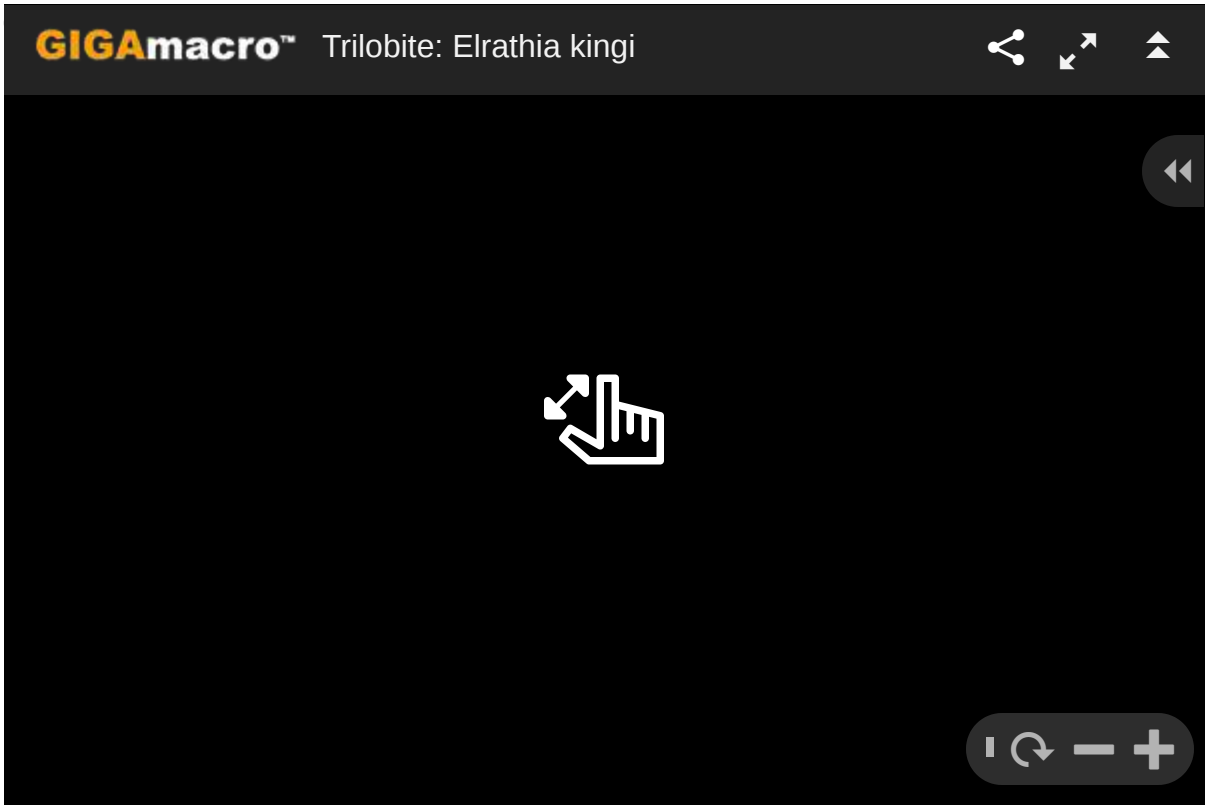
Graptolites

Glossograptus, lower to middle Ordovician:

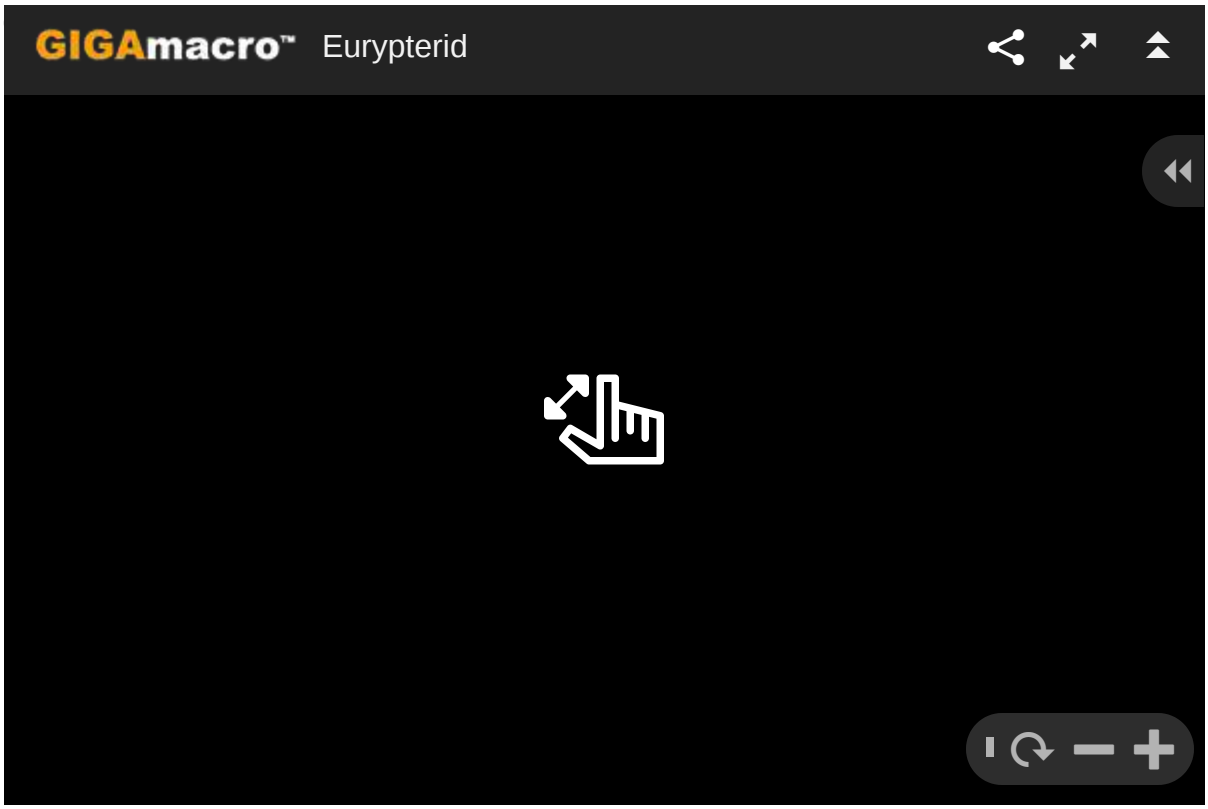


Arthropods

Elraithia, trilobite, Cambrian:



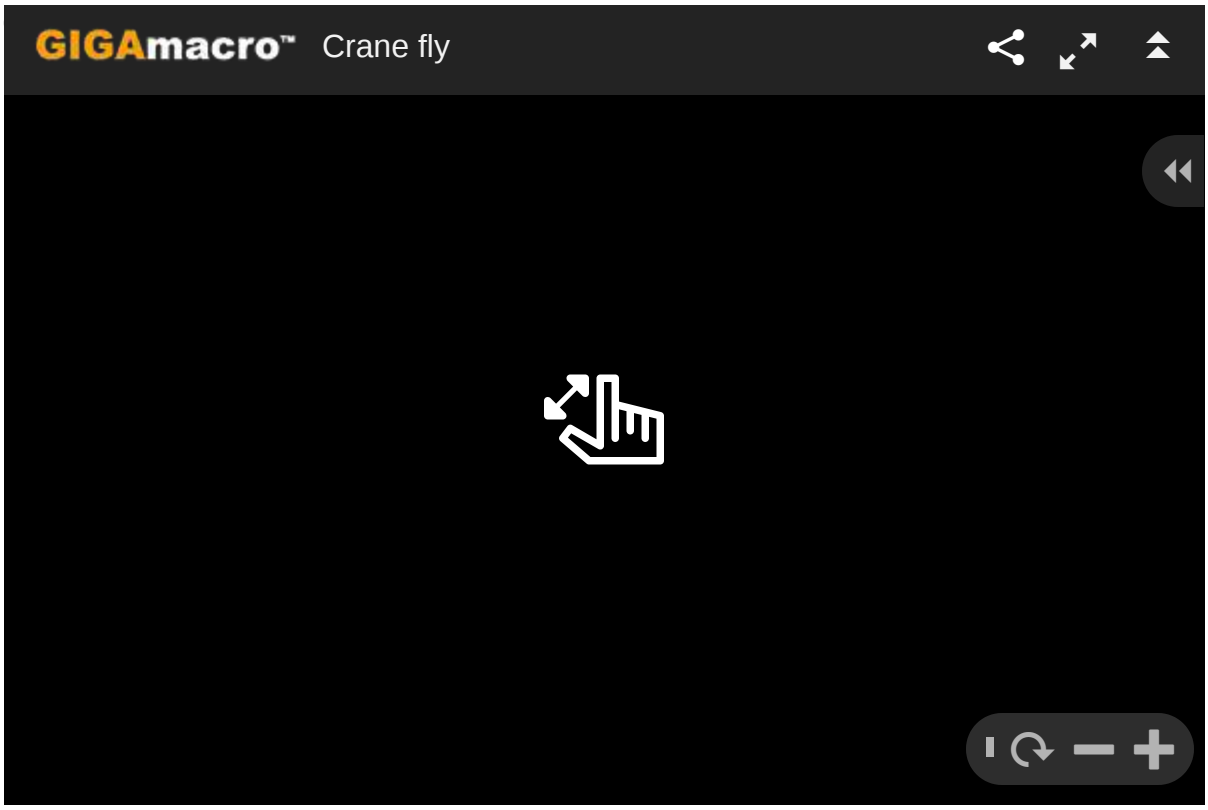
Eurypterus, eurypterid, **note: cephalon (head) only**, Silurian:



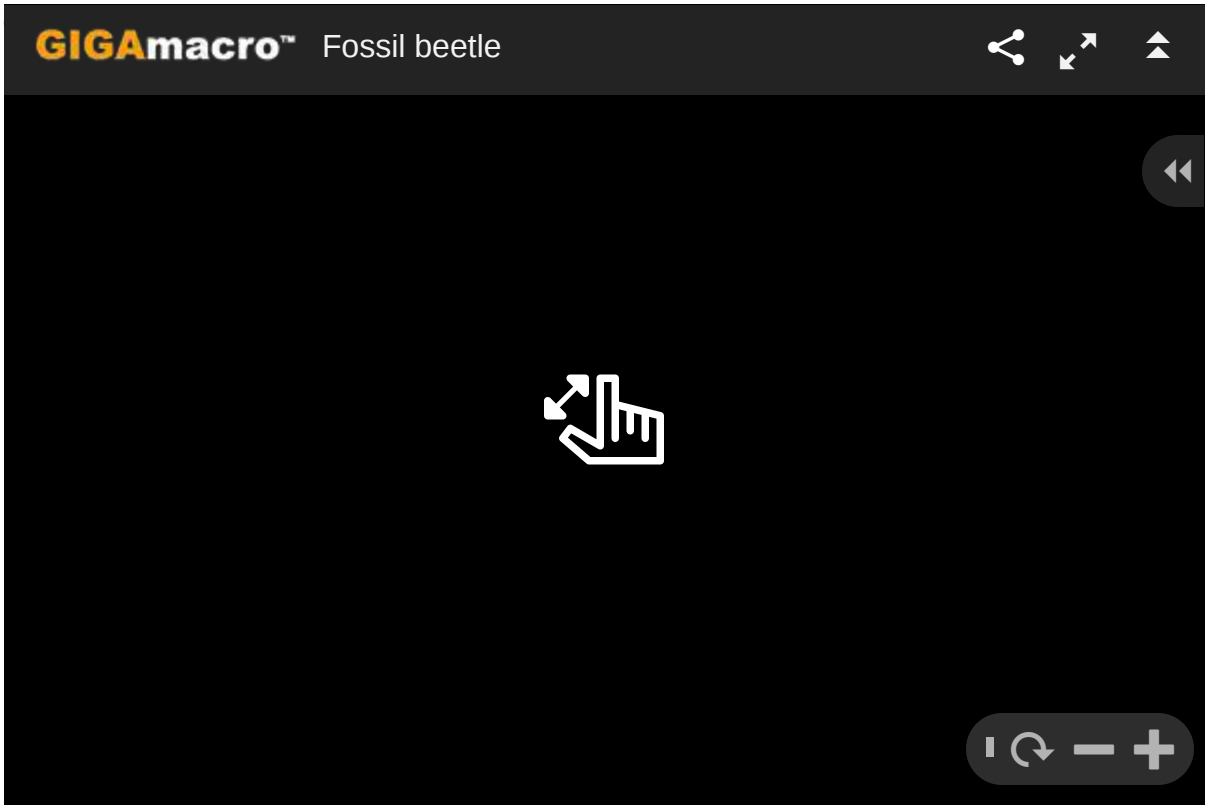
Krausella, ostracode, Ordovician to Silurian:



Crane fly, insect (family Tipulidae), Eocene to modern:



Beetle, insect (family Coleoptera), Pennsylvanian (Carboniferous) to modern:



Mollusks

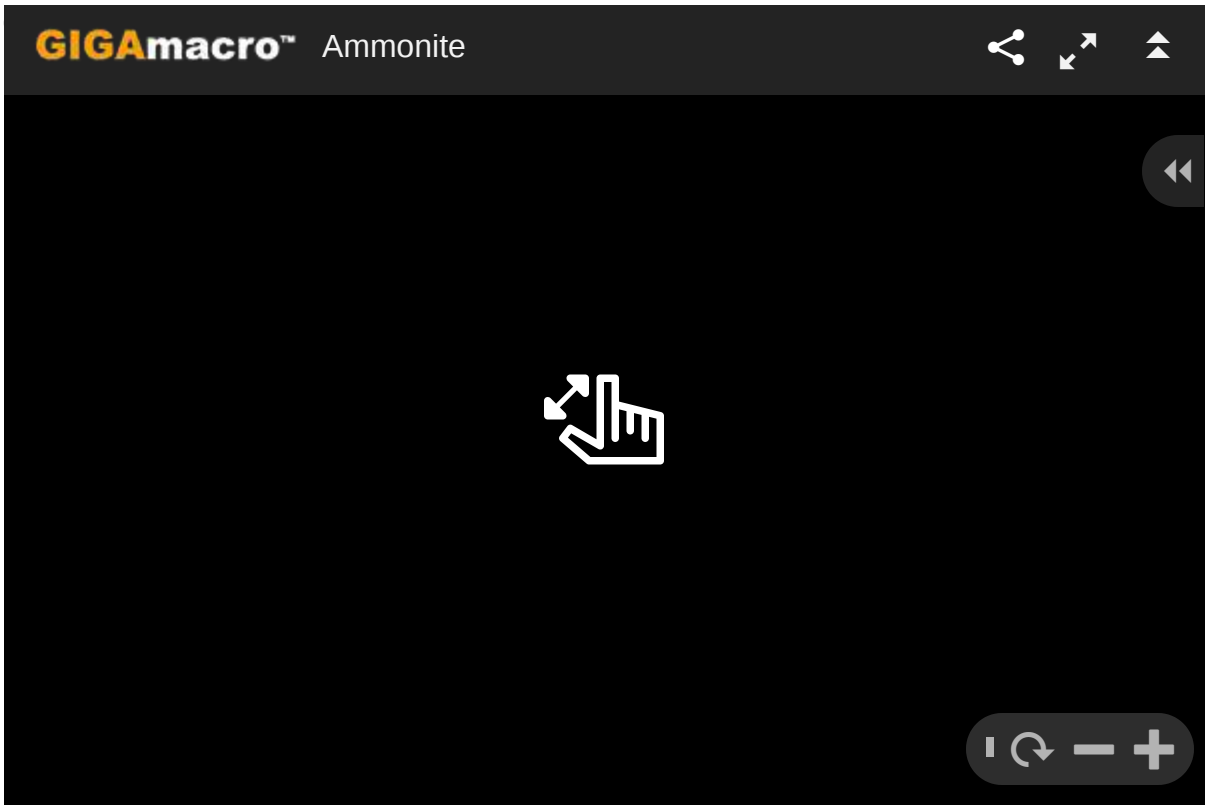
Turitella, gastropod, Cretaceous to modern:



Nautiloid, cephalopod, Devonian:



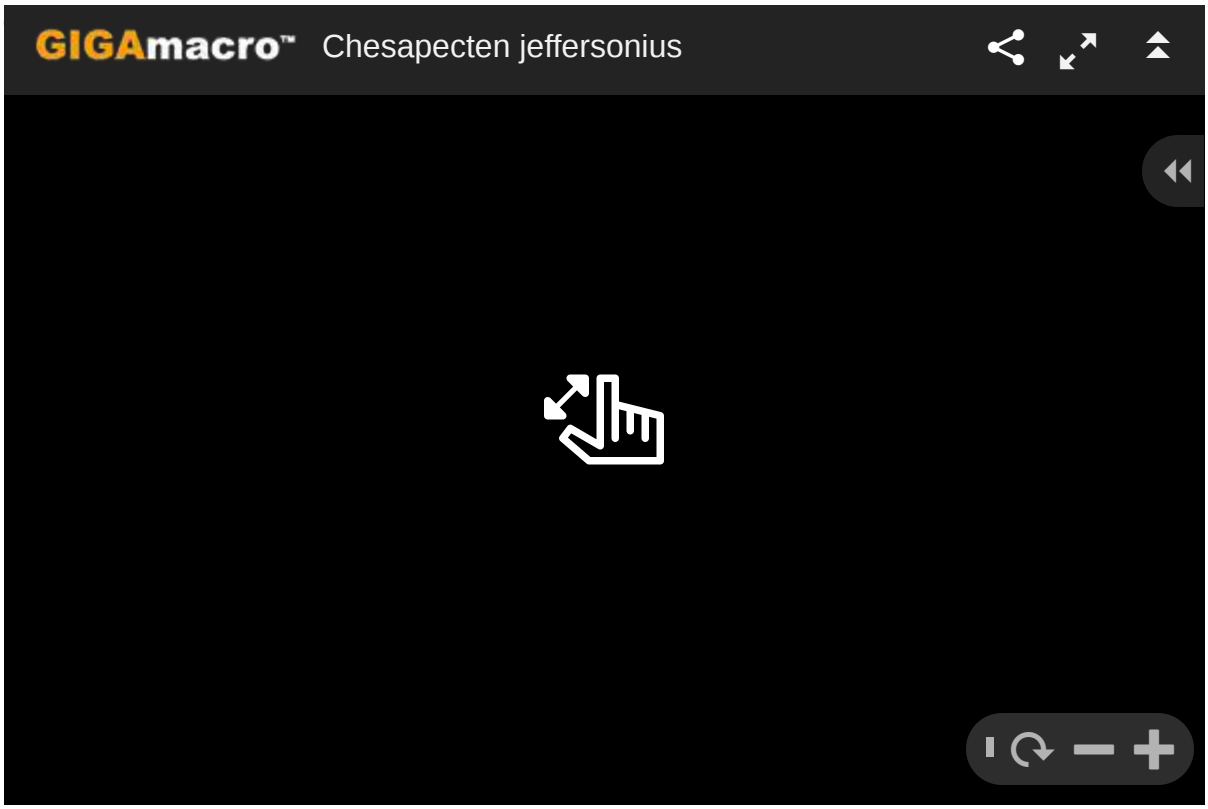
Ammonite, cephalopod, Jurassic to Cretaceous:



Chesapecten nefrens, scallop (bivalve), Miocene:



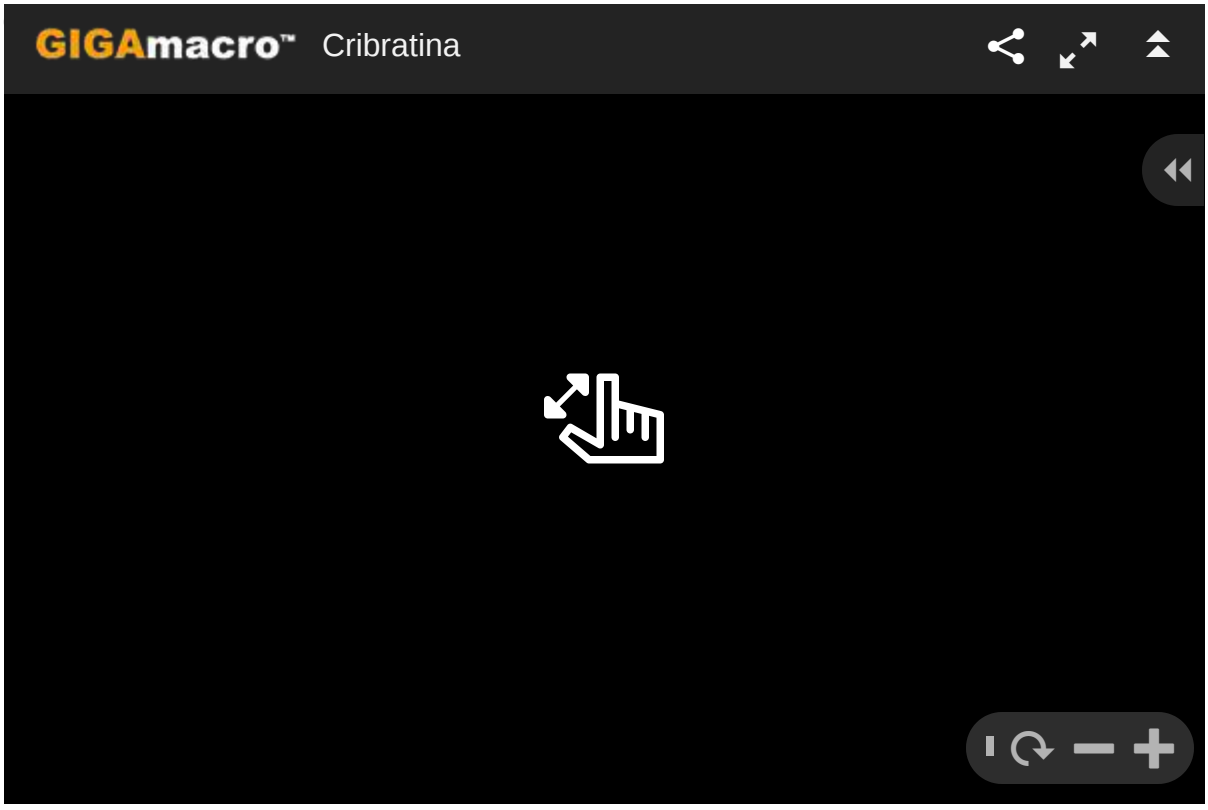
Chesapecten jeffersonius, scallop (bivalve), Pliocene:



Schwagerina, fusulinid foraminiferid, Permian:



Cribratina, textularid foramaminiferid, Cretaceous (Albian):



Brachiopods

Mucrospirifer, articulate brachiopod, Devonian:



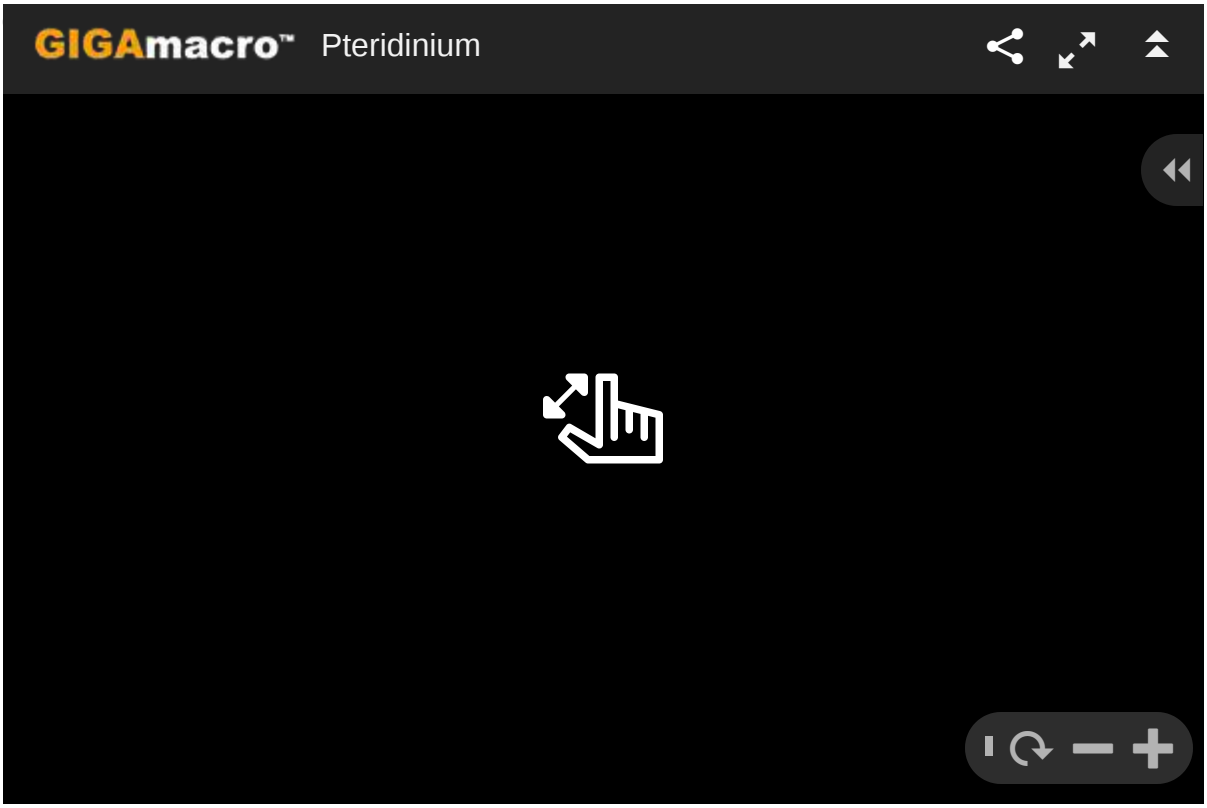
Vertebrates

Knightia, fish (vertebrate), Eocene:



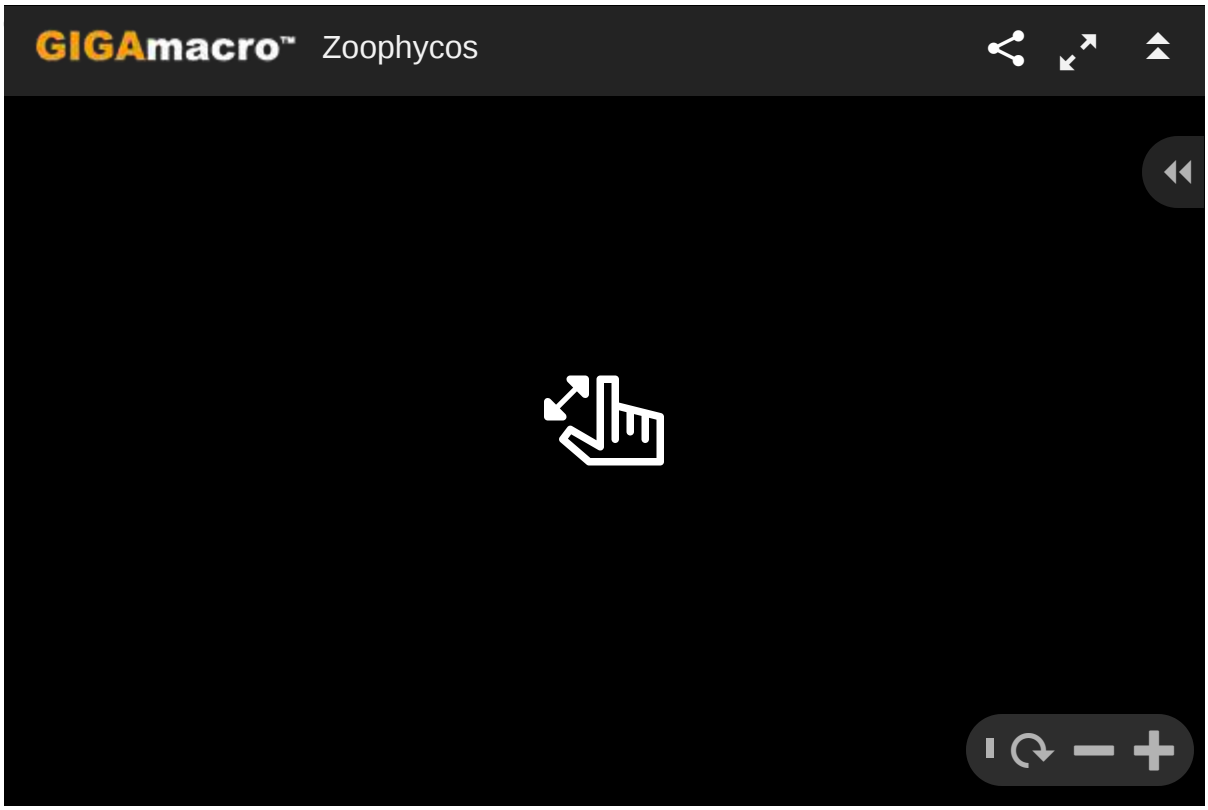
Ediacaran fauna

Pteridinium, unknown group (sea pen?), Precambrian (Ediacaran):

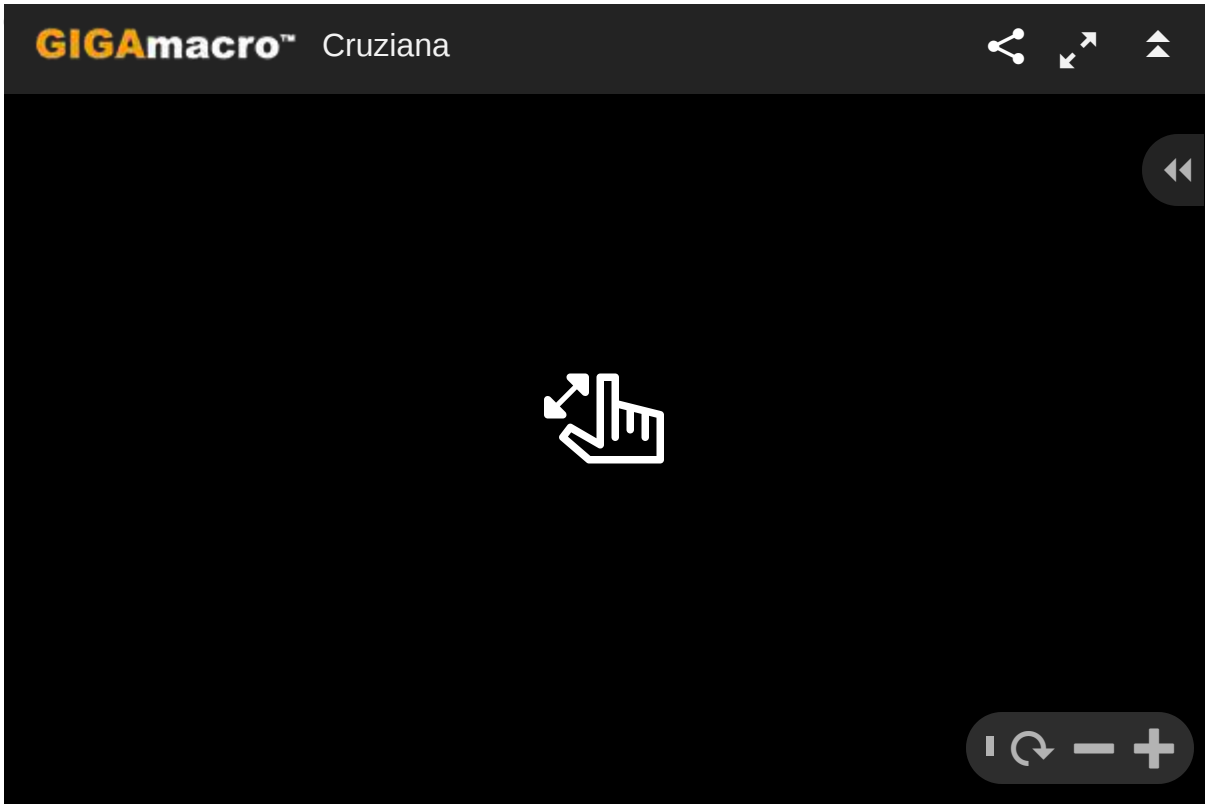


Trace fossils

Zoophycos, trace fossil (feeding trace of a worm?), Paleozoic:



Cruziana, trace fossil (trilobite crawling trace), Cambrian-Permian (Paleozoic):



Skolithos, trace fossil (vertical “worm” burrow), Cambrian-modern:



Okay, now that you have learned the fossils, **it’s time for Part 2**, where you will apply your knowledge to a stratigraphic cross-section of a hypothetical region…

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62.2: Fossil identification and use in correlation, part 2

Fake Field Trip: Fossil correlation exercise

Now that you've learned to recognize the various fossils in this VFE in part 1, it's time to apply your new knowledge to a hypothetical region.

Part 2: On the basis of their fossil content, **figure out the geologic ages of the strata** in this cross section. The cross section shows the position and relation of the various geologic units, "A" through "H," as well as the unconformities that separate some of them ("1" through "4"). Links appear below the cross section that will allow you to examine fossils from each geologic unit.

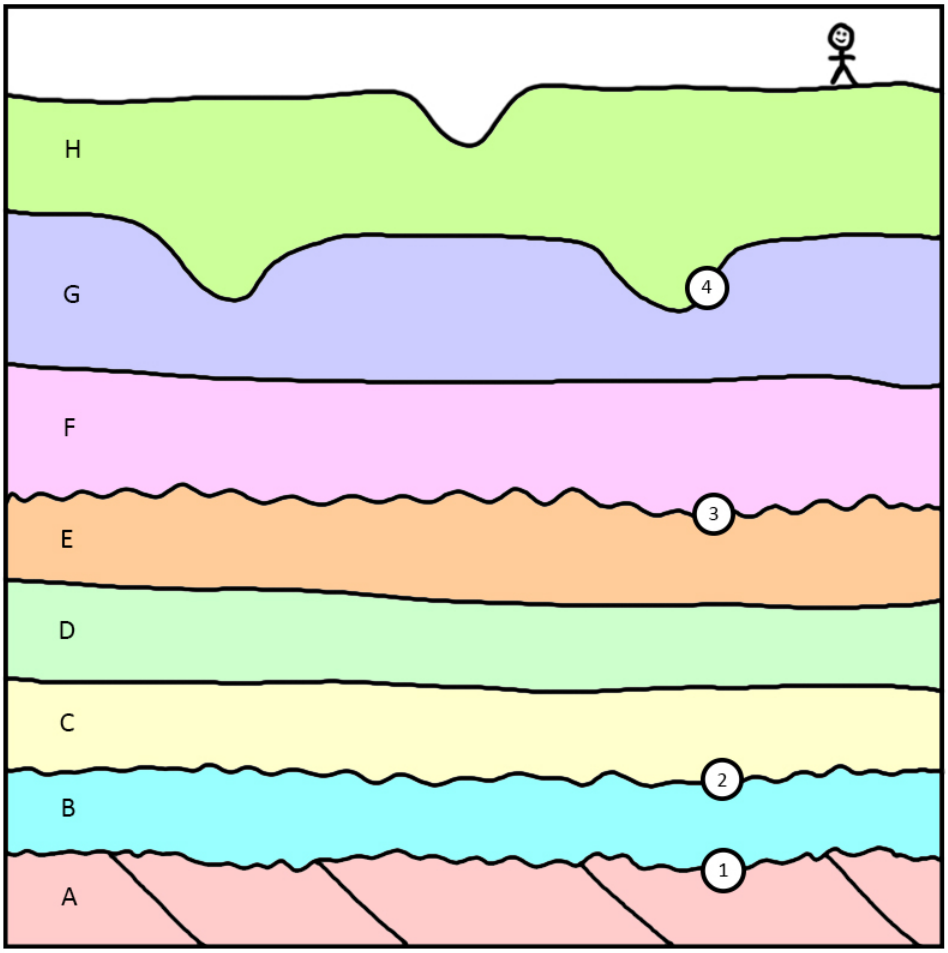


Figure 62.2.1: Cartoon geological cross-section through a hypothetical region.

Exercise 62.2.1: Unit A fossils

Two hand samples of the one kind of fossil found in **Unit A**:


GIGAmacro™ Fossil example 1 from Unit A


GIGAmacro™ Fossil example 2 from Unit A

Age of Unit A: _____

Exercise 62.2.2: Unit B fossils

One sample of the one kind of fossil found in **Unit B**:



Age of Unit B: _____

Exercise 62.2.3: Unit C fossils

One sample of the one kind of fossil found in **Unit C**:

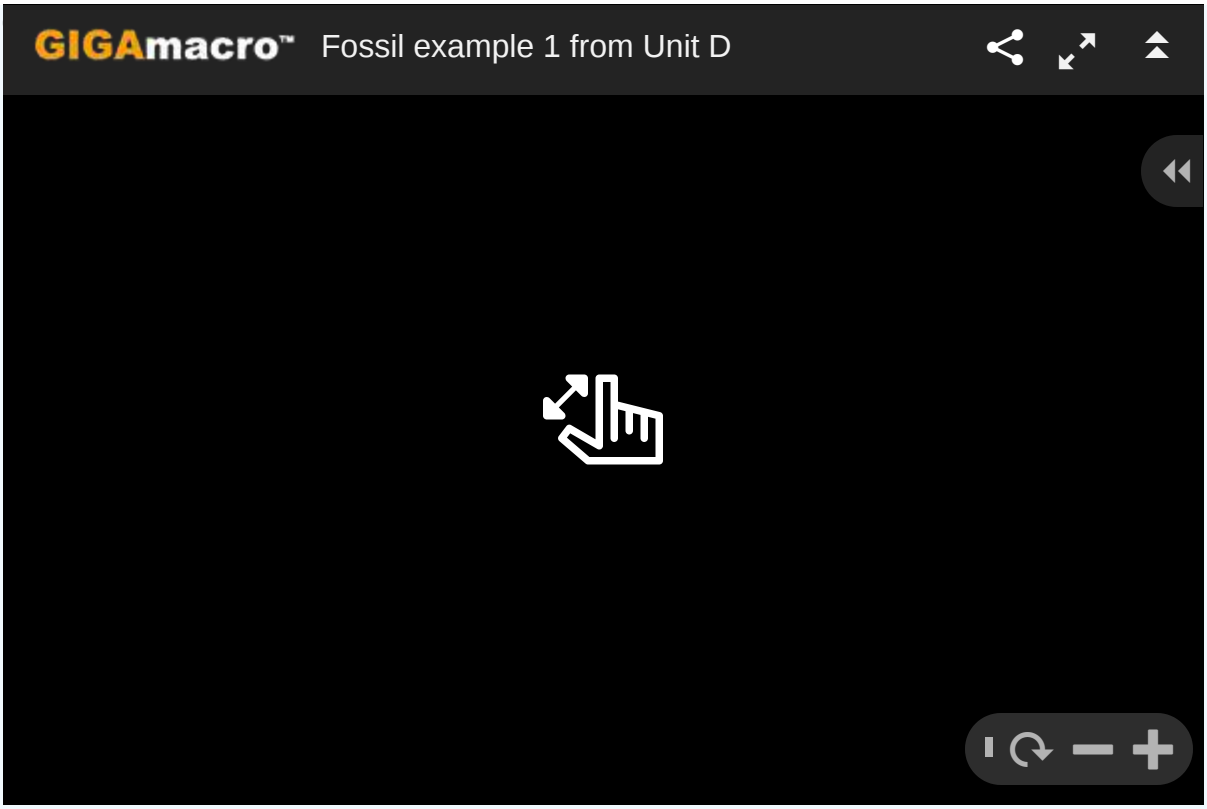


Age of Unit C: _____

? Exercise 62.2.4: Unit D fossils

Three total fossils found in **Unit D**.

First fossil:



Second fossil (two outcrops):





Third fossil (one outcrop and one sample):

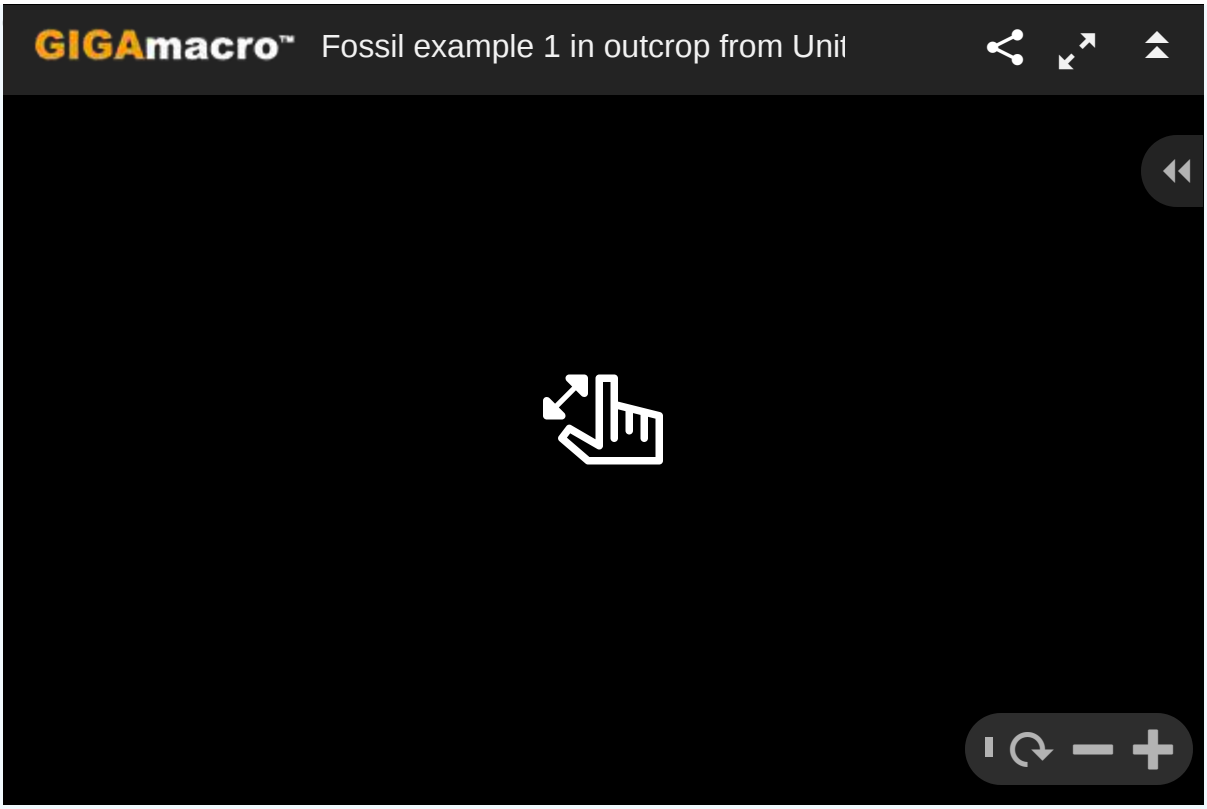




Age of Unit D: _____

? Exercise 62.2.5: Unit E fossils

One useful fossil found in **Unit E** (one outcrop and one sample of fossil-rich sediment from the site of the outcrop):


GIGAmacro™ Fossil example 1 in outcrop from Unit


GIGAmacro™ Fossil example 1 in sediment sample

Age of Unit E: _____

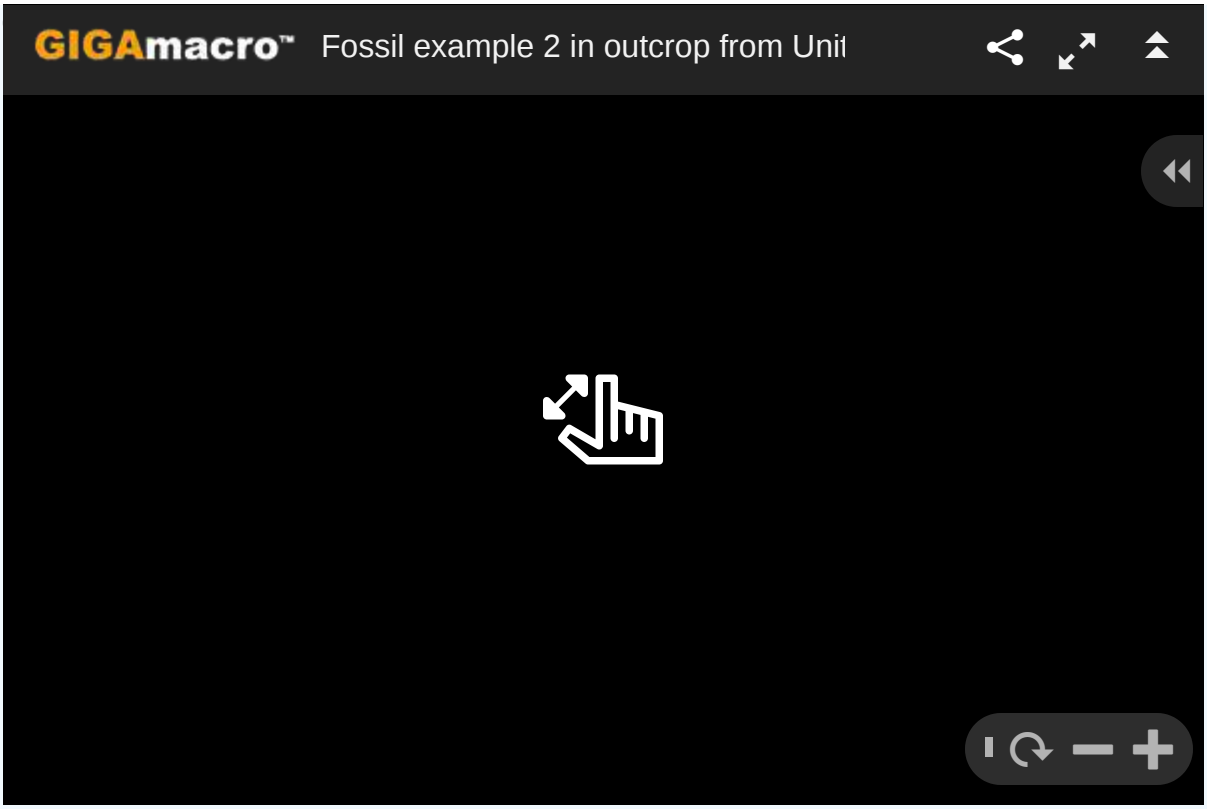
Exercise 62.2.6: Unit F fossils

Three fossils found in **Unit F**.

First fossil:



Second fossil:



Third fossil:



Age of Unit F: _____

Exercise 62.2.7: Unit G fossils

Two total fossils found in **Unit G**.

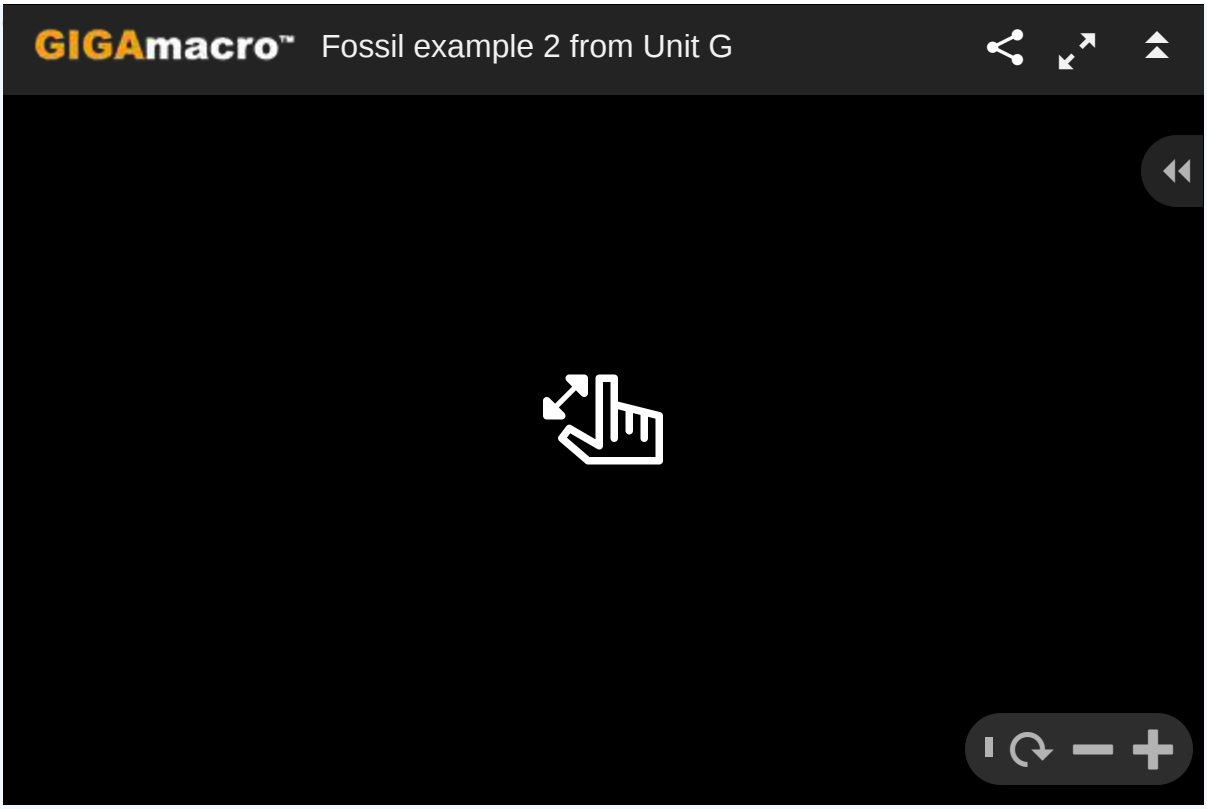
First fossil (three specimens):







Second fossil (one sample):



Age of Unit G: _____

Exercise 62.2.8: Unit H fossils

Here are two outcrops of **Unit H**. The view should load with the most useful fossil at the center of the view.





Age of Unit H: _____

How much time is missing along unconformity “1?” _____

How much time is missing along unconformity “2?” _____

How much time is missing along unconformity “3?” _____

How much time is missing along unconformity “4?” _____

Back to “Part 1” fossil guide

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62.3: Glaciers, glaciated landscapes, and glacial sediments

Glaciers, glacial sediments, and glacial geomorphology VFE

GigaPans

These GigaPans are the basis for the questions in the first part of the lab:

Exercise 62.3.1

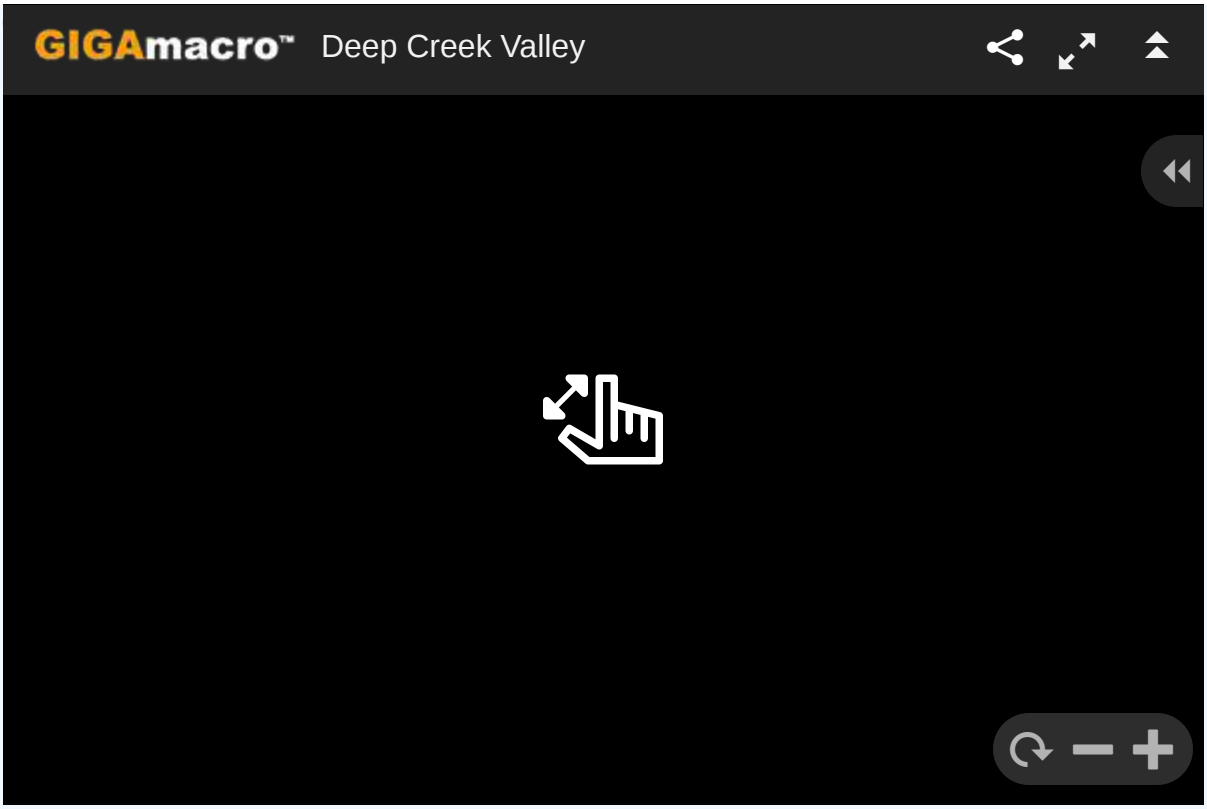
What type of glacier is this?



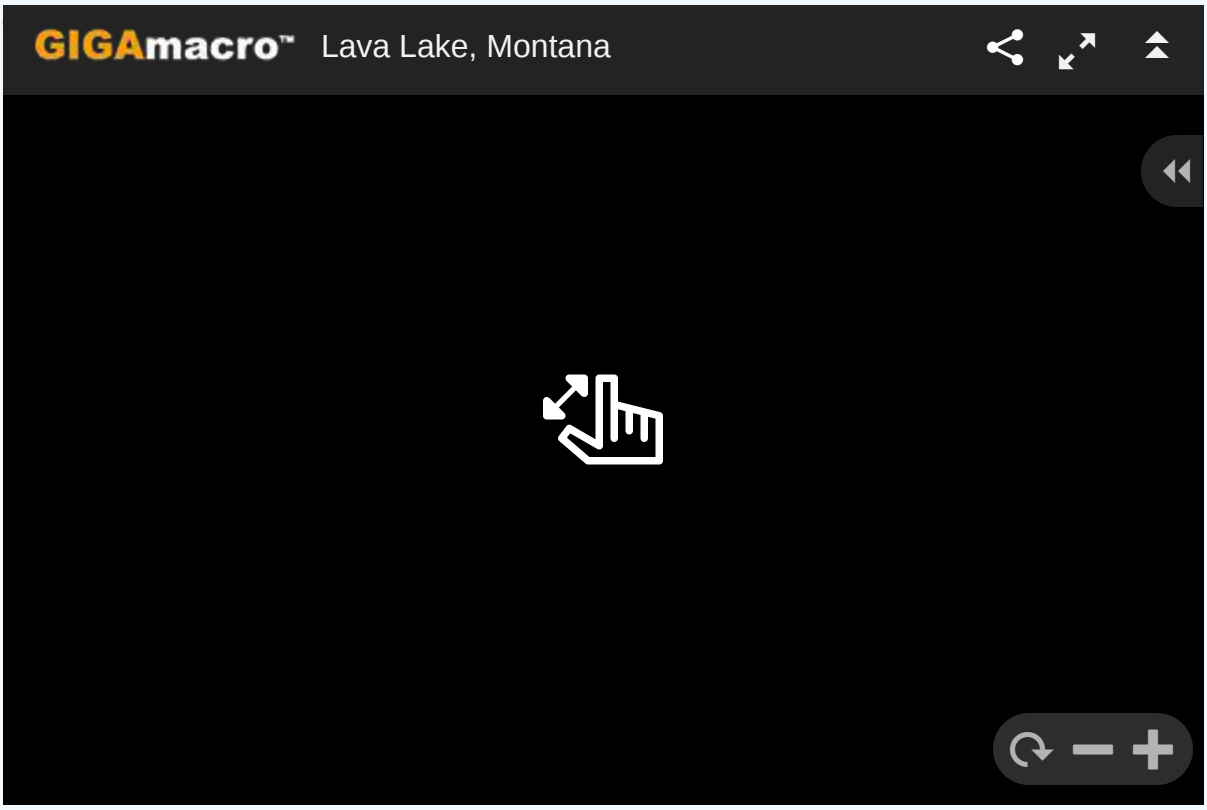
Exercise 62.3.2

Here are two valleys: Which one is glacial in origin? Why?

Valley X

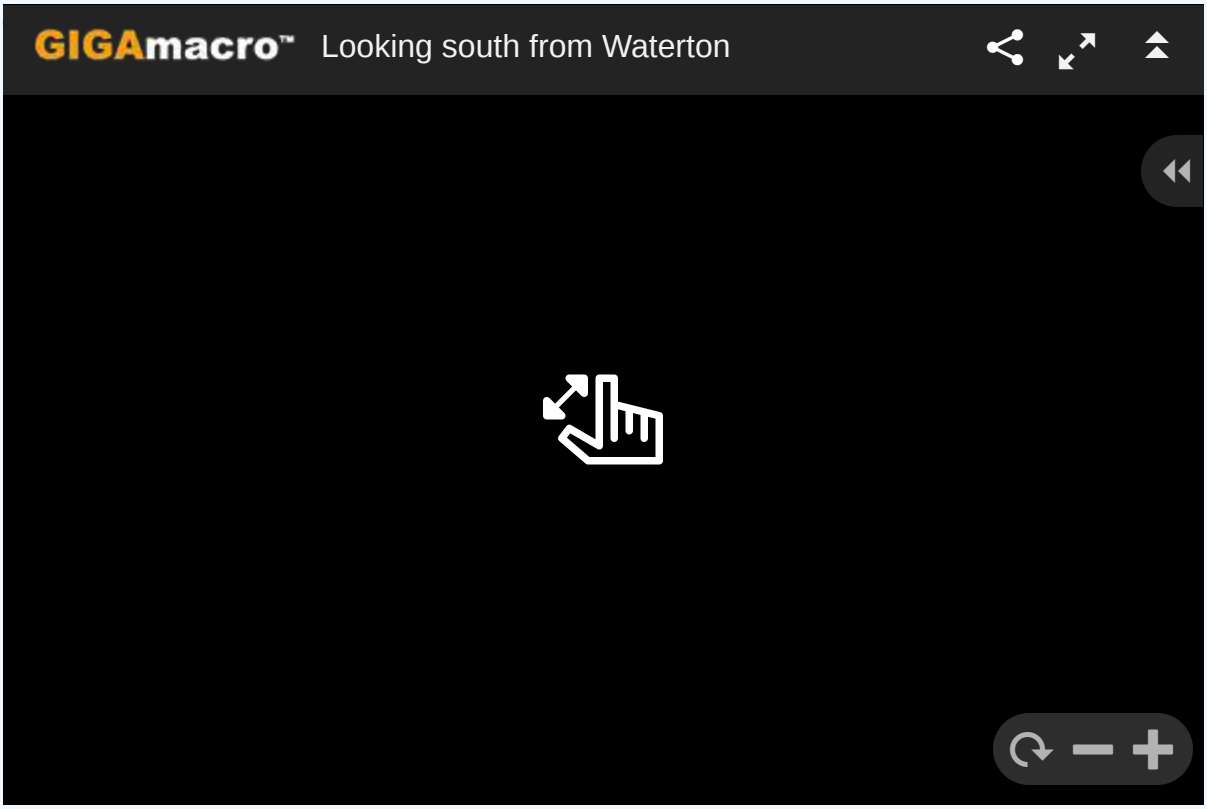


Valley Y



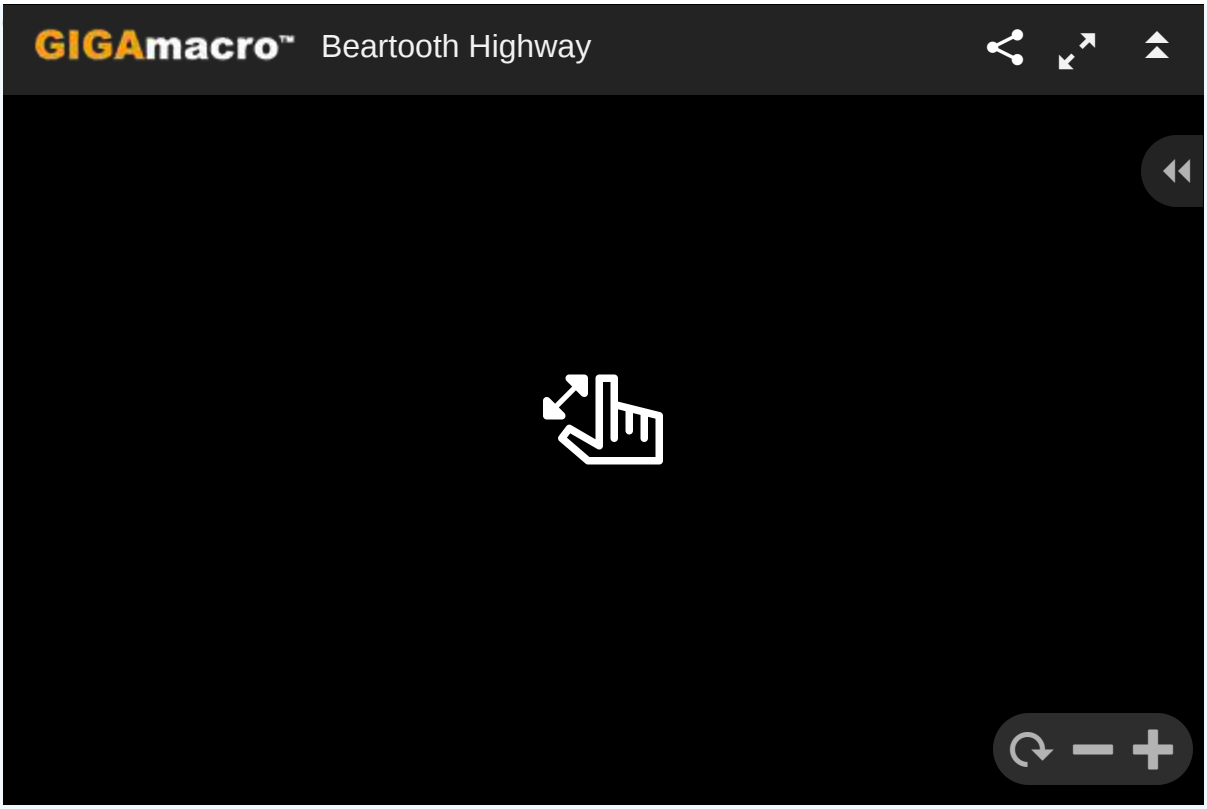
Exercise 62.3.3

What kind of lake is this?



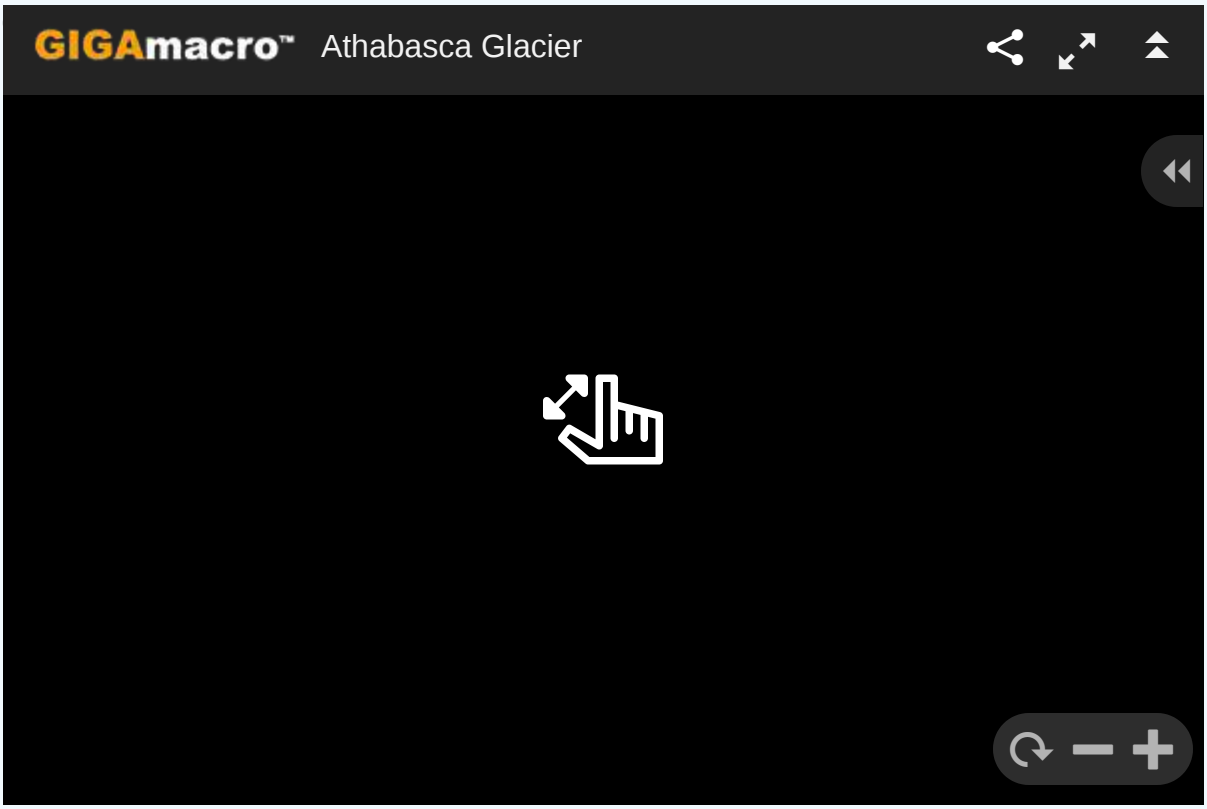
Exercise 62.3.4

Examine the annotations on this image. **First (yellow) annotation:** What kind of lake is this? **Second annotation (red):** What is the name for this landform? **Third annotation (blue):** Is this valley glacial or river-carved in origin?



? Exercise 62.3.5

Examine the annotations on this image. **First (yellow) annotation:** What kind of landform is this? **Second and third (blue) annotations:** What kind of landforms are these?



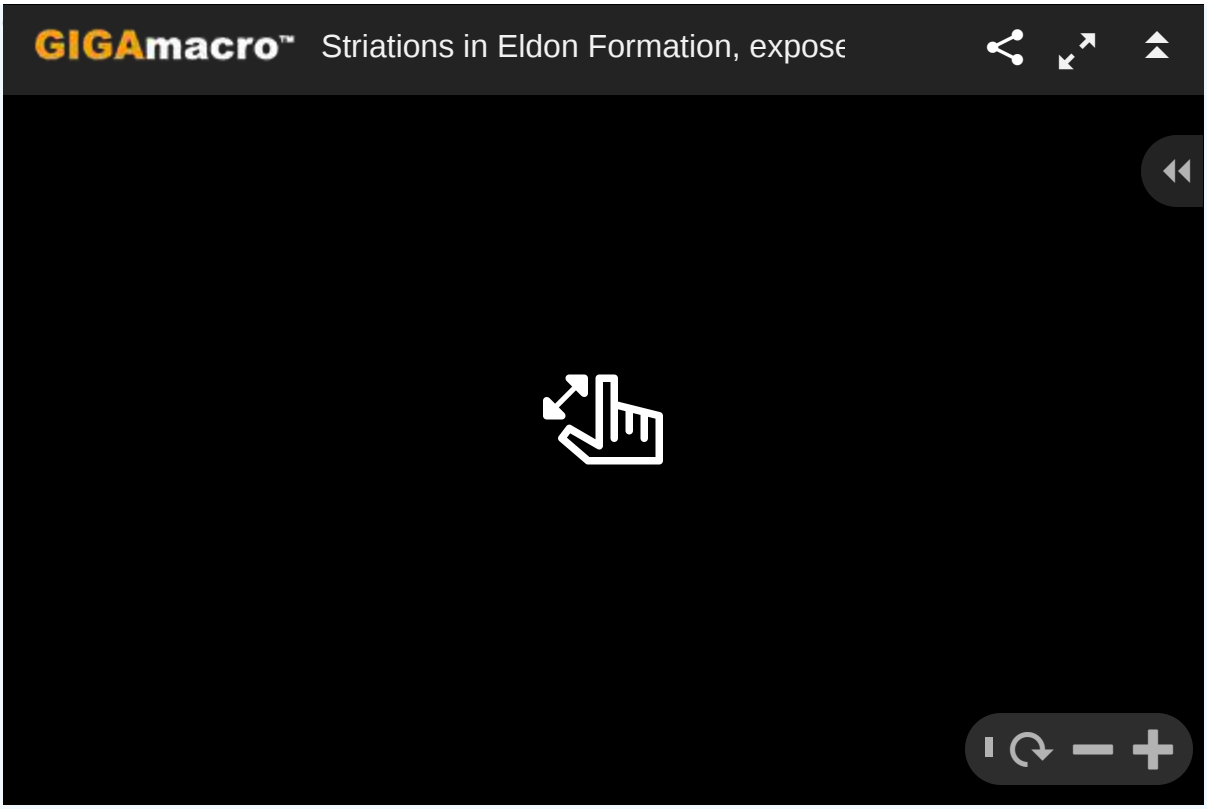
Exercise 62.3.6

Click on the annotations at lower left. **First annotation (yellow):** What feature can be seen here? **Second annotation (red):**What feature can be seen here?



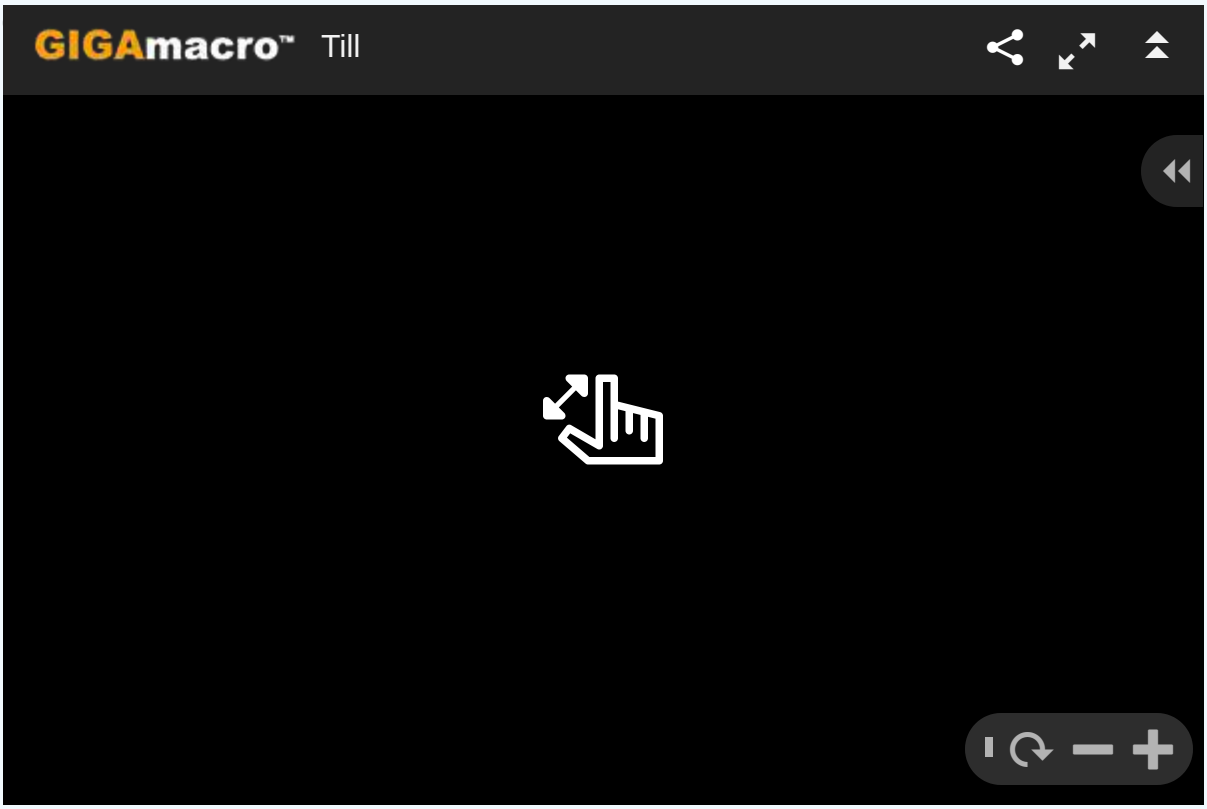
Exercise 62.3.7

This GigaPan was taken looking from north to south. Several features are visible, including sedimentary bedding (gray and tan), mineral veins (orange, perpendicular to bedding; and white, diagonal to bedding), and glacial striations. Identify the glacial striations and determine in which direction the glacier was flowing.



Exercise 62.3.8

This is a deposit of glacial till. Identify the key characteristics of this sedimentary deposit.



Exercise 62.3.9

Here are two sedimentary rocks, A and B: Which one is glacial in origin (lithified till, or “tillite”)? Why?

Sample A



Sample B



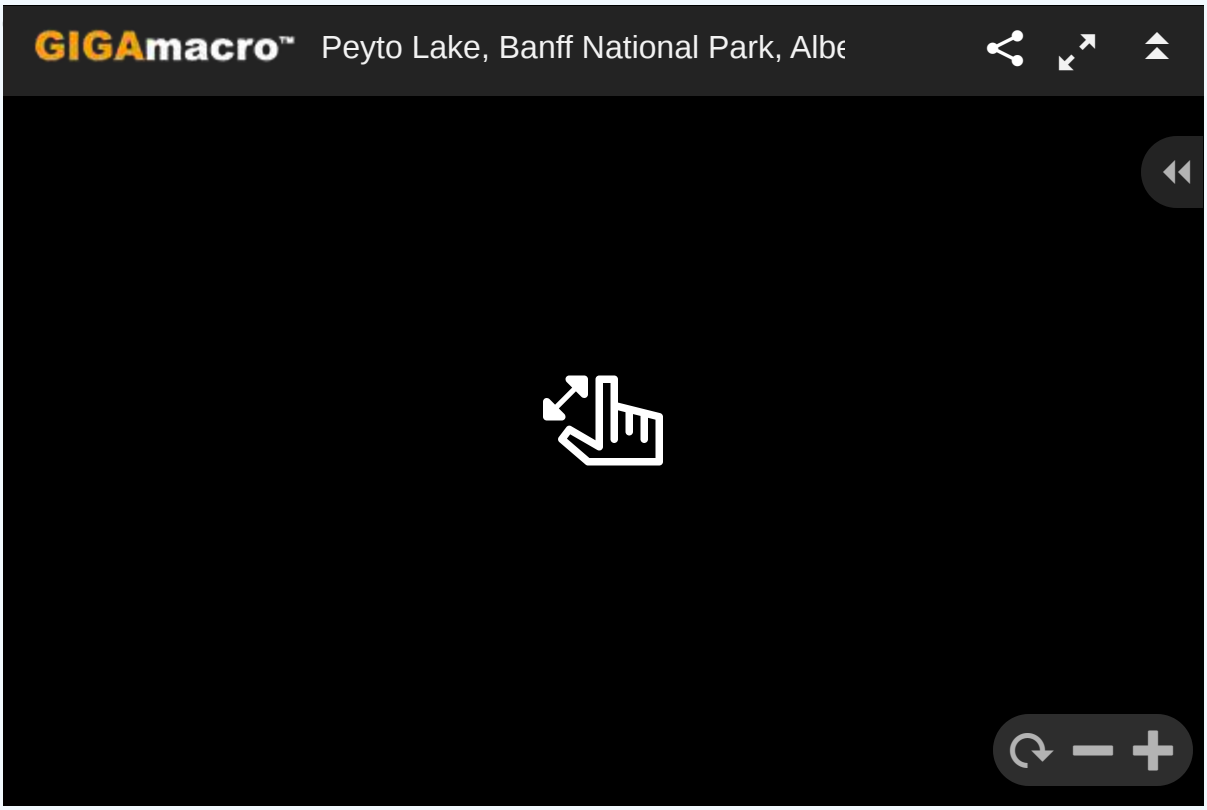
? Exercise 62.3.10

Click on the annotations at lower left. **First annotation (yellow):** What is the origin of this sediment? **Second annotation (red):** What is the origin of this sediment? **Third annotation (blue):** What is the origin of this sediment?



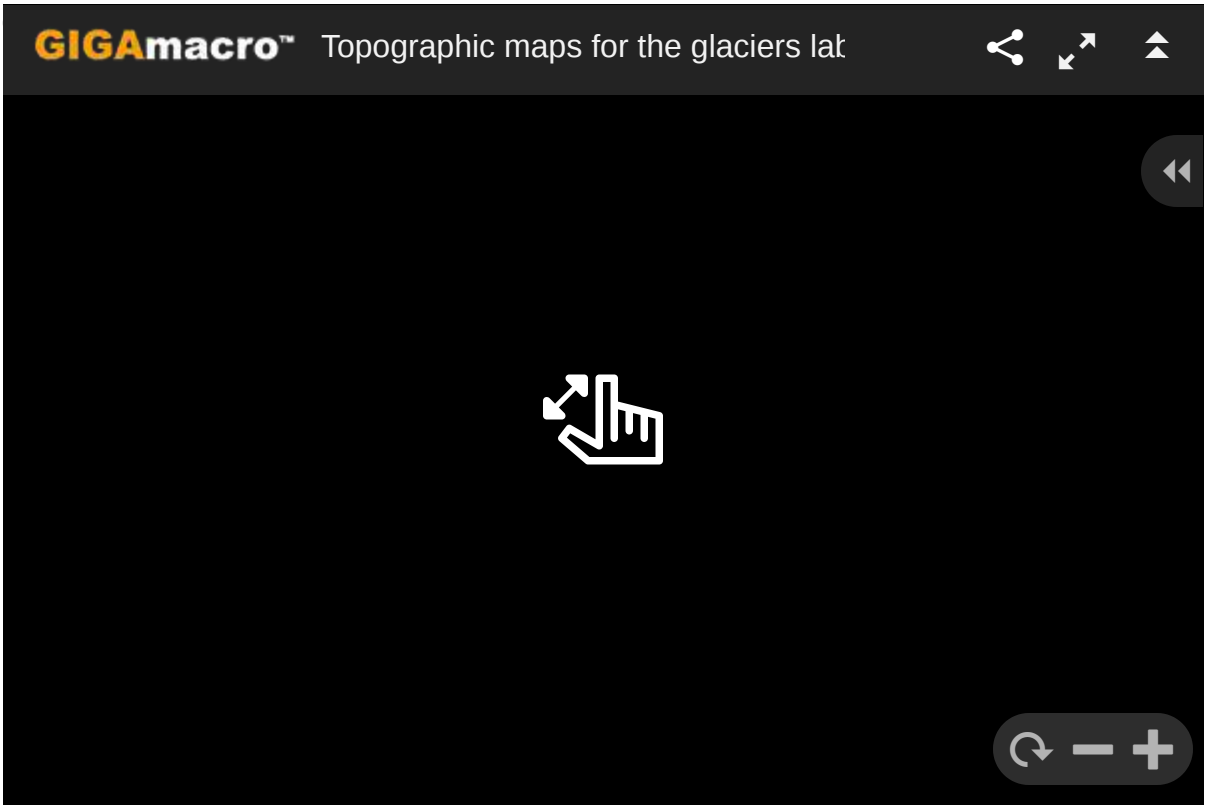
Exercise 62.3.11

Click on the annotations in this image. **First annotation (yellow):** What feature can be seen here? **Second annotation (red):** What feature can be seen here?

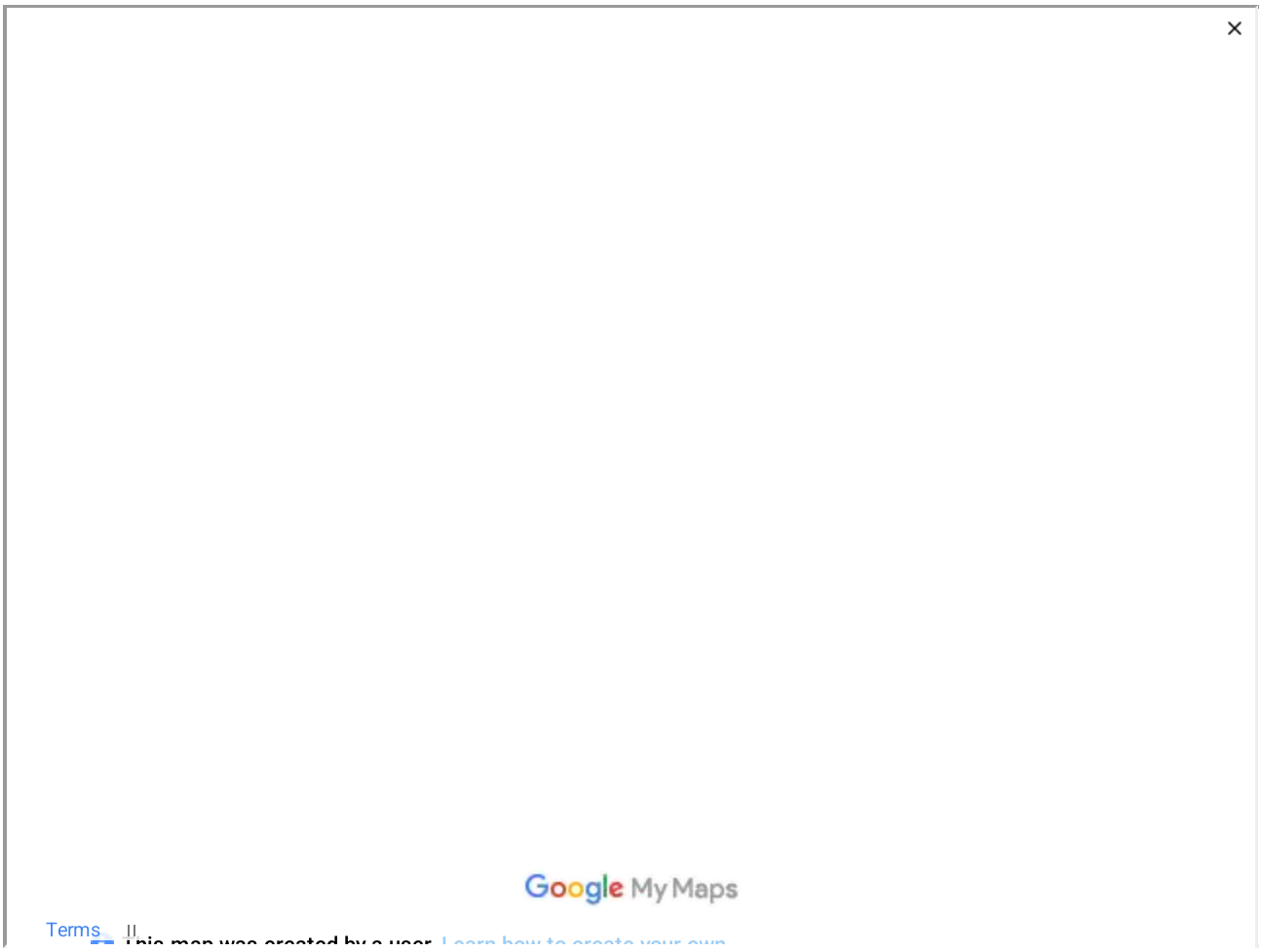


Topographic maps

This GigaPan shows the 5 maps you will need for the second part of the lab:

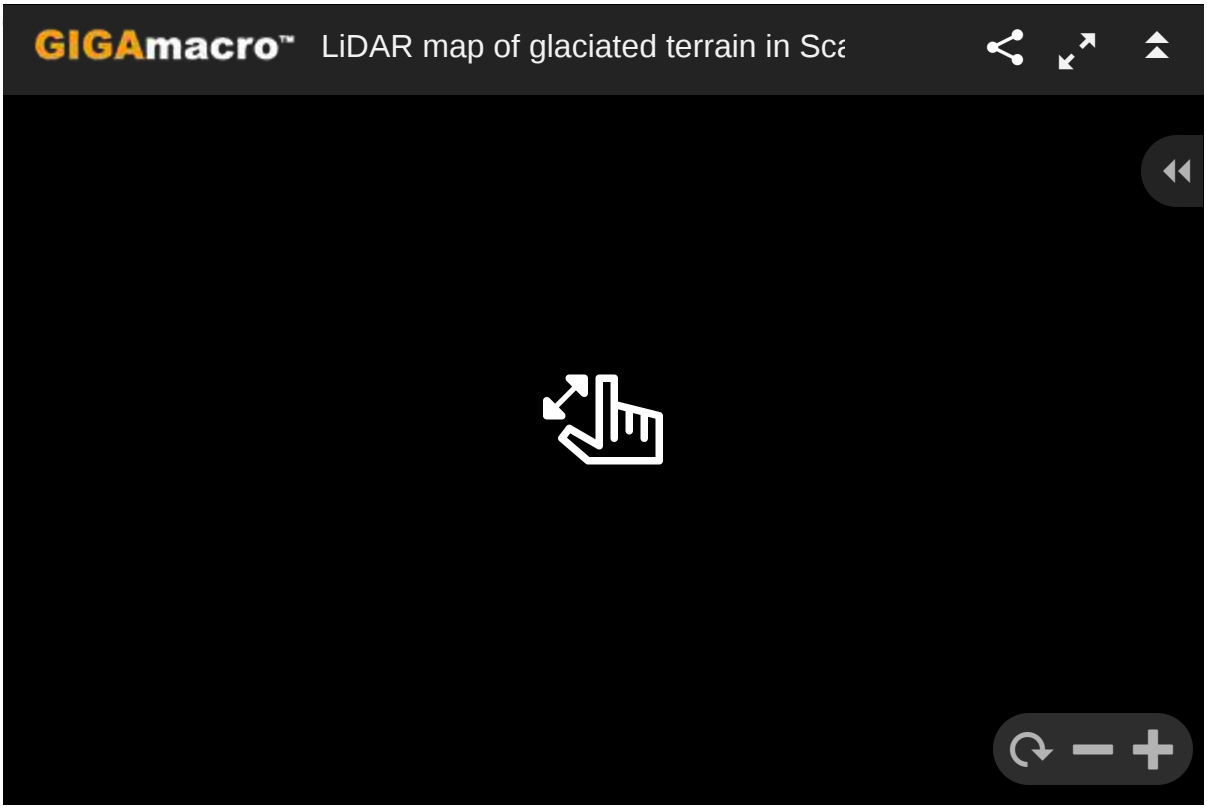


It may also be useful to examine these same sites in a seamless Google Map “terrain view” of the North American landscape (note the drop-down menu in the upper left corner):



Continental glaciation in LiDAR

The final set of questions for the lab deal with a LiDAR image of Scandanavia. This map shows several excellent examples of glacial geomorphology. LiDAR is a super-high-precision topographic mapping technique that uses lasers on aircraft to figure out exactly how high the land surface is. This LiDAR has been color-coded to show elevation: blue is low (but not sea-level); and the colors grade upward from green to yellow to orange to white (highest).



Jouko Vanne

Exercise 62.3.12

Identify one of each of the following (take a screenshot of an example of each, and submit those 5 images with your lab, labeling key features of the image):

- glacial striations
- terminal moraine
- outwash delta
- esker
- kettles

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62.4: Glacier National Park, Montana

Overview

Glacier National Park, Montana, is a superlative place, full of scenic grandeur, inspiring wildlife, and challenging recreational opportunities. Its rocks, structures, and landscape features tell an epic geological story that spans one and a half billion years of time. Going a bit further afield into the surrounding area gives more context to Glacier's story, so this virtual field experience (VFE) also includes visits to sites on the adjacent lands of the Blackfeet Nation as well as Swift Dam and Sun River Canyon, further south in the Sawtooth Range, and Little Prickly Pear Canyon, north of Helena.

The VFE examines the evidence for different modes of (1) sedimentation through the Mesoproterozoic, Mississippian, and Cretaceous, (2) a Mesozoic episode of mountain building, and (3) recent "Ice Age" glaciation. The cumulative result of all these processes is the park we can explore today. Each of the pages covering these distinct processes includes embedded self-quizzes, so students can assess their own understanding of key concepts as they read.

We have also prepared an informal "quiz" that instructors can use to review some of the relevant concepts with their students.

Enjoy exploring this fascinating place!

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62.5: Glacier National Park, Montana - Stratigraphy

The sedimentary sequence

There are three principle packages of sedimentary bedrock that we will encounter in our virtual field experience. From oldest to youngest, they are:

1. The Belt Supergroup: Mesoproterozoic-aged shallow-ocean sediments of mixed character

2. Mississippian carbonates: limestone and dolostone full of coral fossils, laid down during the late Paleozoic

3. Cretaceous Western Interior Seaway siliciclastic sediments: sand and mud deposited in a vast linear sea that stretched from the Gulf of Mexico to Alaska.

There are also surficial deposits of glacial sediment that we will devote some attention to at the end of our VFE. But first, the focus should be on the really old stuff; the bedrock.

Did I Get It? - Quiz

? Exercise 62.5.1

Which of these bedrock sedimentary packages in the Glacier National Park region is the **oldest?**

a. Siliciclastic strata of the Western Interior Seaway

b. the Belt Supergroup

c. Mississippian carbonates

Answer

b. the Belt Supergroup

The Belt Supergroup

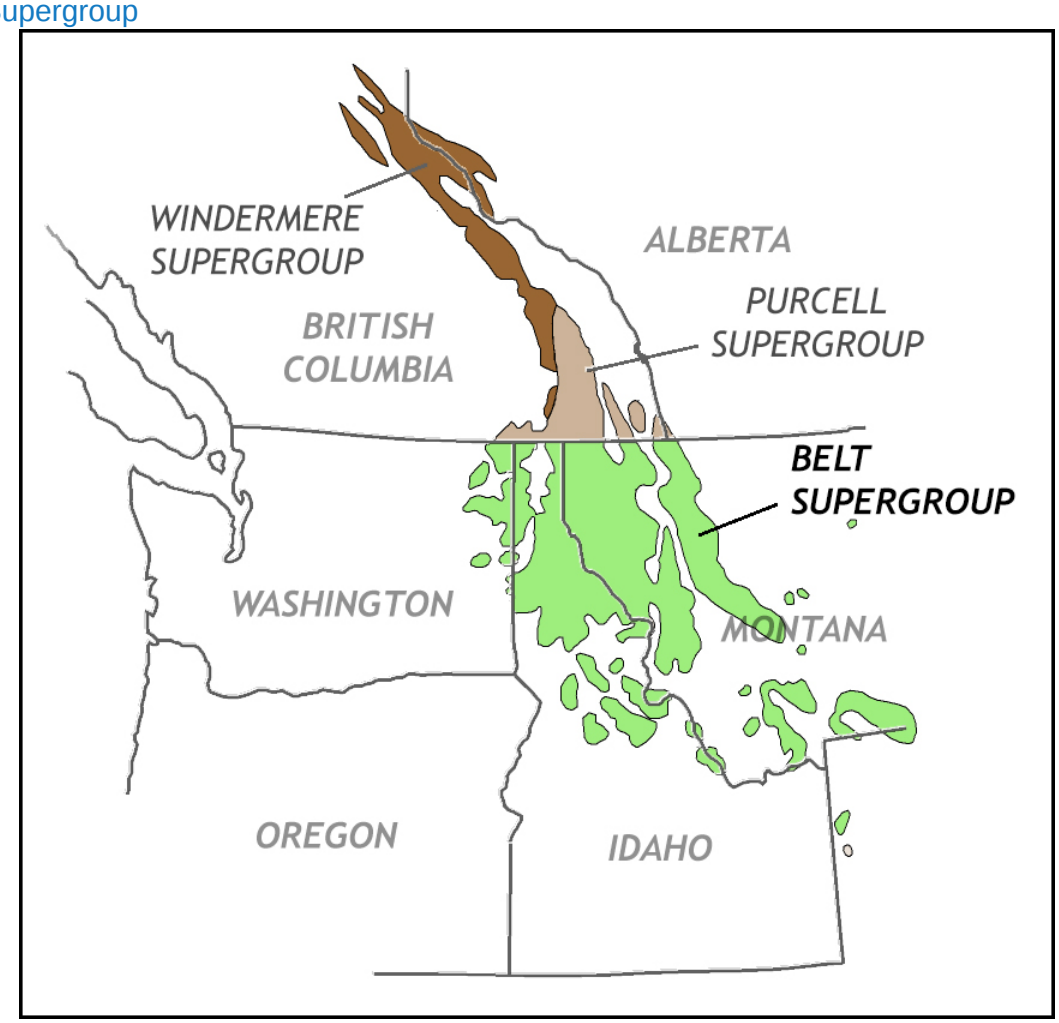


Figure 62.5.1: Map showing the distribution of the Belt Supergroup and related Mesoproterozoic rocks in Canada. (Modified from an original by Tapanila and Link (n.d.).)

The Belt Supergroup is a collection of related sedimentary groups, which are each collections of sedimentary formations. For such a voluminous, inclusive definition, you might expect the Belt to be substantial in volume. You would be right: the whole supergroup is more than *15 km thick*. That is 10 miles from top to bottom! It covers an area of some 200,000 km^2 (77,220 square miles), including most of western Montana, much of Idaho, and portions of neighboring Oregon, Washington, Wyoming, as well as British Columbia and Alberta in Canada. It is named "Belt" for the place it was first described, the Big Belt Mountains in Montana, as well as the adjacent Little Belt Mountains. (Note that in Canada, they call it the Purcell Supergroup, but it is the same thing. Further north from the Purcell outcrops are similarly aged rocks, dubbed the Windermere Supergroup.)

Sedimentary rock types

Belt strata come in many varieties, but most are sedimentary. There are deposits of gravel and grit, sand, and clay. There are carbonates too. All have been subjected to low levels of burial metamorphism. The resulting rocks are conglomerates, quartzites, and argillites, limestones and dolostones. You will notice that some of these are sedimentary names and some metamorphic names. This reflects the Belt's primary sedimentary nature, with a very light metamorphic overprint (greenschist facies). Belt quartzites are slightly metamorphosed quartz sand deposits; Belt argillites are slightly metamorphosed clay deposits. To further complicate things, some formations consist of one main lithology, while others show substantial internal variety.

There are also a few mafic volcanic units, and Belt rock hosts some impressive sills (of much younger age) as well.

Age and tectonic context

The strata of the Belt Supergroup were deposited in the Belt Sea, a Mesoproterozoic shallow sea that existed in roughly the same place (perhaps a bit to the west of) where we find Belt rock exposed today. This sea formed as a consequence of rifting, probably of a block of continental crust that went on to be included in the heart of Siberia. We can make such a claim on the basis of sediment deposited in the Belt Sea. Along the margins of the Belt Sea, coarser sediment dominated, with finer-grained material making it out to the center of the basin. The Lahood Formation in Jefferson River Canyon near Three Forks (or in the northern Bridger Range, for instance in Corbly Gulch) is a coarse sandstone to matrix-supported conglomerate, containing chunks of rock from the Wyoming Craton basement complex. Further west, the sediment is also coarse, and can be traced to now-Siberian source outcrops. But since the Mesoproterozoic, Siberia has rifted away and has left the vast Pacific between it and its original position on the margins of ancestral North America.

The sedimentary sequence at Glacier National Park

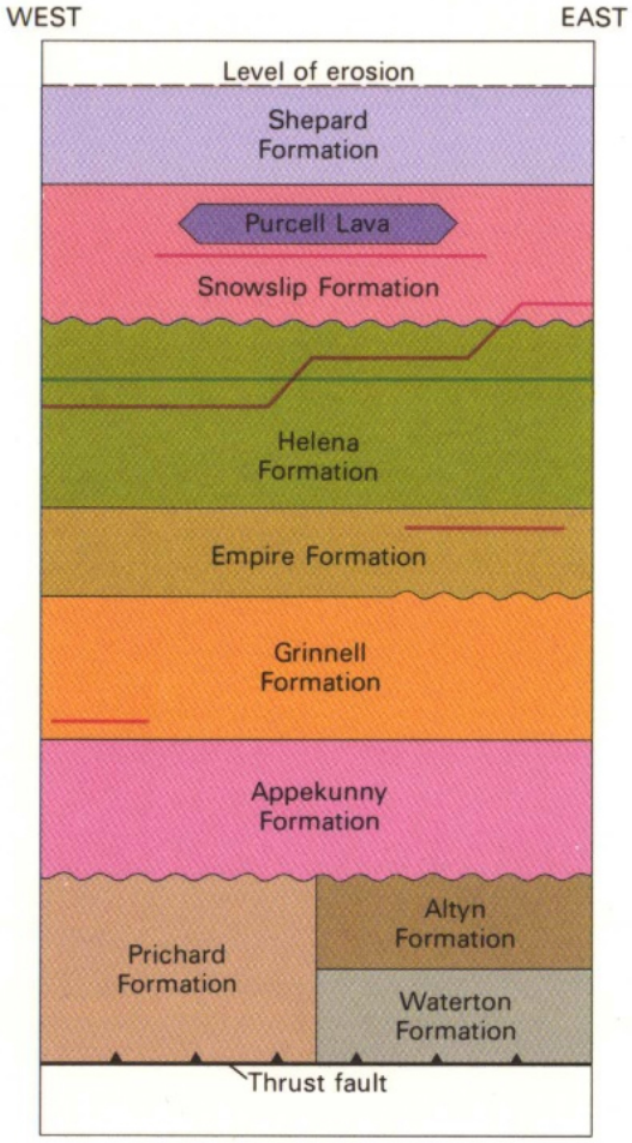


Figure 62.5.2: A stratigraphic column for Belt strata in the Logan Pass region of Glacier National Park. (Modified from Whipple (1992).)

At right is a stratigraphic column for the Belt rocks in Glacier. Diagrams like this one summarize the vertical stack of different sedimentary layers in a location.

Let us start at the bottom and work our way up, since chronologically that is the order in which nature formed these layers (*i.e.*, the principle of superposition). At the bottom is one of two formations: either the Pritchard Formation or the Altyn Formation. The Pritchard appears on the west side of the park, and the Altyn appears on the east side. The Pritchard is a mix of black and gray argillite with some white quartzite and some limestone, while the Altyn is a gleaming white dolostone containing only small amounts of quartz sand. In Waterton Lakes National Park in southwestern Alberta, another formation is found below the Altyn, the suitably named Waterton Formation. It is mainly a dolomite, with some sand and silt in it, to varying degrees, but it does not crop out especially well in Glacier. So we will focus here on the Altyn Formation, which makes up the prominent cliffs that rise suddenly from the plains east of the park. It is about 250 m thick, and very resistant to erosion.



Figure 62.5.3: Appekunny Formation green argillite with minor layers of white quartzite (weathers brown). (Photo by Maddy Rushing.)

Overlying the Altyn is the Appekunny Formation. The Appekunny is 530 to 690 m thick, mostly green argillite in thin beds. There are some quartzite layers too. The green results from the mineral chlorite, which is a low-grade metamorphic mineral and one of the few signatures that the Belt rocks at Glacier experienced burial metamorphism. In other words, this green color is not original; it is a diagenetic feature that resulted from deep burial of the sediments.

Above the Appekunny is the Grinnell Formation. The Grinnell Formation is mostly bright red argillite, but there is a lot of quartzite in it too – a higher proportion of quartz sand than we saw in the Appekunny. The argillite in the Grinnell is red because the iron in it is oxidized. The inference is that it was deposited under higher-oxygen-level conditions than was the Appekunny. Both formations' argillites have identical proportions of iron (less than 3% apiece) but in the Appekunny the iron is reduced, and in the Grinnell it is oxidized. The difference between them is not iron content, but oxygenation state. The Grinnell is 530 to 790 m thick.



Figure 62.5.4: Grinnell Formation (mostly red argillite) exposed in the side of Red Rock Canyon, Waterton Lakes National Park, Alberta. (Callan Bentley photo.)

The bright, bold colors of mint green Appekunny and crimson to burgundy Grinnell are striking aspects to the landscape of Glacier. Even if they do not understand what these colors mean (because they don't speak Rock), almost every park visitor will revel in their glorious color.

The thick package of Grinnell transitions to another thick package, the Helena Formation, via a thinner transitional formation called the Empire Formation. (It is only 125 to 150 m thick.) Some geologists feel that the Empire is not distinctive enough to warrant its own formation name, but it is a well-displayed unit on the Siyeh Pass hike, so it is included it here. The Helena Formation is mostly limestone, so the Empire shows a mix of red argillite and quartzite (typical of the Grinnell) and limestone (typical of the Helena).



Figure 62.5.5: Stromatolitic horizons shown in cross-section within the Helena Formation's Conophyton Zone, downhill of the threshold of the Grinnell Glacier cirque. (Callan Bentley photo.)

The Helena Formation proper is 750 m to 1 vertical km of bluish limestone that weathers to an appealing honey-colored tone. The Helena makes up much of the distinctive rocks in the central part of the park, including at Siyeh Pass itself and in the cirque of Grinnell Glacier. It features many extraordinarily well-developed stromatolites in a 30 m thick band called the *Conophyton* Zone.

This band is very well exposed in the area recently uncovered by the melting Grinnell Glacier, where the massive *Conophyton* Zone has resisted erosion and defines the threshold of the cirque.

In the central part of the park, the Helena Formation also hosts the intrusive diorite sill that makes a distinctive black stripe on many mountain peaks in the region. The diorite sill is not a primary feature of the Helena; it was intruded much later. The Helena's limestone strata are about 1400 Ma (1.4 Ga), and the sill likely dates to 750 Ma. The sill intruded as magma into the solid rock layers of the Helena Formation 650 million years after they were originally laid down as sediment.

The Helena is overlain by the Snowslip Formation, which is 360 to 603 m of mostly siliciclastic strata: quartzites, siltstones, and argillites of various colors, thicknesses, and characters. There are some thin stromatolitic limestones, too, particularly lower down near the contact with the underlying Helena. 95 m of the Snowslip are the basalt lava flows called the Purcell Lava.

The Shepard Formation is the youngest Belt layer exposed within immediate view of the Going-to-the-Sun Road in Glacier; it can only be seen at the highest peaks around Logan Pass. The top of the Shepard is not exposed because it has been removed by erosion, but of the portion that remains, we can say it is at least 400 m thick. It consists of dolostone, siltstone, and argillite, with some limestone and quartzite thrown in too. Like the Snowslip, the Shepard is a bit of a grab-bag of rock types.

There are a few more formations above the Shepard that crop out in high peaks in less accessible parts of the park (*see Whipple's 1992 map for all the details*), but we do not need to focus on those here.

Here is a video summary of the different formations exposed along the Going-to-the-Sun Road:



Did I Get It? - Quiz

? Exercise 62.5.1

The Grinnell Formation overlies the _____ Formation, and is overlain by the _____ Formation.

a. Helena; Altyn

b. Altyn; Helena

c. Shepard; Snowslip

d. Appekunny; Empire

e. Snowslip; Appekunny

Answer

d. Appekunny; Empire

? Exercise 62.5.2

What is the significance of the color red in Belt strata at Glacier?

a. It is an indication that those layers were deposited under low oxygen conditions.

b. It is an indication of carbonate sedimentation, not siliciclastic sedimentation.

c. It is an indication that those layers were deposited under high oxygen conditions.

d. It reveals the influence of active glaciation at this spot continuously over more than a billion and a half years.

Answer

c. It is an indication that those layers were deposited under high oxygen conditions.

Primary sedimentary structures

Besides bedding, Belt strata show a variety of primary sedimentary structures. Primary sedimentary structures are patterns in the sediment that form at the time of deposition, revealing information about depositional conditions at Earth's surface in the past. *(They should not be confused with deformational structures, which are imposed much later due to tectonic stresses.)* In Glacier National Park, most of the Belt strata's primary sedimentary structures are consistent with a shallow basin.

They include:

RIPPLE MARKS

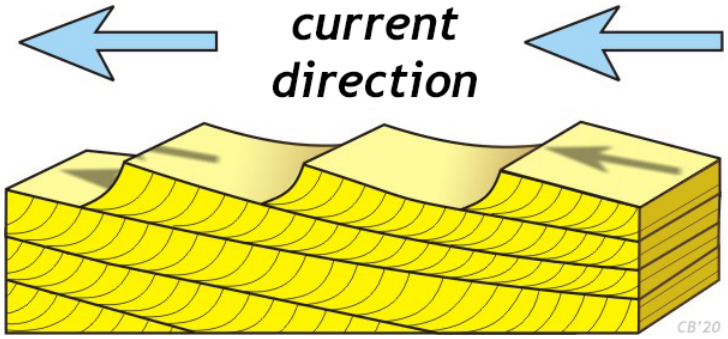


Figure 62.5.6: Cross-bedding and ripples are part of the same depositional setting, forming in a flowing directional current. (Callan Bentley cartoon.)

Ripples, or ripple marks, come in two principal varieties: Asymmetric ripples form in a directional current, and symmetrical ripples which form in oscillation (wave dominated) settings. In the Belt supergroup, there are both types, but asymmetric ripples dominate. These are the ripples associated with cross-bedding. How are ripples related to cross-bedding? The ripple creates a difference in water energy upstream and downstream of it. It is an obstacle to water flow. This means that the downstream side of each ripple is slightly calmer than the upstream side. The upstream side might experience erosion due to its higher-energy conditions, and the leeward side can receive that liberated sediment in a setting that is calm enough for it to settle out. The deposited sediment accumulates as a drape over the ripple, but only the concave-up "smiley face" portion of the drape is likely to survive.



Figure 62.5.7: Ripple marks in Grinnell Formation argillite/quartzite. (Photo by Shelley Jaye.)

In other words, ripple marks are the 3D expression of this single depositional bedform. Cross-bedding is the preservation of the "internal anatomy" of a migrating ripple.

Ripples are are found in many Belt formations in Glacier, but are most prominent in the Grinnell and Empire Formations.

CROSS-BEDDING



Figure 62.5.8: A cross-sectional view of cross-bedding in multiple quartz sandstone layers of the Empire Formation in Glacier National Park, Montana. (Callan Bentley photo.)

Cross bedding is a primary sedimentary structure that forms in a moving current of water or air. In the strata we observe on this virtual field experience, there are not windblown (aeolian) deposits, so all cross-beds you observe are ones that formed in flowing water. As the water moves in one direction over the sediment/water interface, ripples form. In many circumstances, the "line" of the ripple will form perpendicular to the water flow direction. The ripple creates a difference in water energy upstream and downstream of it. It is an obstacle to water flow. This means that the downstream side of each ripple is slightly calmer than the upstream side.

The upstream side might experience erosion due to its higher-energy conditions, and the leeward side can receive that liberated sediment in a setting that is calm enough for it to settle out. The deposited sediment accumulates as a drape over the ripple, but only the concave-up “smiley face” portion of the drape is likely to survive.

If two sets of cross-beds sit atop one another but “lean” in opposite directions, it creates a “herringbone” pattern, and should be interpreted as two different current flow directions in succession. Such bidirectional flow indicators suggest shallow water, and potentially tidal flux.

Examine this sample of Grinnell Formation quartzite + argillite to note the cross-bedding in the middle, the rippled upper surface, plus the presence of something else – chunks of red argillite surrounded by white quartzite. These are mud-chip rip-up clasts...



MUD CHIP RIP-UP CLASTS

When current strength is higher, clay particles remain entrained in the flow of water, and do not settle out, but because sand grains are larger (and heavier) they can settle out to be deposited. When current strength drops, clay can be deposited instead. So clay deposits (now argillite) represent relatively calm water conditions, and sand deposits (now quartzite) represent relatively energetic water conditions. But what happens when these are combined in sequence?

If energetic water precedes calm water, then you would expect a deposit of sand topped with a deposit of clay. If calm water precedes energetic water, then an initial deposit of clay may get scoured by the new, high-energy currents. Clay particles tend to stick together, while sand particles tend to disaggregate readily if blasted by high-energy water. The clay will be ripped up in chunks, which we call mud-chip rip-up clasts. These will be tumbled along in the current, potentially getting rounded as they travel. Any current strong enough to rip up mud chunks is also strong enough to transport sand, so there may also be sand grains tumbling along in the same flow. Finally, when the current ebbs, the sand and the ripped-up mud chunks will be deposited together.

Here is an example of such a deposit: showing red argillite at the bottom (high oxygen deposition), followed by green argillite (low oxygen deposition), followed by sand with rip-up clasts of green argillite. Zoom in on the rip up clasts to examine their shapes:



DESICCATION CRACKS (“MUD CRACKS”)

When wet, muddy sediment is exposed to the air, it can dry out. As water evaporates, there is less “stuff” in the mud deposit, and it no longer takes up quite as much volume. As a result, it shrinks. A flat deposit of mud shrinks across its upper surface, and the amount tension induced by the shrinkage is greater than the cohesive strength of the mud layer. The mud stiffens, and cracks. These cracks are widest at the top, where evaporation is most intense, and they narrow downward.

Examine this gallery to see what Glacier’s mud cracks have in common, and what distinguishes them:



Figure 62.5.9: Mucracks in white argillite, possibly of the Snowslip Formation. (CC licensed; Photo by James St. John.)

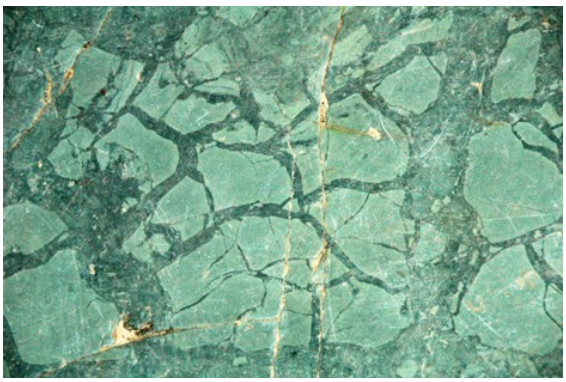


Figure 62.5.10: Mud cracks in green argillite, probably within the Appekunny Formation. (CC licensed; Photo by James St. John.)



Figure 62.5.11: Mud cracks in red argillite, probably the Grinnell Formation. Hiking pole tip for scale. (Photo by Shelley Jaye.)

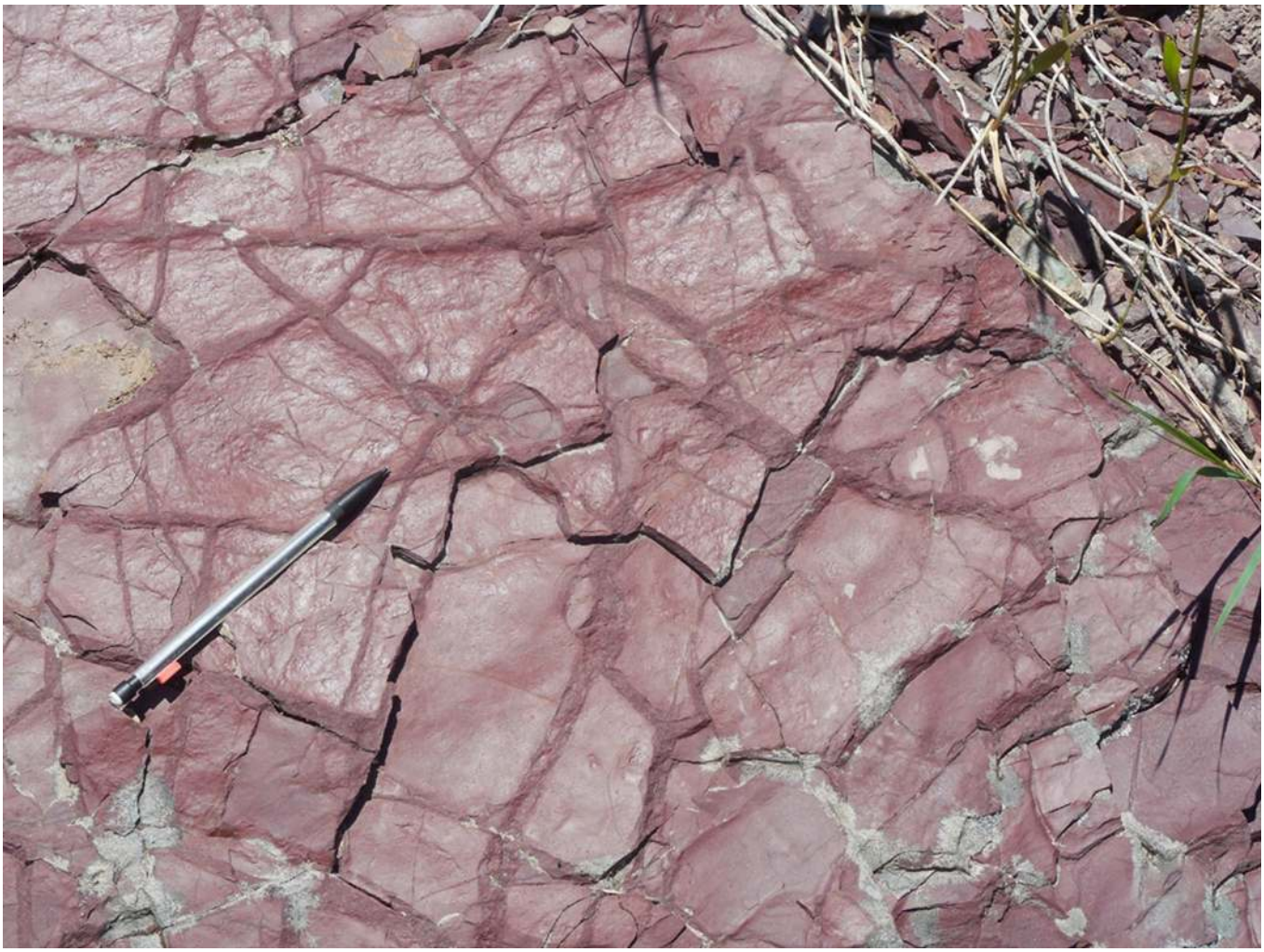


Figure 62.5.12: Mud cracks in red argillite of the Grinnell Formation.

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Figure 62.5.13: Mud cracks in green argillite of the Appekunny Formation.



Figure 62.5.14: Pleistocene mud cracks between sand dunes at Mesquite Dunes, Death Valley, California.

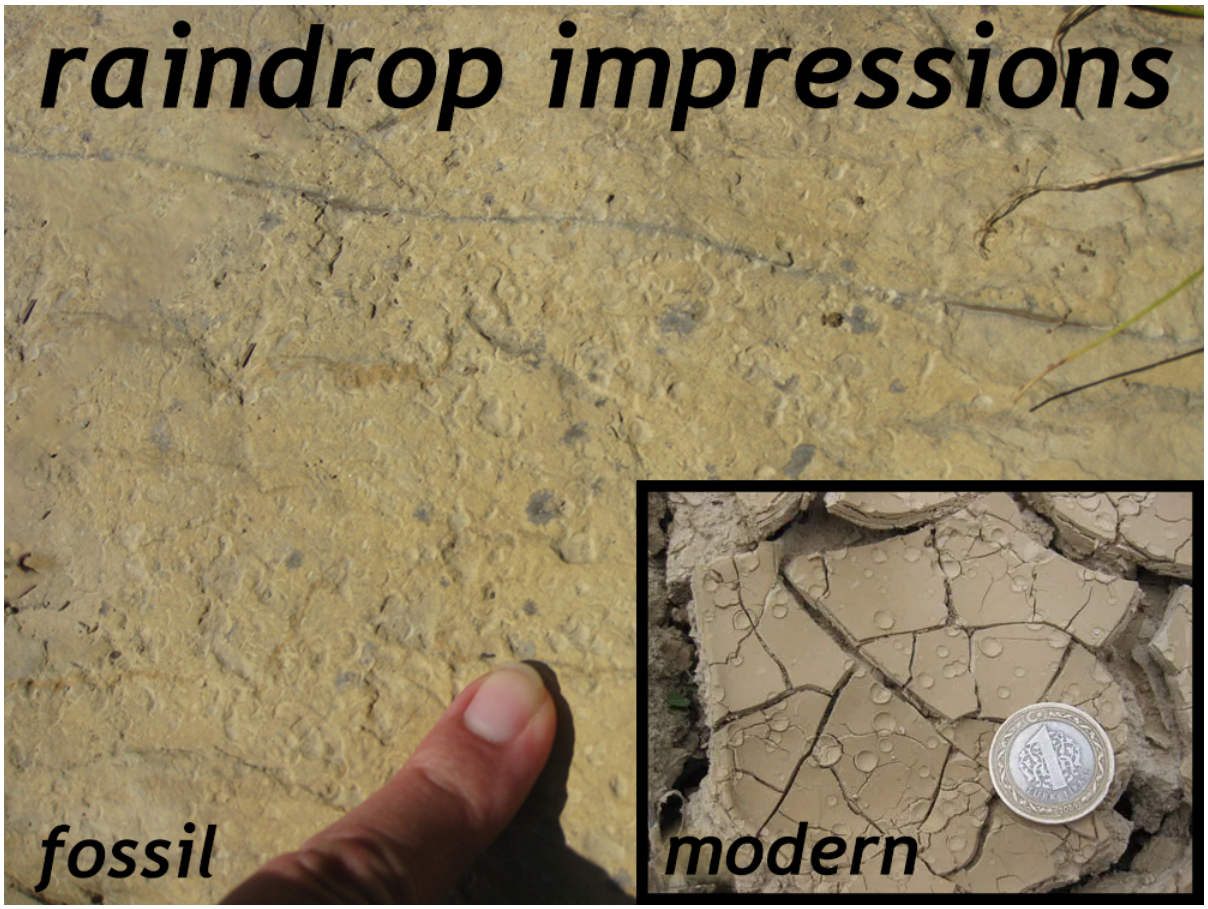


Figure 62.5.15: Raindrop impressions from the Proterozoic of Montana and a few years ago, among mud cracks in Turkey. (Large photo by Shelley Jaye; inset by Callan Bentley.)

When mud dries out in exposure to the air, it is also subject to another distinct, uncommon, but intriguing primary sedimentary structure: little tiny craters carved out when raindrops impact. Obviously, the rain must not have been especially hard or voluminous – just a little sprinkle, like virga trailing from a passing cloud. Those delicate little divots can be preserved in the geological record for us to marvel at, billions of years later. When you find raindrop impressions in the field, it is a special moment: we have all felt a few cold drops hitting our skin as the storm passes to the south, and here is the geological evidence that the same thing happened in that exact spot, an unimaginably long time ago.

HALITE CASTS



Figure 62.5.16: Halite casts can be preserved as 3D cube-shaped forms that protrude from the surface of the bed, as with this slab of Appekunny argillite. (Callan Bentley photo.)

When seawater dries out, it often gets saltier. The solutes in the seawater do not leave as the water does (through evaporation). When they become sufficiently concentrated, they will precipitate out. The mineral halite (NaCl; familiarly known as table salt) is one such mineral that records high evaporation conditions. The issue with salt, though, is that it is so soluble. It is not likely to persist through the geologic record in small amounts in wet sediments. However, we can still see the record of past salt crystals when they are "cast" in mud.

Here is how this works: The water dries out and gets salty. A salt crystal nucleates on the seafloor, growing in size as it adds Na+ and Cl- ions add on to the outside of the existing crystal structure. Given sufficient space, it will grow into the classic cubic habit that is typical of halite. The edges of the cube will grow faster than the center of each square face, and so some crystals may take on a "hopper" structure. Then the whole salt-encrusted layer gets buried by a new deposit of sediment: it could be clastic sediment (mud) or it could be carbonate. Either way, the salt crystals are now underneath a layer of fresh, wet sediment. The salt dissolves away in these wet conditions, and the sediment slumps inward, filling the salt-crystal-shaped hole, mimicking the shape of the original crystal with high fidelity. This "cast" has the shape of a salt crystal, but it would not taste good on popcorn.

Halite casts ("salt casts") can be found in several Belt formations at Glacier, including the Appekunny and the Helena.

Did I Get It? - Quiz

Exercise 62.5.1

Interpret this slab of rock.



a. It formed when photosynthetic microbial mats grew upward in shallow water.

b. It formed when the water got very salty.

c. It formed when a strong directional current moved sediment over the seafloor.

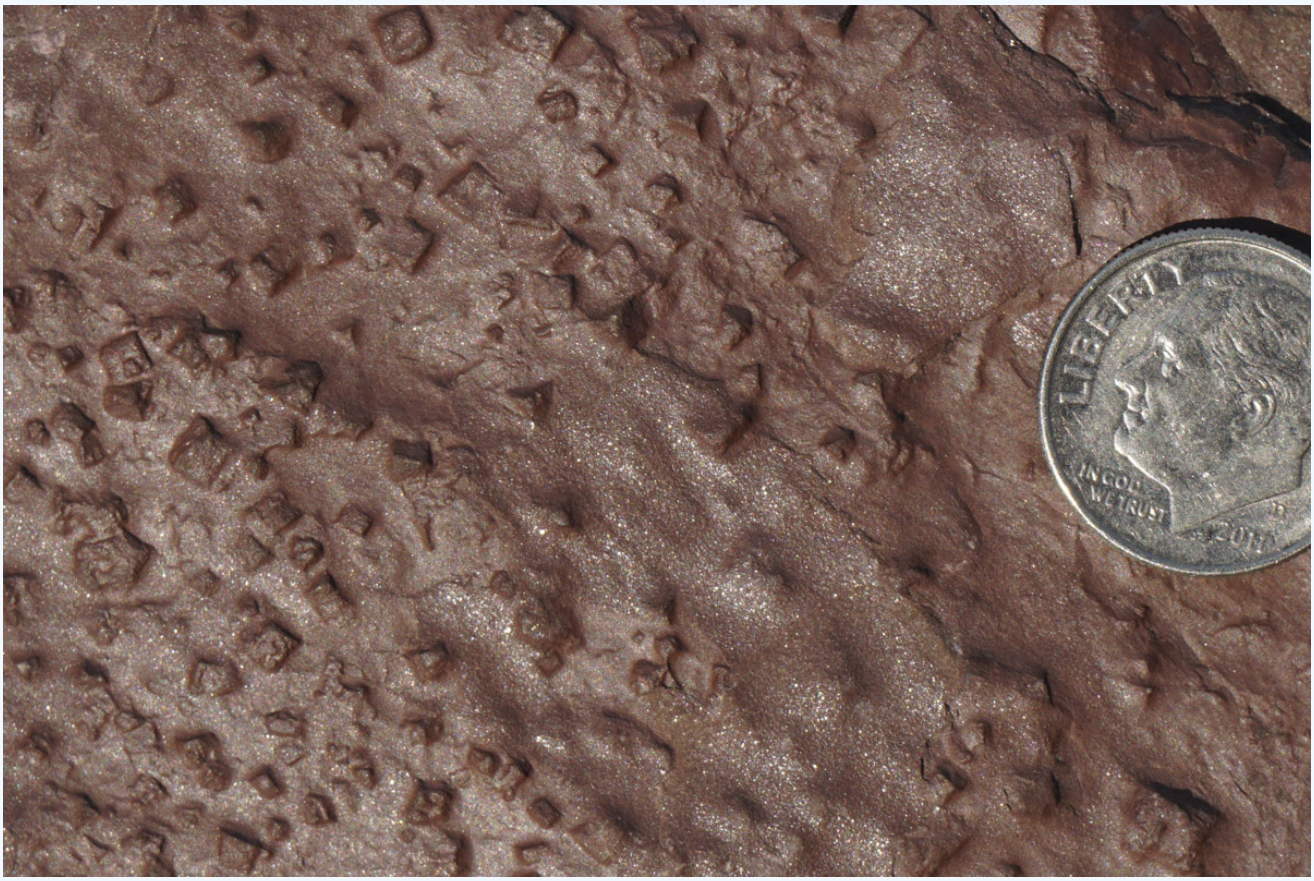
d. It formed when wet mud dried out in the open air.

Answer

d. It formed when wet mud dried out in the open air.

Exercise 62.5.2

Interpret this slab of rock.



a. It formed when the water got very salty.

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b. It formed when a turbid slurry of sediment and water slowed down and settled, with heavy grains settling out more rapidly than lightweight grains.

c. It formed when a strong directional current moved sediment over the seafloor.

d. It formed when photosynthetic microbial mats grew upward in shallow water.

Answer

a. It formed when the water got very salty.

OOIDS

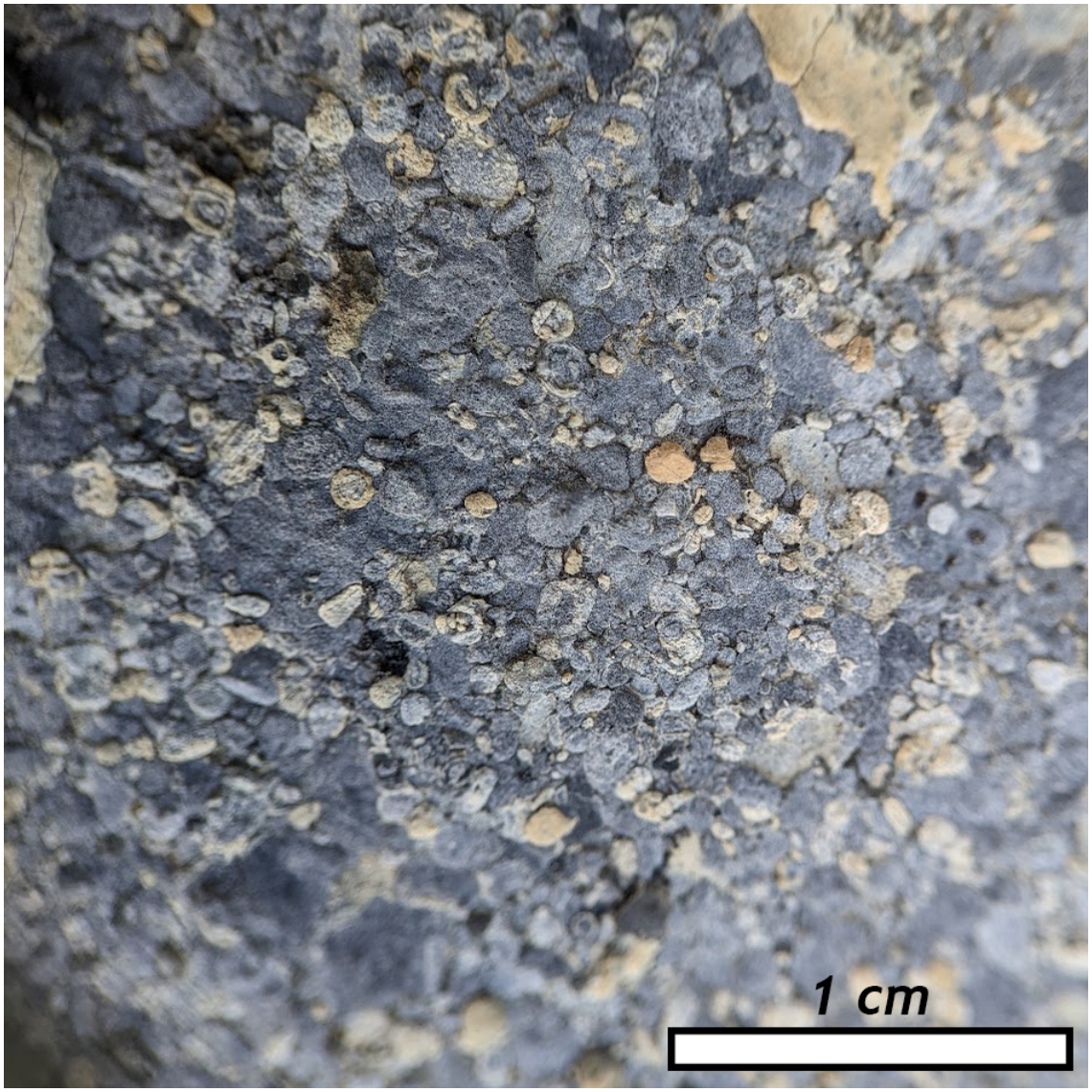


Figure 62.5.17: Ooids in Helena Formation, as viewed through a hand lens. (Callan Bentley photo)

Ooids are small spheres of calcite, deposited in thin layers that coat some small initial "nucleus" object. They are roughly sand sized, and internally layered. Their smooth, round outer appearance makes them look like fish eggs (think caviar or the orange stuff in sushi), which is where the name comes from (*ōon* is Greek for *egg*). Ooids form in shallow tropical seas, on a bottom that is within reach of the waves. A great place to see them forming in the modern world is the shallows surrounding the islands of the Bahamas. Gently turned over by the sloshing waves, they get an even coat of calcite on all sides, resulting in their hailstone-like shape. Ooids get cemented together to make oolitic limestone.



Figure 62.5.18: Pavement outcrop (bedding-plane-parallel) of stromatolites in the Conophyton Zone of the Helena Formation, Grinnell Glacier cirque. (Callan Bentley photo.)

Stromatolites are fossil microbial mats, typically found in carbonate strata like the limestone of the Helena Formation. They are much more common in Precambrian strata (Archean and Proterozoic) than in post-Cambrian strata, for an ecological reason: Once snails had evolved their grazing lifestyle with their rasping radula, they began to mow down microbial mats wherever they encountered them. The only places that stromatolites survive in the modern era are snail-free locations: places where the water is too salty, or the tidal currents too strong, or other circumstances that result in water good for growing stromatolites but bad for snails' survival.



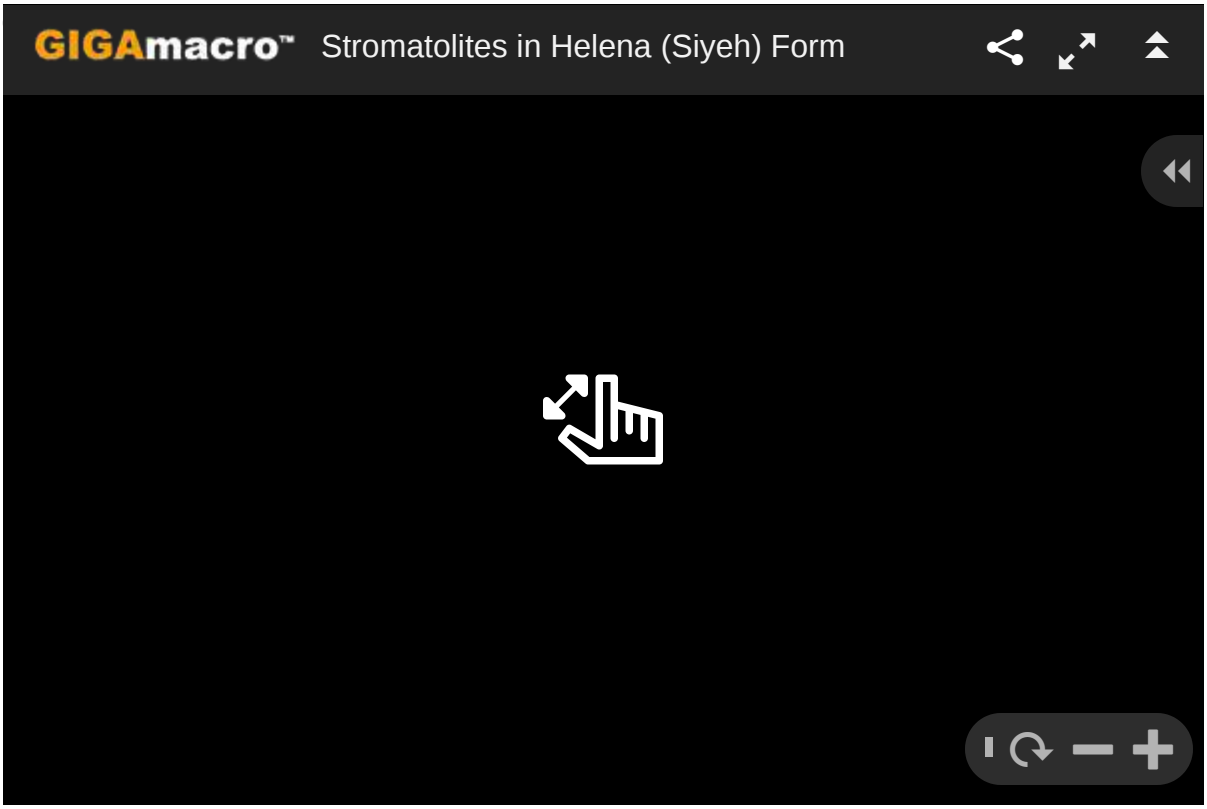
Figure 62.5.19: Stromatolite researcher Dr. Erica Suosaari swims above modern stromatolites in Shark Bay, Australia. (Photo by Justin Gilligan; reproduced with permission.)

Modern stromatolites are survivors, but their ancestors were vital to producing the world we have now: a world where animals (yes, including snails) can survive. Specifically, they are responsible for pumping out enormous quantities of oxygen into Earth's atmosphere. They did this because the microbes that build stromatolites were largely cyanobacteria. They were Earth's first photosynthesizers, in other words, using their own biochemical apparatus to shred water and carbon dioxide molecules from their surroundings, building simple sugars and dumping the leftover oxygen atoms into the surrounding water. Some of that oxygen exsolved from the water, building up in the atmosphere. And some of it reacted with chemicals such as iron in the water. The Archean was an anoxic eon, but oxygen levels began to rise going into the early part of the Proterozoic. By the time of the mid-Proterozoic (when the Belt rocks of Glacier were deposited), stromatolites' oxygen waste had accumulated to great levels in the atmosphere, allowing oxidized sediments such as the red argillite of the Grinnell Formation to be deposited.

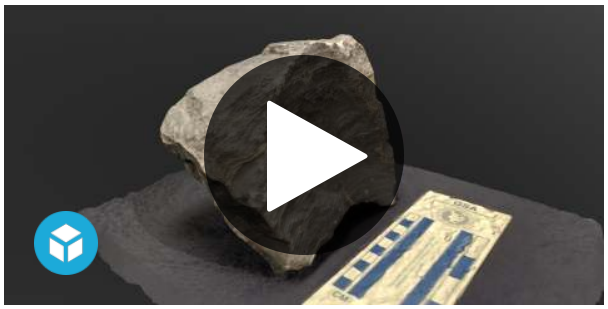
So when you contemplate the exemplary stromatolites of Glacier National Park, a key moment of perspective should emerge: you should thank these ancient microbes for the world they built!

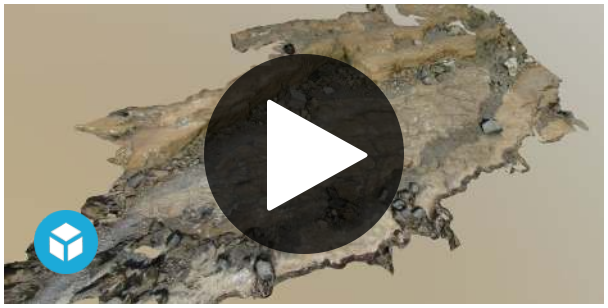


They typically show a domed shape in three dimensions and have many fine internal layers. Viewed in cross-section, they show a convex-upward shape to these layers. Here is a GigaPan of a stromatolite-bearing limestone that shows convex-up doming shapes in its internal laminations (very fine layers). Zoom in and explore this sample, and use the rotation tool to see what these shapes look like if they are flipped up-side-down:



Viewed in map view, they show round, semi-circular forms, with the layers showing as concentric rings. Compare these two 3D models to get a better sense of their shape.

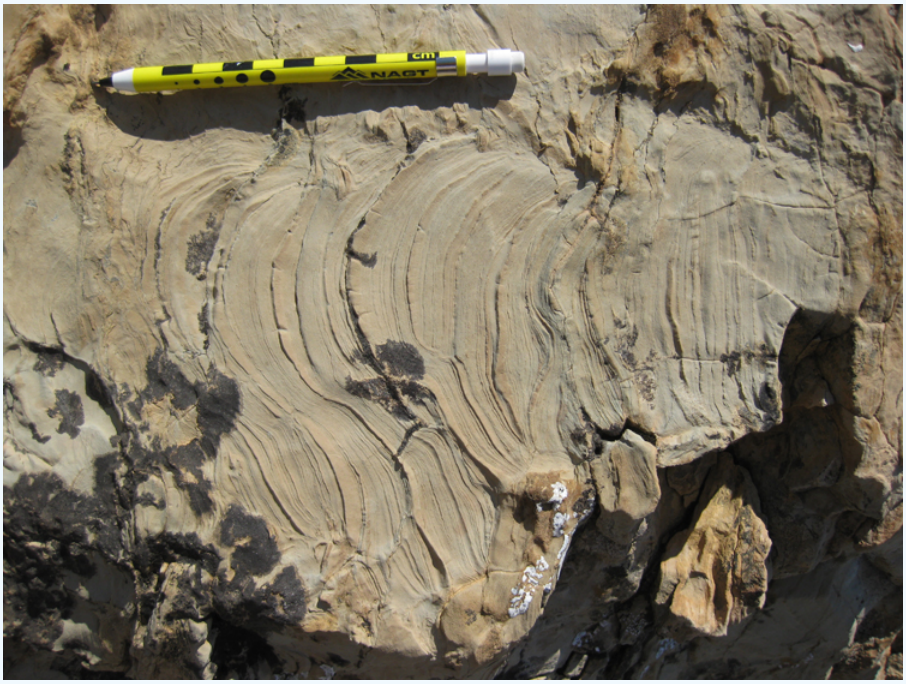




This up-directed doming is very useful for figuring out “younging direction,” or which way was “up” when the strata were deposited, since often times strata get rotated during tectonic deformation. Sometimes they are even flipped up-side-down! Such geopetal applications make stromatolites interesting fossils as well as useful tools.

? Exercise 62.5.1

Which way is stratigraphic "up" in this rock, and why?



a. The bottom of the photo, because the graded bedding gets coarser in that direction.

b. The right of the photo, because the stromatolite's layers are concave in that direction.

c. The top of the photo, because the graded bedding gets finer in that direction.

d. The left of the photo, because the stromatolite's layers bulge in that direction.

Answer

d. The left of the photo, because the stromatolite's layers bulge in that direction.

OTHER ODD FOSSILS

The Belt strata at Glacier are older than the oldest known animal fossils (the Ediacaran fauna). As we have seen however, these rocks do contain fossil evidence, in the form of stromatolites, but those are interpreted as cyanobacterial mats. What, then, are *Horodyskia*? These intriguing “string-of-beads” features are found in several locations in Glacier National Park, all within the Appekunny Formation. Because of how old they are, and how *big* they are, they have been interpreted as some of the oldest evidence of eukaryotic life on this planet. This interpretation is controversial within the scientific community, and some workers think it is instead a pseudofossil or macroscopic manifestation of prokaryotic microbes (that is, akin to a stromatolite). Possibly it is fungal in origin, as a recent paper claims.

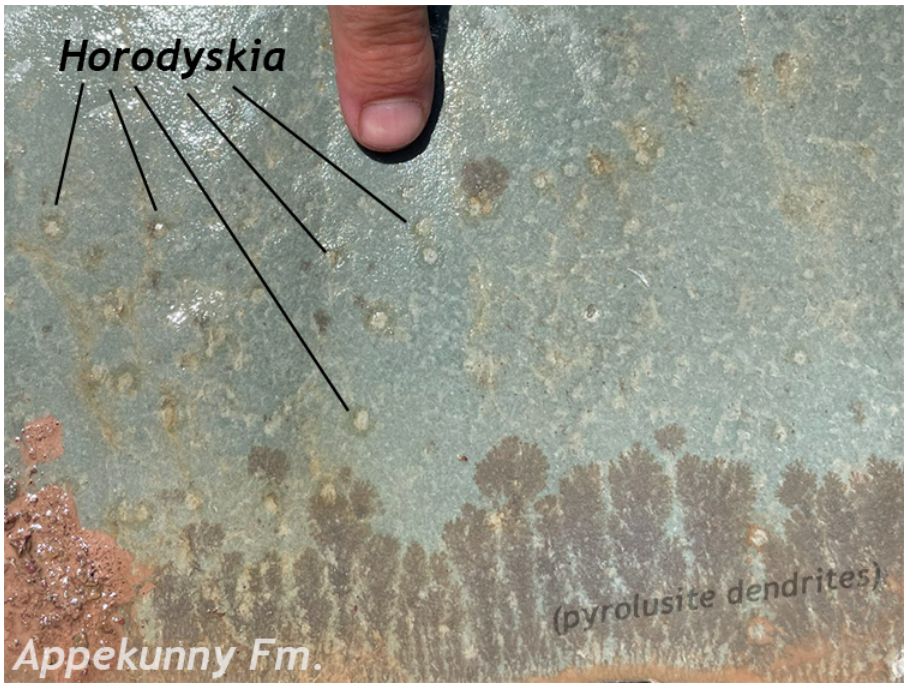


Figure 62.5.20: Horodyskia is an enigmatic primary sedimentary structure. It's probably a fossil, and perhaps a eukaryotic fossil. (Callan Bentley photo.)

There are still mysteries yet to be solved in Glacier's sedimentary stack!

MOLAR TOOTH STRUCTURE



Figure 62.5.21: Molar tooth structure in the Helena Formation. (Callan Bentley photo.)

Molar tooth structure is the name given to an odd feature that is especially common in the carbonate strata at Glacier (i.e. the Helena Formation). It is defined as “arrays of closely spaced, sharply defined, upright veins, and subordinate horizontal sheets and spheroids composed of calcite microspar. Veins are vertically to obliquely oriented, discontinuous, typically strongly squashed or crumpled, and often brecciated” (Pratt, 2011). It looks very messy in cross-section, and something like the upper surface of a molar when viewed on the bedding plane.

Molar tooth structure is pretty much an extinct structure; it no longer forms on this planet. The principle reason for this appears to be that animals and other organisms disturb the sedimentary substrate too much these days. This makes explaining it rather problematic. Is it related to organisms modifying the substrate in an old, no-longer-used way? Is it due to trapped gasses from decomposition of organic matter? As with *Horodyskia*, we do not really know. With molar tooth structure, we can make the observation that it exists, but it is hard to interpret with any confidence.

Did I Get It? - Quiz

? Exercise 62.5.1

Which primary sedimentary structure forms in the open air?

a. Mud cracks (desiccation cracks)

b. Ripple marks

c. Cross-bedding

d. Graded bedding

e. Stromatolites

Answer

a. Mud cracks (desiccation cracks)

A bit of lava



Figure 62.5.22: Exposure of basalt pillows, indicating underwater eruption, at Granite Park, Glacier National Park. (Photo by Shelley Jaye.)

The vast majority of Belt rock is sedimentary. However, within the Snowslip Formation, there are some mafic volcanic rocks too – basalts that show both evidence of lava flow in the open air and of eruption underwater. Evidence for underwater eruption comes in the form of the primary volcanic structure “pillows” (rounded forms of chilled lava with a quenched crust, radiating vesicles, and often a distinctive upside-down teardrop shape). As for subaerial flow, pavement exposures show pahhoehoe texture, another primary volcanic structure. This “ropey” surface forms as the “skin” of basalt flows wrinkles, perpendicular to the flow direction. There are analogues in modern day Hawaii, Iceland, and the Galápagos.

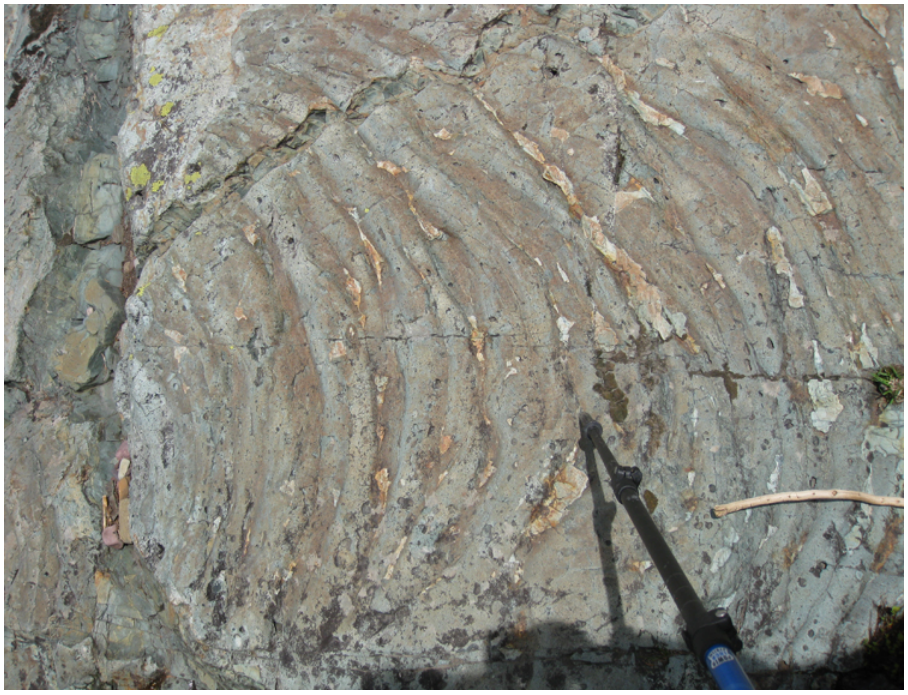


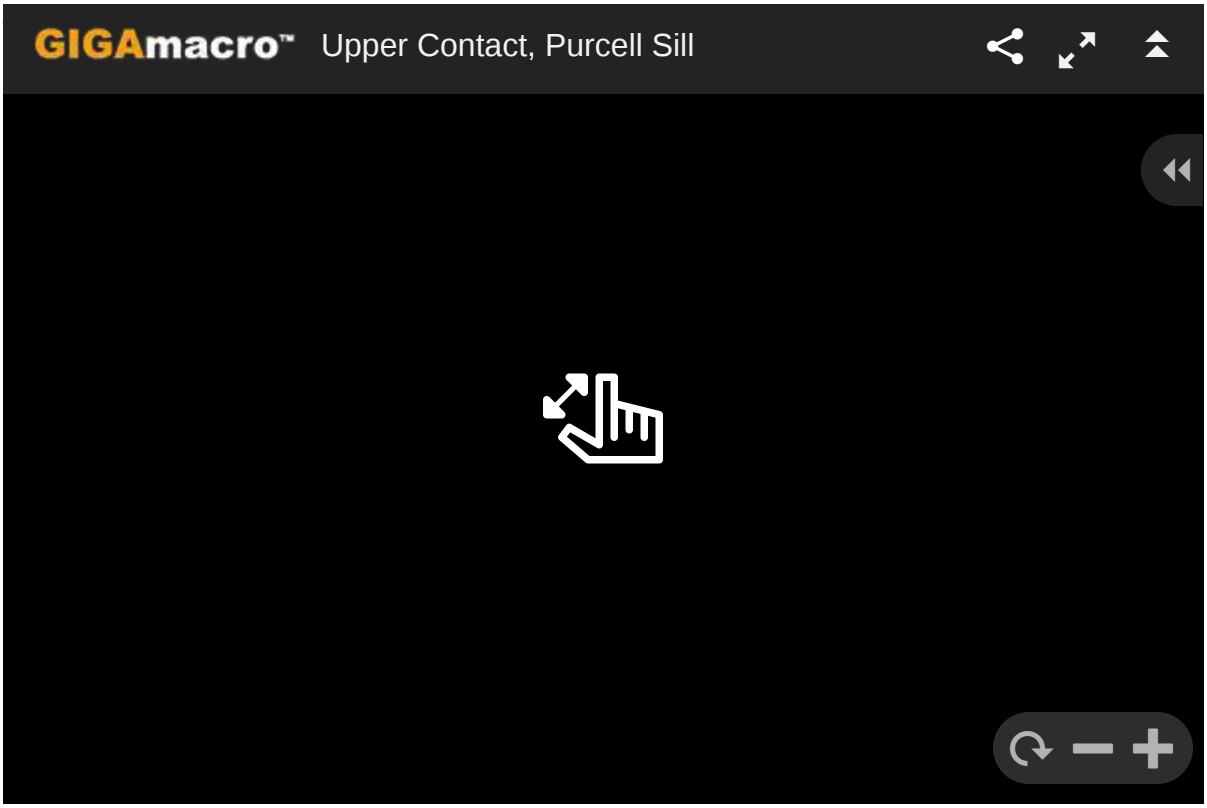
Figure 62.5.23: Pahoehoe texture in Belt lava flows at Granite Park in Glacier National Park. (Photo by Shelley Jaye.)



Figure 62.5.24: Between the summit of Going-to-the-Sun Mountain and the Sexton Glacier is a cliff that shows the diorite sill prominently as a black stripe bracketed by white margins. (Callan Bentley photo.)

Much later, more magma enters the story, intruding between layers of the Helena Formation to make a huge sill. This sill, which is diorite (intermediate) in composition, baked the surrounding rock as it squeezed in. This altered the adjacent limestone to a much lighter color. This “reverse Oreo” pattern of white/black/white shows up very prominently on the mountains around Glacier National Park.

Here is a gigapixel panorama showing the sill’s upper contact with the surrounding Helena Formation limestone, as exposed on Going-to-the-Sun Road. Explore it to note the change in the color of the Helena Formation as it gets closer to the top edge of the sill:



Similar mafic sills and dikes in the region have been dated at 750 Ma (Neoproterozoic). Of course, based on the relative dating principle of cross-cutting relationships, we would expect the sill to be younger than the rocks into which it intruded, but the isotopic date confirms this.

The dark sill and its flanking zones of white altered limestone are very prominent features high on mountain peaks throughout the central part of the park:

Did I Get It? - Quiz

? Exercise 62.5.1

True or false: "All igneous rocks at Glacier National Park are about the same age as the sedimentary rocks in which they are found." Explain.

a. TRUE. "The Purcell Lava and the diorite sill are different names for the same thing."

b. FALSE. "While the Purcell Lava was erupted at Earth's surface at the same time the Snowslip Formation was being deposited, the diorite sill was intruded many hundreds of millions of years later."

Answer

b. FALSE. "While the Purcell Lava was erupted at Earth's surface at the same time the Snowslip Formation was being deposited, the diorite sill was intruded many hundreds of millions of years later."



Figure 62.5.25: Colonial tabulate corals (probably Syringopora) weather out of the Mississippian limestones at Swift Dam. (Callan Bentley photo.)

There is a huge jump in time between the deposition of the Belt Supergroup's strata and those of the next major package of sediment in our story. Elsewhere in Montana, the stories of the intervening times are well recorded and exposed. But in the vicinity of our virtual field experience, we are going to jump over those intervening times and drop straight into the Mississippian period of geologic time, about 340 Ma. This is the mid-to-late Paleozoic Era, which means it is well recorded by fossil animals, particularly marine invertebrates.

During the Mississippian, global sea level was high, and western Montana was low. Accordingly the spot that is today Glacier National Park was then flooded by shallow tropical seas. Shoreline must have been far away, as the resulting sediments are clean limestones lacking significant mud or sand content. They suggest that during the Mississippian, Montana would have looked a lot like Australia's Great Barrier Reef, or the Dry Tortugas in Florida.

Montana was not alone in being below a shallow tropical sea during the Mississippian. The famous Redwall Limestone in the Grand Canyon was deposited during this same transgression, as were carbonate strata in many locations, even as far away as the Appalachians.

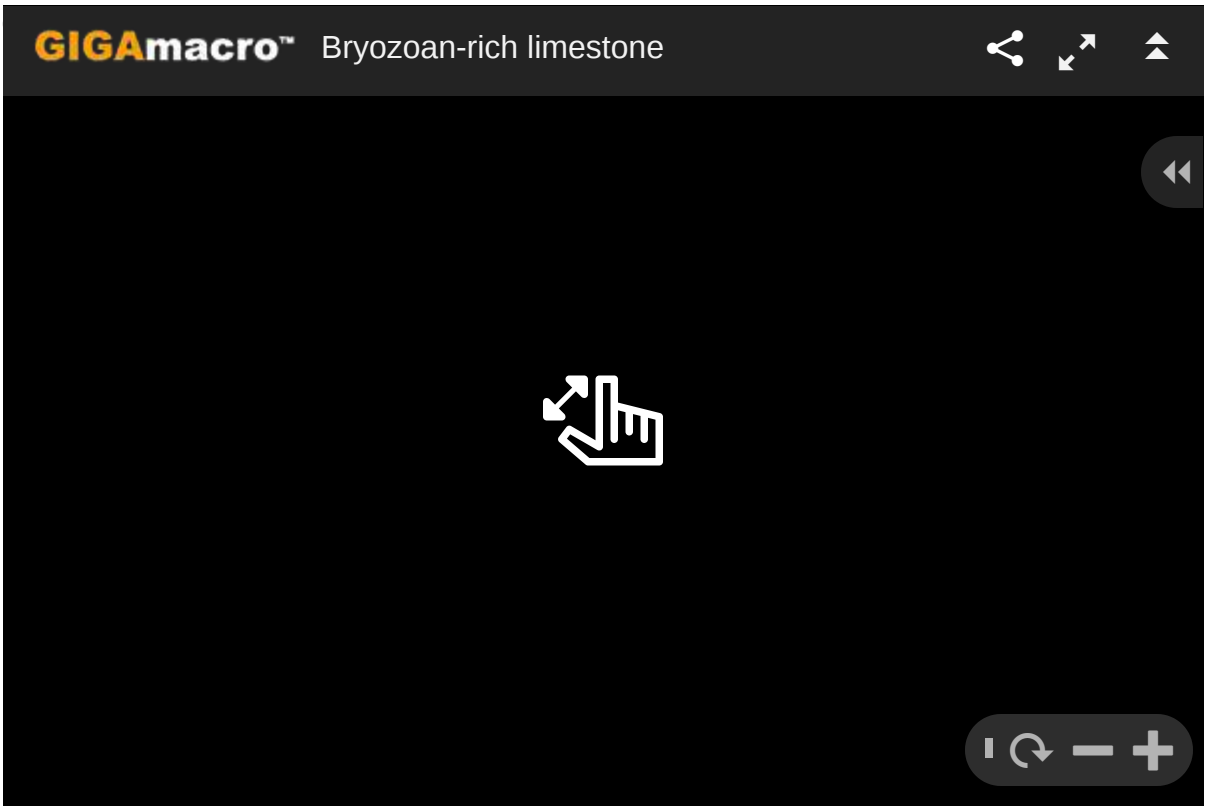


Figure 62.5.26: Solitary rugose corals can be surprisingly large. (Callan Bentley photo.)

Lots of fossil animals are found in these layers, ranging from crinoids and brachiopods to corals and bryozoans. These are all marine-dwelling invertebrate animals, most of which are filter-feeders. They made their hard skeletal material from $CaCO_3$, and when they died, fragments of those skeletal remains built up in vast thickness of limestone.

The limestones are known as the Madison Group limestones. Because of the arid climate of Montana, they tend to weather more slowly than sandstones or shales, and become etched out as prominent ridge-forming units in the landscape. This climatological influence on landscape development is important: similar limestones exposed in the more humid climate of the American east are etched away to form valleys.

Here are two examples of fossils to be found in Mississippian-aged limestones in Montana:





On our VFE, we will see these carbonate strata in the Sawtooth Range (at Sun River Canyon and Swift Dam), but not at Glacier National Park itself.

Did I Get It? - Quiz

? Exercise 62.5.1

Limestone with coral fossils is an indication of _____.

a. deposition in a shallow sea that covered the northwestern part of North America after the rifting from Siberia during the Mesoproterozoic eon of geologic time

b. deposition in a shallow tropical sea that covered much of North America during the Mississippian period of geologic time

c. deposition in a foreland basin that covered the Great Plains states of North America during the Cretaceous period of geologic time

d. deposition at the terminus of a glacier that covered much of northern North America during the Pleistocene epoch of geologic time

Answer

b. deposition in a shallow tropical sea that covered much of North America during the Mississippian period of geologic time

Cretaceous sediments of the Western Interior Seaway

Jumping ahead again in time, we arrive at the youngest package of bedrock in the Glacier region, a collection of Cretaceous-aged sand and mud deposits.

Tectonic context

In the Triassic, the supercontinent Pangaea began to break apart. In the Paleozoic, North America had been heading east to make Pangaea (compressing the rocks on the east coast and building the Appalachian Mountains), but by the Mesozoic, the tectonic motion of the continent shifted into reverse. The east coast became a divergent boundary, and North America began moving westward. This shift in motion opened up the Atlantic Ocean in the "wake" of the continent, with the east coast becoming a passive margin on the trailing edge of the plate. The west coast was now the focus of all the tectonic action. By the Jurassic, North America had started smashing into other landmasses and accreting new rocks along its western edge, now an active margin. Those mountains generated sediment, and those sediments were transported down and into a foreland basin to their east. The Sevier Orogeny had begun.

By the Cretaceous (say, around 100 Ma), the loading of new terranes onto the western continental margin had thickened the crust sufficiently that it subsided a bit into the mantle, both directly underneath the mountains, but also the neighboring crust too. This was a time of vigorous seafloor spreading around the world, so sea level was already high. It was also a time of extreme global warming, with no evidence for glacial ice anywhere on the planet. So there was lots of water in the oceans, but there was also a lot of oceanic ridge volume, displacing water, and now crustal loading of earliest Sevier mountains caused the continent to the east of the range to flex downward. This allowed the development of a substantial sea to develop, stretching from the Gulf of Mexico to the Arctic Ocean. This was the Western Interior Seaway.

Examining this virtual globe will give you a sense of the incredible extent of the Western Interior Seaway:



Figure 62.5.27: Earth as it appeared in the middle Cretaceous period of geologic time. Find ancestral North America and note the band of water running from the Yukon to Texas through the middle of the continent. This is the Western Interior Seaway. This dyanmic visualization was created by Ian Webster using tectonic and paleogeographic maps by C.R. Scotese's PALEOMAP Project, and is embedded here with permission. Grab it and spin it around! Zoom in and out! Explore!

Sedimentary facies

The Western Interior Seaway changed size as sea level rose and fell, so sometimes it was wider and sometimes it was more narrow. In the vicinity of Glacier National Park, if the seaway was in a "wide" phase, then the water was deep at Glacier (*i.e.*, it was an offshore marine facies), and that was principally recorded in the form of lots of black mud. When the seaway was "narrow," the Glacier region hosted the seaway's shoreline, a circumstance recorded by sandier, higher-energy, higher-oxygen deposits. At times when sea level was especially low, the area around Glacier was exposed above sea level (making it terrestrial). Fortunately, plenty of rivers were draining the Sevier Mountains to the west, and dumping plenty of the resulting sediment on their way downhill toward the Western Interior Seaway.



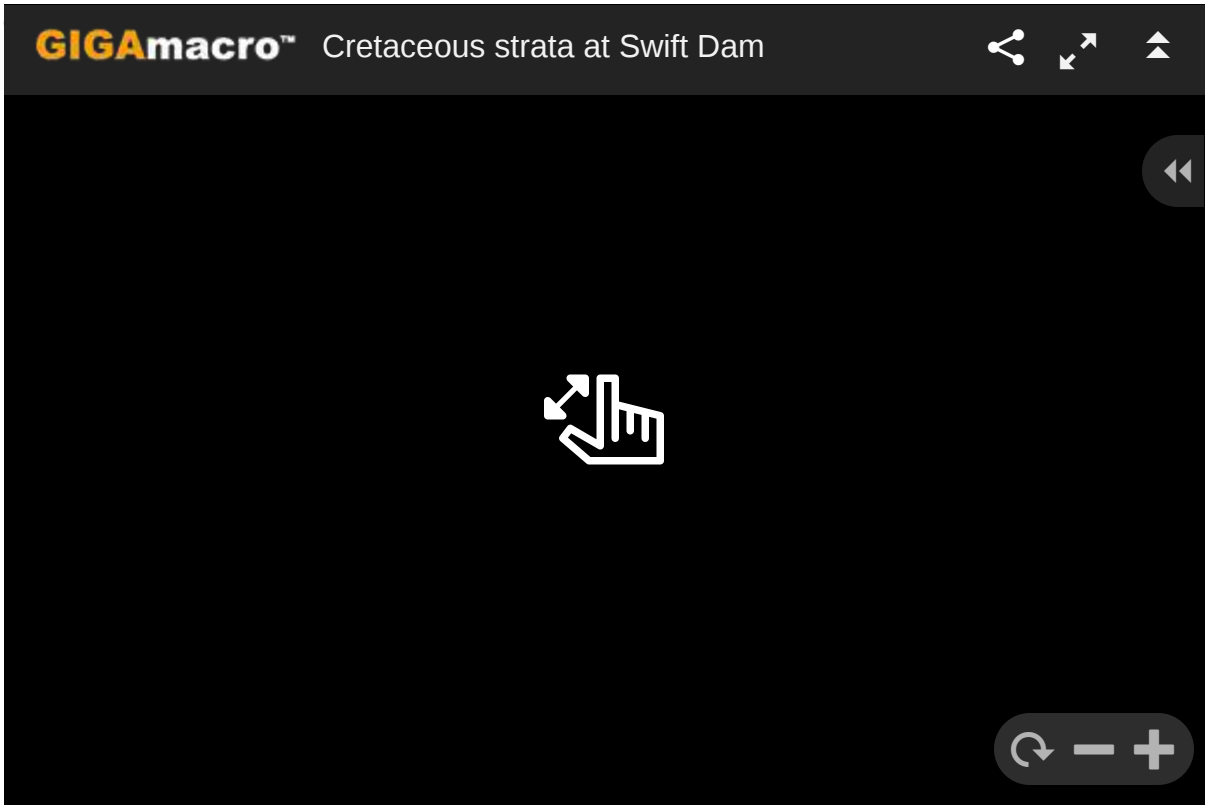
Figure 62.5.28: Ammonites (left) and inoceramid clams (right) from the Bearpaw Shale, a deepwater deposit from the Western Interior Seaway. Collected near Glendive, Montana. Craft beer for scale. (Callan Bentley photo.)

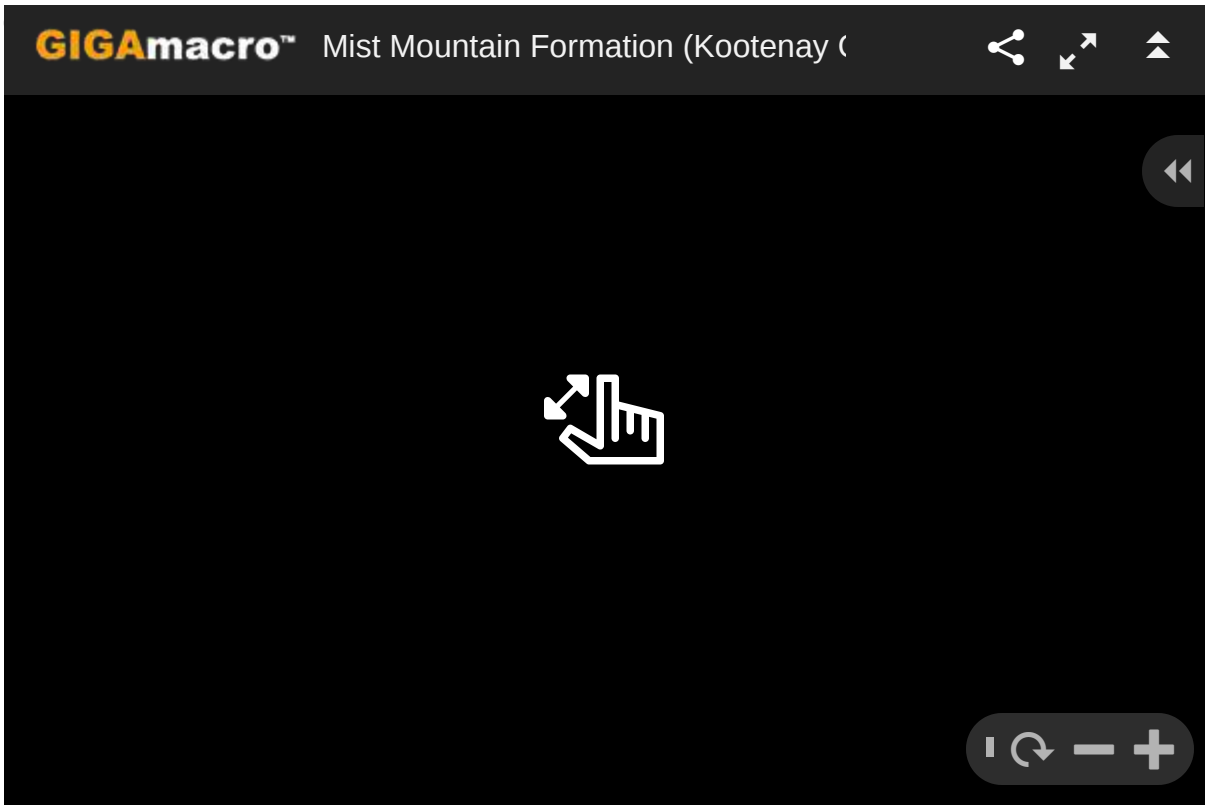
The deepwater deposits of the Western Interior seaway host fossils of marine organisms, including oysters, squid relatives called belemnites, ammonites, and massive marine reptiles such as mosasaurs and sea turtles such as *Archelon*. At Swift Dam, exposures of the deepwater shales include the "Devil's Toenail" oyster more formally known as *Gryphea*, as well as belemnite fragments that look like translucent brown bullets.

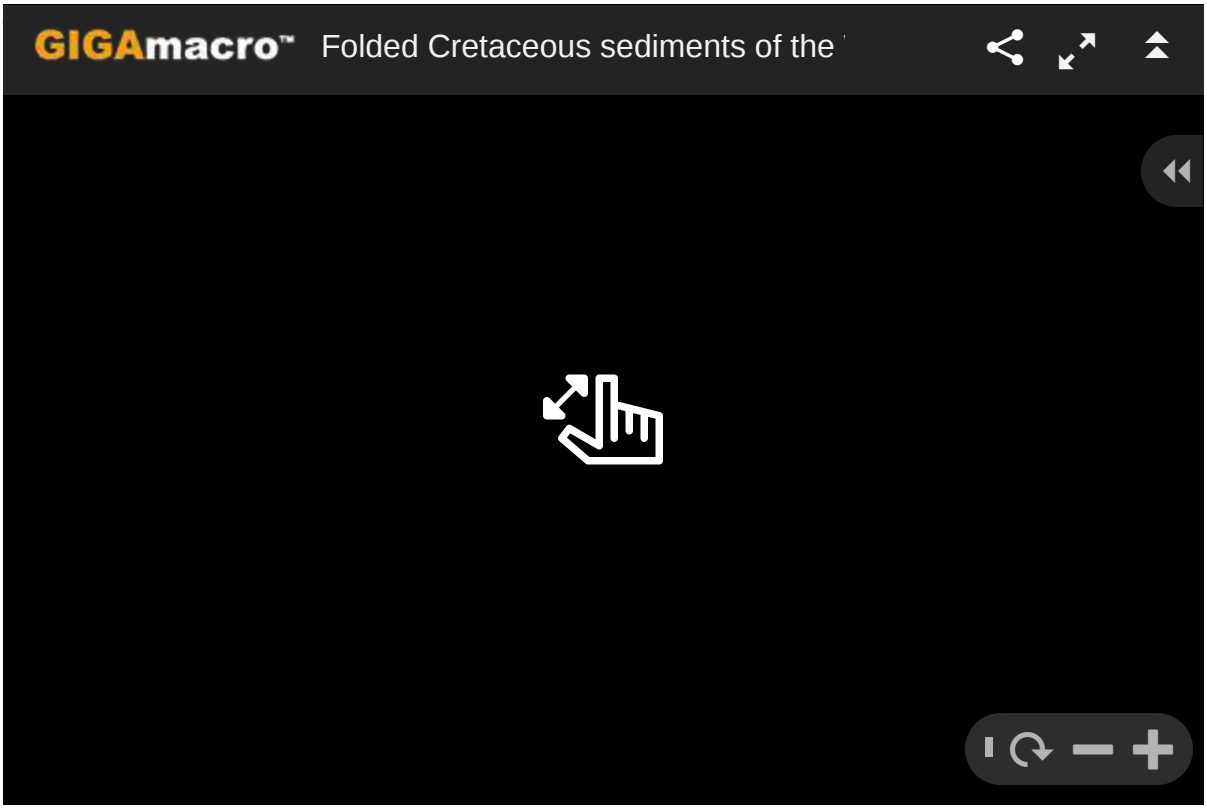
The shoreline facies deposits of the Western Interior Seaway host the fossil remains of terrestrial organisms, most notably dinosaurs. Sites like Egg Mountain near Choteau or Dinosaur Provincial Park in Alberta are terrestrial, and retain an astounding record of these fascinating creatures.

Most of the Cretaceous strata cropping out at the locations visited by this VFE is black shale, but about a third of it is sandstone. You can tell the two apart, even at a distance, by two variables: (1) shale is darker colored (sandstone is lighter) and (2) shale weathers into crumbly slopes (while sandstone weathers in blocky ridges). Formation names include the Blackleaf Formation, the Kootenai Formation, the Marias River Shale, the Bearpaw Shale, the Virgelle Formation, the Two Medicine Formation. For the purposes of this VFE, we will not distinguish among these, lumping them all together instead: a lot of sandstone and black shale.

Here are three examples of representative outcrops, placed in order of most-shale-dominated to most-sand-dominated... And therefore more-offshore/deepwater to more-nearshore/shallow water. Explore each to estimate the proportion of shale to sandstone:


GIGAmacro™ Cretaceous strata at Swift Dam


GIGAmacro™ Mist Mountain Formation (Kootenay (



You will also have noticed in each of those GigaPan images that the strata are not in their original horizontal orientation, implying serious tectonic deformation after they formed. We will cover that in the next major segment of this VFE.

Did I Get It? - Quiz

? Exercise 62.5.1

How big was the Western Interior Seaway?

a. Pretty big: it stretched from the Arctic to the Gulf of Mexico.

b. Petite: It was big enough for about 2 goldfish, but not roomy enough for 3.

c. Enormous: it covered the entire Earth.

d. Modest: About the size of Lake McDonald in modern-day Glacier National Park.

Answer

a. Pretty big: it stretched from the Arctic to the Gulf of Mexico.

? Exercise 62.5.2

What were the principal two rock types deposited in the Glacier National Park region by the Western Interior Seaway?

a. quartzite and diorite

b. limestone and argillite

c. till and colluvium

d. sandstone and shale

Answer

d. sandstone and shale

Pleistocene sediments



Figure 62.5.29: Glacial outwash sediments (stratified drift) exposed on the Looking Glass Highway between Kiowa and East Glacier. (Callan Bentley photo.)

Glaciation during the recent Pleistocene ice ages served not only to sculpt the landscape by removing rock, but then the resulting debris was deposited as sediment. Glacial "till" was deposited directly by glacial ice, and glacial outwash sediment called "drift" was deposited by rivers flowing away from the glacial termini. Where these rivers were impounded by moraine dams, lakes formed, and in these lakes, glacial silt settled out. Glacial icebergs floating in these lakes rafted dropstones out to calm-water sites. You can learn more about this sequence of sediments in the final excursion on this virtual field experience. For now, it is enough to know that there are patchy Pleistocene deposits on top of the Belt strata, the Mississippian carbonates, and the Cretaceous siliciclastic strata of the Western Interior Seaway.

Did I Get It? - Quiz

? Exercise 62.5.1

How does Pleistocene sedimentation fit into the stratigraphic sequence at Glacier National Park?

a. Both till and stratified drift from melt outwash are found in places on top of all the other three major sedimentary rock types.

b. It underlies all the other sedimentary rock units in the park region, as the original material atop which all the other layers were deposited.

Answer

a. Both till and stratified drift from melt outwash are found in places on top of all the other three major sedimentary rock types.

Further reading

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62.6: Glacier National Park, Montana - Structural Geology

Structural signatures of the Sevier Orogeny

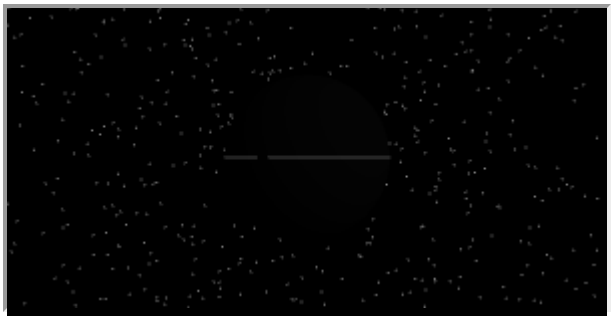
In addition to the sweeping cliffs and brilliant colors of Glacier, there is also a profound story of mountain-building recorded in the rocks of the park. The mountains of Glacier formed during the late Mesozoic and early Cenozoic due to the Sevier Orogeny. This episode of mountain-building shoved older Belt rocks and Mississippian carbonates atop Cretaceous siliciclastic strata in an impressive inversion of their original stratigraphic order.

Tectonic context

The Northern Rocky Mountains of Montana were impacted by mountain-building processes during the Cretaceous. The Sevier Orogeny shoved stacks of sedimentary rocks (including Mesoproterozoic Belt strata, Mississippian carbonates, and Cretaceous siliciclastics) from their original positions deep in the crust and further to the west to new positions, upward and to the east. These displaced layers of rock arched upward and broke, with the slabs of older rock traveling along thrust faults up on top of younger rocks.

The cause of the Sevier Orogeny was the accretion of numerous terranes from the ancient Pacific Ocean, brought eastward as the Farallon Plate subducted beneath the North American Plate. The Farallon Plate used to exist between the Pacific Plate and the North American Plate. The Farrallon's western boundary with the Pacific Plate was divergent: it was a spreading center (an oceanic ridge). The Farallon's eastern boundary (with the North American Plate) was convergent: a subduction zone. There, the Farallon was recycled into the mantle. Any island arcs, microcontinents, or oceanic plateaux it was carrying were too thick, buoyant, or protruding to subduct. When they reached the subduction zone, these "terranes" were scraped off and pushed onto and into the North American continent. As they were pushed eastward, they acted like a bulldozer, scraping and shoving the sedimentary rocks that were layered atop the continental crust.

Consider the nascent mountain-building evident in western North America in this paleogeographic depiction from 105 Ma:



Those mountains were driven by compression from the west. As new terranes accreted to the continent, they pushed the pre-existing sedimentary layers out of the way, so that direction was to the east. This process initiated during the Jurassic period of geologic time, but reached its zenith during the Cretaceous. Most of the Farallon Plate was recycled during that time, but two remnants persist into the present day: the Cocos Plate, currently being subducted beneath Central America, and the Juan de Fuca Plate, currently being subducted beneath the Pacific Northwest states of Washington, Oregon, and northern California.

Did I Get It? - Quiz

? Exercise 62.6.1

What was the name of the mountain-building event that deformed the rocks of Glacier National Park and the surrounding area?

a. the Sawtooth Orogeny

b. the Sevier Orogeny

c. the Rockyrogeny

d. the Laramide Orogeny

e. the Lewis Orogeny

Answer

b. the Sevier Orogeny

Exercise 62.6.2

What was the tectonic cause of the Sevier Orogeny?

a. Collision between continents as the North American continent moved eastward into the Iapetus Ocean basin. Once ancestral Africa arrived at the edge of the continent, the two landmasses collided, and as Africa's western edge overrode North America's eastern edge, it shoved the existing sedimentary layers out of the way, to the west.

b. Accretion of blocks of crust called terranes as the North American continent moved westward into the Pacific Ocean basin. The blocks of crust were "delivered" to the edge of the continent by a subducting plate called the Farallon Plate. Once these island arcs, microcontinents, and oceanic rocks arrived at the edge of the continent, their accretion shoved the existing sedimentary layers out of the way, to the east.

Answer

b. Accretion of blocks of crust called terranes as the North American continent moved westward into the Pacific Ocean basin. The blocks of crust were "delivered" to the edge of the continent by a subducting plate called the Farallon Plate. Once these island arcs, microcontinents, and oceanic rocks arrived at the edge of the continent, their accretion shoved the existing sedimentary layers out of the way, to the east.

Exercise 62.6.3

When did the Sevier Orogeny happen?

a. Triassic to Jurassic

b. Paleogene to Neogene

c. Permian to Triassic

d. Jurassic to Cretaceous

e. Cretaceous to Paleogene

f. Pennsylvanian to Permian

Answer

d. Jurassic to Cretaceous

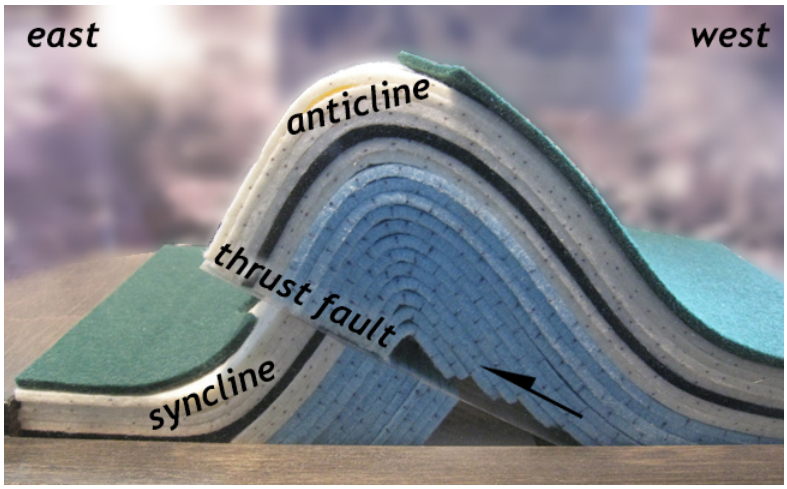
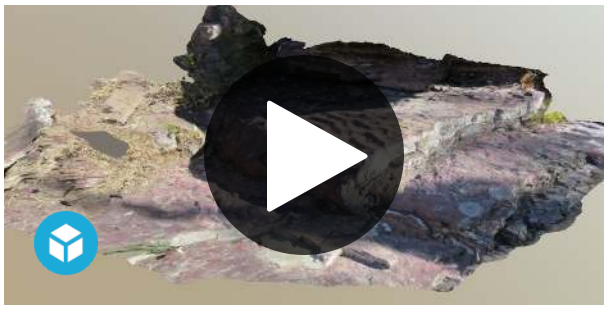


Figure 62.6.1: This felt model of Turtle Mountain, Alberta is on display at the Frank Slide Visitor Center. The structure shown is very similar to the structure of the faults and folds at Sun River Canyon and Swift Dam.

When the sedimentary layers of western Montana were compressed, they buckled and broke, shearing to the east as they deformed. The result was an immense number of asymmetric folds with the short limb on the east and the long limb on the west, frequently breaking at the hinge of the fold, and initiating a fault. These distinctive structures are the typical "structural ingredient" in compressional mountain belts around the world, and they are on great display in the region around Glacier National Park.

An illuminating outcrop on the north side of the Going-to-the-Sun Road shows this structure in three dimensions:



It shows quartzite and red argillite of the Grinnell Formation, with prominent ripple marks on the upper surface of the bed, folded and overturned on the leading edge. This asymmetric, recumbent fold is representative of the general shape of the various thrust sheets in the Disturbed Belt. Minor deformation of the same character occurs within the Lewis Thrust sheet, in a variety of orientations.

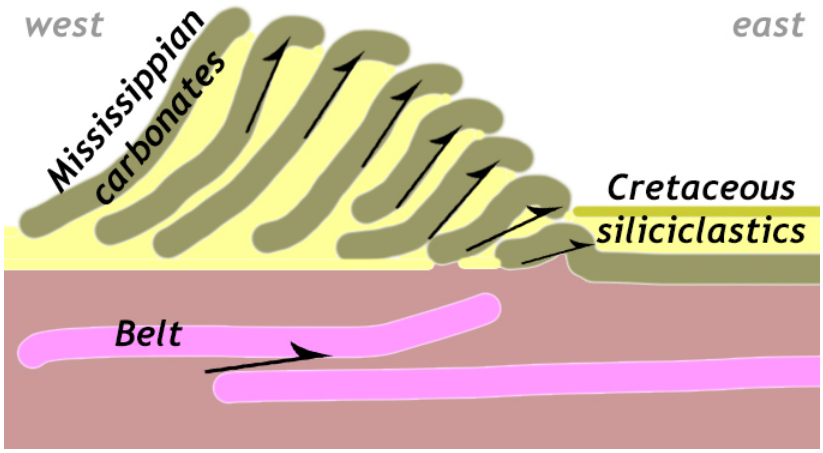


Figure 62.6.2: Cartoon cross-section through an imbricate stack of thrust sheets, as seen at Sun River Canyon.

In general with faults, the footwall rocks are those below the fault, while the rocks above the fault are called the hanging wall. However, additional terminology is used when it comes to thrust faults, with their significant lateral offsets, and their orientations so close to horizontal. The block or slab or rock above the thrust fault is sometimes called a thrust sheet, or by the term nappe.

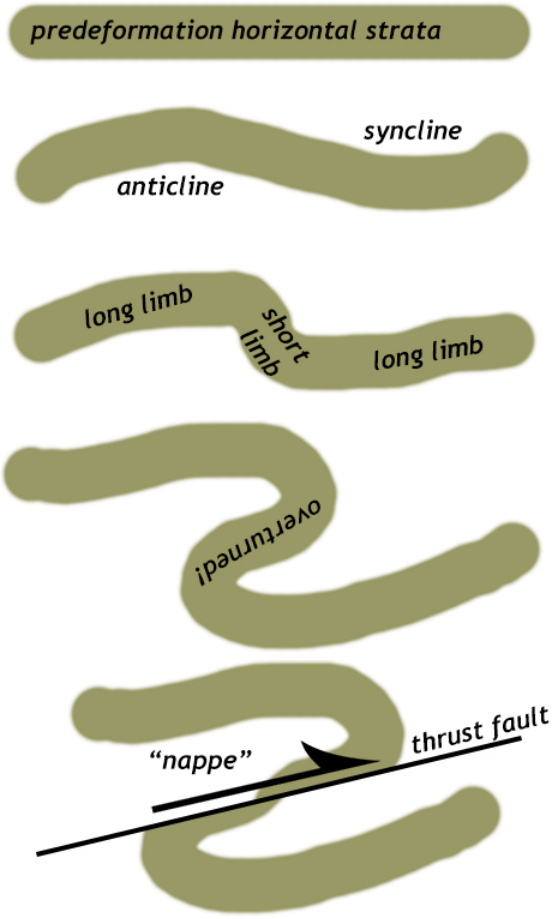


Figure 62.6.3: A cartoon cross-section showing 5 steps in the folding and thrust faulting of an asymmetric anticline to make a "nappe."

Nappes form when horizontal rock layers are compressed laterally (*i.e.*, from the side) and first buckle into asymmetric folds. The short "limb" of these anticlines and synclines forms in the transport direction (the direction toward which the strata are being thrust). Frequently, this short limb rotates back on itself and becomes tectonically inverted. It also frequently is the place where the strata rupture, and a new thrust fault initiates. Once the mechanical discontinuity of the fault exists, it's the "weakest link," and further deformation is accommodated mainly through sliding along that fault surface, not additional folding. Eventually, the fault stops moving, and a new fault forms "outboard" of the first (*i.e.*, further away from the source of the compression). In the case of the Sawtooth Range at Sun River Canyon and Swift Dam, this "more outboard" direction is to the east.

As deformation proceeds, the older thrust sheets get shoved backward and out of the way as new thrusts form beneath them, rising up to the surface. The result is that the older, more western thrust sheets get rotated to steeper dip angles as deformation proceeds. This whole "stack" of nappes is said to be "shingled," or "imbricated."



Figure 62.6.4: Sandbox deformation experiment showing how layers of sand fold and fault due to lateral compression. This closely mimics the evolution of the Disturbed Belt at Sun River Canyon (as viewed from the north, looking south). (Model and source video by Marco Martins-Ferreira.)

Watch the animated GIF here of a sandbox deformation experiment to get a better sense of this process. It's useful to watch it many times, focusing on a different region each time and watching how it changes with increasing deformation. Imagine that the perspective here is the same as you would see if you were (a) able to go back in time and (b) hover above the crust north of the Sawtooth Range, looking south, and (c) watch geological processes happen at much faster rates than they really do, and (d) had a magical pair of X-ray glasses that allowed you to peer into the Earth and see these complicated 3D structures evolve in a 2D cross-section.

Did I Get It? - Quiz

? Exercise 62.6.1

What kind of deformation is recorded by the sedimentary layers that experienced the Sevier Orogeny?

a. Symmetric folds only.

b. Both normal faults and symmetric folds.

c. Thrust faults only.

d. Asymmetric folds only.

e. Normal faults only.

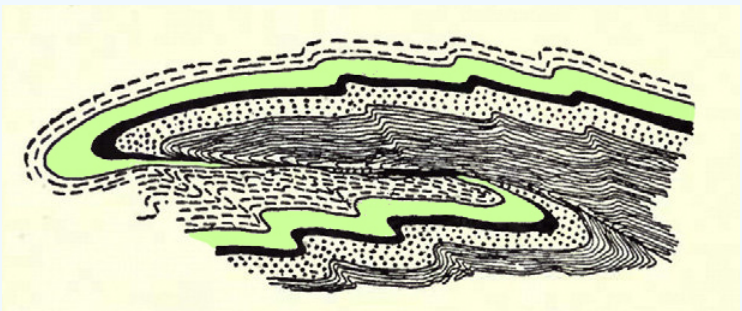
f. Both thrust faults and asymmetric folds.

Answer

f. Both thrust faults and asymmetric folds.

? Exercise 62.6.2

If you saw this structure in outcrop in the Glacier National Park area, in what direction would you be facing?



a. NORTH: the short limb and tectonic transport direction is to the right, and that therefore must be east.

b. EAST: the short limb and tectonic transport direction is to the top, and that therefore must be east.

c. WEST: the short limb and tectonic transport direction is to the bottom, and that therefore must be east.

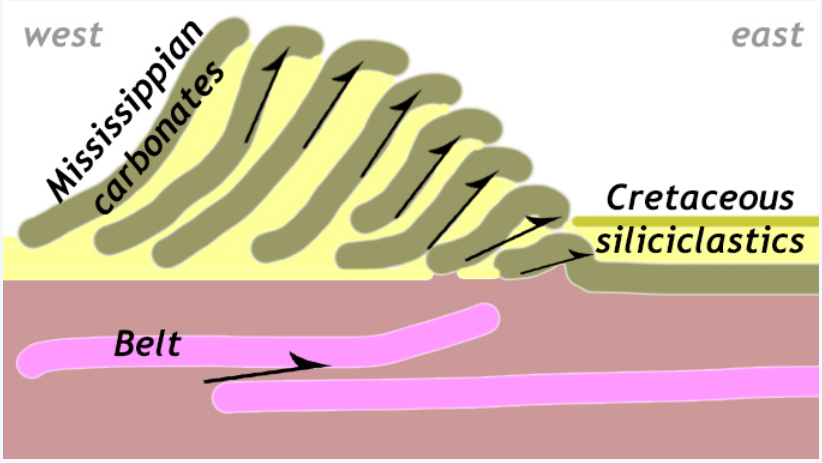
d. SOUTH: the short limb and tectonic transport direction is to the left, and that therefore must be east.

Answer

d. SOUTH: the short limb and tectonic transport direction is to the left, and that therefore must be east.

Exercise 62.6.3

The Mississippian carbonates in this cross-section are _____.



a. imbricated

b. napping

c. undeformed

Answer

a. imbricated

Exercise 62.6.4

Based on color, weathering, and topography, in which stratigraphic unit does the deformation in this photograph occur?



a. Limestones of the Madison Group (Mississippian)

b. Quartzites and argillites of the Belt Supergroup (Mesoproterozoic)

c. Sandstones and shales of the Western Interior Seaway (Cretaceous)

d. Glacial till and outwash (Pleistocene)

Answer

c. Sandstones and shales of the Western Interior Seaway (Cretaceous)

Sevier vs. Laramide deformation

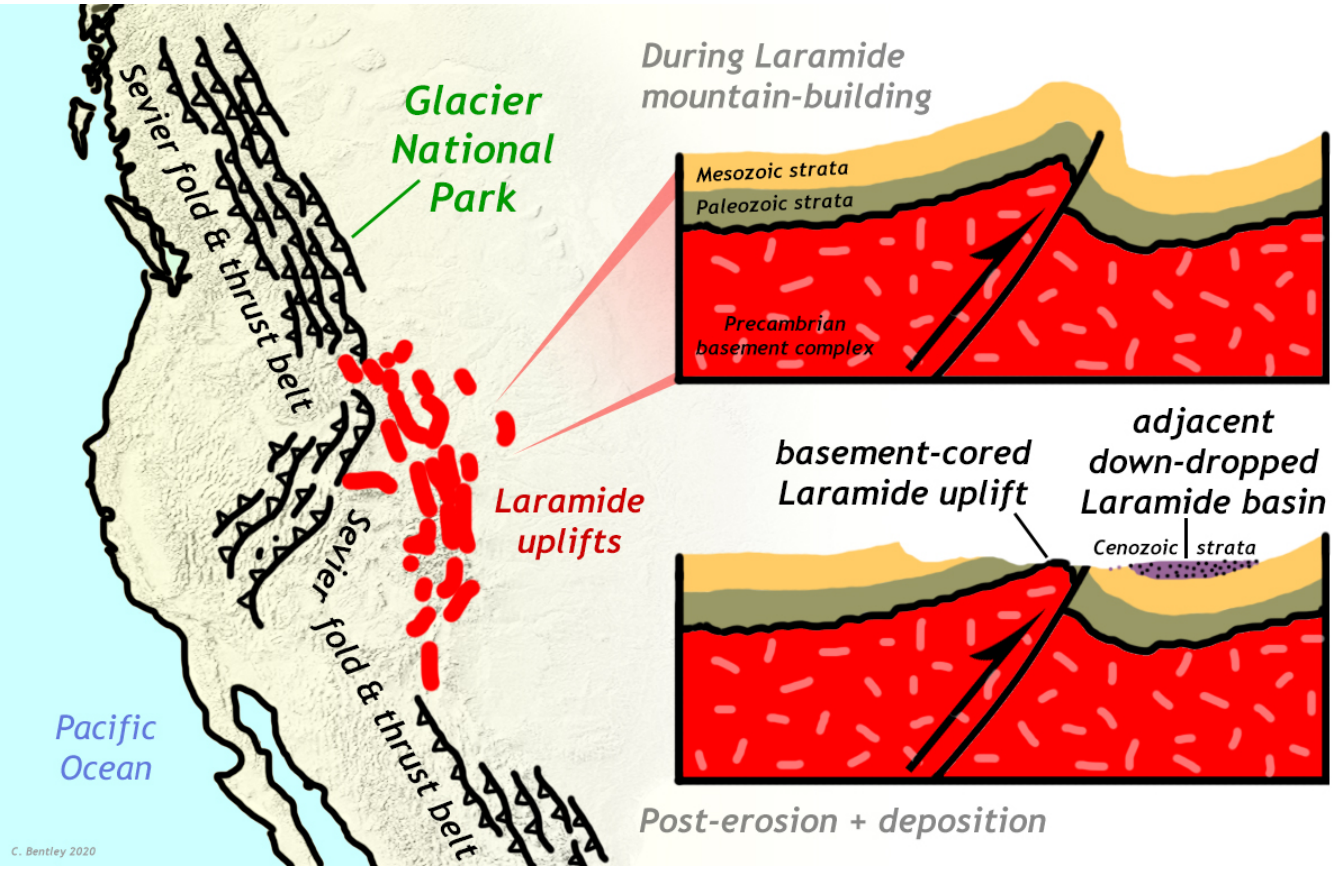


Figure 62.6.5: Map showing the geography of the Sevier fold and thrust belt relative to Laramide uplifts. Cross-sectional views show the structure of the basement-cored Laramide uplifts.

There is another mountain-building event that took place in the Rockies during the late Mesozoic and early Cenozoic, related to but distinct from the Sevier Orogeny. This is the Laramide Orogeny. Ultimately, the Laramide too was caused by the subduction of the Farallon Plate. Compared to the Sevier, however, (1) the deformation style was different, (2) the deformation was more recent, and (3) the deformation took place further to the east. These three variables allow the two orogenies to be distinguished.

Deformation style



Figure 62.6.6: Knightia, a freshwater relative of sardines, is the most common fossil in the Eocene Green River Formation of Wyoming and Utah.

The Sevier Orogeny resulted in a classic fold-and-thrust belt of sedimentary layers. This is an example of "thin-skinned deformation," where the sedimentary cover is deformed, but not the underlying continental crust (the "basement rock" of metamorphic and plutonic rocks). In contrast, the Laramide Orogeny's style is "thick-skinned," which is a way of noting that the crustal "basement" is very much involved. Laramide deformation is characterized by very large anticlines and synclines that involve basement rock. The anticlines generated highlands and mountain ranges, which were weathered and eroded. The synclines were so significant that they warped the surface of the Earth, generating lowlands. These sedimentary basins that were filled with huge amounts of cast-off sedimentary detritus from the eroding mountains nearby, making new Cenozoic-aged sedimentary

deposits that filled the depressions. In some cases, these young sedimentary layers reach thicknesses of 5000 meters! The economically important coal and natural gas deposits of Wyoming formed in Laramide basins, as did the fossil-rich lake deposits of the Green River Formation.

Timing

Because the Sevier Orogeny was caused by accreting terranes pushing sedimentary layers eastward, out of their way as they collided with the western edge of the continent, it began first. It began in the Jurassic, around 160 Ma, and the last of the thrusting was concluded by around 50 Ma. Because the Laramide Orogeny was driven from beneath, once flat-slab subduction had been achieved, it came later. The onset of Laramide mountain-building was in the late Cretaceous, 80 to 70 Ma, with the episode wrapping up in the window of 55 to 35 Ma. So the two orogenies overlapped in time, but the Sevier started first and ended first.

Location

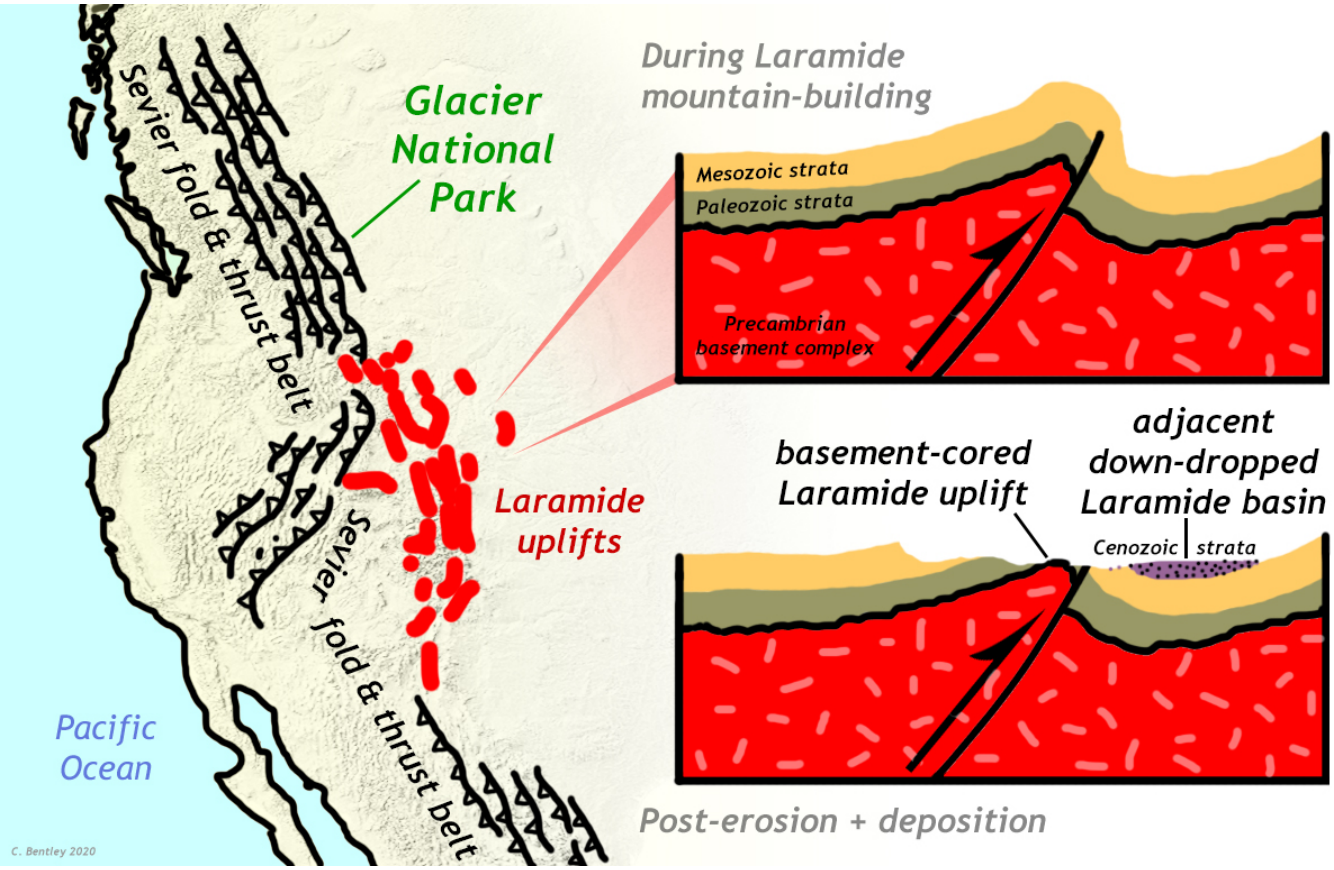


Figure 62.6.7: Map showing the geography of the Sevier fold and thrust belt relative to Laramide uplifts. Cross-sectional views show the structure of the basement-cored Laramide uplifts.

Laramide deformation is not in the same place as Sevier deformation. Generally, Laramide deformation is further east. The Black Hills of South Dakota, the Bighorn Mountains of Wyoming, and the Colorado Front Range are all examples of basement-cored Laramide deformation. In contrast, the Sevier fold and thrust belt is further west, in a sinuous arc that runs from the Alberta/British Columbia border through western Montana, Wyoming, Utah, and Nevada, and all the way to Chihuahua, Mexico. The Sevier orogen is a much longer feature, too running from Canada through the U.S. to Mexico, while the Laramide uplifts are all clustered in the mountain west of the United States alone.

With these distinctions between Laramide and Sevier deformation now clarified, we will return our attention fully to the Sevier, which is the only orogeny recorded in the structures and strata of the Glacier National Park region.

Did I Get It? - Quiz

? Exercise 62.6.1

The Laramide Orogeny was _____ in its timing than the Sevier Orogeny.

a. earlier

b. later

Answer

b. later

? Exercise 62.6.2

The Laramide Orogeny was _____ in its location than the Sevier Orogeny.

a. further east

b. further north

c. further south

d. further west

Answer

a. further east

? Exercise 62.6.3

The Laramide Orogeny was _____ in its style, in comparison to the Sevier Orogeny.

a. "thick-skinned" (involving basement rock)

b. "fish-skinned" (because the scaly nature of the overlapping thrust sheets)

c. "thin-skinned" (only involving sedimentary cover layers

Answer

a. "thick-skinned" (involving basement rock)

Thrust faulting

Reverse faults are faults that form in compressional settings, with the fault's upper block of rock (the "hanging wall") moving up relative to the lower block of rock (the "footwall"). The motion of the hanging wall over the footwall of a reverse fault creates a repetition of strata in vertical sequence. Thrust faults are reverse faults that dip at a low angle. Some are pretty much horizontal! There are two basic thrust fault varieties in the vicinity of Glacier National Park: (1) an imbricated stack of relatively small thrusts, such as at Sun River Canyon, placing Mississippian carbonates on top of Cretaceous siliciclastics, and (2) large coherent thrust sheets of Belt Rock, again climbing on top of Cretaceous siliciclastics, such as (2a) the Eldorado Thrust at Little Prickly Pear Canyon and (2b) the Lewis Thrust at Glacier National Park itself.

The geologic map below gives some context to these different styles of thrust faulting along the Rocky Mountain front. At Glacier National Park, Belt rock (lavender colored) is thrust atop Cretaceous siliciclastic strata (green) along the Lewis Thrust. Notice how the outcrop pattern of that lavender rock makes a northeastward-directed "bulge." This is called a "salient." South of that location, at Sun River Canyon, west of Augusta, the Belt rock is much further west of the mountain front. Instead, at Sun River Canyon (and also at Swift Dam) the leading edge of the fold and thrust belt shows Mississippian carbonates (blue on the map) are thrust atop Cretaceous siliciclastics (green) along dozens and dozens of smaller thrust faults. Further south still, at Little Prickly Pear Canyon, another eastward salient of Belt rock is thrust atop Cretaceous siliciclastics along the Eldorado Thrust.

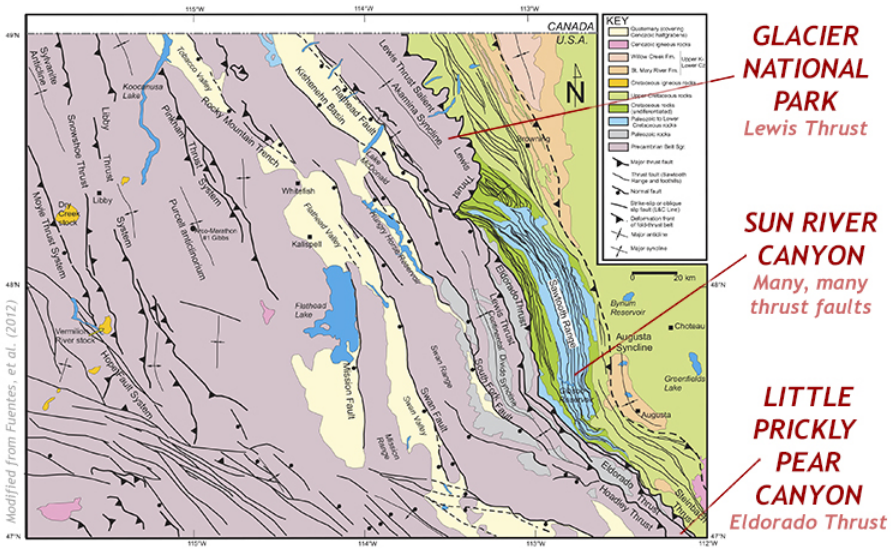


Figure 62.6.8: A modified version of the structural map of northwestern Montana that appears as Figure 3 in Fuentes, et al. (2012). Major thrust faulting locations examined on this VFE are highlighted in red at right.

Where the Mississippian carbonates appear, they appear between the Belt rock and the Cretaceous rock, a reflection of their original stratigraphic order. Where the Mississippian carbonates do not appear, the displacement of the Belt rock is greater. Tectonic transport has displaced it further, overriding the Mississippian package completely, and coming to rest atop much younger strata.

The Disturbed Belt at Sun River Canyon

West of Augusta is Sun River Canyon, a terrific place to view an imbricate stack of thrust faults from the Sevier Orogeny. Collectively, the mountains around Sun River Canyon are called the Sawtooth Range. Let's begin with a Google Maps overhead view:

Each of the north-south-trending white stripes in that Google Map are mountain ridges that are held up by Mississippian carbonates. In between each of them, the dark green valleys are floored by Cretaceous shales and sandstones. The plains at the right (east) side of the map are also Cretaceous siliciclastic sediments from Western Interior Seaway.

In this geologic map, showing roughly the same field of view, the Mississippian-aged carbonate rocks are shown in blue and purple colors, while the Cretaceous siliciclastics are shown in green and brown. Separating them (and oftentimes within them, too!) are thrust faults, marked as black lines with triangular “sawteeth” coming off them to the west (the hanging wall side of the fault):



In cross-section, east-west along the line of Sun River's course, we can see dozens of undulating west-dipping thrust sheets, with the blue/purple Mississippian carbonates underlying the many mountain ridges and green/brown Cretaceous siliciclastic strata cropping out in the intervening valleys (colloquially called "gulches" here):

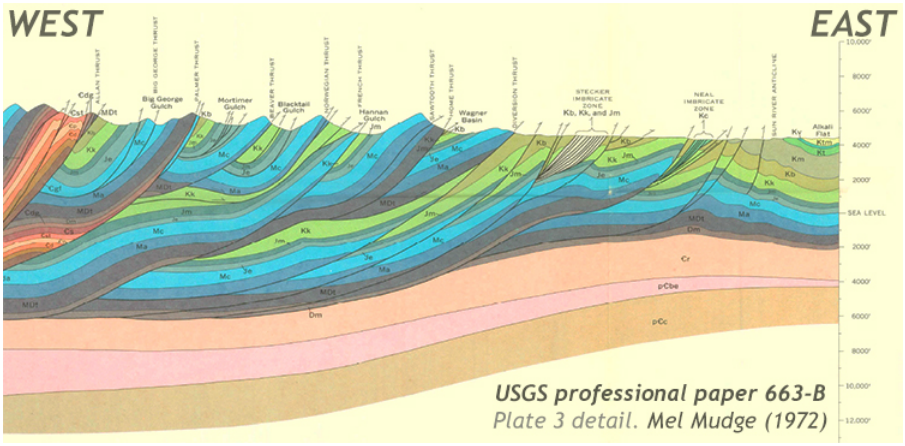


Figure 62.6.9: A geologic cross-section through the strata exposed at Sun River Canyon, from Mudge (1972).

One thing that should jump out at your eye from this map and cross-section is that the strike of the strata and the trace of the faults are parallel. This tells us that the sedimentary layers themselves are controlling the style of deformation: strong, stiff layers are riding up atop weak, pliable layers. The tectonic compression is "accommodated" by displacement mainly along weak shale layers. The over-riding slabs of stiff rock are the layers of Mississippian carbonates.

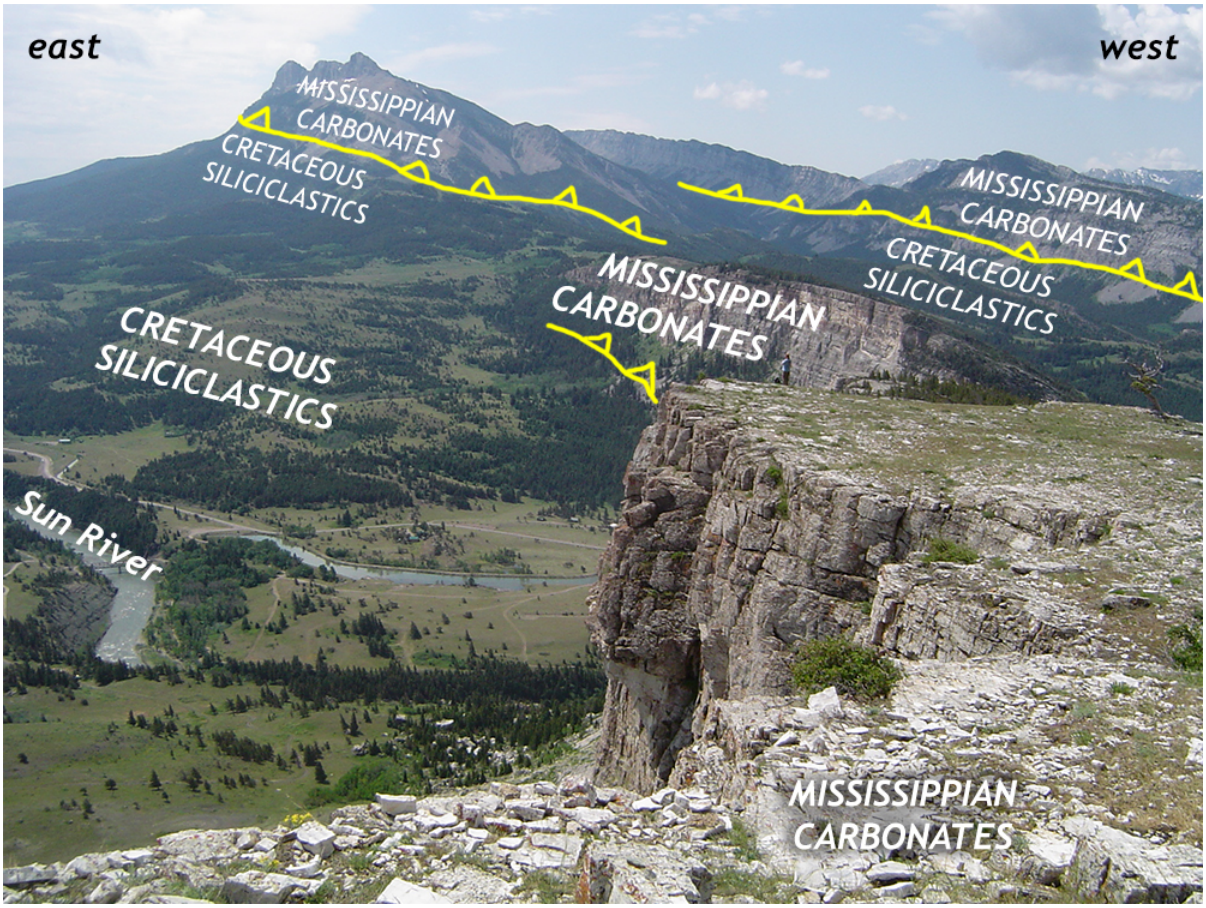


Figure 62.6.10: Photograph looking south across the mouth of Sun River Canyon. Photo by Matt Affolter.

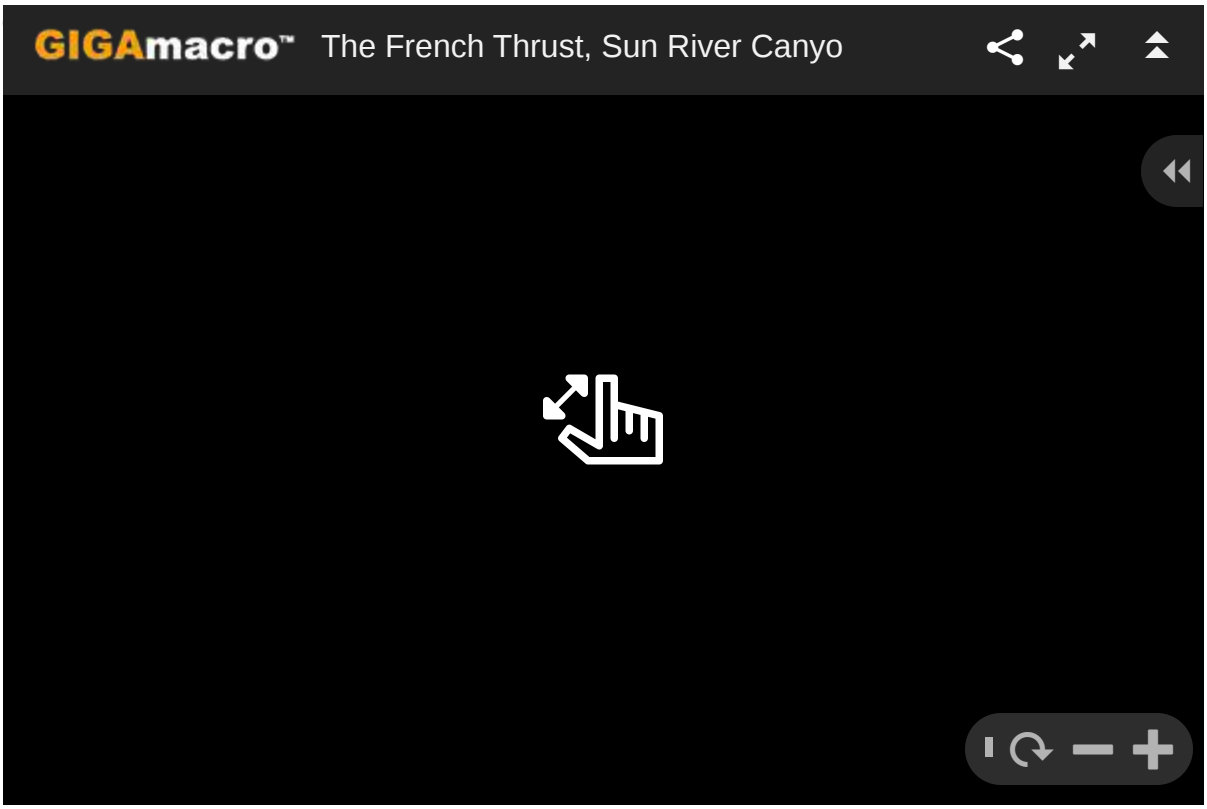
Structural geologists who have mapped these rocks along a 145-km-long transect across the mountain front have estimated that these rocks experienced about 135 km of tectonic shortening. In other words, the stack of sedimentary layers (Belt, Mississippian, & Cretaceous) has been horizontally compressed 135 kilometers east to west. This has resulted in vertical thickening, mainly via thrust faulting and stacking of stiff layers atop weak layers. That tectonic thickening has built the Rocky Mountain front. The mountain front is so topographically extreme for a related reason: Not only are the Mississippian carbonates stronger than the Cretaceous siliciclastics, they are also more stable in terms of weathering. They stand up tough to mechanical weathering and chemical weathering and erosion, and so form topographic "highs," resistant features in the landscape. In contrast, the Cretaceous shales and sandstones are readily removed by erosion, the shales even more rapidly than the sandstones. These strata form topographic "lows" and gently rolling terrain.

Most shortening in the Lewis Thrust system, Sawtooth Range, and foothills occurred roughly between late Cretaceous and Early Paleogene time (ca. 75–52 Ma), yielding a shortening rate of ~5.9 mm/yr. Even though these mountains are very impressive today, they grew slowly: your fingernails grow faster than that!

Here is an example of some imbrication in the Cretaceous siliciclastic strata immediately east of the mountain front at Swift Dam. The sandstone layers get repeated along a set of parallel thrust faults:



One of the best places to examine the fault contact between the two different rock units in the Sawtooth Range is where the French Thrust crosses the road in Sun River Canyon. Here is a GigaPan that allows you to explore this contact. At the left (east) is Cretaceous black shale of the Kootenai Formation, and at the right (west) is Mississippian carbonate of the Castle Reef Dolomite.



(It should be noted that although our focus at the moment is on the structural geology of this site, the outcrop above also features two surficial sedimentary deposits: till (from glaciers) and colluvium (from mass movement), both at the left (east) side of the image. Can you tell which is which? Check your answers with this field sketch, drawn on-site last year.)

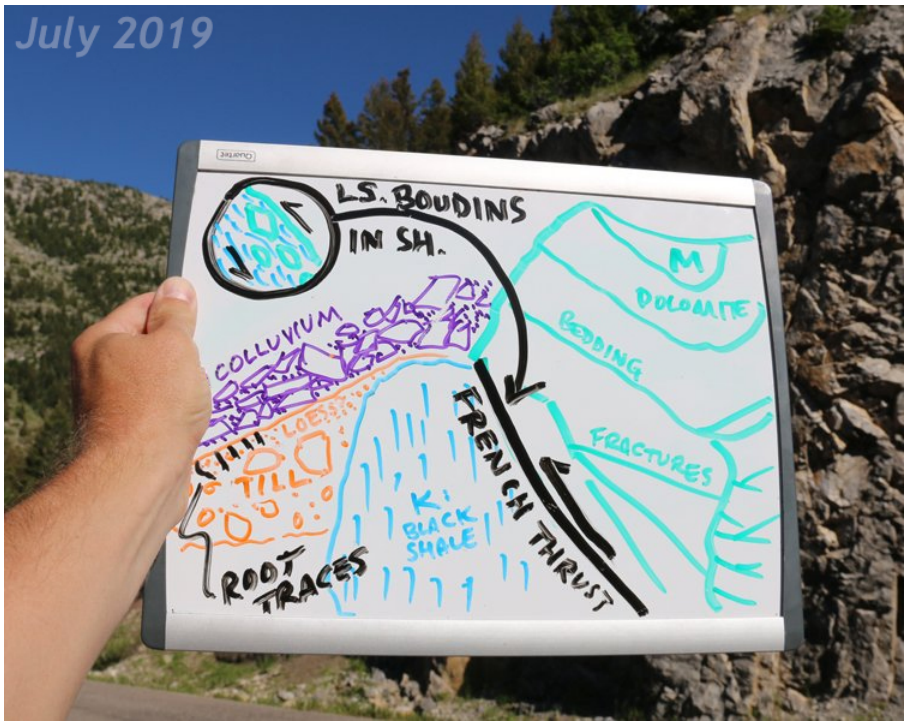


Figure 62.6.11: A field sketch of key geologic aspects of the French Thrust as it crops out roadside in Sun River Canyon.

Swift Dam

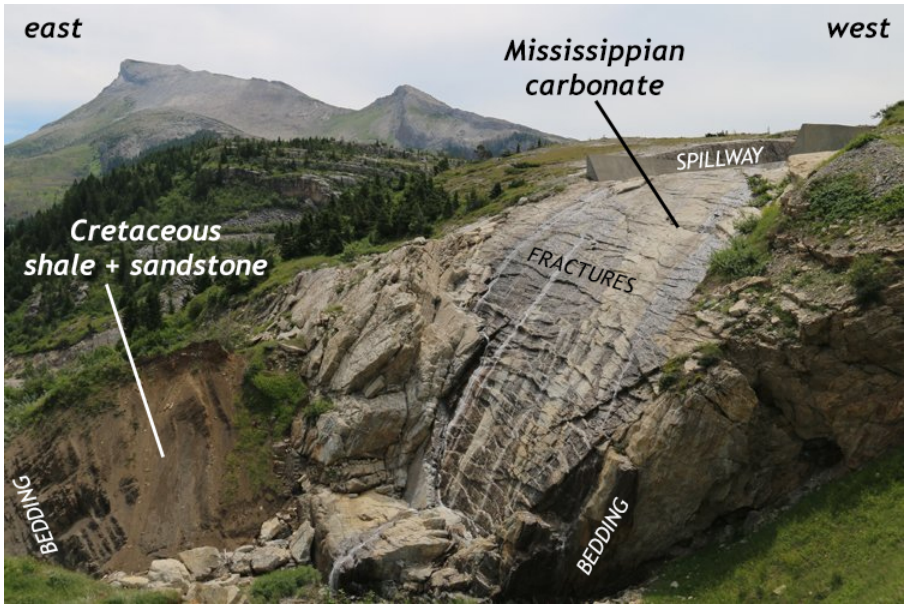
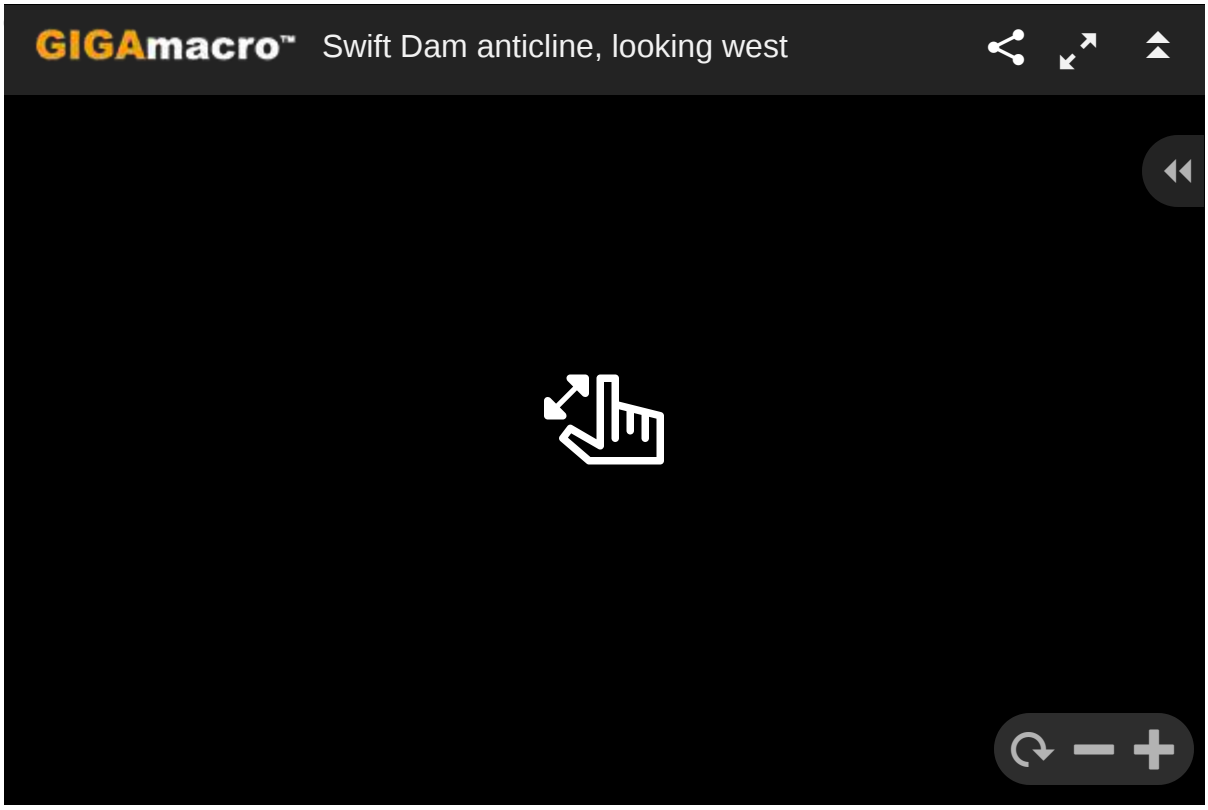


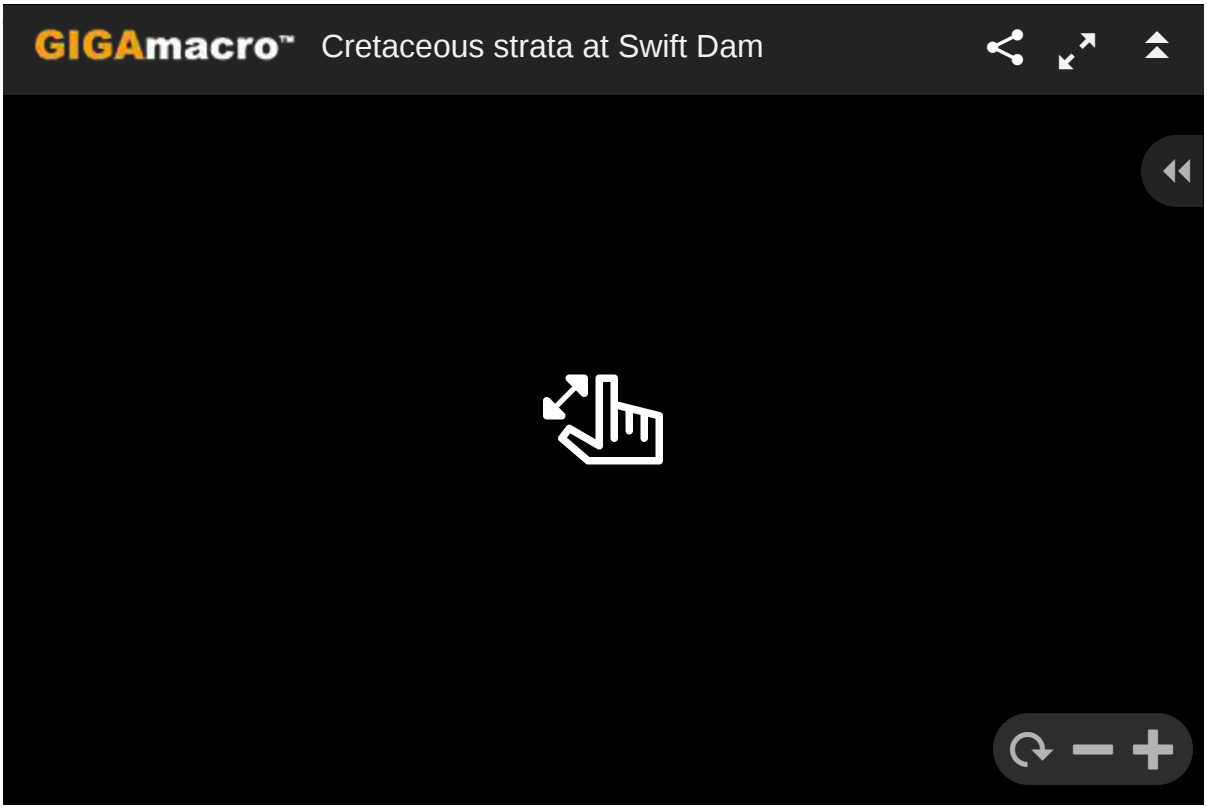
Figure 62.6.12: Swift Dam's spillway showcases a 3D view of the contact between folded Mississippian carbonate and Cretaceous siliciclastic strata. This is the steep, short, eastern limb of the fold.

Swift Dam is a nearby site that adds perspective to this deformation. There, you can see the "nose" (hinge) of the lowermost (newest) anticline bulging out of the mountain front. Cross-sectional views are available both looking north from Swift Dam itself and looking south from near the Swift Reservoir spillway. The spillway actually drains over top of it! The gently curving shape of the overturned anticline is plainly exposed. Additionally, many fractures cross the carbonate rock there (the Castle Reef Dolomite) in an orientation that is parallel to the hinge of the fold. Just to the east of the front of this anticline "nose" are decent exposures of

the strata it has ridden over: the Cretaceous-aged shales and sandstones of the Kootenai Formation and the Blackleaf Formation. Let's look in detail at both sides of this contact:



The weaker siliciclastic strata of the Cretaceous package are intensely sheared and folded on smaller scales. If you climb the dolostone of the Swift Dam spillway and look eastward, this is your view of these sheared footwall sandstone and shale:



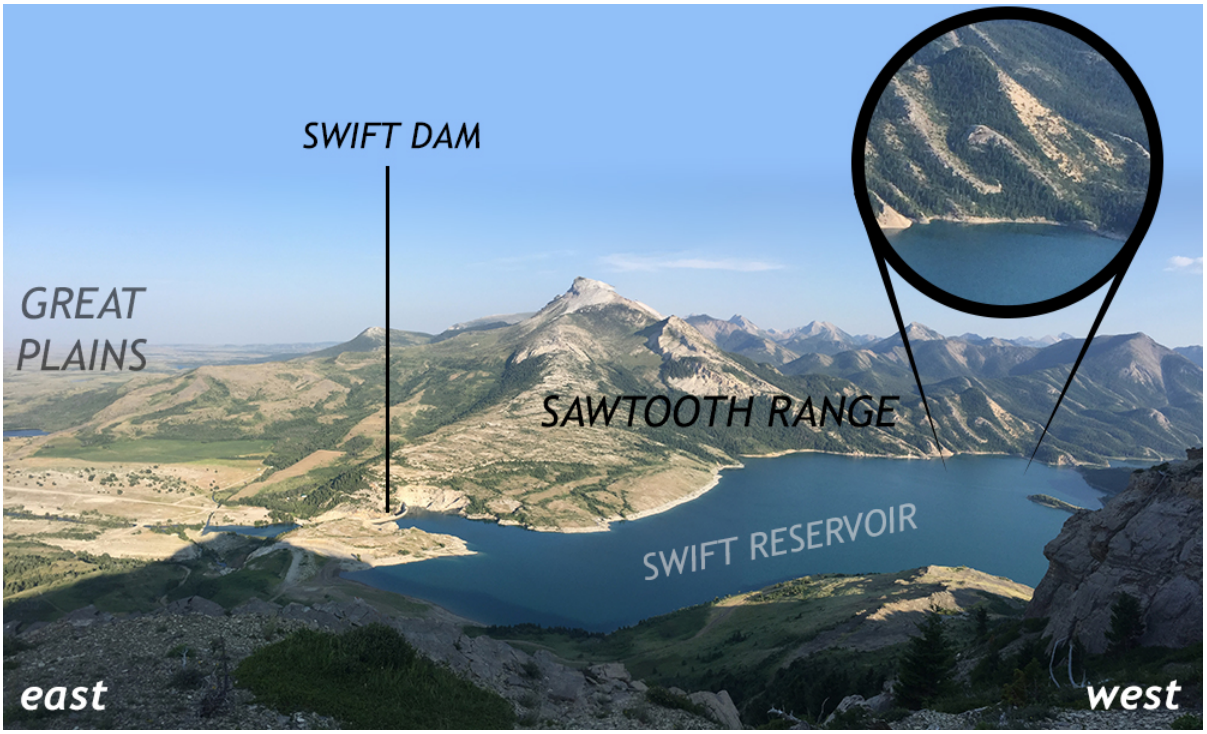


Figure 62.6.13: A view looking south over the mountain front at Swift Dam. Note the “yin-yang” pair of ridges highlighted to the west. Photo by Matthew Koenig.

Zooming out a bit shows that, as at Sun River Canyon, there are many fold/fault pairs at Swift Dam. As you head west into the Sawtooth Range, you encounter more of the same, repeated packages of Mississippian carbonates asymmetrically folded and thrust over Cretaceous shale. As at Sun River Canyon, the dip angle of the thrust sheets gets steeper to the west.

Below is a geologic map of the Swift Dam area: Examining it, you can see the lavender Mm (the abbreviation for the Mississippian Madison Group carbonate; specifically the Castle Reef Dolomite) thrust over the olive Kk and light yellow Kb (Cretaceous Kootenai Formation and Blackleaf Formation, respectively). Note the many northwest-southeast trending faults, again with the black triangular “sawteeth” on the west side, indicating (a) that is the hanging wall and the fault dips to the west, meaning (b) that the compressional force came from the west.



The Eldorado Thrust

The Eldorado Thrust is the name for the fault at the leading edge of the southern Belt salient in our region. (The Lewis Thrust is the fault at the leading edge of the northern Belt salient.) Driving Montana 287 (also Interstate 15) through Little Prickly Pear Canyon takes vehicular traffic through the Eldorado thrust sheet between Helena and Wolf Creek.

In this geologic map, of a location just west of Little Prickly Pear Canyon, you can see the Eldorado Thrust prominently labeled, placing Belt rock (pЄes, the Empire Formation) atop various Cretaceous siliciclastic and volcanic units.

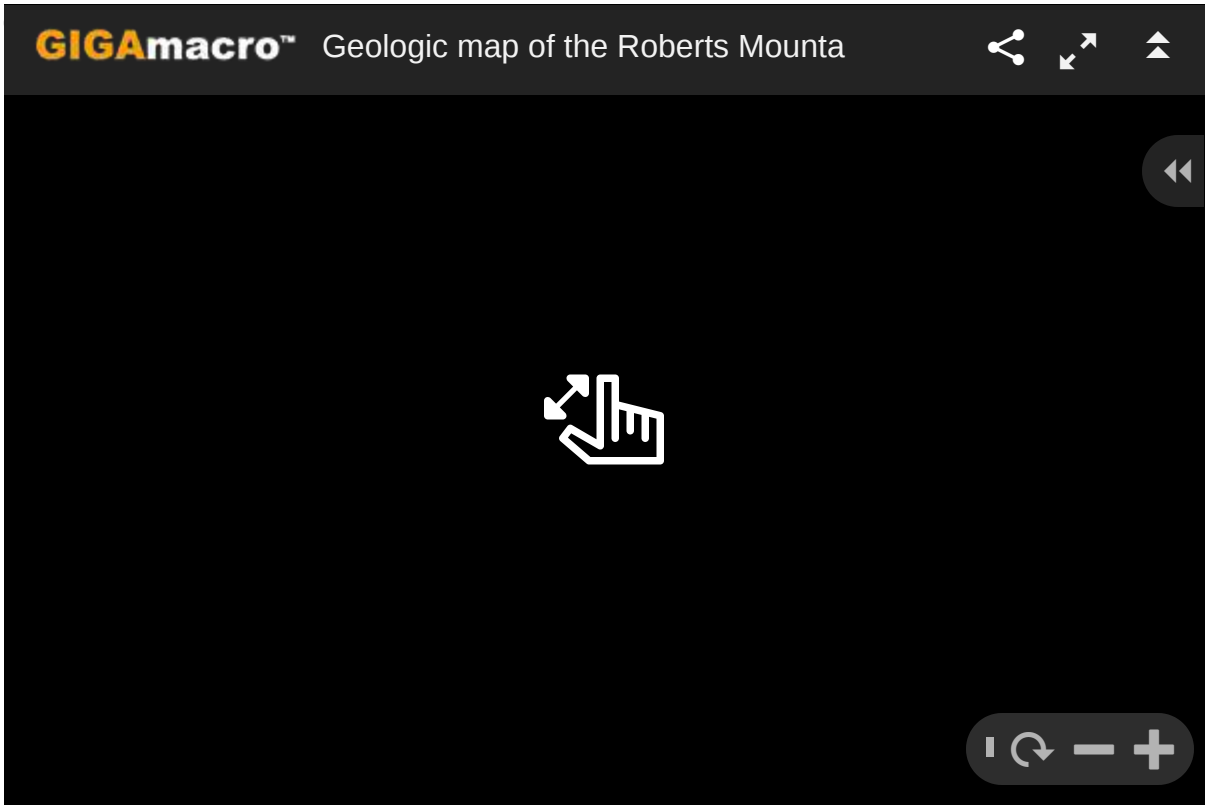


Figure 62.6.14: Schmidt, R.G., and Strong, C.P., 1972, Geologic map of the Roberts Mountain quadrangle, Lewis and Clark County, Montana: U.S. Geological Survey, Geologic Quadrangle Map GQ-977, scale 1:24,00. Explore more here.

Driving north from Helena, one crosses a normal fault at the southern end of Little Prickly Pear Canyon, and enters the hanging wall of the Eldorado Thrust, a southern equivalent to the famous Lewis Thrust that underlies Glacier National Park. The whole drive along route 287 through this canyon is spectacular: cliffs of maroons and mint green rise on either side of the road. Check it out in this Google Maps "Streetview," which you can manipulate and explore:

In places, additional structures can be observed, like smaller-scale faults...

https://geo.libretexts.org/@go/page/22905

And intrusive dikes:

The Lewis Thrust



Figure 62.6.15: “The Head,” a mountain ridge south of the Two Medicine Valley, shows a thrust fault where an upper package of Belt sedimentary rock (at an angle dipping to the right) has been shoved atop a lower package of Belt sedimentary rock (that lies close to horizontal).

This brings us finally to Glacier National Park itself, where the Lewis Thrust has emplaced an even more coherent hanging wall thrust sheet atop a footwall package of sheared-out siliciclastic strata of the Western Interior Seaway. The mountains of the park are

the Lewis Range, from which the fault takes its name, and the mountains were so dubbed in honor of Meriwether Lewis, explorer and co-leader of the Corps of Discovery expedition.

Though the Belt sequence is more or less intact throughout Glacier National Park, smaller-scale deformation can be seen, in particular along the eastern edge of the Lewis Thrust sheet. There, recumbent folds and subsidiary faults can be seen.



Figure 62.6.16: Spot Mountain, north of the Two Medicine Valley, shows a recumbent fold on its eastern edge.

As at Little Prickly Pear Canyon with the Eldorado Thrust, the principle of superposition has been upended, with Mesoproterozoic Belt strata coming to rest atop much younger Cretaceous aged rock. However, Glacier is much more impressive for the exposed thickness of the Belt rocks in the hanging wall that are exposed, as well as their impressive glacial sculpting, a later modification that is missing at Little Prickly Pear Canyon.

Did I Get It? - Quiz

? Exercise 62.6.1

Which thrust fault places Mississippian carbonates over Cretaceous shales?

a. the Lewis Thrust at Glacier National Park

b. the Eldorado Thrust at Little Prickly Pear Canyon

c. the French Thrust at Sun River Canyon

Answer

c. the French Thrust at Sun River Canyon

? Exercise 62.6.2

These layers of Grinnell Formation quartzite and red argillite were compressed from which direction?



a. from the right, judging by the color of the sediment

b. from the left, judging from the asymmetry of the fold

c. from the top, judging by the offset on the fold

d. from the bottom, judging by the Swiss Army knife

Answer

b. from the left, judging from the asymmetry of the fold

Exercise 62.6.3

Which of the following allows for an exception to the stratigraphic principle of superposition?

a. weathering

b. eruption of lava

c. erosion

d. burial metamorphism

e. thrust faulting

Answer

e. thrust faulting

Exercise 62.6.4

What is distinctive about the outcrop at Swift Dam?



a. It is the only nappe that places Mississippian **carbonate** atop Belt rock.

b. It is the only **Laramide** thrust in the Sawtooth Range.

c. It shows the "nose" of a nappe in **three dimensions**, whereas this brittle leading edge of the overturned anticline is eroded away in most other Sevier range-front thrust faults.

d. It is among the **oldest** faults in the region

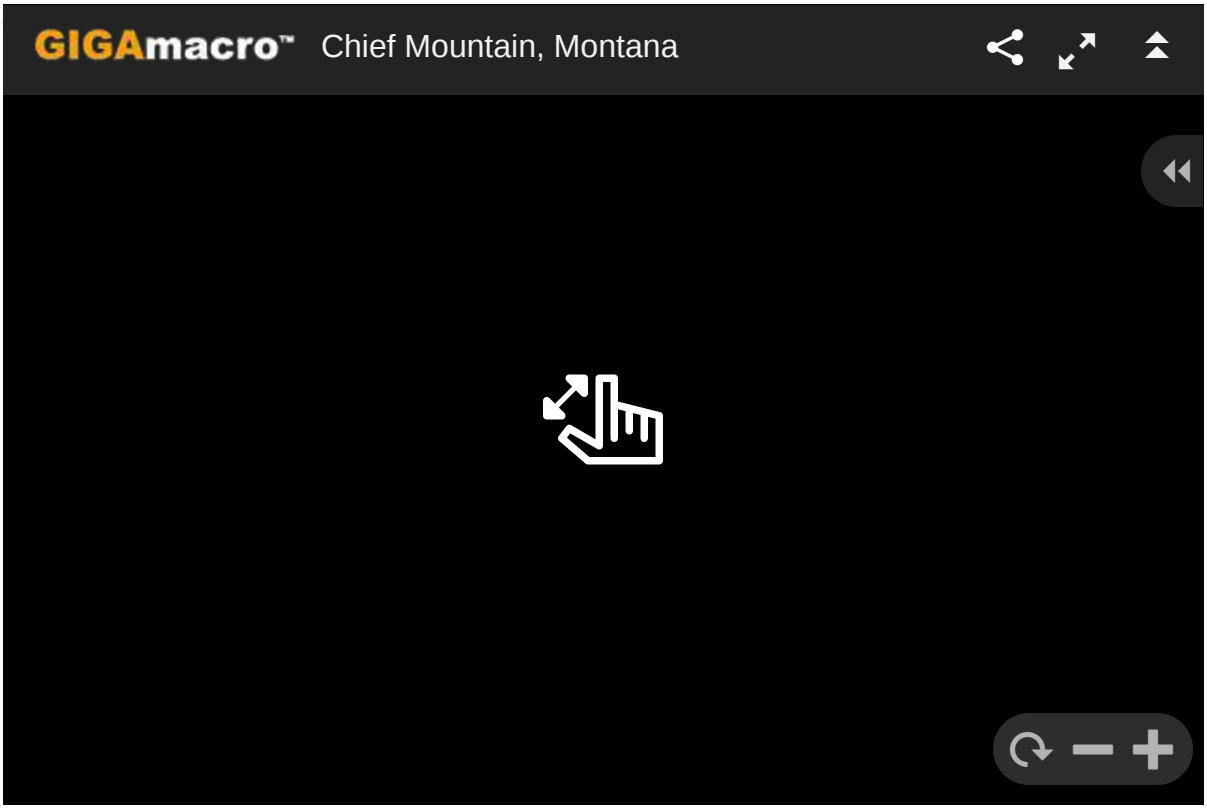
Answer

c. It shows the "nose" of a nappe in **three dimensions**, whereas this brittle leading edge of the overturned anticline is eroded away in most other Sevier range-front thrust faults.

Klippen

When a thrust sheet such as the Belt rocks above the Lewis Thrust is subject to erosion, sometimes it gets sufficiently chewed away so that only small remnants are left, like “islands” surrounded by the footwall rocks below the thrust. This situation results in a feature called a klippe. If you have more than one, they are *klippen*, plural.

The classic “textbook” example of a klippe is Chief Mountain, on the park’s northeastern boundary with reservation lands of the Blackfeet tribe:



The rolling, vegetation-covered slopes in the foreground are Cretaceous strata below the Lewis Thrust, mantled with a thin covering of Quaternary-aged landslide deposits. The Lewis Thrust itself is almost perfectly horizontal here, and runs across the base of Chief Mountain, cropping out at the bottom of the cliffs. Above the Lewis Thrust are Mesoproterozoic Belt rocks of the Altyn Formation. There's actually another ~horizontal thrust running through Chief Mountain, too, about 2/3 of the way up. It can be detected by looking at the change in angle of the Altyn strata within the mountain: horizontal at the top, dipping moderately northward at the bottom.

Here is what that looks like on the geologic map, with Altyn Formation (brown, Ya) sitting atop Cretaceous strata (green, Ku), topped with landslide deposits (orange, Qls).

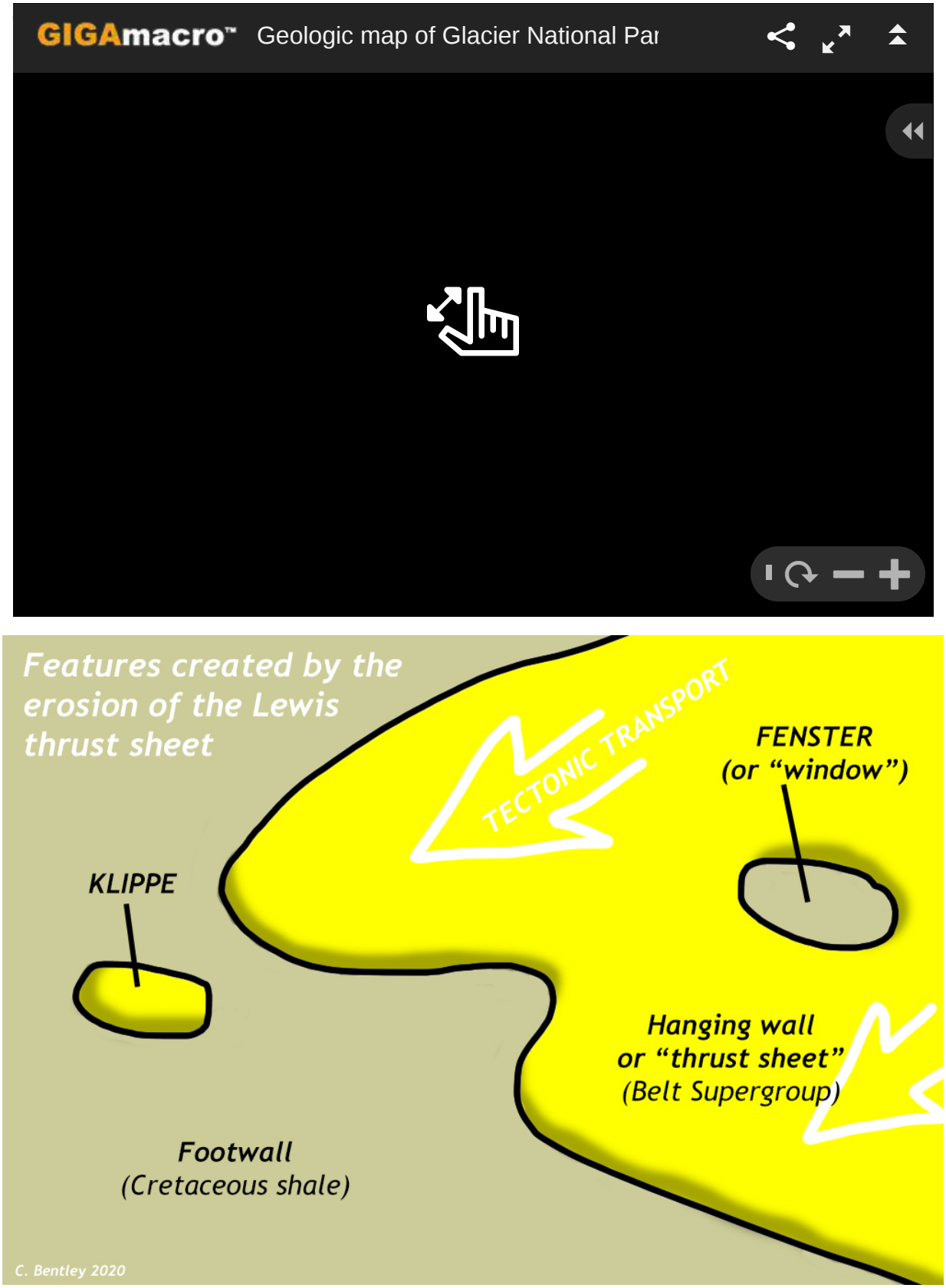


Figure 62.6.17: Oblique view of a thrust sheet, showing how erosion can generate isolated klippen and also tectonic "windows" through the thrust sheet. Yellow = hanging wall rocks (Belt, at Glacier NP); Gray = footwall rocks (Cretaceous siliciclastics at Glacier).

Klippen form when erosion removes the connection between a salient and the rest of the thrust sheet. In the case of Chief Mountain, the erosion was accomplished by glaciers, which flowed here during the Pleistocene, gouging away at the Cretaceous-aged thrust sheet, including its Mesoproterozoic-aged Belt rocks and the Cretaceous siliciclastic strata that comprised the footwall of the fault. It's also possible for erosion to remove some of the hanging wall rock so a "window" (German: *fenster*) is opened up, and a small zone of footwall rock peeks out from beneath the thrust sheet. Klippen and fensters are "opposite" features of one another.

Did I Get It? - Quiz

? Exercise 62.6.1

An erosional hole through a thrust sheet so that the footwall of the thrust fault is exposed is a _____.

a. orogeny

b. nappe

c. fenster

d. klippe

Answer

c. fenster

? Exercise 62.6.2

The Acropolis in Athens, Greece, is a hill made of ~93 Ma limestone sitting atop ~68 Ma shales in the surrounding lowlands. The contact between the two geologic units is a thrust fault, making the Acropolis a(n) _____.



a. interdimensional portal

b. klippe

c. syncline

d. fenster

e. window

f. anticline

Answer

b. klippe

Folding on several scales

Overall, the Belt rocks of Glacier have been folded into a vast trough-like fold, the Akamina Syncline. This results in the overall southwest-to-northeast symmetry of the outcrop pattern across the park: pinks and lavender in the middle, flaked by maroon, then

olive green, then pumpkin orange and brown headed out in either direction:



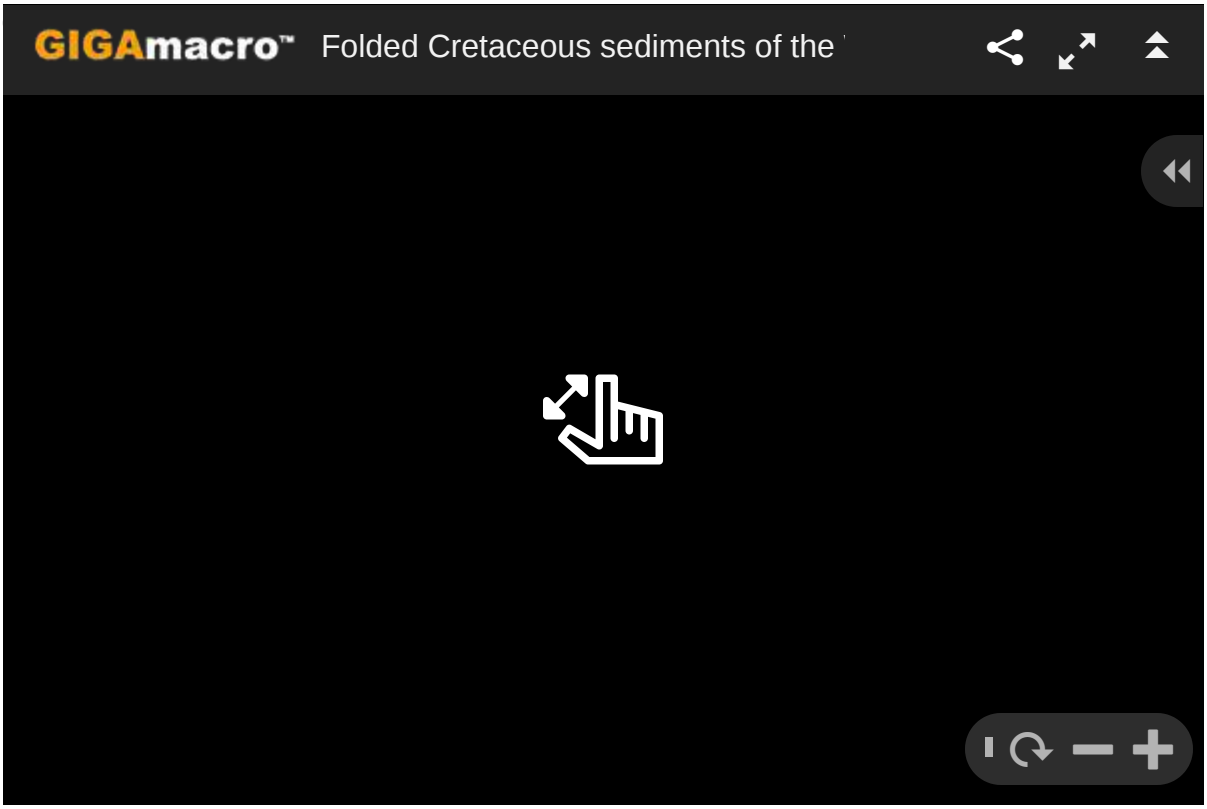
It is a fold with a wavelength of more than 20 km! That's big.

The axis of that syncline (the line of maximum inflection) passes under some of the tallest peaks in the park, and is highlighted by outcrops of the youngest Belt strata. It is reasonable to infer that this same syncline extends through all the Belt strata down to the Lewis Thrust itself at the bedding-defined base of the hanging wall.

That is the interpretation shown on the Whipple (1992) map, which shows a scoop-shaped profile to the fault, reaching its deepest nadir beneath the axis of the syncline:



Beneath the Lewis Thrust, in the footwall rock, there's plenty more folding to be observed, but it occurs on much smaller scales: folds with wavelengths of km to m to cm, as the tectonic "banana peel" of shale and sandstone smeared out underneath the massive overriding Lewis Thrust sheet. While the thick stack of burial-metamorphosed Belt rock was stiff and strong, these thinly-layered sandstones and shales were never buried deeply, and are quite weak. As a result, they were pulverized as thrusting brought the Belt rock eastward over their position. Good outcrops showing this deformation can be found along the Looking Glass Highway, north of East Glacier:



Did I Get It? - Quiz

? Exercise 62.6.1

In the Glacier National Park area, an outcrop of deformed shale and sandstone showing folds with wavelengths of 10 cm to 10 m is likely to come from _____.

a. above the Lewis Thrust, in the Akamina Syncline

b. below the Lewis Thrust, is sheared-out Cretaceous strata

c. Mars

Answer

b. below the Lewis Thrust, is sheared-out Cretaceous strata

Conclusion

Tectonic compression impacted the sedimentary strata at Glacier National Park during the Cretaceous due to the Sevier Orogeny. The accretion of exotic terranes at a subduction zone far to the west shoved extant rock layers out of the way, piling them up on one another to the east. Older rock, such as Belt Supergroup strata and Mississippian carbonates, climbed atop younger rock, in particular the siliciclastic strata laid down by the Western Interior Seaway during the Cretaceous. The structure of the rock layers changed from a “layer cake” with oldest at the bottom and youngest at the top into a new arrangement dominated by asymmetric folds and thrust faults, with nappes of ancient rock transported eastward and upward by astonishing distances. In places where the Belt rock made it to the surface (Glacier National Park and Little Prickly Pear Canyon), we find more or less coherent thrust sheets, though later erosion has carved klippen out from the leading edge in several places. In places where Mississippian carbonates were instead the rocks on the leading edge (Sun River Canyon and Swift Dam), we instead find a densely-imbricated stack of smaller thrust sheets, older and steeper on the west, and younger and more shallowly dipping on the east.

Further reading

Anne DeVaney (n.d.) "Structural Geology of the Rocky Mountain Front," *Montana Geoheritage Project*, hosted on the SERC website.

Facundo Fuentes, Peter Decelles, and Kurt Constenius. (2012).

David R. Lageson (1987). "Structural geology of the Sawtooth Range at Sun River Canyon, Montana Disturbed Belt, Montana," in Beus, S.S., ed., *Centennial Field Guide Volume 2: Rocky Mountain Section*, Geological Society of America, p. 37-39.

Omar B. Raup, Robert L. Earhart, James W. Whipple, and Paul E. Carrara (1983). *Geology along Going-to-the-Sun Road, Glacier National Park, Montana*. Glacier Natural History Association.

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62.7: Glacier National Park, Montana - Glacial geomorphology

During the Pleistocene epoch, Earth experienced an overall cooler than average climate, which alternated between glacial and interglacial conditions several dozen times. Repeated oscillations between ice-building and ice-melting conditions correlate with variations in Earth's orbit around the Sun. In the northern hemisphere, both mountain glaciers and ice sheets have waxed and waned in response. In some places, the flow of these glaciers has sculpted the landscape through erosion, and modified it elsewhere through the addition of glacially-transported sediment.

Examine this physiographic map of the region to see the distinctive topography that results from the action of these moving "rivers of ice" on a mountain landscape:

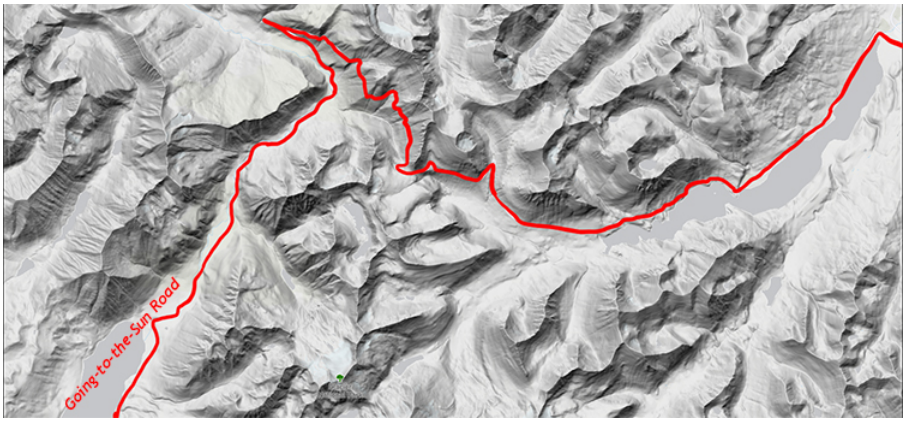


Figure 62.7.1: Alpine glacial geomorphology in the Logan Pass region between Lake Macdonald and St. Mary.

If you click on that map, it will take you to a dynamic map where you can zoom in and out and explore the topography in more detail.

Pleistocene glaciation

The last 2.1 million years of geologic time have been cold relative to the bulk of global climate history. Starting shortly after the Isthmus of Panama formed, diverting the Gulf Stream northward in the Atlantic, and oscillating in rhythm with slight variations in Earth's orbit around the Sun, the Pleistocene epoch is characterized by repeated advances and retreats of glacial ice, particularly pronounced in the northern hemisphere. Nominally, the Pleistocene ended 10,000 years ago with the start of the Holocene, but from a zoomed-out perspective the Holocene looks an awful lot like the 30 or so interglacial periods that preceded it, every 100,000 years or so, throughout the Pleistocene.



Figure 62.7.2: Glaciers create new landforms by modifying the landscape into new shapes (sculpting) to make erosional landforms and also by adding the resulting eroded-away particles to other places on the landscape, in piles (depositional landforms). (Callan Bentley graphic)

The glaciers of the Pleistocene were important agents of change for the landscape. They reshaped it in two major ways: (1) by removing rock from some areas (this scraping generated particles we call sediment) and (2) by depositing that sediment somewhere else. There are glacial landforms that result from the removal of rock and other glacial landforms that result from the deposition of sediment in another location.

Erosional landforms of alpine glaciation

Processes of glacial erosion

Above the high elevations of the mountains, air masses are forced upward, which cools them and triggers condensation and deposition (the processes by which water vapor turns to liquid water or ice). All else being equal, precipitation falls in greater amounts over mountains than surrounding lowlands. High elevation land is also cooler than low elevation land, and any snow that falls there is more likely to persist. As a result, snow tends to accumulate in temperate mountain regions. During an ice age, the amount of snow that builds up is greater than the amount of snow that melts away. This is the precursor to glacial ice.

When the snow gets to be about 50 m (150 feet) thick, the pressure of the overlying snow compacts the snow below into ice, and this ice begins to flow. The flow is a solid-state flow, akin to putty or warm wax. Gravity pulls the ice mass downward, and so the glacier begins to flow downhill. It will utilize pre-existing drainages (valleys) on its path downhill. Moving downhill into open air is easier than moving downhill into solid rock, so the pre-glacial valleys end up being retained as pathways even after they are filled with ice. This BBC video shows the process very well.

Glaciers erode the bedrock in two major ways: (1) abrasion and (2) plucking. Abrasion is a grinding, sandpaper-like process, where rock particles of many sizes are entrained in the glacial ice and scraped against the bedrock as the ice flows over it, gouging the solid rock on many scales. This gouging results in the liberation of new particles, which then also become agents of abrasion.

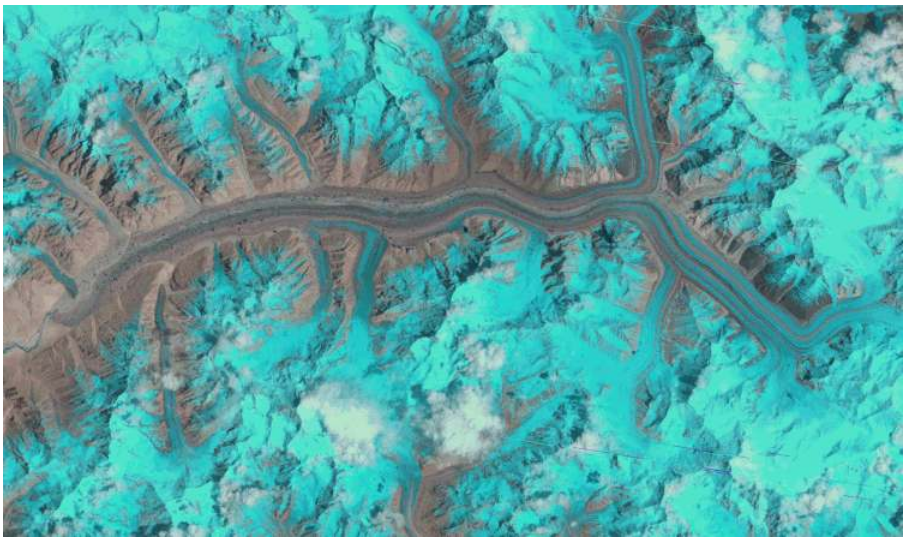


Figure 62.7.3: Alpine glaciers flow downhill through valley, and have tributary glaciers just like rivers do. Karakorum Range, Pakistan. (Animation from Landsat photography by F. Paul, USGS)

Plucking is a bit more complicated. Ice is a mineral, and when it is subjected to higher pressures, it dissolves (liquid water takes up less volume than ice does). When it encounters lower pressures, it refreezes. When glacial ice flows against some obstacle of bedrock, it dissolves on the upstream (uphill) side, flowing through cracks in the rock, and refreezes on the downstream (downhill) side. As it refreezes, it rejoins the glacial ice mass and potentially surrounds blocks of bedrock. In Belt strata, these blocks of bedrock are defined by the bedding plane and joints that are perpendicular to bedding. The ice-turned-high-pressure-water infiltrates these planes of weakness and, once refrozen, liberate the blocks from their original place.

Did I Get It? - Quiz

? Exercise 62.7.1

Which epoch saw the most recent episodes of widespread glaciation in Earth history?

a. the Pleistocene

b. the Permian

c. the Neoproterozoic

d. the Triassic

Answer

a. the Pleistocene

? Exercise 62.7.2

How do glaciers modify the landscape?

a. Both by carving away at existing rock outcrops (sculpting or erosion) and by adding sediment elsewhere (deposition).

b. Only by adding sediment elsewhere (deposition).

c. Only by carving away at existing rock outcrops (sculpting or erosion).

Answer

a. Both by carving away at existing rock outcrops (sculpting or erosion) and by adding sediment elsewhere (deposition).

? Exercise 62.7.3

What are the two principle methods of glacial erosion?

a. Abrasion and deposition.

b. Deposition and plucking.

c. Plucking and abrasion.

Answer

c. Plucking and abrasion.

The shape of the landscape

Abrasion and plucking change the shape of the mountain landscape they flow over. The glaciers gouge downward and outward, resulting in scoop-shaped valleys that show a U-shape in profile. This is in sharp contrast with the V-shaped valleys that rivers carve. When these valleys intersect in various ways, they generate a variety of distinctive landforms. When the glaciers melt away, the landforms are left as testimony to their former occupancy of the land.

U-shaped valleys



Figure 62.7.4: U-shaped valleys (sometimes called “troughs”) are characteristic of erosion by alpine glaciers. (Photo by Tom Biggs.)

Much of the grandeur in Glacier National Park comes from its extraordinary valleys, which are wide and then curve gently into near-vertical cliffs on either side. Unlike rivers, which focus their erosion downward in a relatively focused zone, and may be surrounded by broader floodplains, glaciers can erode on their bottoms simultaneously with erosion on their edges. While rivers cut V-shaped valleys, alpine (mountain) glaciers cut U-shaped valleys, and even after the glaciers are gone, the mountains retain this distinctive signature of their former presence.



Figure 62.7.5: U-shaped valleys like the Boulder Creek Valley (east of Siyeh Pass) are evidence of alpine glaciation. (Callan Bentley photo)

Cirques

Each U-shaped valley has a scoop-shaped head, as if an ice-cream scoop was plunged downward and outward from the upstream end of the glacier.

Arêtes

When two U-shaped valleys run parallel to one another, and gradually eat away at the bedrock ridge between them, a thin fin of rock is etched out. Glacier has many examples of arêtes, including the Garden Wall. Fusillade Mountain lurks prominently in the background of the most photographed scene in the park, Wild Goose Island. Viewed end-on, it might look at first like a horn, but it's really an arête, as a side-view shows!



Figure 62.7.6: Horns are pyramid-shaped mountain peaks, with each side being defined by a U-shaped valley. This is Reynolds Mountain, as viewed from the trail to Siyeh Pass. (Callan Bentley photo)

When several concave-out cirques point away from a central, isolated peak, it gives the peak a shape like a pyramid. There are dozens of examples of horns in Glacier National Park.

An excellent example is Flinsch Peak:

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Hanging Valleys

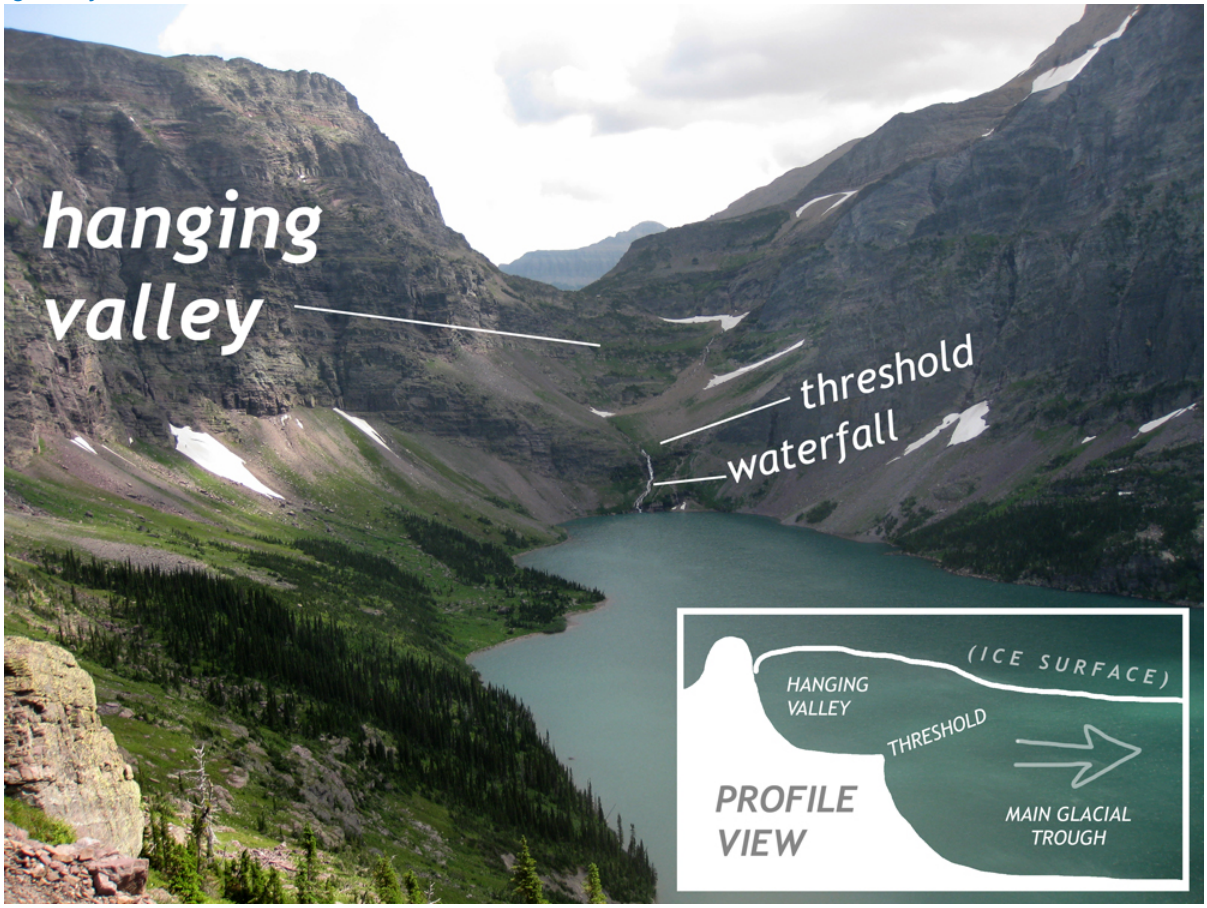


Figure 62.7.7: Hanging valleys are the former sites of tributary glaciers.

Like rivers of flowing water, glaciers have tributaries and main "trunk" ice streams. Unlike rivers, however, these tributary glaciers flow into the main glacier in a way that merges the two ice streams at the top rather than at the lowest point. This means that when the ice has all melted away, what's left is a tributary valley "hanging" above the main valley, where the valley floor of the tributary glacier is far higher in elevation compared to the main glacial trough. Any rain or meltwater collected by the hanging valley coalesces into a stream, which then flows downhill and over the "threshold" of the hanging valley, creating a waterfall. Bridalveil Falls in Yosemite National Park is a classic example of this phenomenon, but there are literally dozens of examples in Glacier National Park, too.

Roche Moutonnée

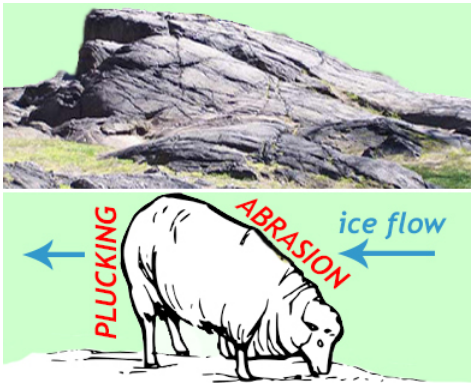


Figure 62.7.8: Roche moutonee translates to something like "rock sheep." The asymmetry of these bedrock knobs shows ice-flow direction, and resembles a grazing sheep.

Roche moutonnée are knobs of bedrock that have had a glacier slide over top of them. On the up-current side of the knob, the ice scours a gently-inclined ramp-like surface into the rock. On the down-current side of the knob, the glacial ice plucks large boulders from a steep downhill-facing cliff. The result is an asymmetric hill of solid rock, shaped something like a grazing sheep, with its head down in the grass and its rump poking up in the air. The name “roche moutonnée” translates from the French as something like “rock sheep” (or “rock ***mutton***“). With this mammal mnemonic in mind, remember that grazing sheep faces *up* the glacial valley, in the direction from which the ice originated.

Striations



Figure 62.7.9: Glacial striations on an outcrop of Helena Formation near Grinnell Glacier. (Callan Bentley photo)

Because the ice is constantly grinding directly against it, the upstream end of a roche moutonnée is likely to be covered in striations (sometimes called “striae”). These are nothing more than scratches, as cobbles and boulders entrained in the glacial ice are dragged slowly but firmly over the bedrock surface. These striations are therefore a record of glacial flow direction, and intersections between striation directions can indicate changes in glacial flow direction through time.

There are also some glacial lakes to be aware of:

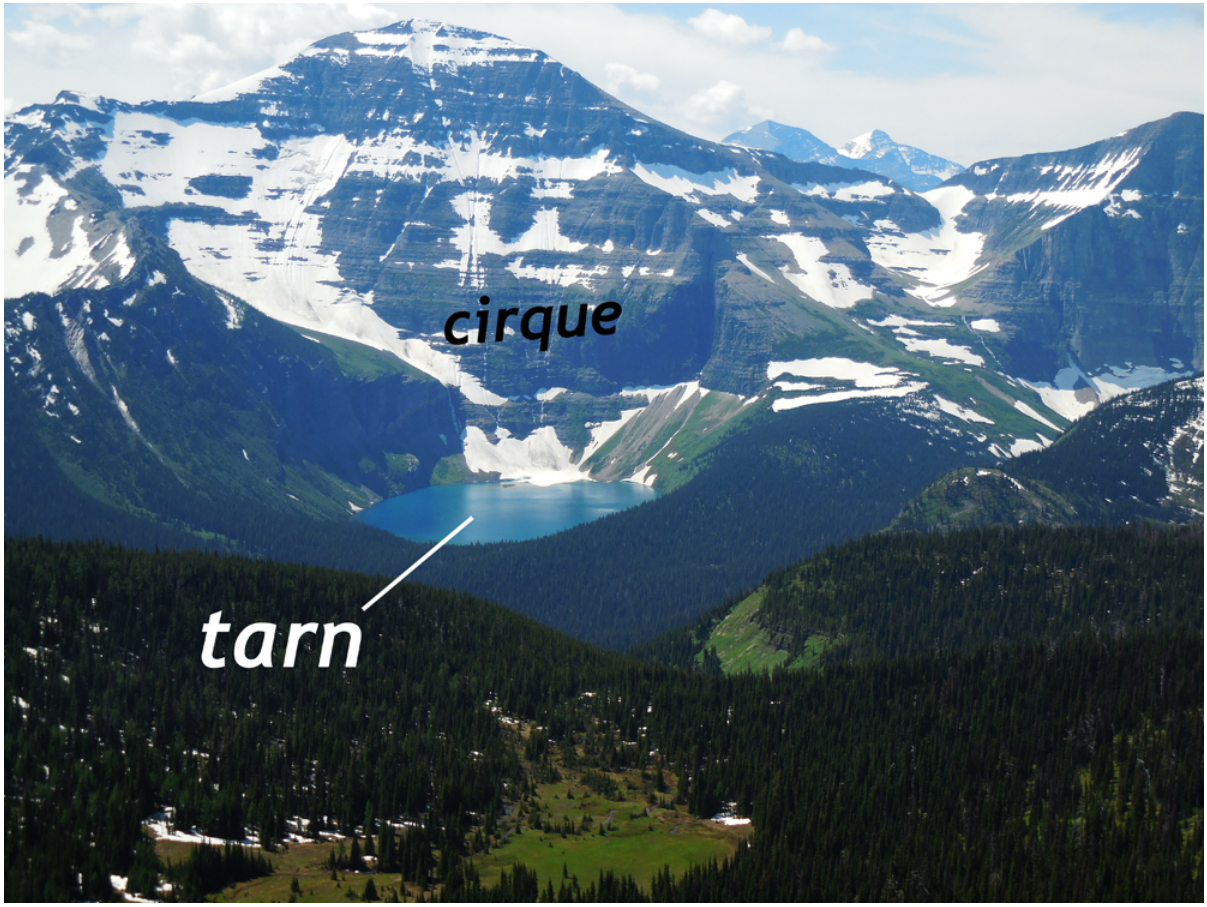


Figure 62.7.10: Tarns are ponds that form in deglaciated cirques. (Photo by Tom Biggs.)

Tarns are glacial lakes that sit at the bottom of a cirque. As glacial ice forms and flows downhill out of the cirque, it moves both down and out, rotating. This “scoops” out the cirque in a way that often leaves a low spot upstream of a threshold. The difference in topography can also be enhanced by the deposition of an end moraine across the edge of the cirque. If this depression in the landscape isn’t well drained, it will act to impound and retain water: a tarn.



Figure 62.7.11: “Pater noster” is the name given to “strings” of connected lakes in a linear drainage like this U-shaped valley.

Along the course of a glacial valley’s trough-like shape, there are various things that can stop and impound water, most particularly a series of recessional end moraines. Add in a tarn at the top of the valley, and you can have a series of lakes, each connected to the next one downstream by a single drainage stream. In map view, then, these lakes look something like beads on a string, resulting in their distinctive name. To some Catholic geologist, this pattern resembled a rosary, and they were dubbed after the first two Latin words of a common prayer refrain in that faith, *Pater noster*, which means “our Father.”

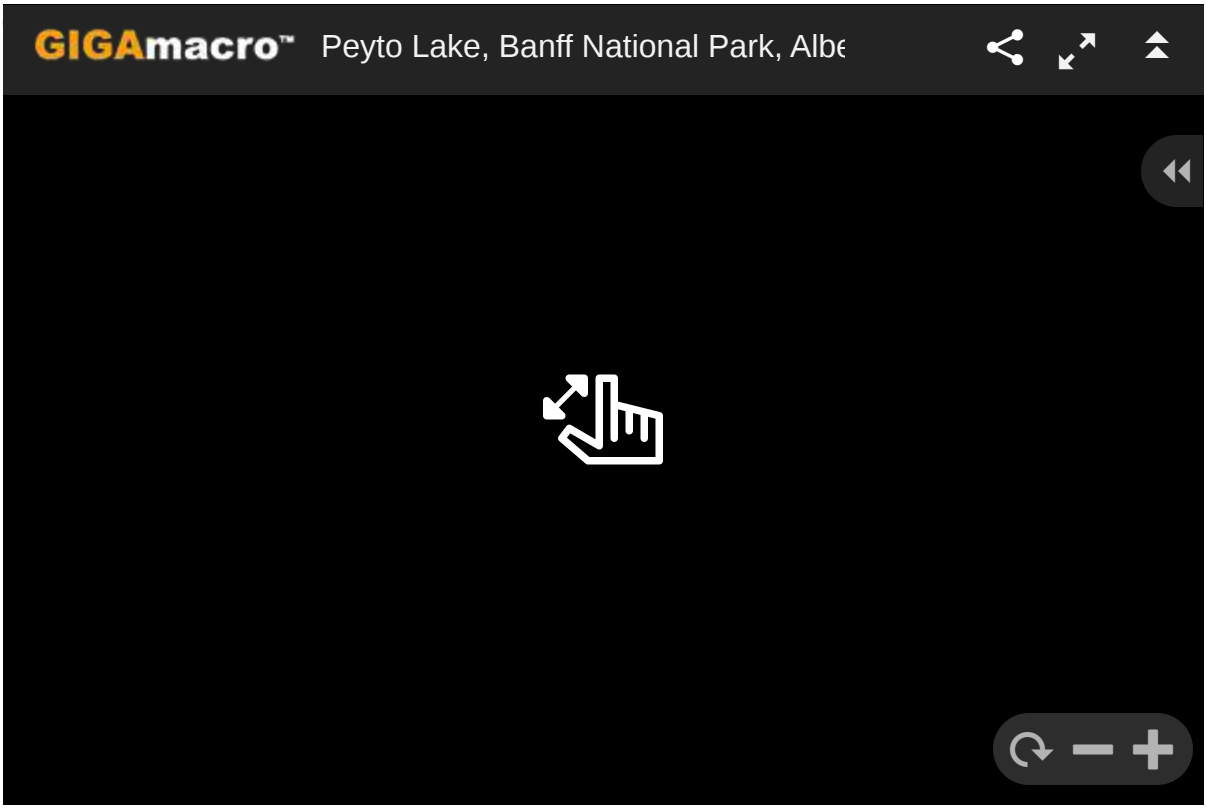
COLOR



Figure 62.7.12: Suspended glacial sediment (“flour” or “milk”) in glacial lakes often imparts an astounding blue color. (Callan Bentley photo)

Glacial lakes high in the mountains (close to their glacial source) often have a striking blue color. This is due to suspended fine particles of rock. Glacial erosion generates many sizes of sedimentary particles but when rushing water flows away from the glacial terminus and enters a lake, it keeps silt in suspension, but will leave behind larger particles such as gravel and sand (frequently dumped where the meltwater stream enters a lake, forming a delta).

Peyto Lake in Banff National Park, just north of Glacier in the Canadian Rockies, shows this dynamic well: The upstream-most quarter of the lake basin is filled with an outwash delta, and the lake downstream of that retains a heavy load of suspended silt, rendering it a peacock blue color.



Did I Get It? - Quiz

Exercise 62.7.1

What kind of glacial lake is shown?



a. Moraine dammed lake

b. Kettle

c. Tarn

d. Pater noster

Answer

d. Pater noster

Exercise 62.7.2

A peak surrounded by cirques flowing away from it has a pyramid shape. This is a _____.

a. kettle

b. arête

c. tarn

d. trough

e. horn

Answer

e. horn

Glacial sediments and depositional landforms

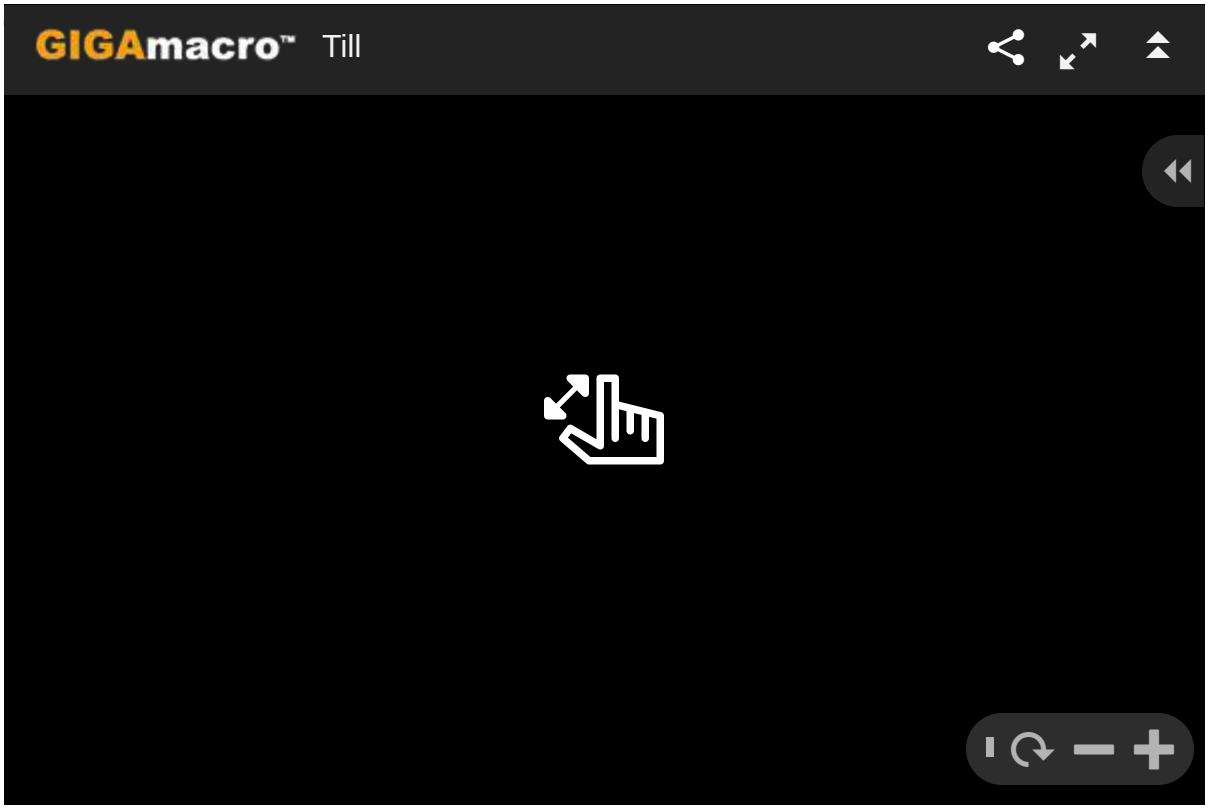
Glaciers leave their mark on the landscape in many ways. One is through the distinctive striations they scratch into the bedrock as they move.



Figure 62.7.13: A glacial terminus is a place where till is deposited to make moraine, and where ice melts to generate outwash. (Photo by Tom Biggs.)

Another is the deposition of extraordinarily poorly sorted sediment, called till, and of water washed sediment, called stratified drift. Till forms directly from the deposit of sediment by glacial ice, while stratified drift is the result of meltwater transporting sediment away from the glacier's terminus, sorting and rounding it along the way.

Here's an example of till:



Note how this outcrop shows substantial boulders dumped amid much smaller particles, including clay. This same effect can be seen from space, as in this Google Earth view of the area between Sun River Canyon and Augusta:

All the white chunks you see in this Google Maps view are erratics, large glacially transported boulders dumped far from their source outcrops. This broad region covered in erratics indicates it is an area of ground moraine.

Individual cobbles or boulders in till will often show glacial scratches on their surface, as with this example:



The lithology (rock type) of the erratics can be used as a clue to trace the source of the sediment. Where did the glaciers come from? In this case, they are mostly Mississippian limestones, so the adjacent Sawtooth Range of the Rockies (*i.e.*, the mountains around Sun River Canyon) are the source. But not too far from here, a short distance to the northeast, there are erratics with a different character. There, they are made of granite and gneiss and other rocks of the Canadian Shield. This suggests the just northeast of Browning is where the Cordilleran ice sheet and the Laurentide ice sheet met, each delivering their own distinctive populations of erratics as a signature of where they came from.

Did I Get It? - Quiz

? Exercise 62.7.1

Till is _____.

a. moderately sorted sediment deposited by rivers that drain away from a melting glacier

b. poorly sorted sediment deposited directly by a glacier

c. well sorted sediment deposited by beach waves a thousand miles away from a melting glacier

Answer

b. poorly sorted sediment deposited directly by a glacier

? Exercise 62.7.2

A large boulder of rock that sits atop a different kind of bedrock is _____.

a. a moraine

b. a till

c. an erratic

d. a striation

Answer

c. an erratic

Exercise 62.7.3

Is this cobble glacial in origin? What evidence tells you the answer?



a. No; it's not scratched.

b. Yes; it's scratched.

c. No; it's gray.

d. Yes; it's gray.

Answer

b. Yes; it's scratched.

Ground moraine, kettles and deranged drainage

Ground moraine is a general term for the collection of landscape features that mark an area of continental glaciation. In addition to erratics, they often show hummocky topography, deranged drainages, underfit streams, and kettles. Let's now examine each.

Typically in this area, areas of glacial melting are marked with hummocky topography, a characteristic lumpy shape to the landscape, where kettles and other low spots are interspersed with low, erratic-covered hills. Here's an example from the Blackfeet Reservation, between Browning and Piegan:

It’s hard for water to find its way smoothly downhill amid such an irregular landscape. The resulting drainage pattern is twisted and gnarled, and sometimes called “deranged.” Deranged drainage patterns are a signature of landscapes where the sedimentary deposits and flowing streams have yet to come to terms with one another.

When a lot of till is deposited rapidly as a glacier rots, big chunks of ice break off and are buried or partially buried. When they eventually melt, it leaves holes in the landscape. Sometimes these are large, and sometimes these are small. If the bottom of the hole is deeper than the water table, then they will be expressed as a pond. We call these ponds “kettles.” (Walden Pond in Massachusetts is a famous example.) In the arid conditions of Montana’s high plains, though, the ponds can’t stay wet for long, and the same glacial set-up gets a different expression: it makes salty playas: ephemeral ponds that dry out though evaporation and infiltration. Any minerals in solution end up being precipitated, giving them a white crusty appearance. Here are three prominent examples, just north of the road into Sun River Canyon:

There is no major through-going drainage into or out of these kettles, so they never get these “evaporite” minerals flushed out of them.

Here’s an example of what happens to modern streams running through this post-glacial landscape:

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In this Google Map, you can see an “underfit” stream running through a broad glacial outwash valley. Where it encounters a recessional end moraine (horseshoe-shaped feature opening to the northwest), it has ponded, making a moraine-dammed lake (the pale blue region). As the stream flows, it continues transporting sediment. When the current of the stream hits the standing water of the lake, the water’s ability to move sediment decreases dramatically. Sediment is deposited at that point, creating a delta (dark green area with vegetation). Eventually this delta will advance southeast, filling in the lake with sand and mud.

The character of the sediment deposited by the meltwater is different than glacial till. The water is much more effective at teasing out fine particles and dropping larger ones, resulting in better sorting. It also tends to produce layered deposits, which is which the drift, the sediment deposited by outwash rivers, is often called *stratified* drift.

Did I Get It? - Quiz

? Exercise 62.7.1

What geomorphic evidence is there for the former presence of Piedmont glaciers east of the Rocky Mountain front near Glacier National Park?

a. Hummocky topography, deranged drainages, kettles, and underfit streams.

b. Drumlins, eskers, kames, and channeled scablands.

c. Horns, cirques, arêtes, and tarns.

Answer

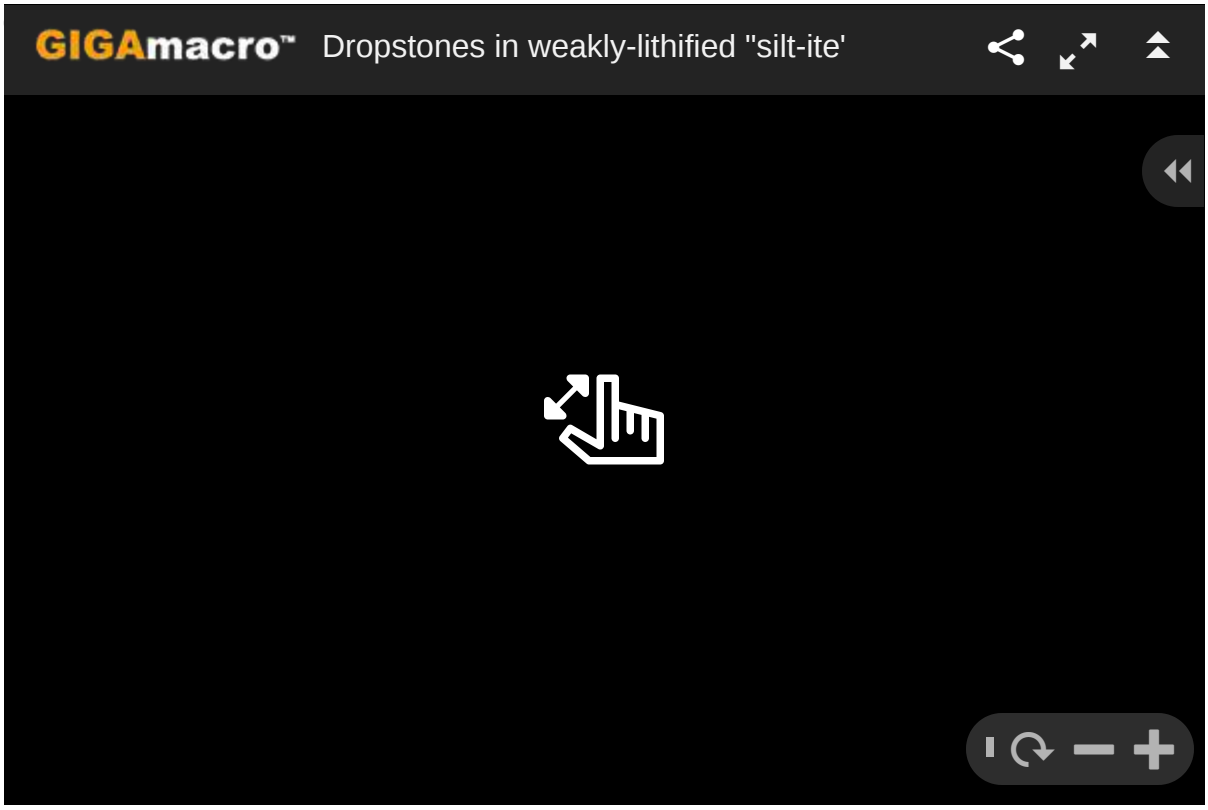
a. Hummocky topography, deranged drainages, kettles, and underfit streams.

Evidence for multiple cycles of glaciation

In addition to till, glacial outwash is another kind of sedimentary deposit that forms downstream of glaciers, as meltwater works through the poorly-sorted till and separates out the finer-grained components from the coarser-grained components. In this outwash setting, stream deposits produce clast-supported conglomerates, like this one:



If the pebbles and cobbles are being dropped from floating icebergs into silty lake sediments, the character of the conglomerate is different: It can be matrix-supported instead:



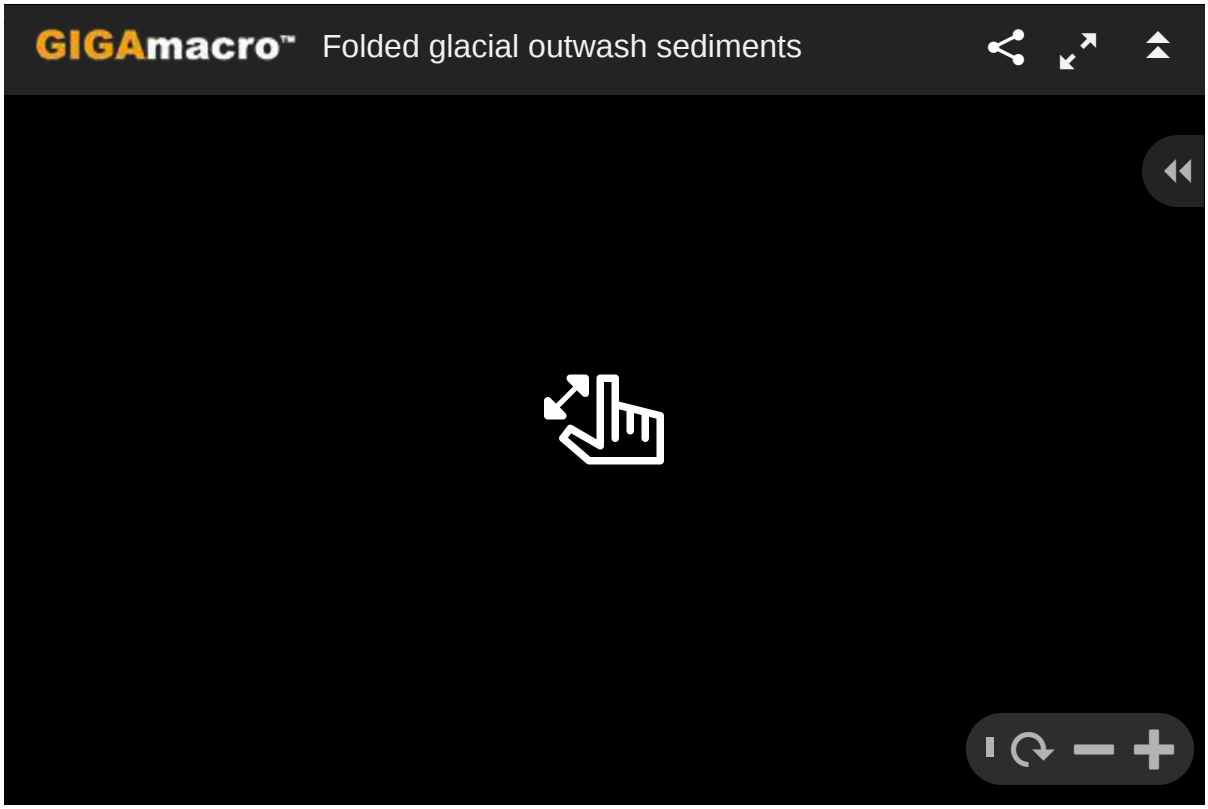
Lake deposits are dominated by silt (the same “glacial flour” that makes glacial lakes such striking colors when it is suspended in the water column), producing a chalky-feeling deposit.



Figure 62.7.14: Glacial outwash sediments exposed on the Looking Glass Highway between Kiowa and East Glacier. (Callan Bentley photo)

At a set of striking outcrops on the Looking Glass Highway (between St. Mary and East Glacier), there are outcrops of these very young outwash sediments, but with an unusual character: Mostly silt, but also with layers of river gravel and dropstones. The familiar colors of red and green argillite jump out, as well as cobbles of diorite, quartzite, and limestone. It's a grab bag of Belt rock types, rendered into cobbles and pebbles, and brought eastward out of the park by Pleistocene glaciers.

But here's the kicker: these deposits of stratified drift are not horizontal. Check out the attitude of the layers here:



That’s quite deformed! Here’s a look at the adjacent outcrop, 10 meters to the south:

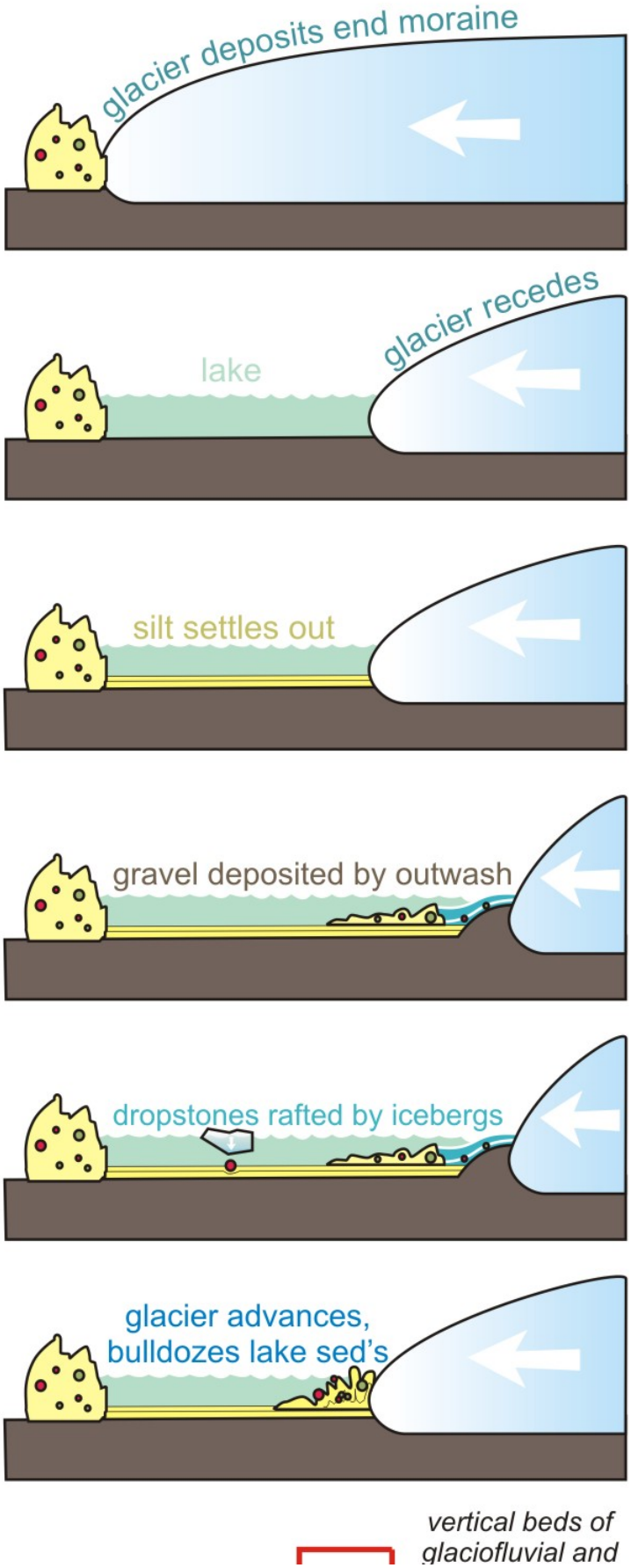




Figure 62.7.15: Near-vertical beds of stratified drift: proglacial lake deposits of silt (glacial flour) interbedded with deltaic gravel layers. (Tom Biggs photo)

These layers are tilted, and tilted ***a lot.*** In fact, they are almost vertical!

How could such a situation come to pass, given that the tectonic compression of the Sevier Orogeny ended tens of millions of years previously? These are young glacial outwash sediments; how could they have been forced into such astoundingly deformed positions, given their extreme youth? This is a puzzle!



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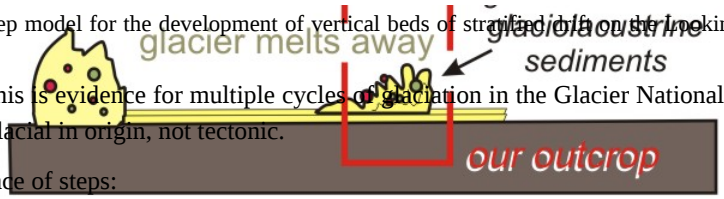


Figure 62.7.16: A seven-step model for the development of vertical beds of stratified drift on the Looking Glass Highway. (Callan Bentley graphic)

The answer seems to be that this is evidence for multiple cycles of glaciation in the Glacier National Park region. The compression that rotated these layers was glacial in origin, not tectonic.

Consider the following sequence of steps:

1. A glacier deposited an end moraine. It could have been a terminal end moraine (representing the maximum extent of the glacier) or it could be a recessional end moraine (representing some spot where the glacial terminus "paused" for a while, reflecting a balance between inputs and outputs).

2. The glacier receded a bit, with a proglacial lake forming between the glacier and the moraine (which acted as a dam).

3. Outwash sediments from the glacier entered the lake, including suspended silt (glacial "flour" that settled from the lake) as well as gravel brought in by outwash streams. Icebergs calved from the glacial front floated out into the lake, melting and dropping their internal sediments into the silt on the floor of the lake (dropstones).

4. The glacial budget changed again, but this time with inputs exceeding outputs. So the glacier advanced, moving out over the area of the lake and "bulldozing" the outwash deposits, folding and tilting them.

5. The glacier finally melts away for good, leaving the deformed outwash sediments high and dry, and available for curious geologists to contemplate... and perhaps even marvel at!

Did I Get It? - Quiz

? Exercise 62.7.1

Why is it unlikely that Sevier Orogeny tectonics tilted the Pleistocene outwash gravels and glacial lake silt deposits on the Looking Glass Highway?

a. Because the Sevier Orogeny didn't cause any compressional forces.

b. Because they are the wrong color.

c. Because the compressional stresses of the Sevier Orogeny had been over for tens of millions of years before these outwash sediments were even deposited.

Answer

c. Because the compressional stresses of the Sevier Orogeny had been over for tens of millions of years before these outwash sediments were even deposited.

Bull Lake & Pinedale glaciations in the northern Rockies

On the North American continent as a whole, glaciologists and Quaternary geologists have identified multiple cycles of Pleistocene glaciation. Each cycle consists of an advance of glacial ice southward from Canada into the lower latitudes of the United States, but also from higher elevation mountain regions downhill into lower elevations. The problematic part of this work, teasing out the history of glaciation, is that each new episode of glaciation tends to destroy the evidence of its predecessors as it rolls over the same territory. This means that, with a few exceptions, we only really know "a lot" about the most recent episode of glaciation, and the only older episodes we can see well recorded are those which reached further south (and downhill) than their successors. Those older but more extensive deposits survive only on their uttermost fringes, with any territory later occupied by younger glaciers overprinted.

The major episodes of North American glaciation that have been identified are named after U.S. states: pre-Illinoian (*consisting of perhaps 11 separate episodes, though this number is controversial*), Illinoian, and Wisconsin. In the Rocky Mountain region, including Glacier National Park, these terms are not applied, as the character of the glaciation we strongly influenced by the mountains, with their high elevations and steep gradients. Most Quaternary glacial geomorphologists identify two principal pulses of glaciation in the Rockies: the Bull Lake glaciation from 200 ka to 130 ka, and the Pinedale glaciation, which spanned from

around 30 ka to 10 ka. As such, the Bull Lake correlates with the Illinoian on the continental scale, and the Pinedale is the Rockies equivalent of the Wisconsin glaciation.

The Pinedale is named for Pinedale, Wyoming, where exquisite landforms associated with that glaciation can be found along the southwestern front of the Wind River Range. Consider this gorgeous example of a moraine-dammed lake, for instance:

There are plenty more in the area – Explore for them using this live Google Map, zooming out and panning southward along the mountain front.

Interaction between Cordilleran and Laurentide ice sheets

As alpine valley glaciers extended downhill and eastward out of the Rocky Mountains during the Pinedale glaciation, they spread out into Piedmont glaciers and merged to make a substantial ice sheet. This ice sheet is called the Cordilleran ice sheet, as it was centered on the North American Cordillera (the “spine” of mountains that runs from Alaska to Panama).

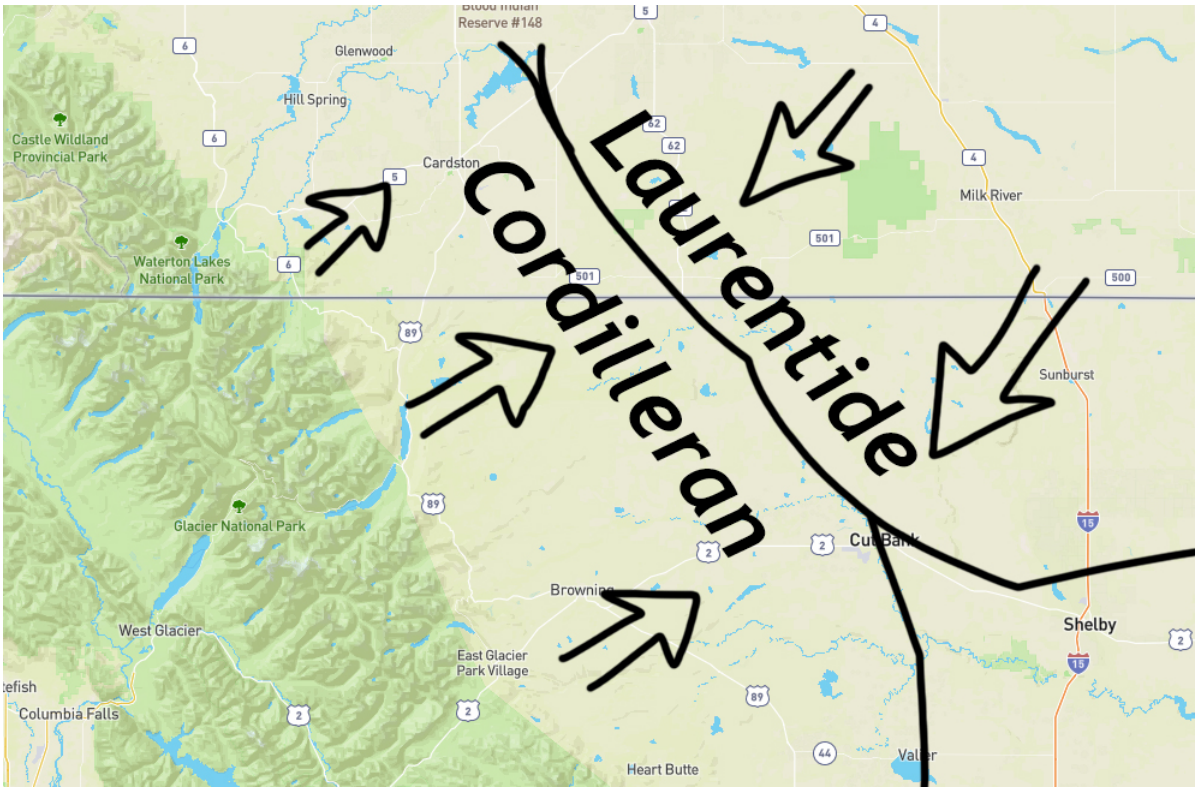


Figure 62.7.17: Just east of Glacier National Park, the Cordilleran ice sheet met the Laurentide ice sheet.

The area immediately east of Glacier National Park is interesting because it is where the Cordilleran ice sheet met the continental-scale Laurentide ice sheet. In other words, the bulky maximum extend of the Pinedale glaciation crept out of the Rockies and pressed up against the edge of the maximum extent of the ice sheet from the Wisconsin glaciation. Consider for a moment how Quaternary geologists might determine where this zone of interaction was. What evidence might they search for?

As seen in the preceding examples of till and glacial outwash in the Glacier National Park region, the rock types making up the glacial sediments of the Cordilleran ice sheet must match the source outcrops in the mountains. Similarly, the Laurentide ice sheet brought its own distinct blend of rocks with it, transporting them south out of the Canadian interior. These rocks of the Canadian shield tend to be high-grade metamorphic rocks like gneiss as well as plutonic rocks such as granite. There are no such lithologies present in the Lewis Range or Sawtooth Range, so it stands to reason that, headed east from Glacier, when you run out of Belt and Mississippian carbonate as particles in the local glacial sediments, and start seeing gneiss and granite instead, that's the indication that you've crossed out of the zone occupied by the former Cordilleran ice sheet and into the "continental" signal showing the footprint of the Laurentide ice sheet.

Did I Get It? - Quiz

? Exercise 62.7.1

What are the two principle episodes of Pleistocene glaciation in the northern Rocky Mountains?

a. Wisconsin and Bull Lake

b. Pinedale and Bull Lake

c. Pinedale and Illinoian

d. Wisconsin and Illinoian

Answer

b. Pinedale and Bull Lake

? Exercise 62.7.2

How can glacial deposits of the Laurentide ice sheet be distinguished from those of the northern Rocky Mountains?

a. The rock types making up the till and outwash are different.

b. The Cordilleran deposits are more poorly sorted.

c. The Rocky Mountains glacial deposits are much older than those of the Laurentide ice sheet.

d. The Laurentide deposits are more crumbly.

Answer

a. The rock types making up the till and outwash are different.

Conclusion

During the Pleistocene epoch of geologic time, repeated episodes of glaciation dramatically changed the region of Glacier National Park. These flowing rivers of ice carved the deformed Belt and Mississippian bedrock into mountain scenery with new shapes, characterized by U-shaped valleys and spiky peaks and fins of rock. In doing this carving, a tremendous amount of sediment was generated, and that sediment was dumped largely to the east of the park, producing another set of landforms made of till and stratified drift. The Rocky Mountain record of glaciation shows two discernible pulses of ice growth, as well as interaction in the northwest Great Plains with the continental Laurentide ice sheet.

Further reading

Eric Karlstrom (1987), "Zone of interaction between Laurentide and Rocky Mountain glaciers east of Waterton-Glacier Park, northwestern Montana and south western Alberta," *Rocky Mountain Section of the Geological Society of America Centennial Field Guide* vol. 2, Geological Society of America, Stanley S. Beus, editor.

National Park Service (n.d.) "Glacial Landforms." Website.

Also consider reading the Snowball Earth case study in this text "book."

"How do glaciers move? TIMELAPSE!" (2018) BBC Earth video.

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62.8: Glacier National Park, Montana - Quiz

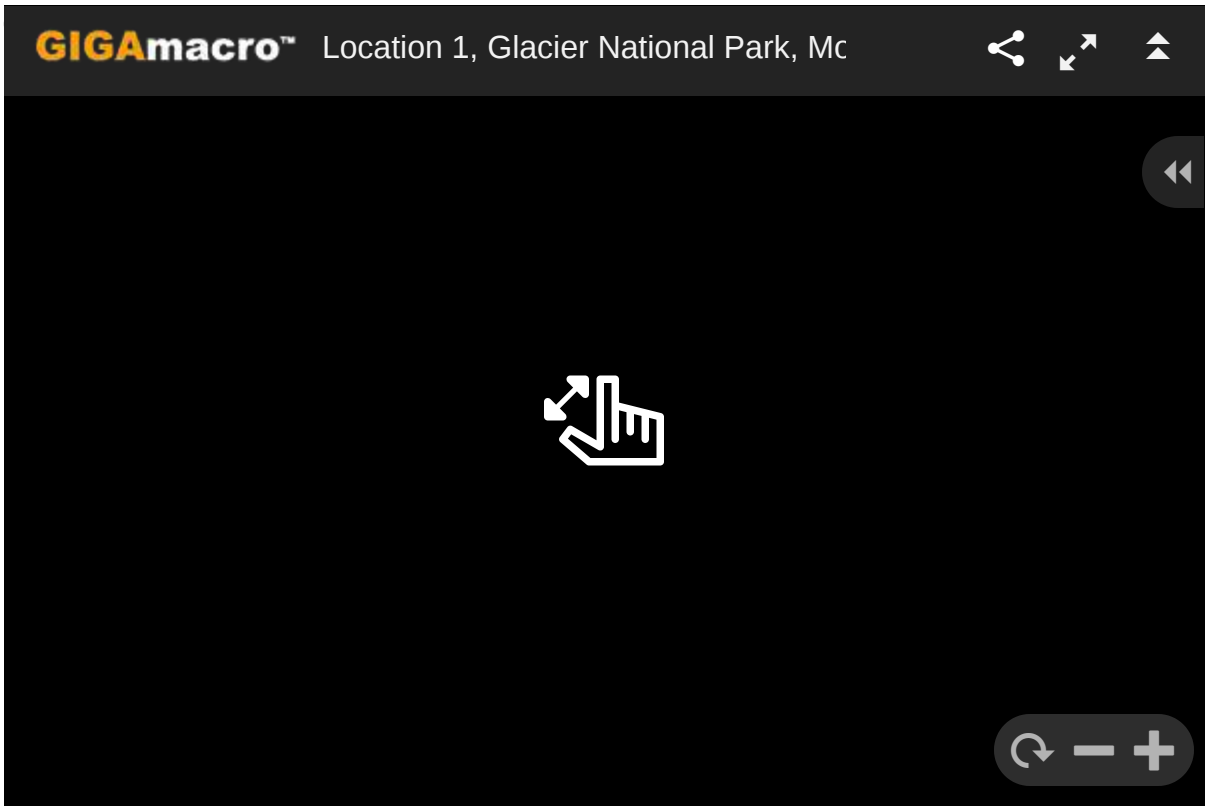
Here is a little review of some of the concepts covered in this virtual field experience, considering gigapixel panoramas of both landscapes and outcrops, as well as hand samples. If you need to review the relevant concepts, here are the four main pages that comprise this VFE:

1. Overview
2. Stratigraphy
3. Structural geology
4. Glacial geomorphology

Work in small groups to answer each question as accurately as possible. Then share the resulting answers in the larger class group, with the instructor facilitating the discussion.

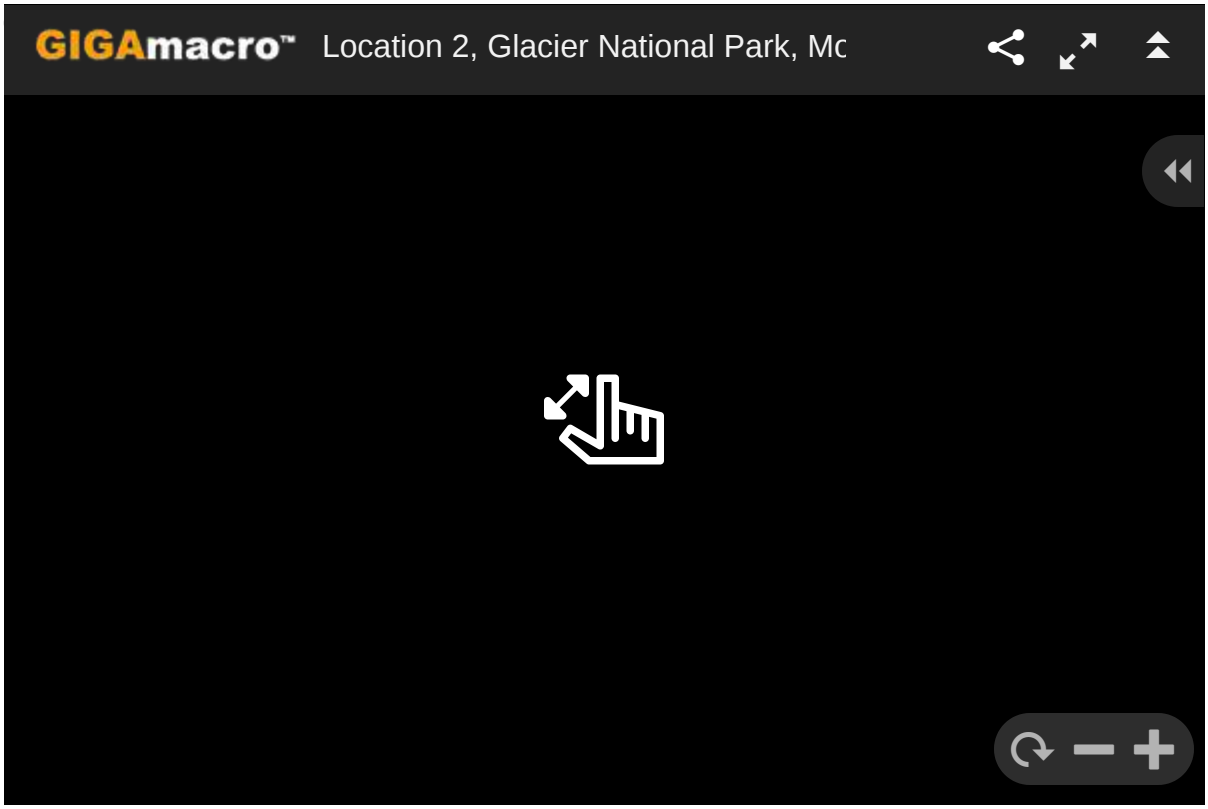
Location 1

Describe the geology present in this view.



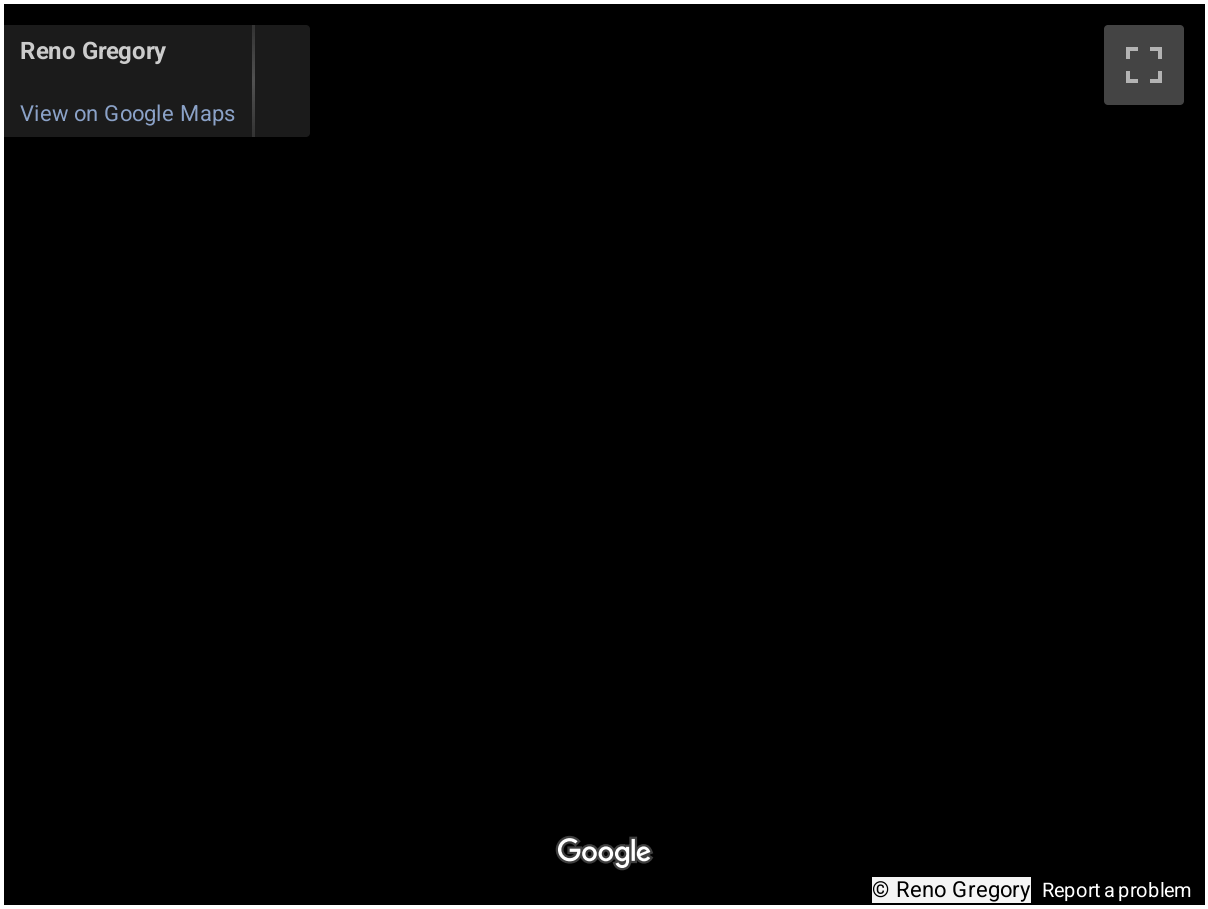
Location 2

Describe the geology present in this view.



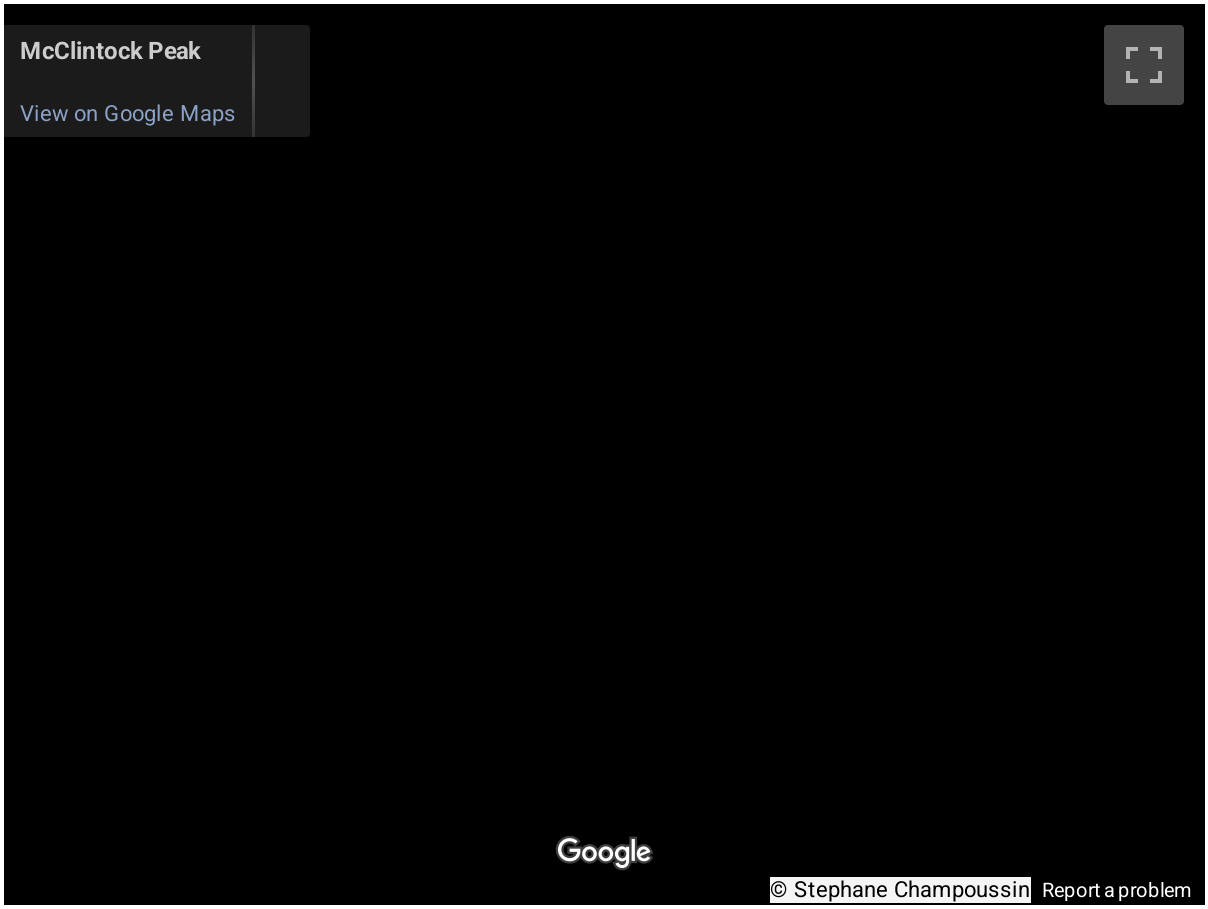
Location 3

Describe the geology present in this view.



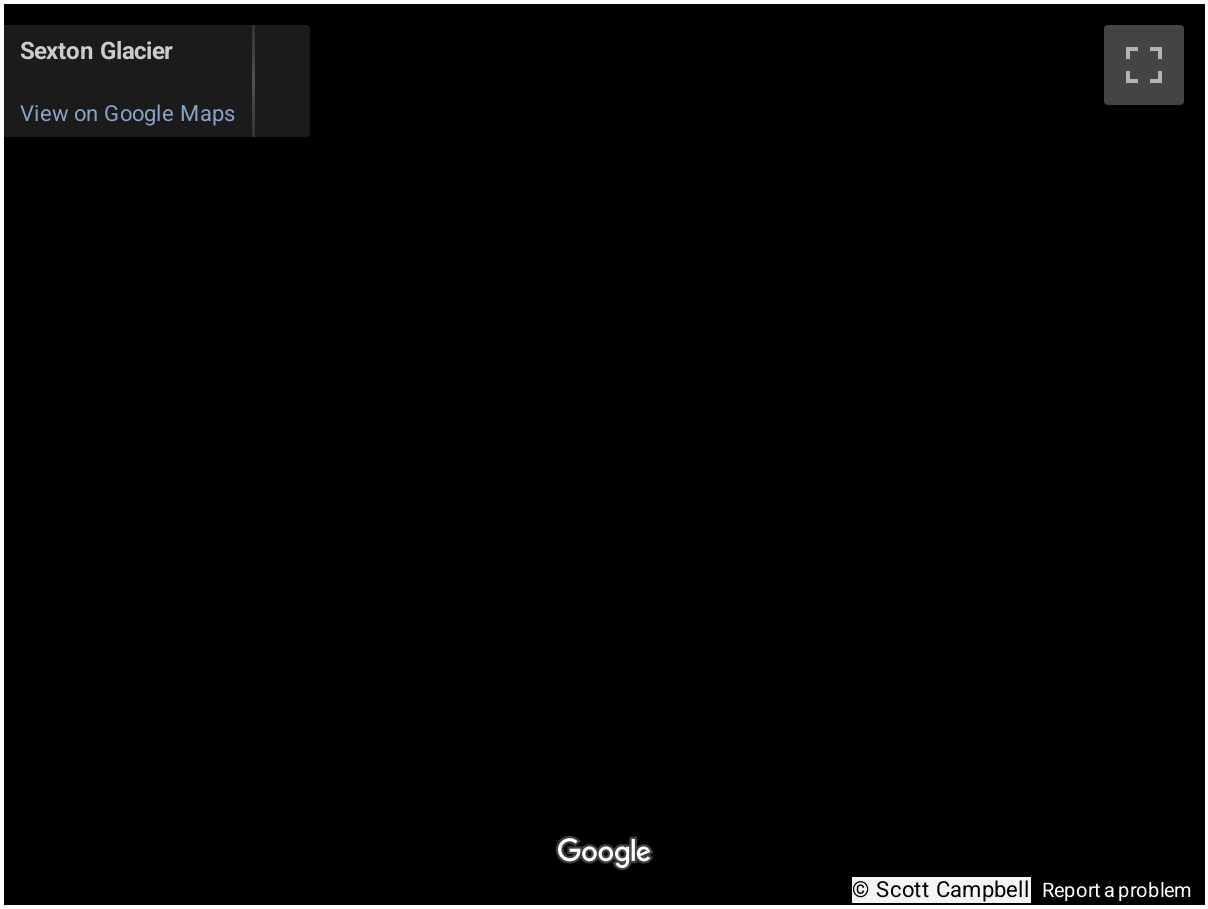
Location 4

Describe the geology present in this view.



Location 5

Describe the geology present in this view.



Sample 1

Do these samples show lithified (a) till, (b) alluvium, or (c) colluvium? Cite your evidence.



Sample 2

Identify and interpret the primary sedimentary structure shown here.

Which way would this model have to be oriented to show “paleo-up” at the top of the screen?



Sample 3

Is this a sample of lithified till, alluvium or colluvium? Cite your evidence.



Sample 4

Is this a sample of lithified till, alluvium or colluvium? Cite your evidence.



Sample 5

Identify and interpret the primary sedimentary structure shown here.



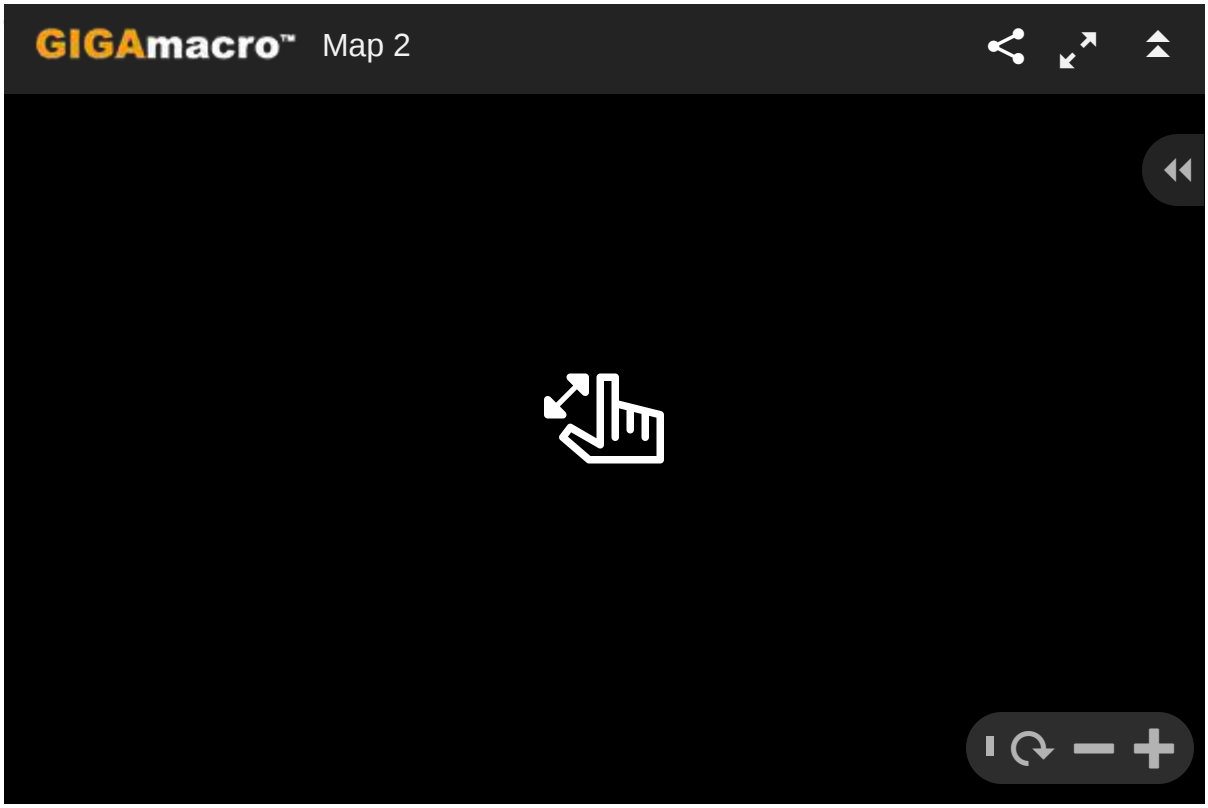
Map 1

Locate the Lewis Thrust on this map.



Map 2

Identify the feature shown at the center of this view of this map and describe how it came to be.



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62.9: The Geodetic Hills Fossil Forests - Unraveling a Paleoclimate Puzzle

An Arctic field project

Field work in remote places is always a challenge, mentally and physically. The Canadian Arctic is no exception. Mentally, because it really is isolated, and the one or two companions you will spend the next 6-10 weeks with should be reasonably accepting of each other's foibles. Frigid winters (minus 40° C and F – this specific temperature is the same on both scales) and snow-drift blizzards mean that most field work is a summer venture (some geophysical work is conducted during winter when fjords and seaways are frozen).



Figure 62.9.1: Geodetic Hills, the location of the middle Eocene fossil forests, is located on east-central Axel Heiberg Island.

Field logistics in places like Ellesmere and Axel Heiberg islands require good planning – getting there and moving around while on site that, in addition to foot-slogging, frequently relies on a helicopter that needs fuel and an engineer. Food, tents, and general field equipment are deemed essential. A source of clean water commonly dictates the location of a camp. One base camp on western Axel Heiberg in 1983 sourced water from an iceberg a few 100 m offshore, retrieved by a helicopter dipping a 45-gallon drum into a melt pond and ferrying it to camp. That the water was probably a few thousand years old was humbling (although there was a distinct paucity of gin and tonics to take advantage of the ice). Equally important is finding a convenient spot for the loo (toilet for North American readers), placed a suitable distance from all other camp activities; one benefit of working summers in the Arctic is that you never need to stumble in the dark in search of it.



Figure 62.9.2: To source fresh water, a helicopter would dip a 45 gallon drum into a melt-water pond on top of this iceberg. Located in Strand Fiord.

Geodetic Hills, Axel Heiberg Island

In 1985 my field assistant and I were examining sedimentary rocks on central Axel Heiberg Island. The project was part of a broader science program being run by the **Geological Survey of Canada**, and my role was to map and interpret the Late Cretaceous to Paleogene sedimentary basins. I had surmised, from some of my earlier work that the deposits in this sector of the island (the Buchanan Lake Formation) had formed in response to tectonic upheaval and mountain building in the region about 40 to 45 million years ago (**middle Eocene**); we were on the look-out for additional information to assess this hypothesis.

Our helicopter had dropped us off at the base of a gentle ridge, known as **Geodetic Hills**. It was a beautifully clear, cold summer day, as so often is the case in the eastern Arctic Islands, but very windy. Working our way up the ridge, measuring rock layers and recording observations as we progressed, we came upon thin layers of coal; at least that's what they looked like at first glance. Beds of coal are common in Arctic geology, testaments to very different environments in ancient times, so initially these beds were of no surprise. Continuing up the ridge, we chanced upon flat ground that, as luck would have it, allowed us to walk across the top of a coaly layer. It was stunning.

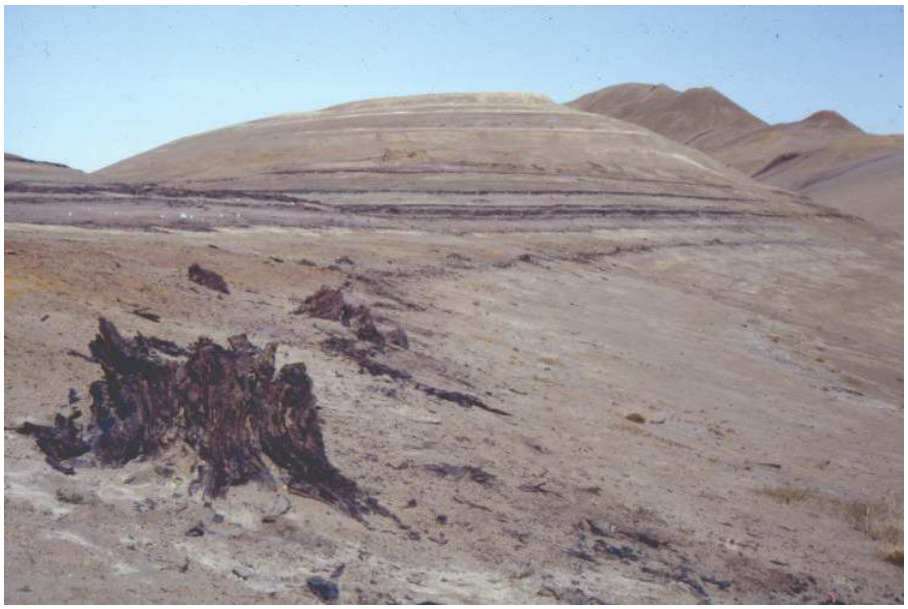


Figure 62.9.3: One of several coaly woody beds containing tree stumps in their position of growth. Brown hues in the mummified wood stand out in the nearest stump. The intervening beds contain fossil soils with abundant roots, and mudstone-siltstone deposited during flooding across the floodplain.

Poking out of the layer were numerous tree trunks; we were basically walking through the remains of a very old forest, around the trees, stumbling on exposed roots, branches strewn across the ground. But the real surprise was that these trees were not coal; they were darkened by age, but there was still wood – mummified wood. And this wood was **45 million years old! Mummified wood** is fossil wood that has not been permineralized, and is relatively unaltered due to rapid burial.

On closer inspection the layers, in addition to tree stumps and logs galore, contained fossil leaves, cones and seeds that could easily be pulled intact from the deposit. It was like extracting plant remains from a recent forest litter. Some ancient spruce or fir cones looked almost as fresh as the day, 45 million years ago, when they fell to the forest floor (today, the nearest spruce tree is about 1600km south). The litter layers were a treasure-trove – here was good evidence for a climate back in the middle Eocene that was much warmer than the modern frigid, desert conditions we were experiencing on that exposed Arctic hillside.



Figure 62.9.4: Slightly flattened Metasequoia cones and stems plucked from one of the forest litters. They still have some of their original brown colors.

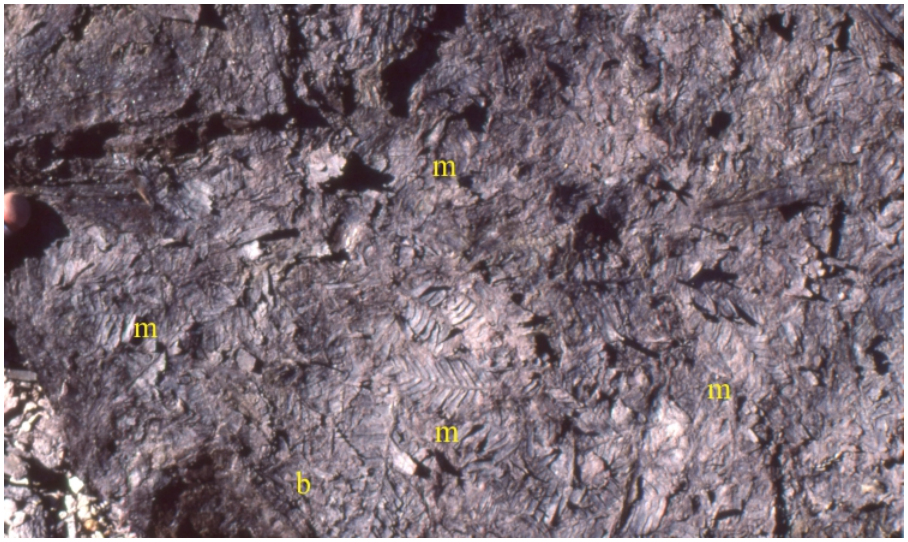


Figure 62.9.5: Metasequoia fronds (m) and smooth-edged broad leaves (b) that can be plucked from this litter with a pair of tweezers. Hickory nuts were also harvested from this sample.



Figure 62.9.6: An Eocene spruce cone, looking as if it has just fallen from its tree. Its preservation is exquisite.

The puzzle continued...

There is another twist to the story. We know from Plate Tectonics theory that over long periods of geological time, continental and oceanic plates move in relation to each other, in some case many thousands of kilometres. Within this knowledge framework we also know that the land area now occupied by the Arctic Islands has basically been in the same position relative to the geographic North Pole for at least the last 100 million years. In other words, the ancient forests were at a similar (high) latitude back in the Eocene as its fossil remains are now. This being the case, the forests and whatever critters that lived therein must have experienced months of continuous summer daylight, and for the opposite season months of darkness. What kind of forests were they; and who lived there?

Back in the Eocene...

What was the rest of the world like 40-45 million years ago?

Continental and oceanic plates (other than the Arctic landmass) were in quite different positions to where they are now. A consequence of this was that the distribution of land masses, mountain belts and ocean currents, all of which contribute or help drive climate change over millions of years, were also different. Significantly, the Geodetic Hills forests flourished at a time of mountain building on Ellesmere and Axel Heiberg islands.

There were no Antarctic or Greenland icecaps, and no Arctic sea ice. The North Atlantic Ocean was quite a bit narrower than it is today. Australia was separating from Antarctica about this time, opening a major seaway that changed southern hemisphere ocean currents; New Zealand had already separated from the Antarctic-Australia amalgam long before this time. All were remnants of the Gondwana supercontinent. And importantly, from the point of view of our Arctic forests, the global climate had veered into carbon dioxide-driven hothouse conditions.

Average global temperatures began increasing in the late Paleocene. This was established initially from foraminifera oxygen isotope data (the proportion of $^{18}O/^{16}O$ is strongly correlated with temperature), and supported by multiple studies of fossil plants, pollen, and vertebrates, plus marine invertebrates.

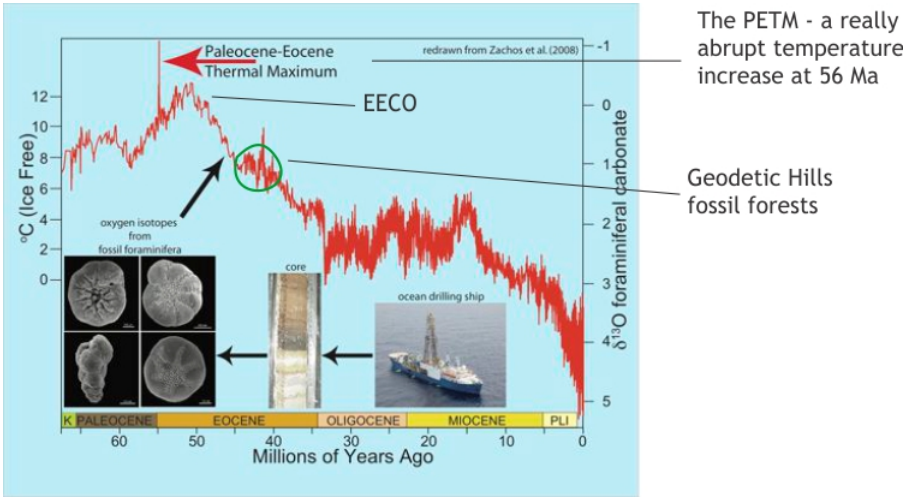


Figure 62.9.7: The chart shows progressive changes in temperature from the beginning of the Paleocene (65Ma) to the present day, based on foraminifera oxygen isotope data (18O) from deep sea cores. The PETM, EECO and Geodetic Hills forests are indicated. (Image credit: Smithsonian National Museum of Natural History)

The data record a thermal maximum at about 56 Ma (the Paleocene-Eocene Thermal Maximum, or PETM), an abrupt, short-lived event that developed over a few thousand years (from a geological perspective this is very quick). The cause for this spike in average global temperatures is thought to have been a sudden introduction of CO_2 into the atmosphere. The source for this sudden infusion of CO_2 is unknown: it may have been volcanic. Interest in the PETM has grown recently because it offers a possible analogy for possible abrupt climatic effects resulting from modern increases in CO_2. The PETM is superimposed on a more gradual increase in temperatures into the Early Eocene – this was the Early Eocene Climatic Optimum (EECO).

The Geodetic Hills fossil forests are a few million years younger than the EECO, and by this time average temperatures had cooled slightly. Note that initiation of the Antarctic Ice sheet was still 10-15 million years in the future (Early Oligocene). This then is the general climate backdrop to the Geodetic Hills' Eocene forests – warm, and probably humid.

Fast-forward to 1986

The 1986 field season was the beginning of a concerted effort for different scientific groups to study the Geodetic Hills forests and other fossil sites in the region; this effort continues today. Earth scientists who study fossil pollen, vertebrate animal life, wood preservation, ancient soils, coals, and sedimentary rocks have at one time or another converged on this site.

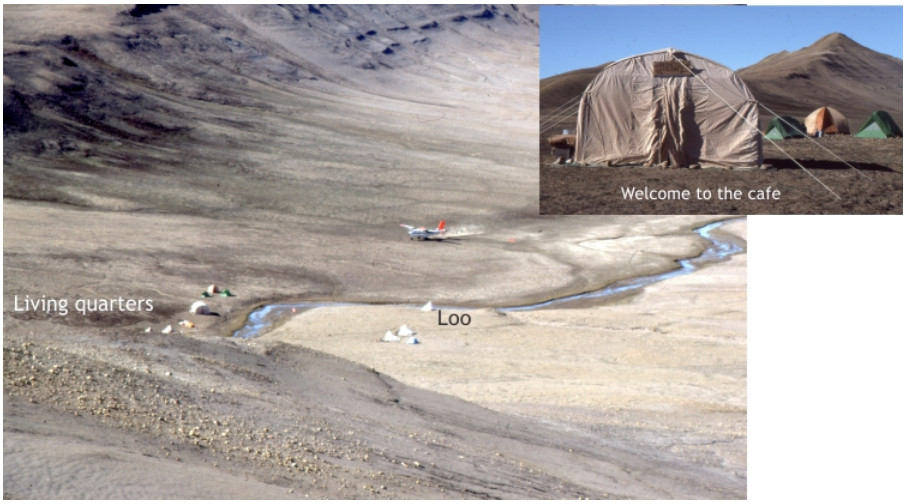


Figure 62.9.8: The 1986 camp site, below Geodetic Hills ridge, where scientists from various disciplines came and went throughout the field season. In this view, a Twin Otter supply plane (centre) is just about to take off. The inset shows the communal tent where one drank coffee and chatted about the day's findings.

The first paleobotanical analyses by Jim Bassinger (University of Saskatchewan), Jane Francis (British Antarctic Survey) and their colleagues focused on the flora and forest dynamics (tree size, tree rings, spacing, canopy). More recent field excursions that

extended to similar-aged deposits beyond Geodetic Hills, have concentrated on plant physiology, leaf sizes and shapes, floral densities, floral associations and wood chemistry to interpret paleoclimatic conditions like precipitation and seasonality.

The list of plants includes ancient conifers, like **hemlock**, **spruce**, **pine**, **larch** and **Dawn Redwood** (*Metasequoia*, once thought to be extinct but subsequently found in China), the broadleaf-conifer relative **Ginkgo**, plus deciduous broad-leaf trees like **birch**, **alder**, **oak** and **sycamore**, and **ferns**. Based on trunk size and spacing, it has been inferred that some of these trees grew to 30 m (100 ft) (Francis, 1991).

The only animal fossils to have been found at Geodetic Hills are the teeth of an ungulate (hooved mammal) that resembled a hippopotamus. However, slightly older Eocene deposits on nearby Ellesmere Island have revealed a diverse group of mammals such as small hedgehog-like animals, various rodents, tapirs, semi-aquatic creatures like turtles, badger-, fox-, and weasel-like carnivores, plus boa-like **snakes,** and a species of **alligator**. Presumably there were insects, but the fossil record of them at this site is poor.

So, what was the Geodetic Hills site like back in the Middle Eocene?

Picture a lowland area, carpeted with mixed conifer and deciduous broadleaf trees and shrubs, interspersed with swamps and bogs. Sluggish rivers meandered through, en-route to a more distant sea. Farther inland new mountains were shedding gravel, sand and mud, some of which was moved by the rivers and deposited on their floodplains.

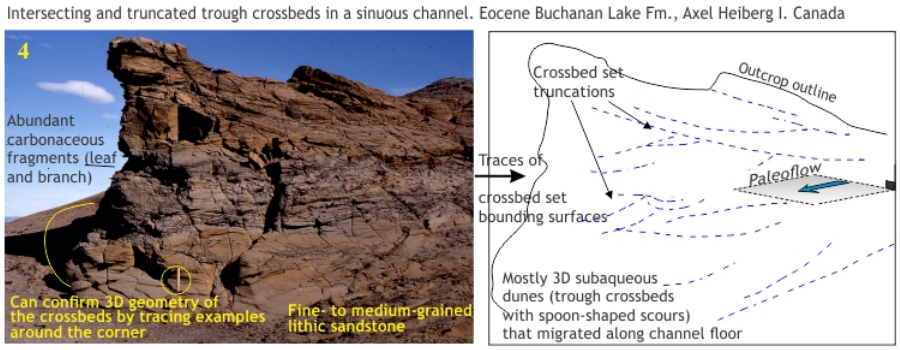


Figure 62.9.9: Meandering river channel deposits of crossbedded sandstone. Plant fragments ranging from leaf-size to large logs derived from an adjacent vegetated floodplain, are occasionally strewn across the channel floor. This record of sinuous, meandering river-channel deposits corresponds to area 4 in the reconstructed paleoenvironment below.

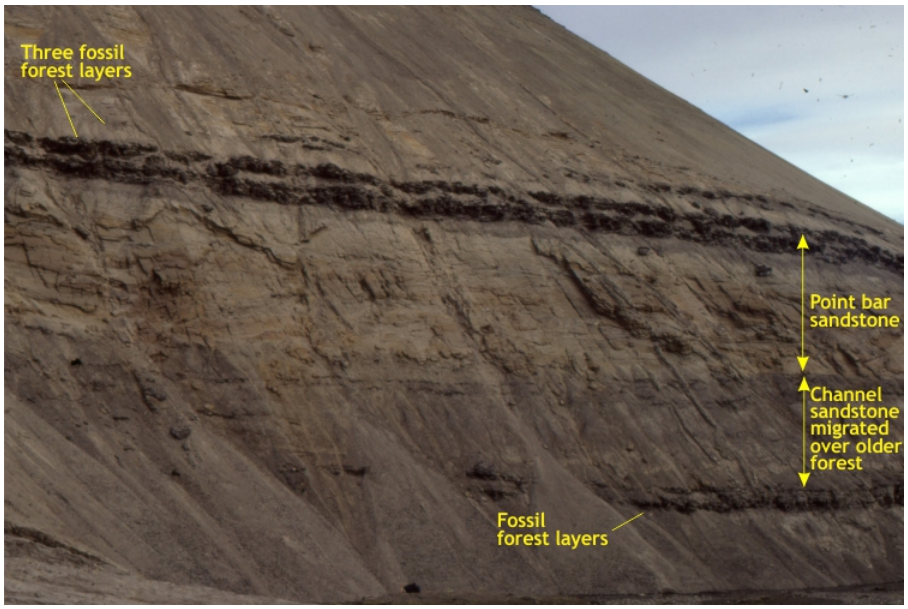


Figure 62.9.10: In addition to channel deposits, fine-grained sandstones also accumulated on point-bars on the constructional side of meander loops. In this outcrop, the point-bar deposits are capped by muds distributed across the adjacent floodplain, and lignite beds that formed in swamps and forested areas beyond the main river channels. This record of sinuous, meandering river-channel deposits corresponds to area 4 in the reconstructed paleoenvironment below.

There are two main schools of thought on what climate prevailed here. Early studies that examined wood chemistry, posited the hypothesis that the climate was wet and monsoonal, much like modern coastal Asia. However, a recent publication on slightly older Eocene deposits from the same area (West et al. 2015) argues the case for a seasonal rain forest based on analysis of leaf physiognomy (size and shape), including leaf area and leaf margin analyses, and comparing these with modern British Columbia coastal conditions.

Three methods commonly used to estimating paleoclimate, using proxies for paleotemperature and paleo-precipitation, from floral assemblages (information from West et al, 2015) are:

1. **Leaf margin analysis** is an empirical, quantitative measure of **mean annual paleotemperature**, calculated from the ratio of toothed leaf margins to complete (smooth) margins. It is based on the observed correlation between temperature and the proportion of toothed, or serrated leaf margins to complete or smooth margins; the higher the mean annual temperature, the greater the proportion of smooth margins. However, this method only applies to woody, dicotyledonous species (one of the main groups of flowering plants).

2. Likewise, **mean annual precipitation** can be estimated by comparing the **leaf areas** of fossil assemblages. Again, this is an empirical method based on modern floral associations where it can be demonstrated that leaves tend to be smaller in drier climates, because it reduces the loss of water via evapotranspiration. One advantage of the leaf margin and leaf area analyses is that taxonomic identification of leaves is not necessary (although it is desirable for forest reconstructions).

3. Comparing your fossil assemblage to **nearest living relatives** is also useful, but is less precise than the leaf margin and leaf area analyses. This method does exactly what the title states – fossil species are compared with extant species. This method requires correct taxonomic identification of fossil leaves.

These wet and warm conditions prevailed during the middle Eocene PETM; mean annual temperatures are estimated to have been **13°-15°C (55-60°F)**. The Geodetic Hills forests are a bit younger and the climate had cooled slightly. But the forests flourished, enough to support various mammals and rodents, and in the swamps, alligators and turtles.

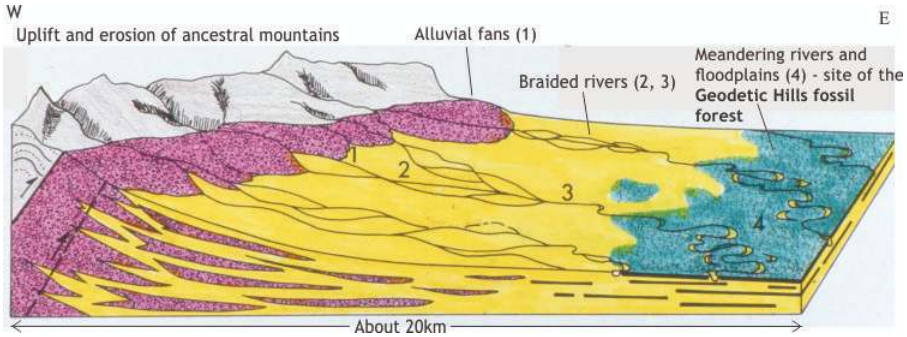


Figure 62.9.11: A 3-D reconstruction of interpreted sediment source area (rising mountains) and depositional environments associated with the fossil forest. There is a progressive change in sedimentology and paleoenvironment from the mountain front alluvial fans (1), to various kinds of sparsely vegetated, coarse-grained (gravelly) braided rivers (2) and (3), to forested and swampy flood plains (4). The channel and point-bar deposits shown above would have accumulated in meander loops like those depicted here (4).

Afterword

Our walk through these ancient forests, although having scientific purpose, unlocks one's imagination – stepping over those fallen logs 45 million years ago, brushing aside spruce and hickory boughs, swatting the odd insect, camera-ready. And depending on the season, a walk at twilight, and even total darkness. Savour the moment, because that cold blast of Arctic air will bring you back to a wind-swept ridge.

The fossil forest is still being studied. It has become an important component of a paleoclimate puzzle that many scientists believe has a direct bearing on the kind of climate changes we are witnessing today. Therein lies the value of scientific enquiry and discovery; the unexpected lies just over that ridge.

All the images used in this chapter are from geological-digressions.com

References and Further reading

R.L. Christie and N.J. McMillan, *(Editors), 1991. Tertiary fossil forests of the Geodetic Hills Axel Heiberg Island, Arctic Archipelago. Geological Survey of Canada, Bulletin 403.* 15 papers on many aspects of the forests.

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K. West, D.R. Greenwood and J.F. Basinger*, 2015. Was the Arctic Eocene 'rainforest' monsoonal? Estimates of seasonal precipitation from early Eocene megafloras from Ellesmere Island, Nunavut. Earth and Planetary Science Letters v. 427 p. 18–30.* Available for download. Has a useful summary of the PETM.

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62.10: Massanutten Synclinorium, Virginia

A virtual field experience via gigapixel panoramic images

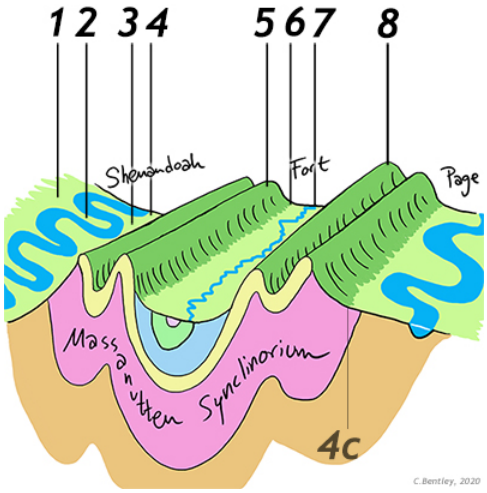
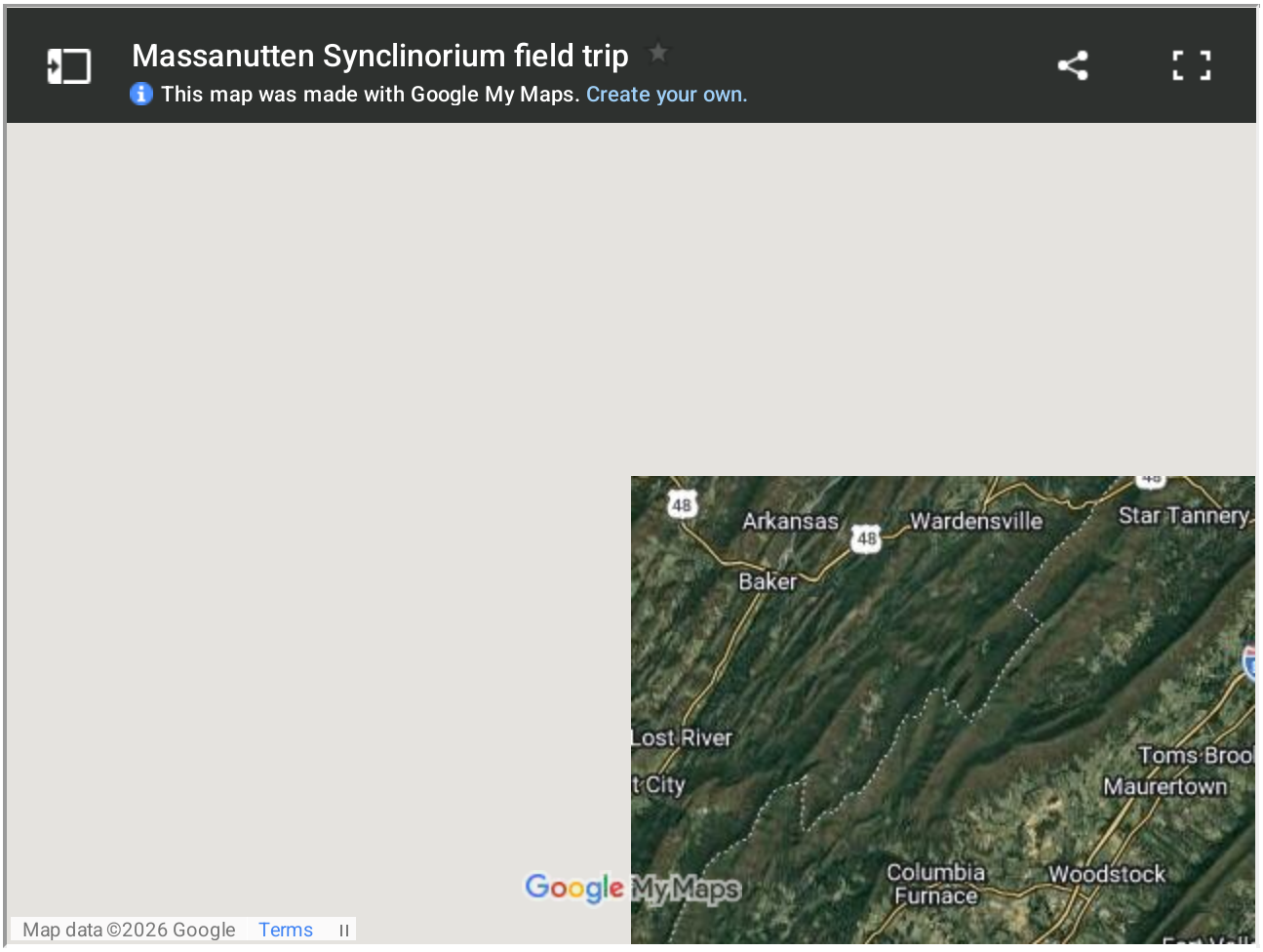


Figure 62.10.1: A cartoon conceptual cross section showing the regional structure of the area, with field trip stops labeled.

Explore these sites in Shenandoah County, Virginia. This virtual field experience will “take you” to the northern Shenandoah Valley near Strasburg, the northern Massanutten Range, and the Fort Valley. All these sites show rock outcrops that are part of the regional-scale Massanutten Synclinorium.

A synclinorium is a large down-fold (where the limbs overall dip inward toward a central fold axis), with smaller-scale synclines and anticlines superimposed on it. The sedimentary strata on the west side of the structure are the same on the east side of the structure, and those are older than the strata exposed at the center of the structure. Differential erosion of the strata has preferentially etched away at the limestones, shales, and graywackes, making low-lying valleys: the Shenandoah Valley and its eastern counterpart, the Page Valley, and the Fort Valley in between. The mountain ridges of the Massanutten Mountain system are all underlain by a resistant quartz sandstone unit that stands up better to the forces of weathering.

Here is a Google Map showing where this VFE’s eight principal stops are located (in blue), plus logistical information if you wanted to visit in person (in red), and supplementary stops (in yellow):



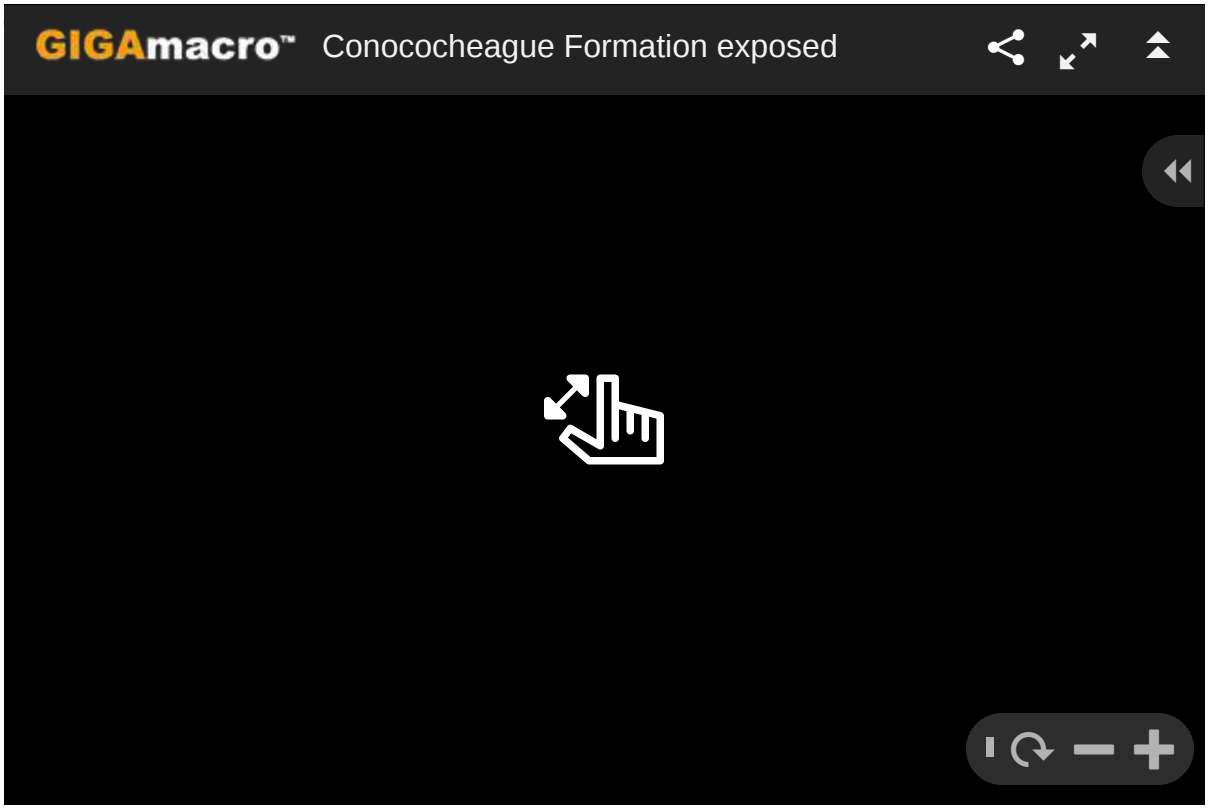
The eight principal stops on this field trip tell a story (1) of shifting sedimentary conditions in the Appalachian sedimentary basin over the course of two episodes of passive margin sedimentation and two orogenies and then (2) all of those same strata being deformed during a third orogeny. The two passive margin episodes are marked by carbonate deposition, while the two early (basin-adjacent) mountain-building episodes that contributed clastic sediment are the Taconian and Acadian orogenies. The final mountain-building episode that turned the basin into part of the mountain belt was the Alleghanian Orogeny.

To explore this story's rock record, the stops below are shown with gigapixel panoramas of outcrops and hand samples. Each is "explorable" by you as the virtual field tripper: you can zoom in, pan around, make observations, pose potential interpretations, and then look somewhere else to test those ideas. The images are organized by stratigraphic and structural sequence. We begin with the oldest strata, work our way up through the available outcrops to the youngest strata exposed in the region (considering what they have to tell us about depositional conditions and active/passive margin conditions through the early to mid-Paleozoic) and then look at deformation associated with the late Paleozoic Alleghanian orogeny.

Stop 1: Conococheague Formation exposed at Mulberry Run

The oldest rock unit on our trip is the Cambrian-aged Conococheague Formation. It is a limestone with some secondary dolostone. It features primary sedimentary structures including ooids, cross-bedding, and stromatolites.

Easternmost portion of the south-facing outcrop:



A little bit further to the west: Can you find the stromatolites?





Figure 62.10.2: Ooids from the Cambrian-aged Conococheague Formation, Shenandoah County, Virginia.

Some of the Conococheague, particularly at the west end of this outcrop, show small ooids. It's oolitic limestone. Ooids forms due to chemical precipitation of calcite from seawater, building up on small nuclei particles, layer on layer, like a hailstone. They get rounder as they get larger, and the reason they have such even layering is they are gently shifted around from moment to moment by waves. If we look at where ooids occur today, we find them in shallow tropical marine settings, such as the carbonate platform of the Bahamas Islands. Therefore, using uniformitarianism, we can infer that when the Conococheague Formation was deposited, Virginia must have been a carbonate platform where the seafloor was within reach of the waves – moderately shallow, in other words.

Here is a Petri dish full of ooids, the spherical building blocks of oolitic limestone:

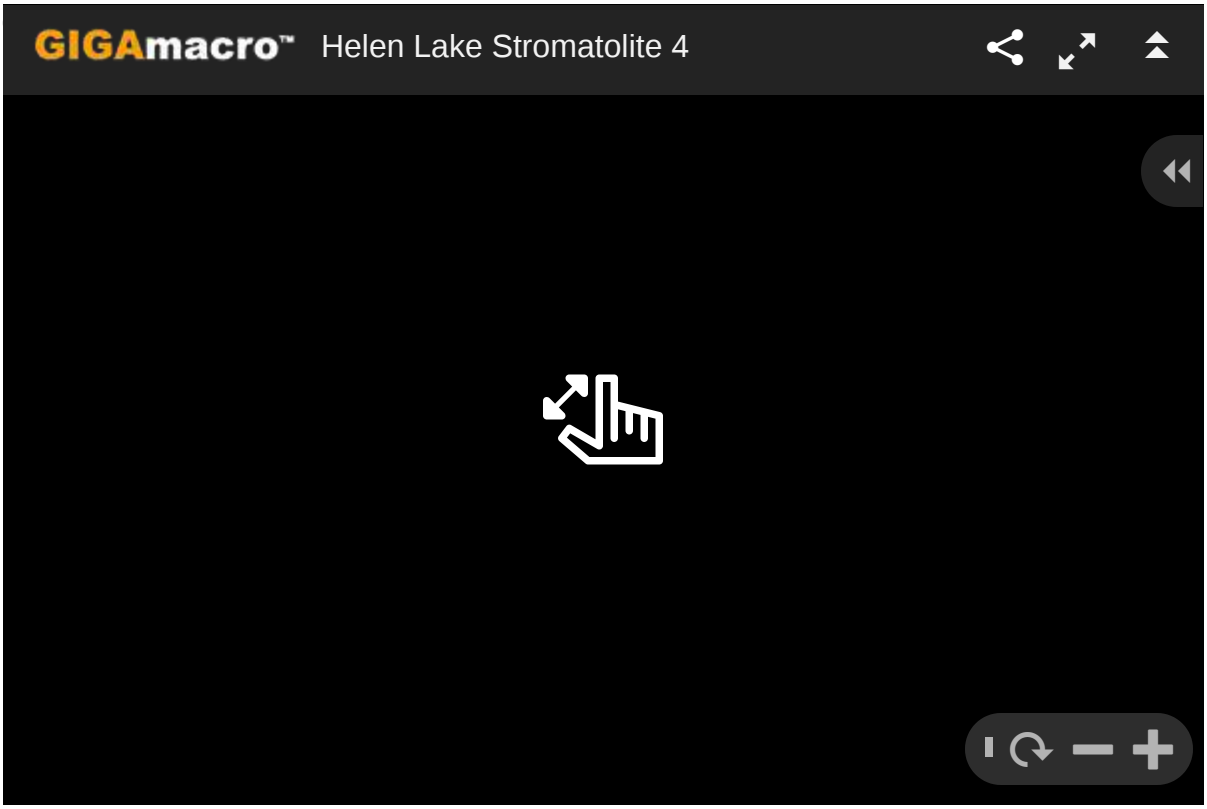


So: ooids only form in moderately shallow water, as their formation requires gentle shifting by the waves. But stromatolites too are an indication of shallow conditions, as the microbial mats were photosynthetic, and needed sunlight in order to grow. Light doesn't penetrate to great depths through the water, so stromatolites "corroborate the testimony" of the ooids, backing up the interpretation of the Conococheague Formation as being indicative of conditions we would expect on a shallow tropical carbonate platform.



Going further: Here are some samples of stromatolites from other sites, similar to those in the Conococheague Formation:




GIGAmacro™ Helen Lake Stromatolite 4


GIGAmacro™ Conophyton

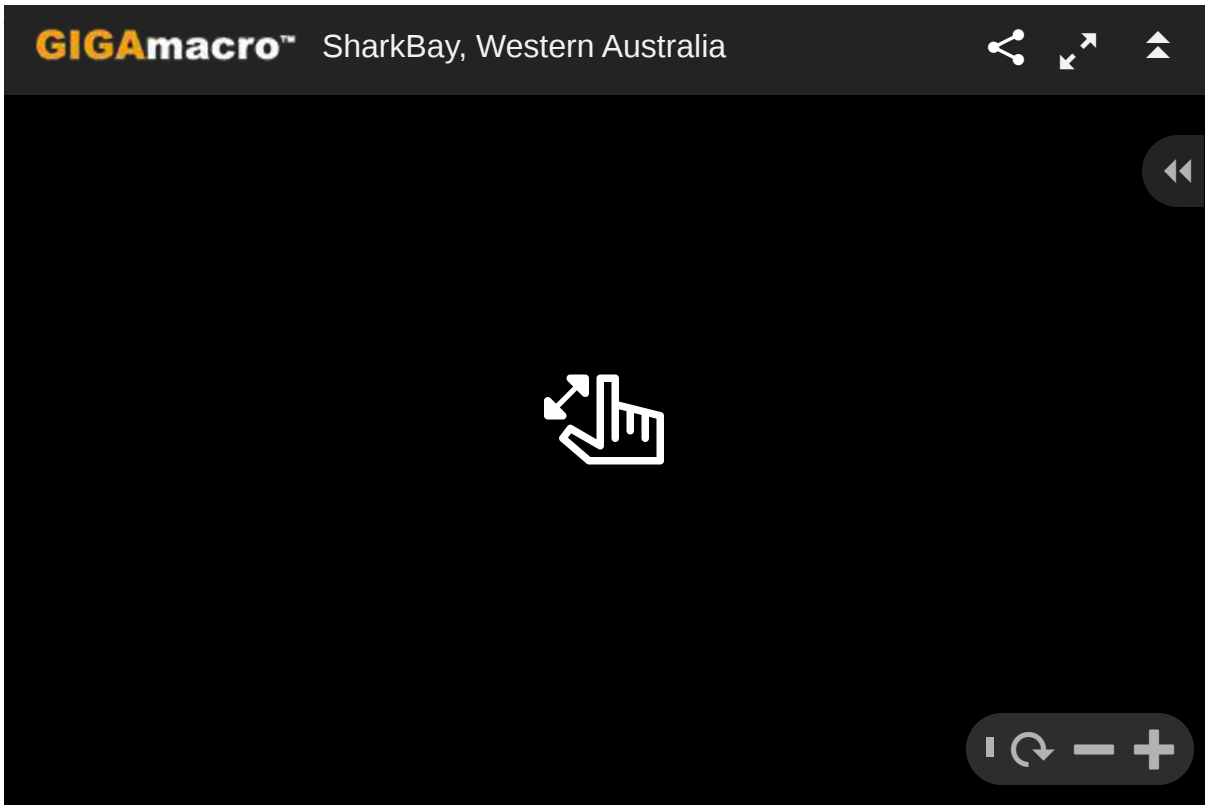




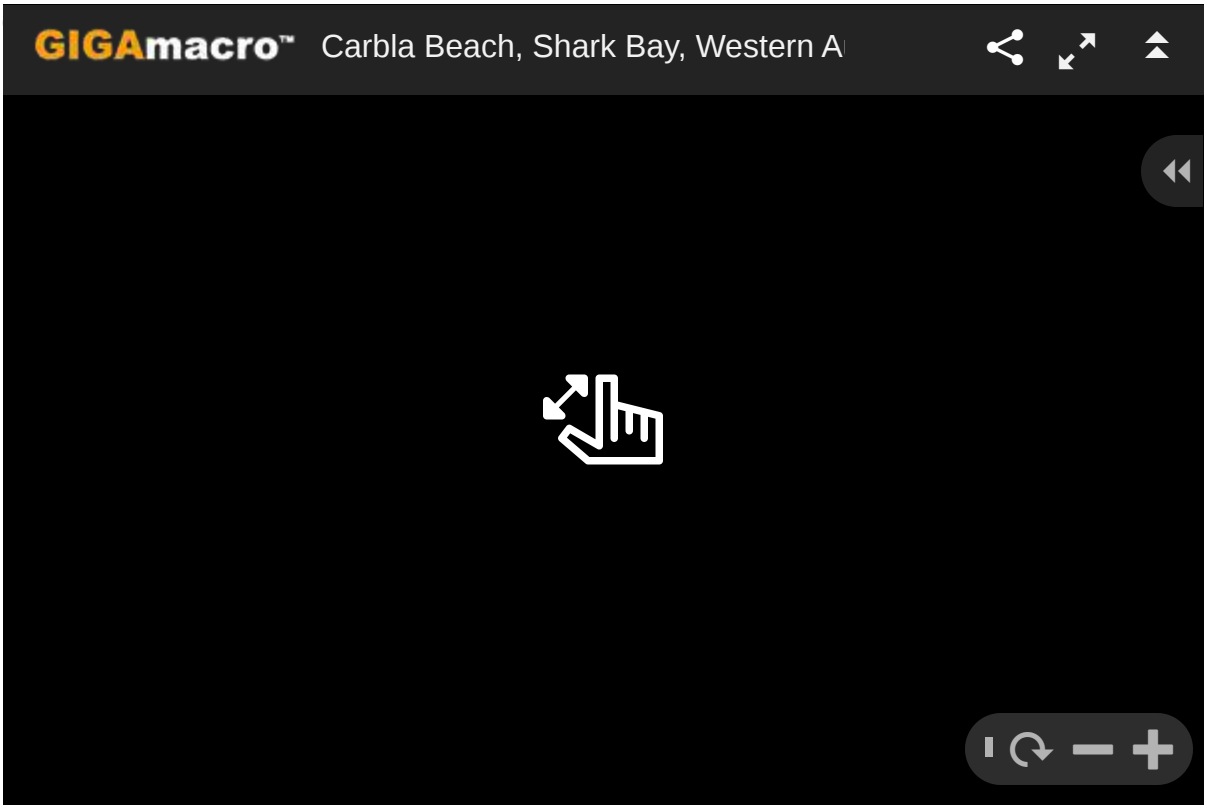
(Credit Ariel Anbar)



Two views of Shark Bay, western Australia, a place in the modern world where stromatolites are thriving; geologists infer the depositional setting of the Conococheague Formation would have been similar:



(Credit Ariel Anbar)

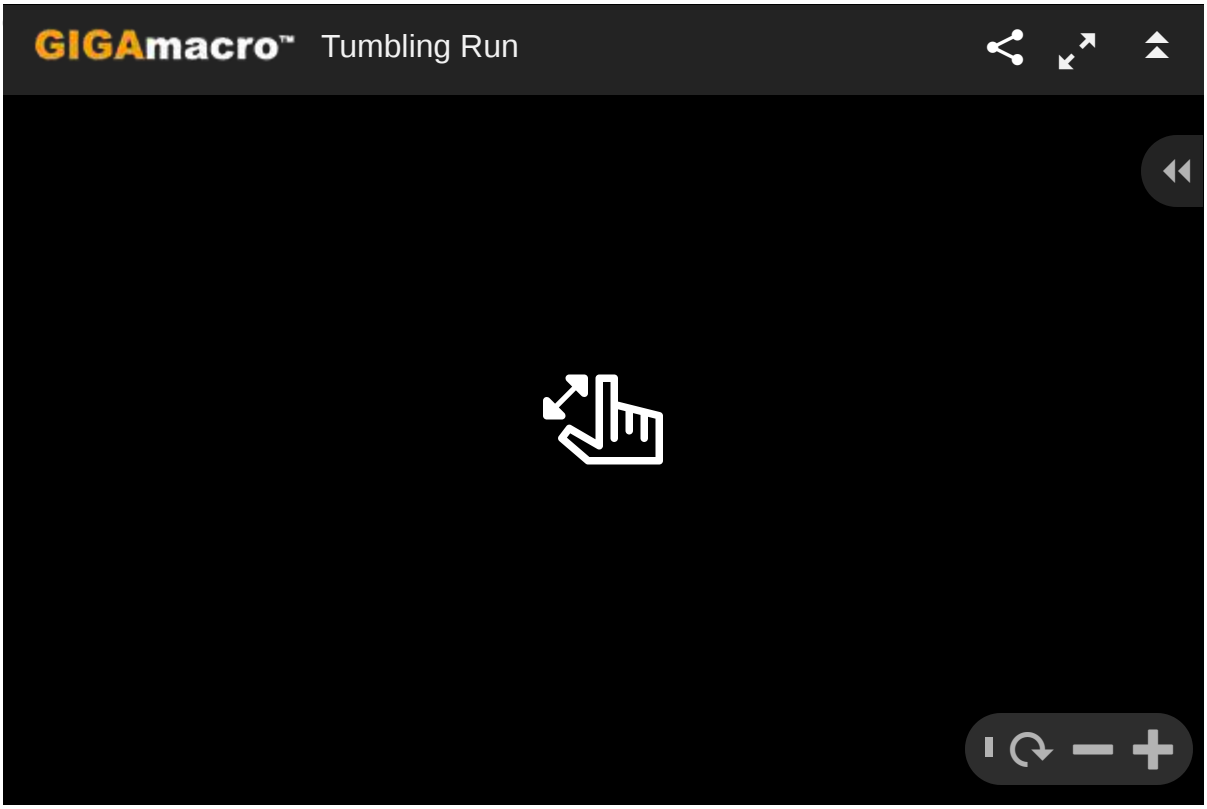


(Credit Ariel Anbar)

> Q: What is the “take home message” regarding our interpretation of the depositional conditions that prevailed during the deposition of the Conococheague Formation and related Cambrian carbonates?

Stop 2: Tumbling Run (New Market, Lincolnshire, & Edinburg Formations)

There are three formations exposed at a famous outcrop, east of Tumbing Run on Fisher’s Hill. The uppermost end of the outcrop (stratigraphically, the lowest/oldest layers) shows the contact between the New Market and overlying Lincolnshire formations:



The strata at Tumbling Run change as you walk up-strata (downhill) eastward through the outcrop: the proportion of carbonate decreases, and the proportion of clay and silt increases. They also get darker.

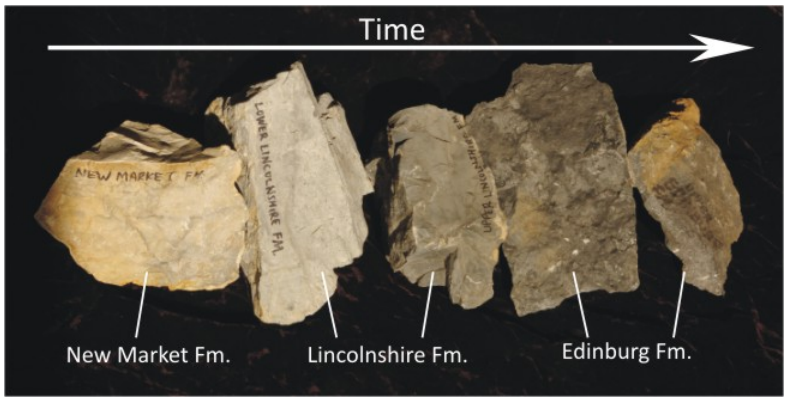


Figure 62.10.3: Over the course of the late Ordovician, light-colored shallow limestones gave way to increasingly dark deep-water limestones and shales.

In the Edinburg formation, there are a couple of layers that weathers away much more easily than the limestone above and beneath them! Upon closer inspection, these layer consists of tan to yellowish clay, very crumbly to the touch. Each is a **bentonite** layer, a layer of clay that has resulted from the weathering of a volcanic ash deposit.

Near here is another bentonite of extraordinary size. In North America, it is referred to as the “Millbrig bed.” In Europe, a similar bentonite layer is referred to as the “Big Bentonite.”



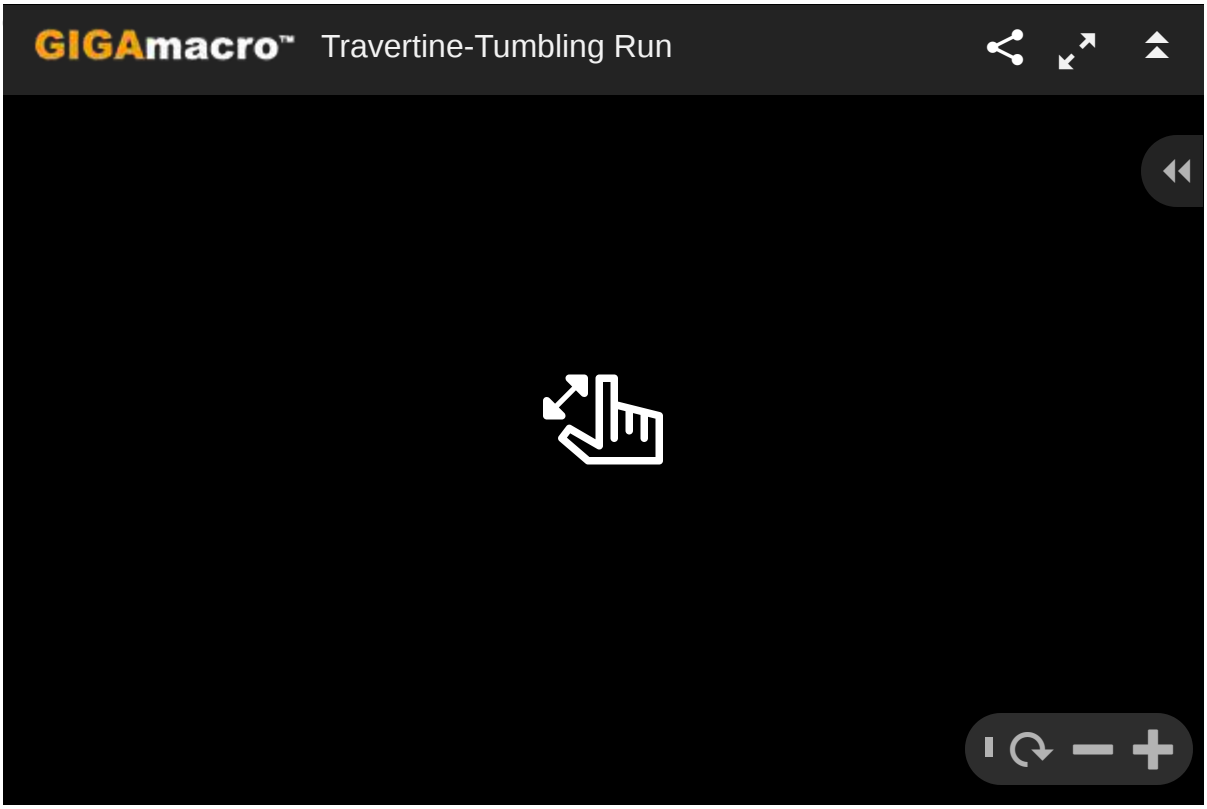
Figure 62.10.4: Late Ordovician bentonite layer between limestone layers in the Valley & Ridge province of northern Virginia.

The presence of bentonites indicates that volcanic ash was periodically being added to the sediments of the carbonate platform.



Q: What is the overall story recorded by the outcrops at Tumbling Run? (Hint: it changes from bottom/older to top/younger.)

As a tangential observation, it's cool to note that in Tumbling Run itself, Ca and CO_3 ions (dissolved from local source rocks such as those displayed in the outcrop) reprecipitate to form fresh deposits of travertine, like this:



Stop 3: Lower Martinsburg Formation on Route 11

This is a quick stop to see an outcrop of rock that is intermediate between the strata of Tumbling Run (Stop 3) and the strata we will see at Stop 4. There is limy shale of the Martinsburg Formation exposed at the intersection of Funk Road and Route 11.

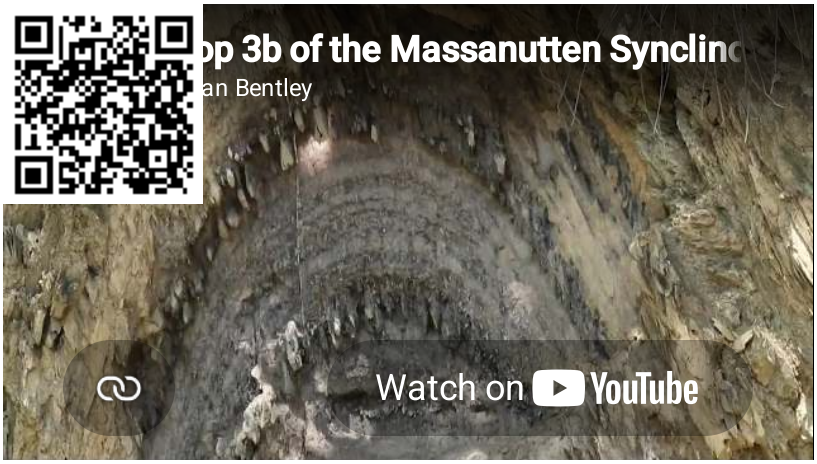


This is stratigraphically higher than the Tumbling Run site's uppermost strata:



This outcrop shows us that the story told at Tumbling Run (of less carbonate and more mud) continues. Along with the bentonites, this shift from carbonate to siliclastic sediment is the first clear signal that the Taconian Orogeny has started to the east.

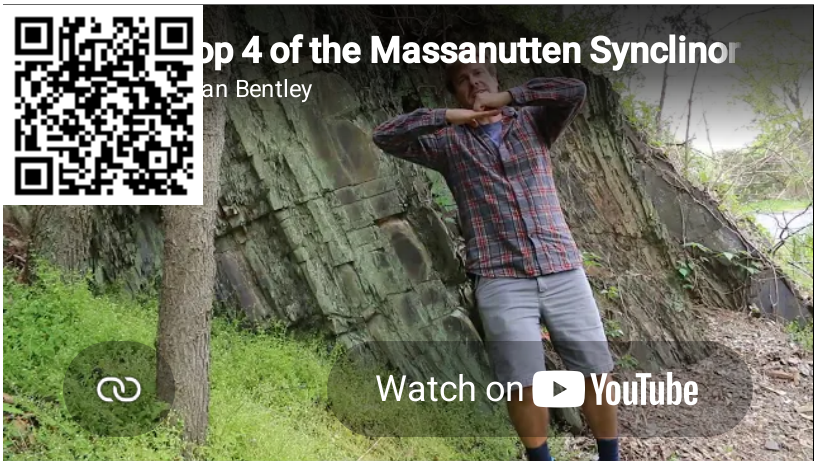
On the other side of the Massanutten Synclinorium, there is a sweet fold and and some very well developed slaty cleavage developed in this same set of strata. If you want a bonus stop, consider visiting the South Page Valley Road with this video:



Stop 4: Upper Martinsburg Formation on Millner Road:

The transition from 100% carbonate to 100% clastic sediment is complete by the time we get to Stop 4. This shift is a signficiant new source of clastic sediment (i.e. a mountain range) exists by the late Ordovician period of geologic time, and is shedding lots of sand and mud into the basin.

The basin itself has deepened, presumably due to the edge of the ancestral North American crust flexing downward upon being loaded with the Taconian mountain belt. In this deep basin, turbidity currents avalanched downward, bringing sand and mud from their mountain source area deep into the water to slowly settle out.



As evidence of this mode of deposition, here we see very steeply dipping strata of graywacke and shale (lacking any calcite), exhibiting graded bedding. This alternation between coarser and finer clay-rich clastic sediment and the characteristic graded bedding is interpreted as being due to deposition by turbidity currents:



Here is a graded bed from the Martinsburg Formation as seen in hand sample:



Another sample of graded bedding from the Martinsburg Formation, but cut into four pieces to show four sections through this single Bouma sequence:



> Q: How can you tell that the two previous GigaPans are shown “right side up?” Explain.

Overall, then, this stop shows us a great example of “flysch” (deep marine clastic sedimentation, where the sedimentary source is an adjacent mountain belt).

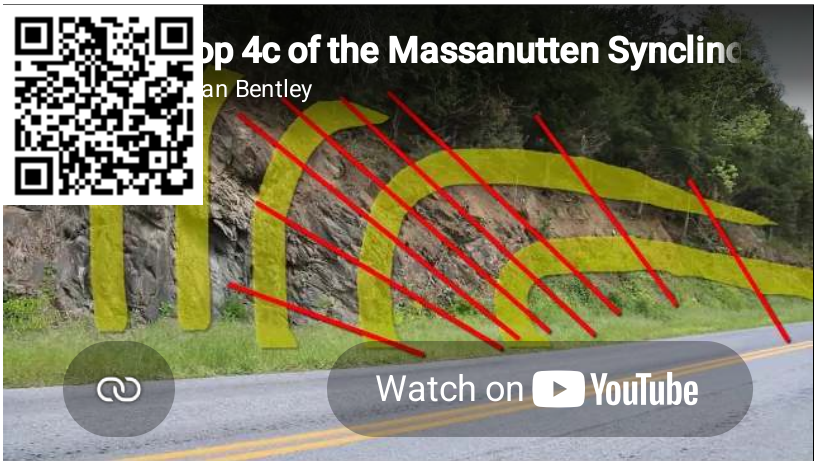
Before we leave, though, let’s stroll down the road for a couple of minutes to check out an outcrop that shows bedding/cleavage relationships in Martinsburg Formation:



Bedding here dips moderately to the right (east), and cleavage dips very steeply to the left (west). Though deformation isn’t our focus at Stop 4, it’s worth it to make note of it when we see it. Both the tilting of the sedimentary layers and the cleavage are both later tectonic signatures that overprint the original sedimentary story. Outcrops like this one tell two stories for the price of one! (They are “palimpsest” in character.) Though right now our focus is on the sedimentary signature of the Taconian Orogeny, the deformational structures should be noted as well. We will return to the deformational “overprint” story when we get to Stop 8.



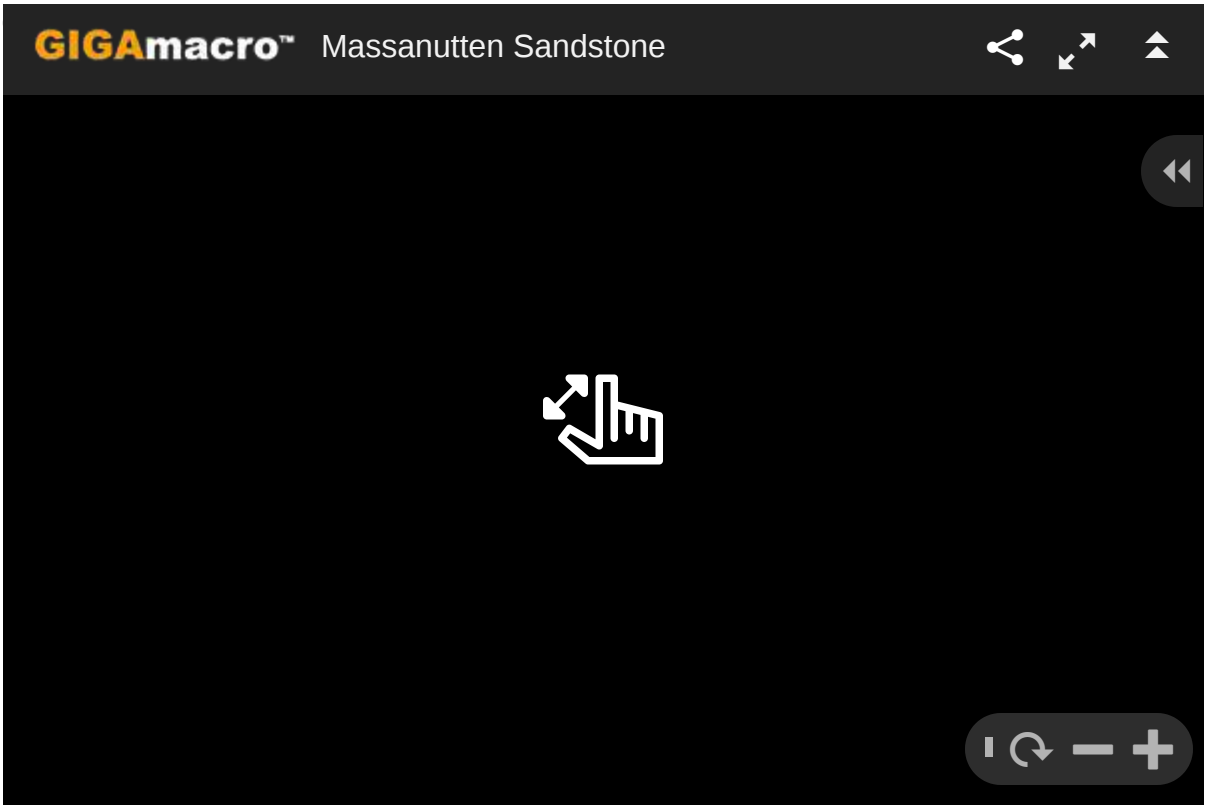
Reiterate key points from stop 4 with a bonus stop on the other side of the synclinorium:



Stop 5: Massanutten Formation at Passage Creek

Now it's time to enter the Massanutten Mountain system and the Fort Valley, by way of the narrow gorge that Passage Creek has carved below Buzzard Rocks. By the time we get to stop 5, the deepwater turbidites of the late Ordovician Martinsburg Formation that we saw at Stop 4 have given way to cross-bedded quartz sandstones of the Silurian-aged Massanutten Formation. The shift in sedimentary rock signals a shift in depositional setting. Let's examine the new rock outcrops and think about what sort of setting they might have accumulated in.

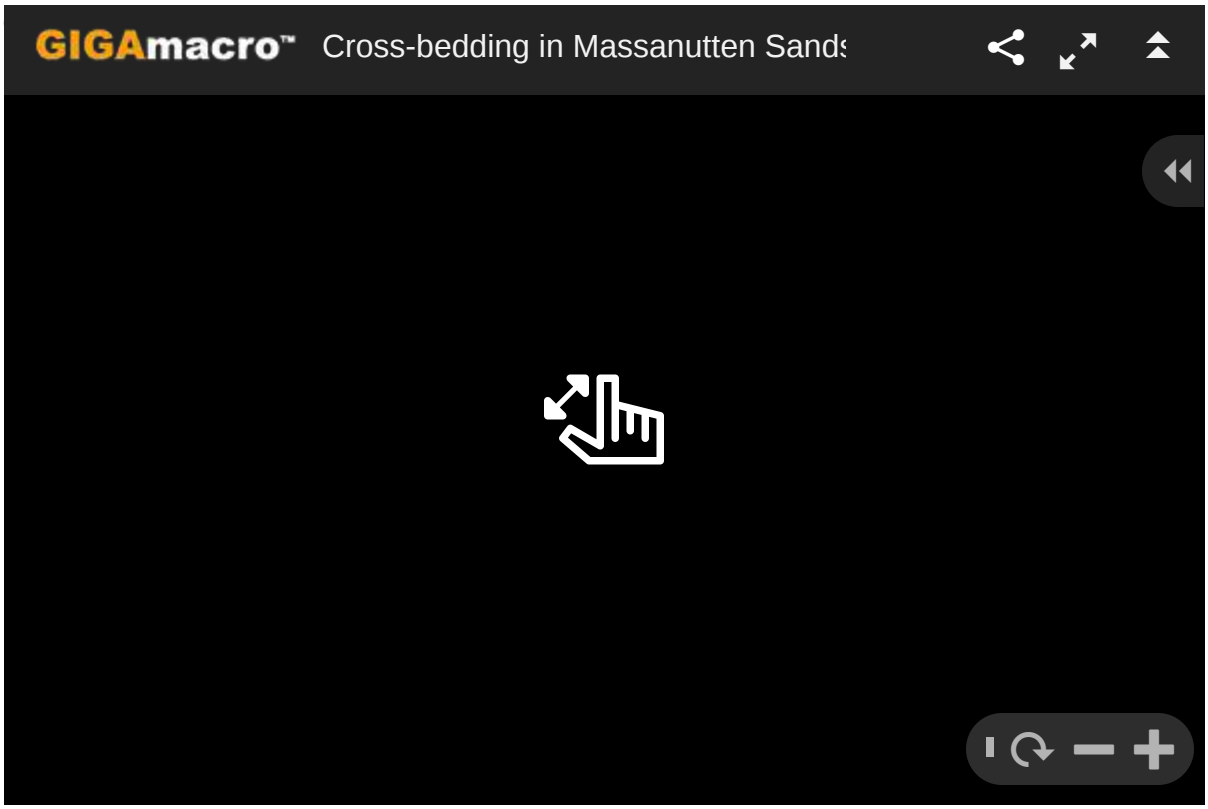
Outcrop #1 adjacent to Fort Valley Road, in the George Washington National Forest:



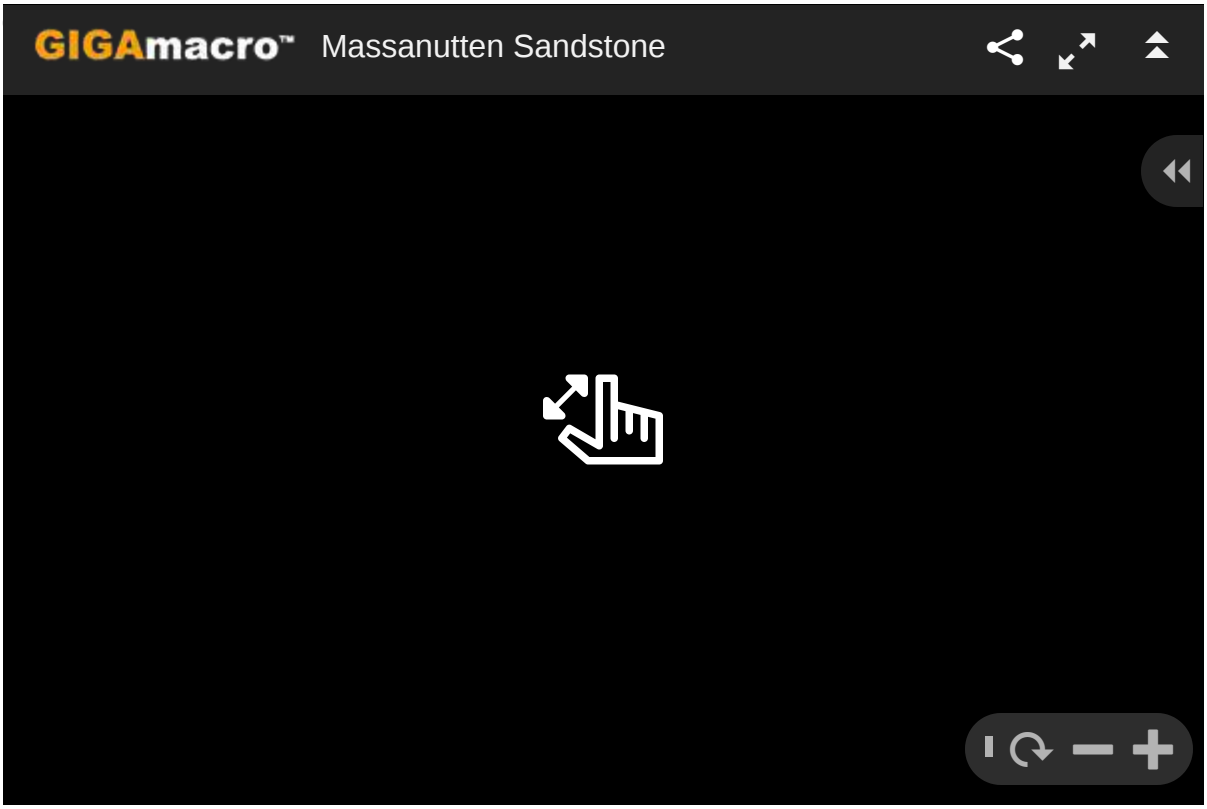
Don't let the brownish color of that outcrop throw you off – that is just surface staining by iron oxide; a thin veneer of rust on the outcrop surface that gives a false impression of the bright white rock's true color. As proof, consider this hand sample #1 of Massanutten Formation quartz sandstone:



So we have a quartz sandstone (a quartz arenite) that shows pretty good sorting and is dominated by the hard, chemically-stable mineral quartz. The individual grains vary between rounded and subangular in shape. So this is a fairly mature deposit of sediment. But what about primary sedimentary structures? Consider sample #2 of Massanutten Formation here, showing cross-bedding:



Outcrop #2, up the hill from outcrop #1: Can you find cross-bedding here?



> Q: Compare Massanutten Sandstone hand sample #2 to outcrop #2. Both show cross-sectional views of beds of sandstone. In each case: what way (what direction) was the current flowing when these strata were deposited (from left to right, or right to left)? What structure tells you this? Explain.

What sort of depositional setting could produce layer after layer of cross-bedded quartz sandstone like this? One interpretation is a vast braided river system, advancing outward over the now-filled-up flysch basin. The Massanutten Sandstone correlates with the Tuscarora Formation sandstone to the west, and so a tremendous amount of quartz sand was required to make such an extensive deposit.

Further west, an intervening formation, the red bed "molasse" of the Juniata Formation, occurs below the Tuscarora, recording advancing deltas, rivers, and floodplains. In the Massanutten Synclinorium, there is no Juniata, implying perhaps a local unconformity between the Martinsburg Formation and the Massanutten Formation.

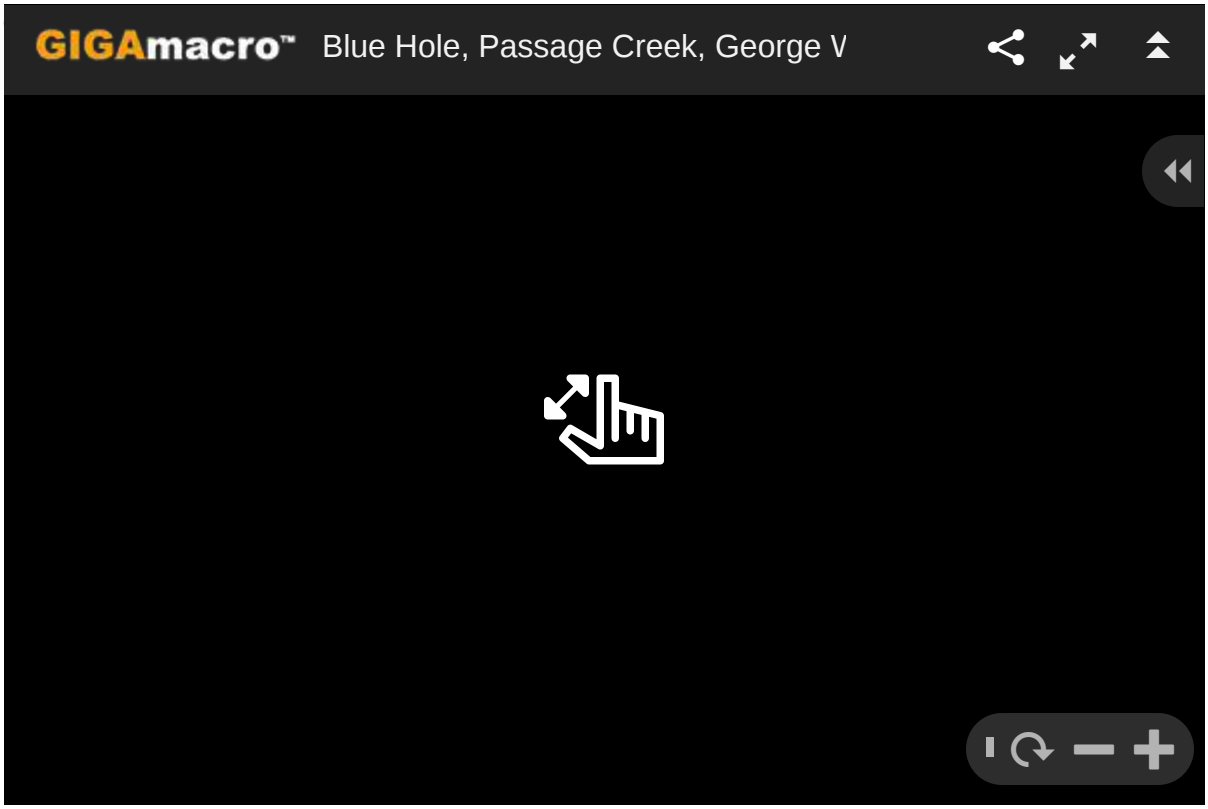
Primary sedimentary structures such as cross-bedding survived long enough to be lithified, but other more subtle structures may have been lost to bioturbation. Trace fossils indicate that there were animals disturbing the sediment, too. For instance, explore these trace fossils on a slab of in the Massanutten Sandstone: You're looking down on the bedding plane here, and the blue lines are annotations showing the trace fossils. They track together; play with it a bit to "train" your eye to see these structures:



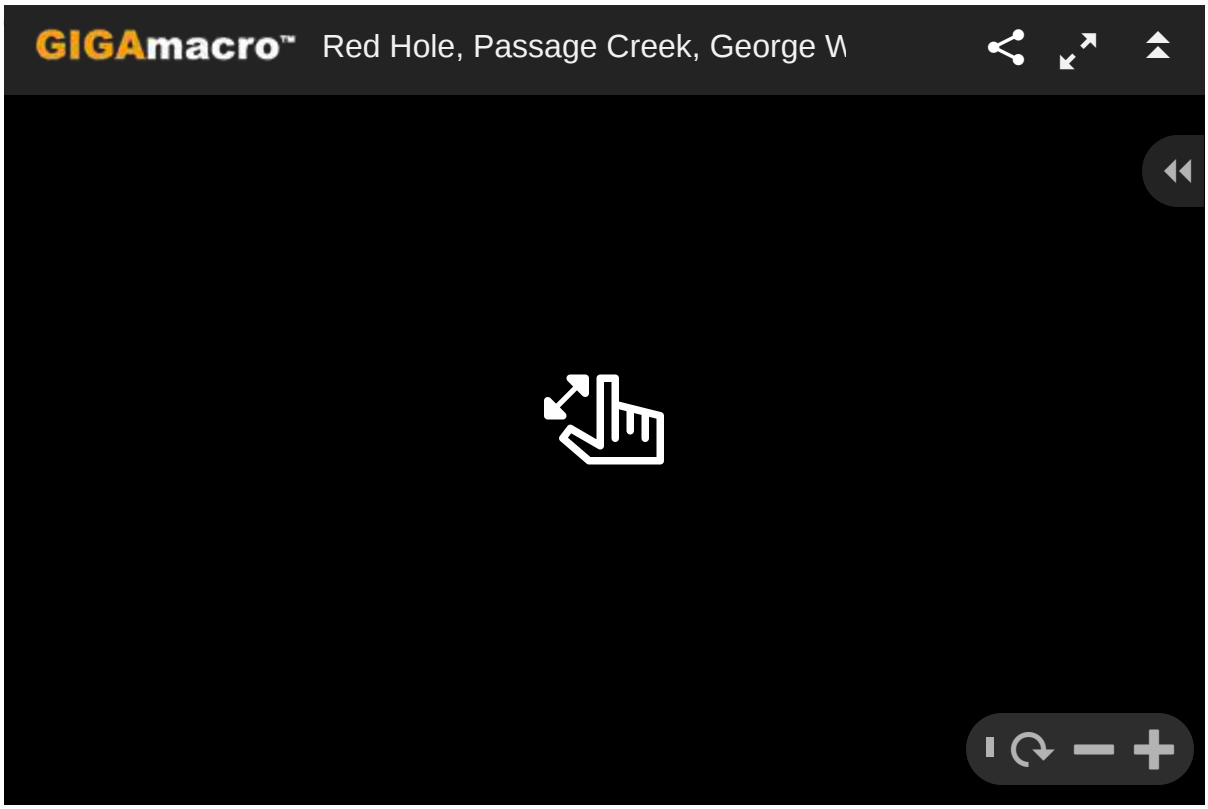
So there were creatures burrowing in the sediment at the bottom of this braided river system, prior to lithification.

Now, once again, let's briefly touch on deformational observations that can be made at this stop. As with the cleavage we observed at the end of Stop 4, these are the result of the late Paleozoic Alleghanian Orogeny. The overall regional structure (the Massanutten Synclinorium) is a result of that final mountain-building event. Its shape is hinted at by these two outcrops in Passage Creek:

On the east side of the syncline, the strata dip to the west:



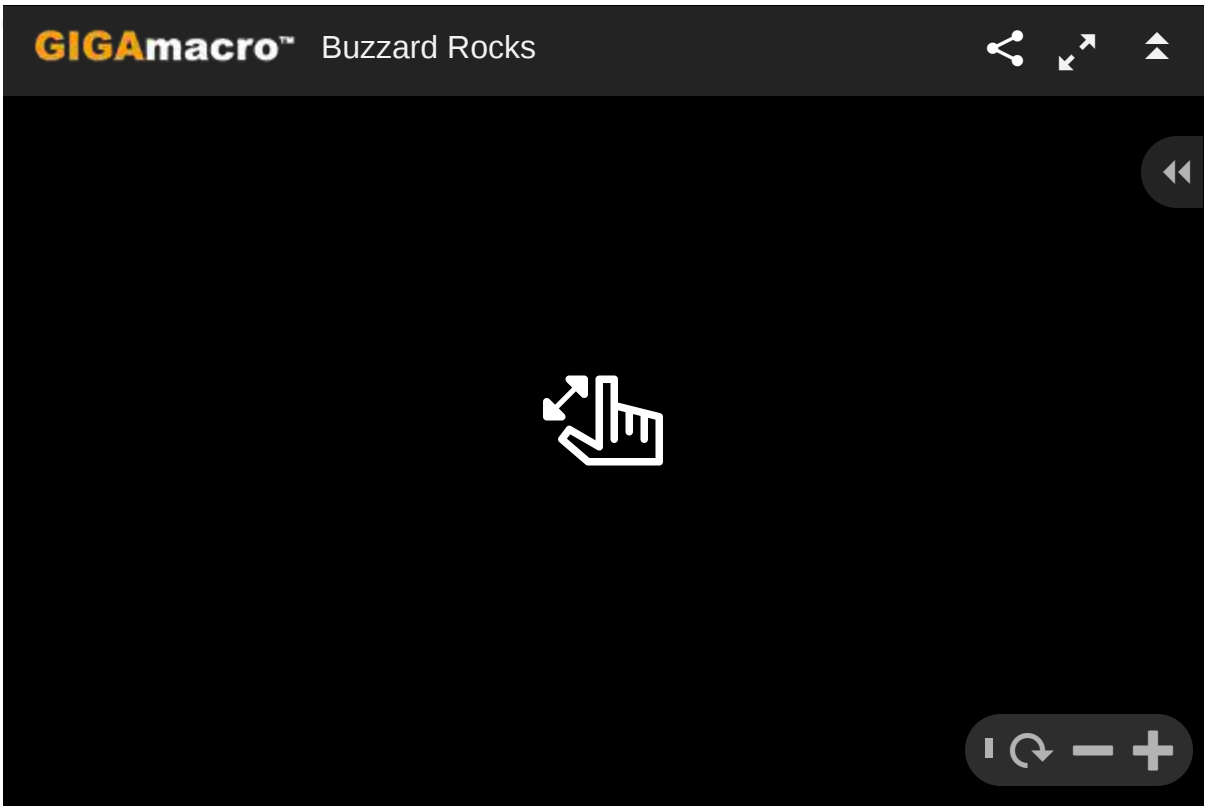
On the west side of the syncline, the strata dip to the east:

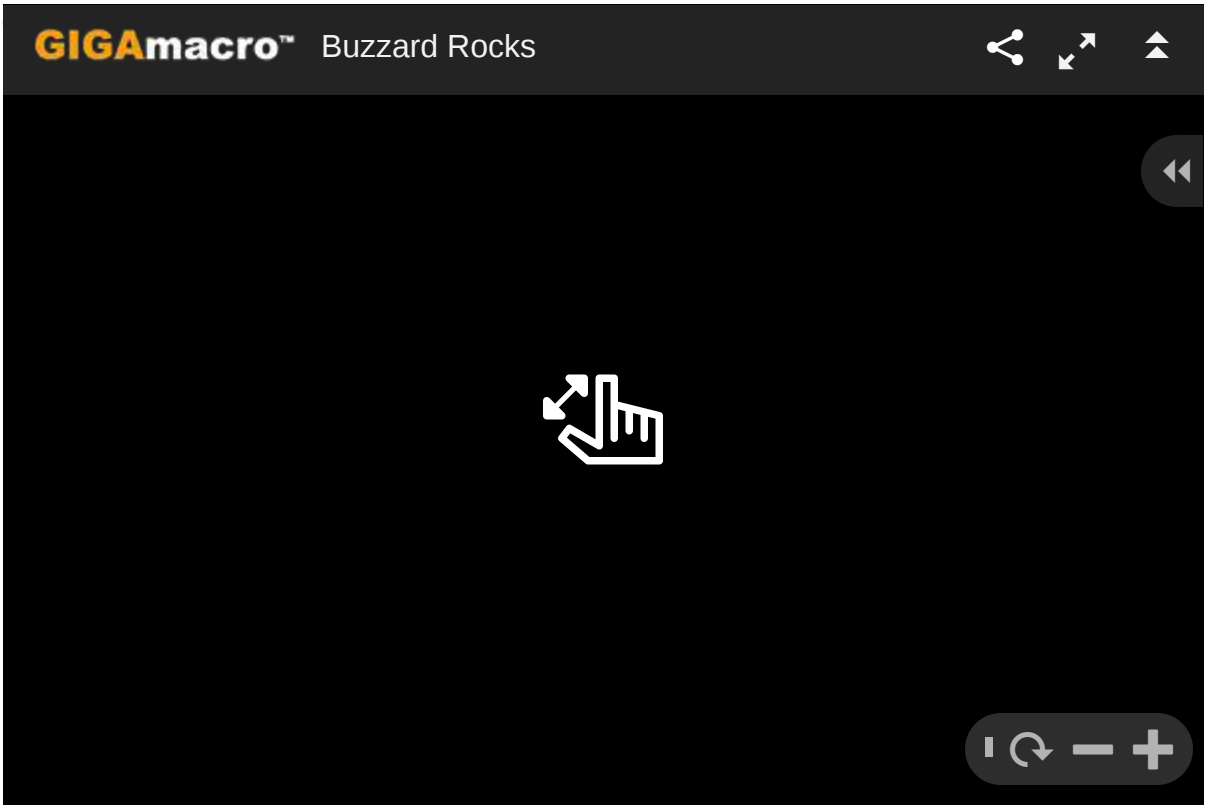


This location is essentially at the axis (or “hinge”) of the Massanutten Synclinorium: the line of greatest flexure of the strata. Those east of us dip downward to the west; those west of use dip downward to the east. Our structural perspective is much like that of an

ant at the bottom of a canoe, somewhere close to the stern: From our position, the layers of the Massanutten Formation move out and up, getting steeper the further away from us we look, right up to the crest of the mountain ridge.

As proof of that, consider these two views of Buzzard Rocks, the cliffs above and to the east of the outcrops we have just examined:



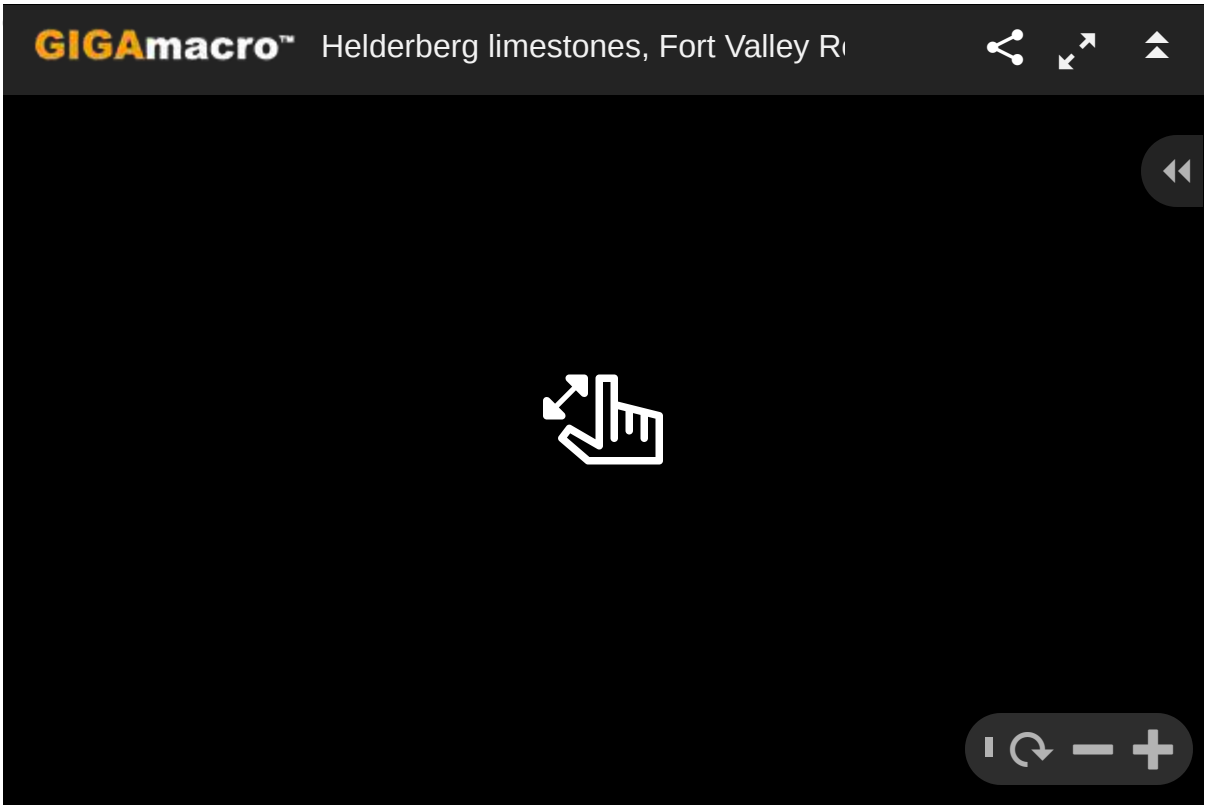


The point of devoting a bit of your attention to Buzzard Rocks is to note how very ***steep*** the beds of Massanutten Sandstone are at that location – almost vertical! (Did you find the rock climber on one of the cliffs in the left image?)

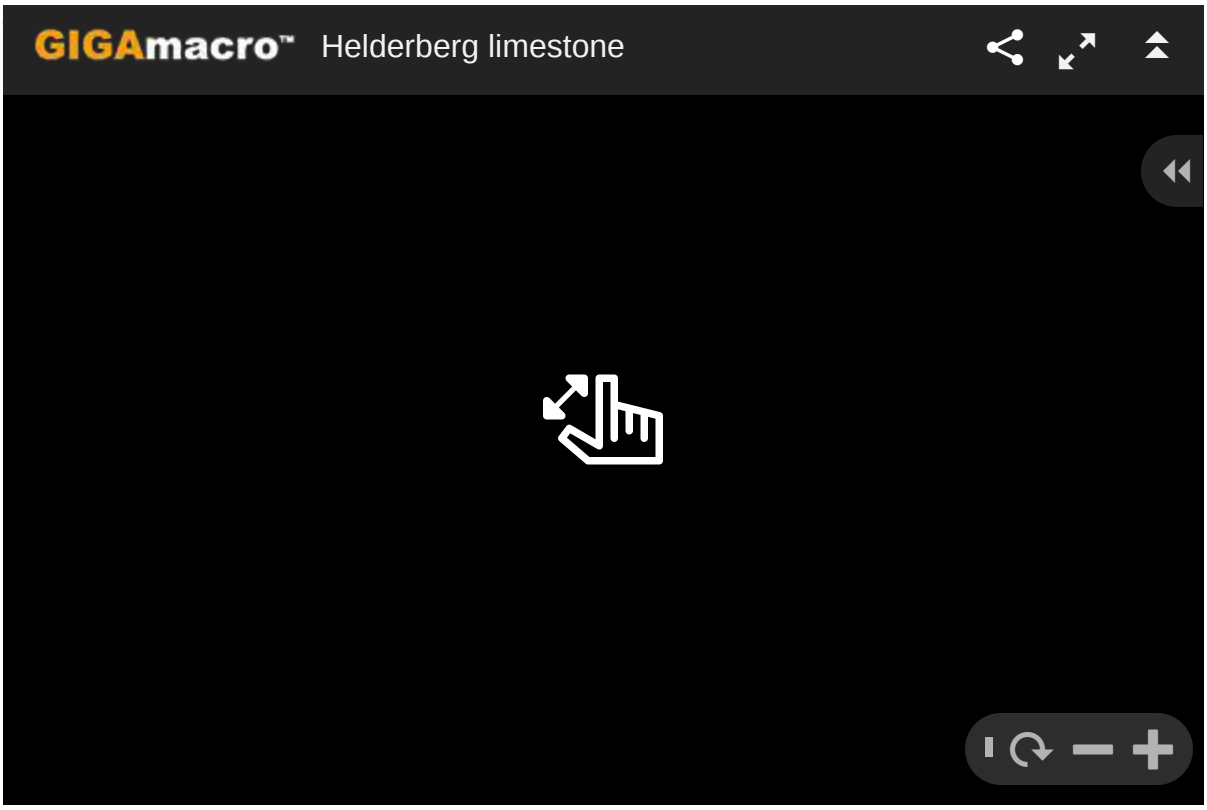
Stop 6: Helderberg Group limestones exposed on Fort Valley Road:

There isn't too much to see at Stop 6, and indeed it's not very safe to actually visit these outcrops in person, as there's nowhere safe to stand on the road. But they are outcrops of important rock layers for our overall story, so we want to make time for them. And there's no danger of being hit by a virtual speeding car on a virtual field experience!

These limestone strata are stratigraphically above the Massanutten Formation, and like the Martinsburg Formation outcrops at Stop 4b, this rock shows prominent cleavage due to post-depositional tectonic squeezing (during the Alleghanian Orogeny):



Next, let's focus on a cleaner slab of the Keyser Limestone, one of the formations within the Helderberg Group. Explore it for copious fossils; what marine invertebrates can you identify here?



These fossils help us establish the relative age of the Helderberg Group as late Silurian to early Devonian. The boundary between the two periods (based on conodont fossils) is found near the top of the Keyser Limestone.

So there doesn't appear to have been much sand or mud being deposited here during the early Devonian... What does that imply about the presence or absence of nearby mountain ranges?

> Q: Taken together, what do the Massanutten Formation and Helderberg Group have to say about depositional conditions during the Silurian and early Devonian? How does this relate to what we saw in the Martinsburg Formation? Explain.

Stop 7: Mahantango Formation

So if a return to carbonate deposition (Stop 6) implies that there are no longer nearby mountains to serve as a source of clastic sediment, what does it mean when clastic sediment begins to show up again in the mid-Devonian? That's the question that the Mahantango Formation should have us pondering....

Here's a black shale outcrop at the intersection of Fort Valley Road and Ramsey Road.



Clearly, the beds are thin, and the sediment is fine, and the color is dark: these are deep-water shales. Where's that sediment coming from, though? It turns out that phase 2 of Appalachian mountain building, the Acadian Orogeny, has gotten underway, and it's producing some sediment that has made it out to this location (in the future Fort Valley) to be deposited.

As with previous formations, organisms are living in this sediment. They left their traces...

Mahantango Formation showing trace fossils (orange-colored tubes), Fort Valley Road:



An outcrop of Mahantango Formation, showing characteristic carbonate nodules:



These nodules can get huge, up to 2 meters across. They are formed through diagenetic processes, and should not be considered as primary sedimentary structures. For whatever reason, they formed in the Mahantango Formation, but not other local shale units. So

they are distinctive marker features, unique to the Mahantango. Here's one of the big ones:



Mahantango Formation showing intersection of bedding and cleavage, with both body fossils and trace fossils, exposed near Trinity Church:



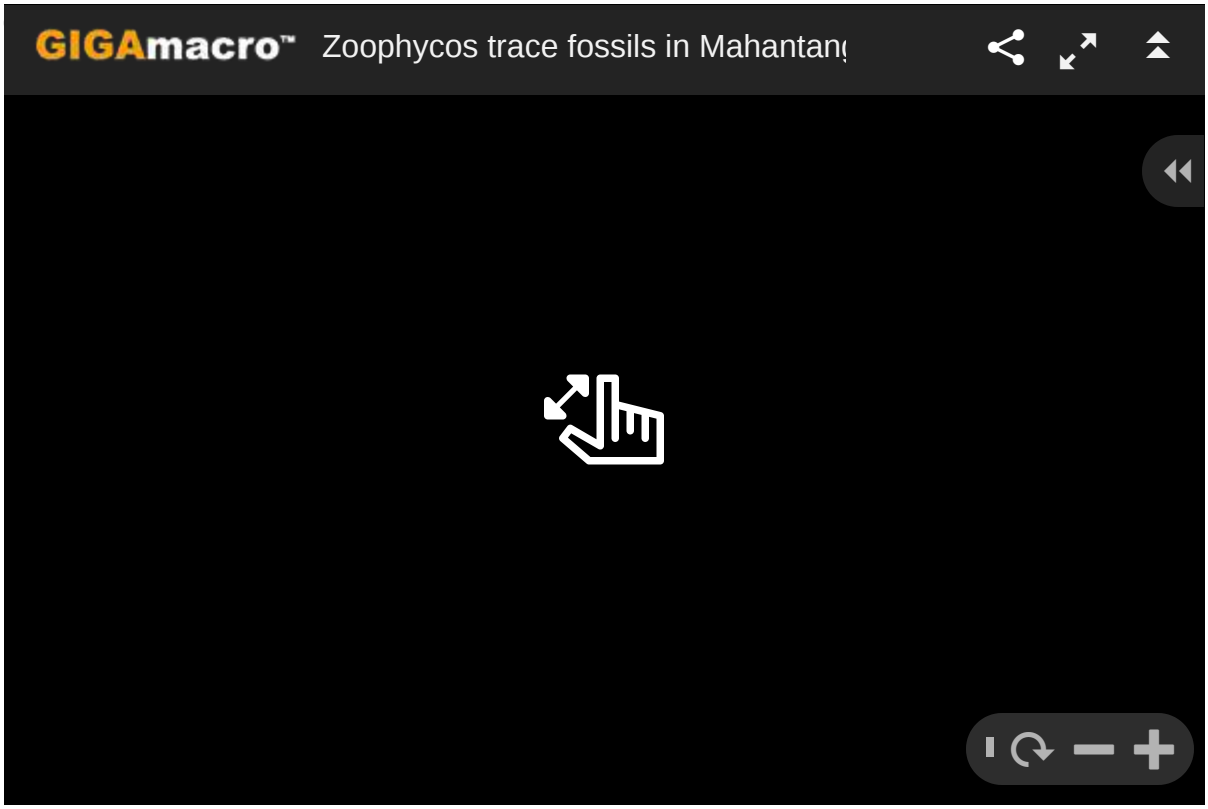
Mahantango Formation sample showing a large straight nautiloid fossil:



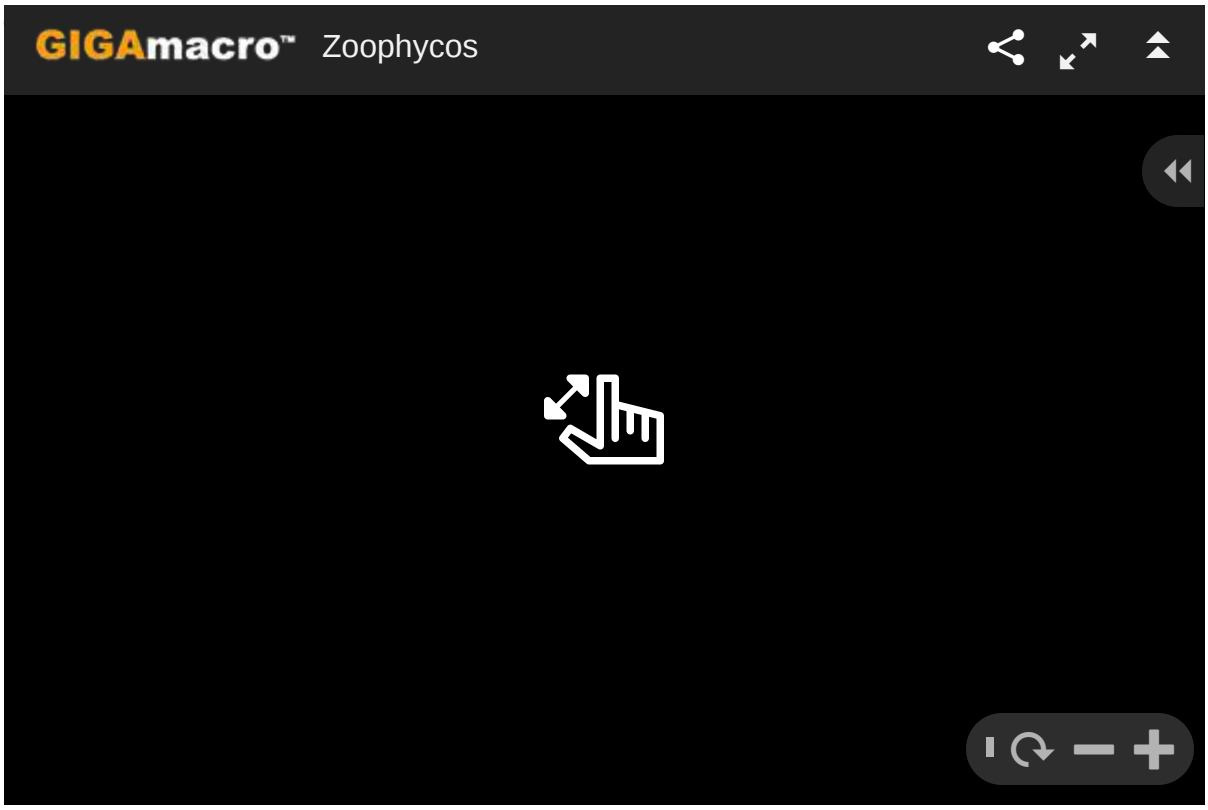
Mahantango Formation outcrop showing *Zoophycos* trace fossils:



Another Mahantango Formation outcrop showing *Zoophycos* trace fossils:



Mahantango Formation sample showing a nice example of the *Zoophycos* trace fossil:



Mahantango Formation sample showing a shell-rich horizon:



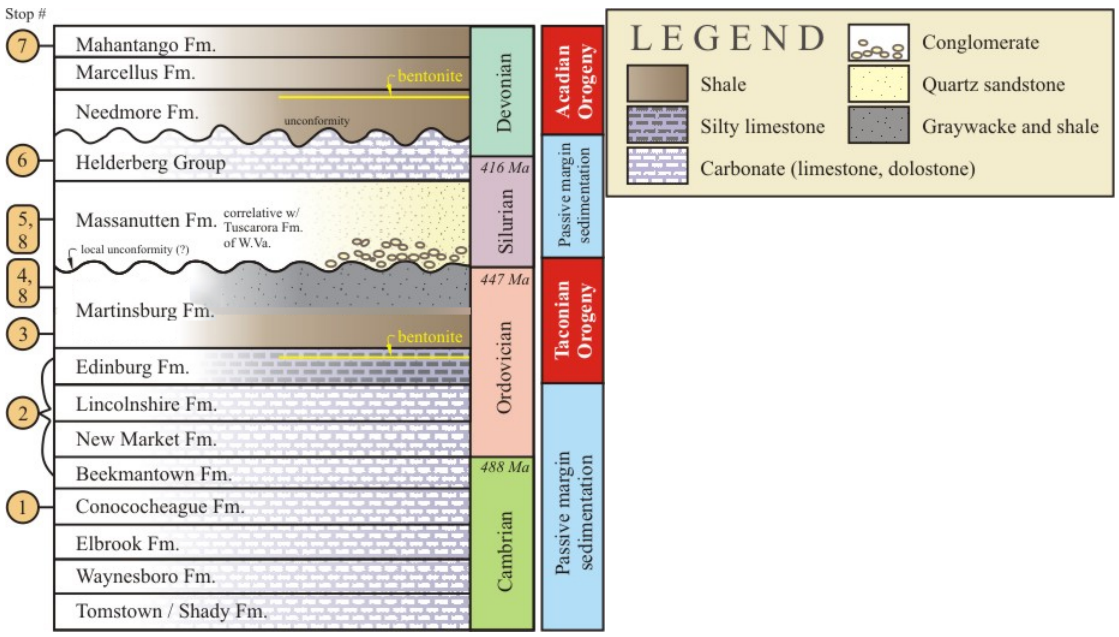


Figure 62.10.5: A conceptual stratigraphic sequence through the strata of this VFE, offering a tectonic interpretation in the right column.

Much of the Mahantango is fairly featureless black mud, with bedding rarely spotted except in shell-rich horizons like the one pictured above. These layers are thought to be storm deposits, where short-lived high-energy conditions took brachiopod shells from shallower places and tumbled them down into the deep, dumping them *en masse* and leaving them in an undisturbed state thereafter.

As a summary: for the second time in this trip, we've seen a transition from passive-margin carbonate deposition to tectonically-induced clastic sediment deposition. A new orogeny had intiated by the mid-to-late-Devonian.

Let us conclude our examination of the stratigraphic of this region by summarizing everything in a simplified stratigraphic column: the sequence, names, and lithologies of the various geologic layers are shown, as well as their timing within the appropriate periods of the geologic timescale, and finally each is marked with long-term tectonic interpretations on the far right.



> Q: How do the strata of the Mahantango Formation relate to the presence or absence of mountain-building during the late Devonian? How does this relate to what we saw in the Massanutten Formation and Helderberg Group? Explain.

Stop 8: Veach Gap, to see deformation induced by Alleghanian mountain-building

Though the overall structure of the region is synclinal (youngest strata in the middle; oldest strata on the flanks of the fold), there are smaller-scale anticlines and synclines superimposed on that larger structure. The anticlines in particular weather out well at a small water gap in Little Crease Mountain. This gap is called Veach Gap.



Here are four views of the anticlines exposed at Veach Gap (all are folds within the Massanutten Sandstone):

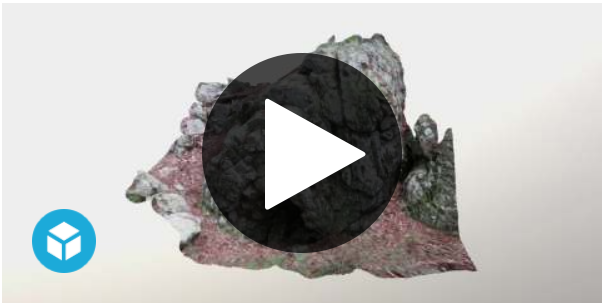

GIGAmacro™ Veach Gap Anticline 1


GIGAmacro™ Veach Gap Anticline 2

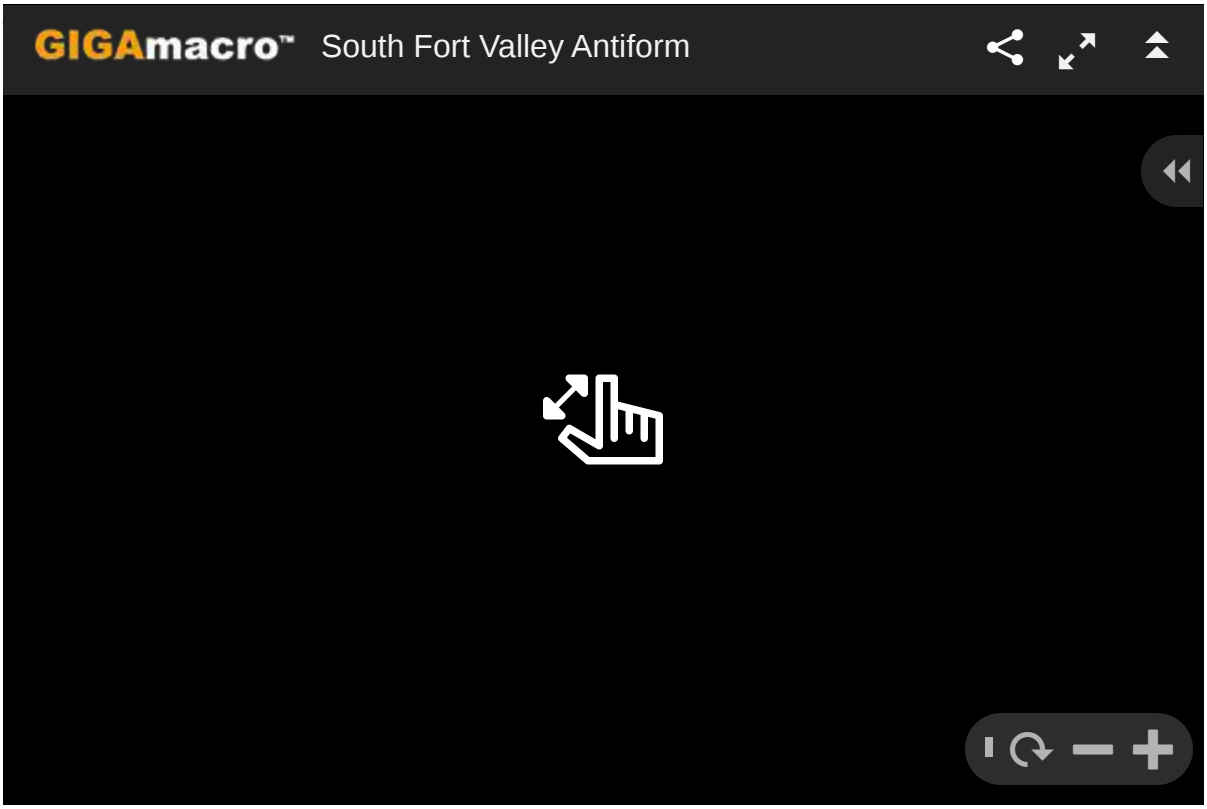




3D model of one of the Veach Gap anticlines:



As a bonus, here's an anticline developed in Needmore Formation shale, further south in the Fort Valley. Note the prominent cleavage "fan" superimposed on the sedimentary layering:



All of these structures challenge us to explain them by invoking the principle of original horizontality – these strata were originally deposited within a few degrees of horizontal. The fact that we see them contorted into such extraordinary positions now implies that they must have been deformed at some time after they originally formed.

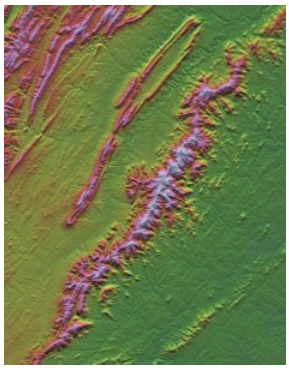


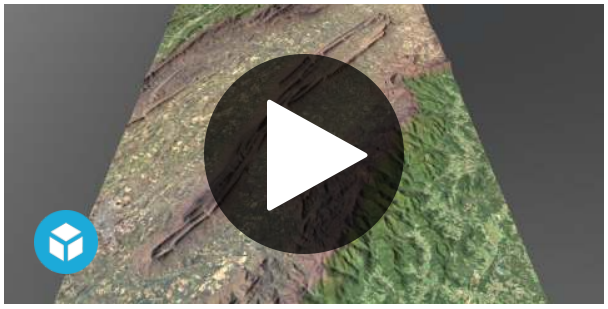
Figure 62.10.6: Differential erosion of different geologic units has resulted in the modern topography.

The folds we see today (as well as the tectonic cleavage, which forms due to pressure solution) is evidence these rocks got seriously squeezed at some time after they were deposited and lithified. The thing that caused the compression was the tectonic merger of ancestral North America with ancestral Africa (then the leading edge of the Gondwana supercontinent).

As ancestral Africa and North America converged, the Iapetus Ocean was snuffed out, and the resulting mountain belt grew upward and laterally, overtaking the sedimentary basins that formed during the earlier Taconian and Acadian events. The former mountain-adjacent sedimentary basin was now incorporated into the Alleghanian mountain belt's fold and thrust belt; *i.e.*, the future Valley & Ridge geologic province. Meanwhile, Blue Ridge rock was arched upward into a vast anticlinorium, and thrust upward and westward. The strata of the eastern Valley & Ridge province were compressed, causing them to fold into the enormous Massanutten Synclinorium, as well as undergoing penetrative pressure solution, particularly in shaley units, giving them a slatey cleavage.

Finally, weathering erosion went to work on this landscape of deformed rock, etching out some rock units preferentially relative to their neighbors. The Massanutten Sandstone, now folded into a crinkly shape, stood up tough against these forces. As the shales and limestones above it and below it were eroded away, it gradually rose above the landscape to make the long, sinuous form of Massanutten Mountain.

Here is a 3D model showing some of the terrain features that resulted from this differential weathering:



Thanks for exploring with us on this virtual field experience!

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62.11: The Moho (crust/mantle boundary)

VFE: the Moho, a trip to the crust / mantle boundary

A virtual field experience examining the crust/mantle boundary

Orientation: Moho Basics

This virtual field experience (VFE) will enable you to explore a unique boundary within the Earth's lithosphere. You will explore the Mohorovičić Discontinuity, the base of the Earth's crust. Ultimately, the goal will be to better understand how this boundary provides insight into the evolution of plate tectonics on our planet.

Jules Verne was wrong. No one has ever been to or seen the Earth's interior. Geoscientists know about its structure in two primary ways. First, through large blocks of rock that have been thrust up onto the outer crust tectonically. Second, through igneous materials that have been erupted volcanically. In this VFE, we will explore the boundary between the crust and mantle, a layer referred to as the "Moho", for short, or the Mohorovičić Discontinuity, more fully. Before you begin, you can read more about the Moho here.

You can get a glimpse of the Moho from the image below, where it is traced out with a green line. Keep in mind that while the Moho averages about 10km deep beneath oceanic crust, it can be as much as 70km deep beneath continental crust.

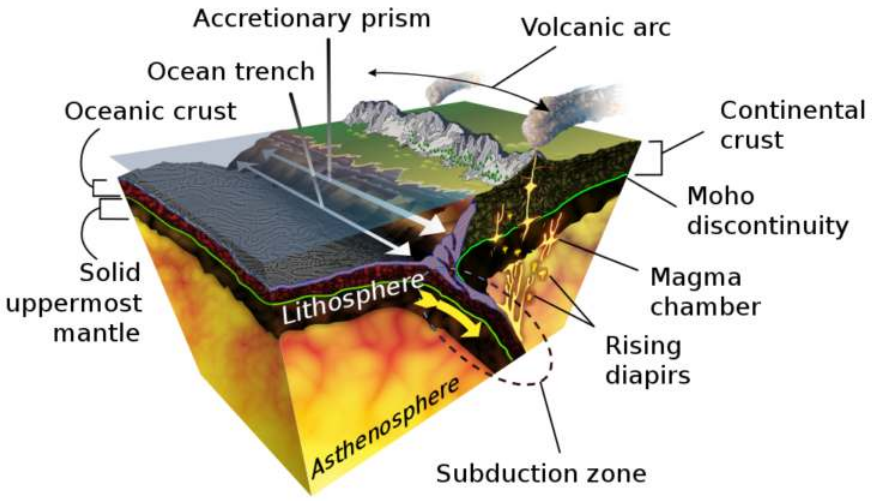


Figure 62.11.1: Earth's crust and mantle, Moho discontinuity between bottom of crust and solid uppermost mantle

The Moho's depth varies depending upon geographic location. In the geophysical map below, the depth of the Moho is evidently much greater in some places than it is in others. Can you see where it is shallowest and where it is deepest?

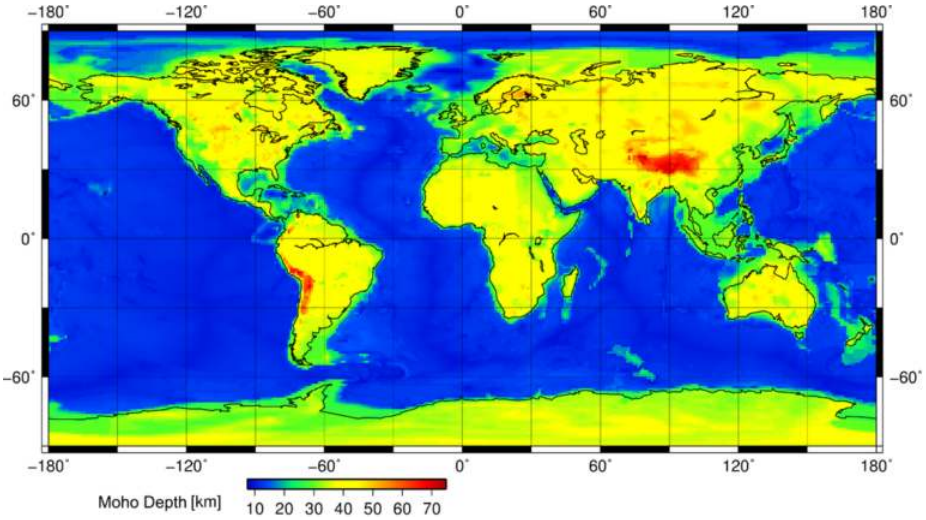


Figure 62.11.2: As shown in the figure, the Moho maintains a relatively stable average depth of 10 km under the ocean sea floor, but can vary by more than 70 km below continental land masses.

Moho: How Deep Is It?

? Exercise 62.11.1

Under what type of lithosphere is the Moho deepest?

a. Continental lithosphere

b. Oceanic lithosphere

Answer

a. Continental lithosphere



Figure 62.11.3: Two paths of a P-wave, one direct and one refracted, as it crosses the Moho.

The Moho is very identifiable using seismic data. Seismic waves are produced by energy released by earthquakes and other disturbances. Seismic waves are not unique to earth, Mars has seismic waves also that travel through the planet, as does our natural satellite, the Moon. On Mars, much of this activity is caused by wind. Wind can also cause seismic disturbances on Earth, but to detect anything at depth like the Moho, earthquake data is necessary. On any of these bodies, we can use seismometers to measure the waveforms and differentiate between them. P-waves, are also known as “Primary Waves” (first to arrive at a seismic station from the focus of an earthquake) or “Pressure Waves” (they push and pull the medium in which they travel as the energy propagates). In the image to the right, the P-wave velocity increases significantly at the Moho, from 7m/s to 8m/s. This sudden velocity change shows us that at the Moho, there is a significant change in the density of the material. The layer beneath the crust (the mantle) is more dense than the crust, and so the P-waves speed up when they enter it.

Moho: What is It?

? Exercise 62.11.1

What is the Moho? You now know it is between 10 and 70 km in depth. You know it represents a sudden change in seismic velocity between less dense material above and more dense material below. You know it is nearly everywhere. You also know that it is entirely contained within the Earth's lithosphere.

Is it a physical boundary or a chemical transition? Or, is it both?

a. Both Chemical and Physical - The Moho represents a disontinuity, or boundary, between two layers. These two layers are distinct both physically and chemically.

b. Physical transition - The Moho represents a discontinuity, or boundary, between two layers of distinct densities.

c. Chemical transition - The Moho represents a discontinuity, or boundary, between two layers of very different chemistries.

Answer

a. Both Chemical and Physical - The Moho represents a disontinuity, or boundary, between two layers. These two layers are distinct both physically and chemically.

A Trip to Oman: The Semail Ophiolite Complex

In this first VFE, our purpose is to introduce some material that we will revisit later, but to do so in a very cursory way. Thus, our main goal is to help you learn to make observations, sketch them, and relate them to actual processes that can be modeled and then explained. So, this may feel like an advanced activity, but try not to let that feeling get in the way of what the questions are asking you to do. Your main focus should be answering each question in a way that is satisfactory in ***your*** mind...not worrying about what you think your instructor wants.

So... Off to beautiful, magical, Oman! Below is a geological map of the Semail ophiolite complex. Please read about this unique rock complex and how it is being used by some researchers to answer important questions about microbial metabolism as it relates to geochemistry (life and geology interacting!). More purely geologically, the Semail ophiolite offers researchers an opportunity to investigate low-temperature metamorphism: how certain minerals interact with water.

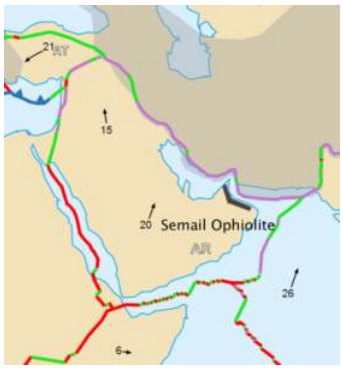


Figure 62.11.4: Location of the Semail Ophiolite Complex in Oman, near the eastern end of the Persian Gulf.

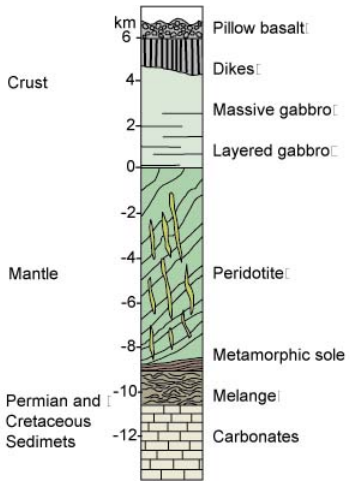


Figure 62.11.5: The Semail Ophiolite as a model for general ophiolite stratigraphy. Mantle rocks are topped by the peridotite and crustal rocks are marked at their base by layered gabbros. Thus, there is both a physical AND chemical differentiation between the mantle and crust.

An “ophiolite” is simply a chunk of oceanic crust thrust up onto land. Please explore how an ophiolite makes its way onto land at this link.

An ophiolite, as a section of oceanic crust, has a particular model of formation resultant structure. At right is a model of the stratigraphic and structural layering of the Semail Ophiolite.

You can see here that there are a bunch of new terms for you. For now, just know that these terms refer to either different kinds of rocks or features. For the moment, the only thing to focus on is that boundary between the peridotite (mantle) and gabbro (base of the oceanic crust).

Field Sketch: The Semail Ophiolite

Begin exploring this image and produce a “field drawing” of the ophiolite. Be sure to include in your drawing:

The “OASIS” method:

Orientation

Annotation

Scale

Information

Sketch what you see!

Here is a nice introductory video on making geological field sketches:



Below, you will see a GigaPan image of the Semail Ophiolite. You can view the image in its native location here. You are looking at a very rare and unusual mountain. Pan around, zoom in, and see if you can find the Moho. Hint: The boundary is located in a place where you can see the lighter-colored lithospheric upper mantle rocks separated from the darker-colored crustal gabbros.

Once you complete this, annotate your drawing by identifying the main boundary between the mantle and crustal rocks and what you see at this field site.

Here is the field site for you to sketch below:



Where's the Moho?

Exercise 62.11.1

About how far from the top of the mountain is the Moho, the boundary between lighter-colored upper mantle rocks and darker crustal gabbros?

a. 1000 ft

b. 100 ft

Answer

b. 100 ft

Field Specimen 1: Peridotite

So, you decide to climb up to the base of this mountain to collect a specimen to compare the rocks below and above the MOHO. While climbing, you pick up this lovely piece of peridotite along the way. Initially, you note that the peridotite sample is "coated" with a film of basalt (kind of like a chocolate-covered peanut):



Make some observations of the specimen above after reading about the rock Peridotite.

Peridotite: Phaneritic or Aphanitic?

? Exercise 62.11.1

After reading the article at the link above, and examining the size of the mineral grains in the hand sample (> 0.5 mm), you determine the rock is coarse-grained (phaneritic). What does this say about its genesis?

a. It cooled rapidly and at a shallower depth, somewhere near the surface.

b. It cooled slowly deep within the Earth, in this case, the upper mantle.

Answer

b. It cooled slowly deep within the Earth, in this case, the upper mantle.

? Exercise 62.11.2

What is the dominant mineral grain in this specimen? Consider the color, the location of its formation, and the rich nature of iron and magnesium at that location within the Earth.

a. Plagioclase

b. Anorthosite

c. Amphibole

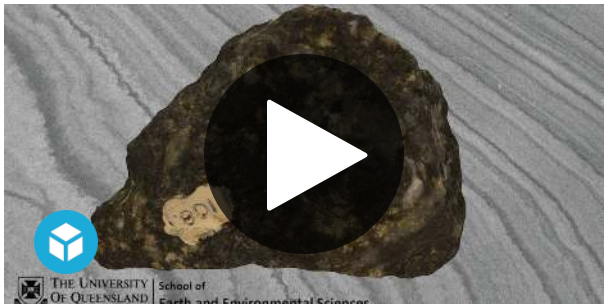
d. Olivine

Answer

d. Olivine

Field Specimen 2: Gabbro

After taking a break and a long drink of water, you continue to venture up the mountain to areas above the Moho. In fact, you climb directly to the top! What a view! Once you're there, you look around for some good specimens. You find this one:



What do you notice right away that's very different? You can click on the numbers on the rock also – they will give you some clues.

Gabbro: The Dominant Minerals

? Exercise 62.11.1

In this speciment of gabbro, what are the dominant minerals?

a. Augite and Plagioclase

b. Olivine and Pyroxene

c. Quartz and Orthoclase

Answer

a. Augite and Plagioclase

? Exercise 62.11.2

Examine the specimen closely. Are the mineral grains clearly visible and larger than about 0.5mm, making the rock coarse-grained (phaneritic)? Or, are the grains much smaller, <0.5mm, and thus fine-grained (aphanitic)?

a. Coarse-grained (Phaneritic)

b. Fine-grained (Aphanitic)

Answer

a. Coarse-grained (Phaneritic)

? Exercise 62.11.3

After examining the size of the mineral grains in the hand sample, you determine the rock is coarse-grained (phaneritic). What does this say about its genesis?

a. It cooled slowly deep within the Earth, in this case, the upper mantle.

b. It cooled slowly deep within the Earth, in this case, the deepest oceanic crust.

Answer

b. It cooled slowly deep within the Earth, in this case, the deepest oceanic crust.

Leaving the Field: Considering the Moho

After visiting the Semail Ophiolite Complex, drawing the Moho as it appears in outcrop, and considering the specimens you collected, you leave with a better understanding of a part of the Earth's interior that is otherwise invisible to you. You have learned that the Mohorovičić Discontinuity is a real seismic boundary within the Earth. It is located at different depths within the Earth's lithosphere. In oceanic lithosphere, it is shallower and in continental lithosphere, deeper. It exists because their densities are different due to changes in physical and chemical attributes, which are a function of their different compositions.

You might be left with some questions:

1) Is this compositional boundary deep within the Earth so distinct everywhere?

Actually, the gigapan you sketched above is a bit misleading! The Moho is considered to be a transitional ***zone*** between the crust and mantle that is, on average, about 500m in thickness. It looks very crisp and neat in the gigapan. How did you draw it in your sketch?

> 2) Has the Moho always existed?

Probably not, but we really do not know.

> 3) Does the Moho exist everywhere?

It does not! It is noticeably absent under volcanic island arcs. This is due to the process of **delamination** **(separation)**, the detachment and descent of lower crust into the asthenosphere, possibly due to lateral flows of mantle material due to subduction. Japan is an example. The Philippines and New Zealand are others. Interestingly, the Moho is nice and crisp under the eastern United States. Considering that this region is composed of island arcs that accreted onto the ancestral continent between 460 Ma and 250 Ma, this might be surprising. One hypothesis that attempts to explain this suggests that once the volcanism ceases, so does the delamination of the lower crust. Under island arcs like Japan, there is no Moho due to delamination due to active volcanism and the processes that feed that. Under the eastern United States, a passive margin, there is no volcanism. So, the lower crust is wedded crisply to the upper mantle.

The Moho in Earth's History

Much about the evolution of plate tectonics remains unknown. It is an active field of study. Seismic data, particularly in the form of seismic tomography (kind of like CAT scans of the Earth), is rapidly filling in our knowledge of the current situation. How we see our planet's interior today, coupled with what we know of its past from the rock record, is giving scientists the ability to look back into these earlier times.

The Moho plays a role in this. Is characteristics tell us important things about the Earth's past and its present, their similarities and differences. As far as we know, there has been a Moho if there was solid crustal material in existence. Its depth and even the rocks on either side of it may have changed with time and the evolution of plate tectonics. What has not changed is the nature of the boundary – it is a disontinuity between two layers of distinct densities that result from relatively sudden changes in physical and chemical properties.

For Further Reading

King, Hobart (2020). Peridotite. Geology.com. Retrieved from: https://geology.com/rocks/peridotite.shtml

King, Hobart (2020). Mohorovičić Discontinuity. Geology.com. Retrieved from: https://geology.com/articles/mohorovicic-discontinuity.shtml

Rock Powered Life (2020). Semail Ophiolite, Oman. University of Colorado, Boulder, Rocker Powered Life. Retrieved from: https://www.colorado.edu/lab/rockpoweredlife/research/theme-i-field-investigations/samail-ophiolite-oman

Volcano World (2020). Ophiolites. Oregon State University Volcano World. Retrieved from: http://volcano.oregonstate.edu/ophiolites

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62.12: A Plate Boundary Odyssey

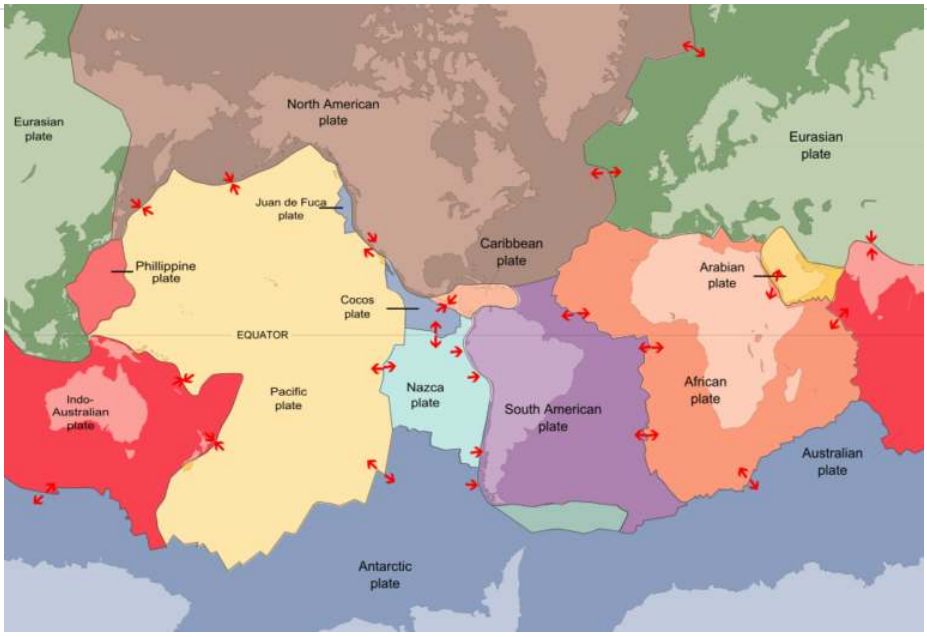


Figure 62.12.1: Plates and plate boundaries with a general sense of motion identified.

Around the World in Six Plate Boundaries

In this VFE, we will virtually travel to examples of all of the major types of plate boundaries, including:

1) Convergent Boundary, Continental-Oceanic (Franciscan Complex, San Francisco, USA)

2) Convergent Boundary, Oceanic-Oceanic (Semail Ophiolite, Oman)

3) Convergent Boundary, Continental-Continental (Himalayas)

4) Divergent Boundary, Oceanic (Iceland)

5) Divergent Boundary, Continental (East African Rift)

6) Transform Boundary (San Andreas Fault)

At each of these sites, you will be asked to use your powers of observation to explore some of the field geology produced at such locations.

Convergent Boundaries

Continental-Oceanic Convergence: The Franciscan Complex

For a more in-depth geological analysis of this plate boundary, read this case study on the Franciscan Complex here.

The first image below is of an artist's rendering of an idealized subduction zone displaying the "Franciscan Complex" in situ. This complex of rock is a part of the accretionary wedge, or mélange, associated with a subduction zone. The subduction zone is no longer active; it shows us rocks that formed during the Mesozoic Era. In this case, the subduction was a classic situation of oceanic-continental convergence. Study the diagram below. Consider what kinds of other rocks you might seek nearby if you wished to confirm your analysis of an outcrop of Franciscan Complex.

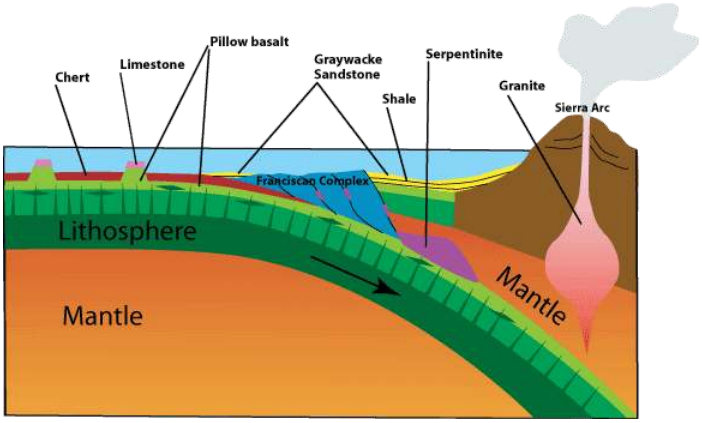


Figure 62.12.2: The geologic setting of the Franciscan Complex. The accretionary prism units, including the mélange, are indicated in blue.

The Franciscan Complex is a body of rock that, as you can see below, is in part made up of a metamorphic rock called serpentinite (formed under very intense pressures when mantle peridotite reacts with water), along with chunks of metamorphically altered ancient basaltic seafloor mixed with shelf and seafloor sediments such as cherts and mudstones. All of this is stirred up and mashed ocean floor sediment metamorphosed as one plate dove beneath another. This is what is referred to with the French term "mélange." Translated, it means "mixture." Mélange is a sheared-out mixture of heterogeneous seafloor scrapings.



One type of rock that makes mélange in subduction zones is serpentinite. Serpentinite is a metamorphic rock that is created through the hydrous alteration of ultramafic rocks under great pressure. Large rounded chunks of seafloor are included within the serpentinite mélange. Study the gigapixel panorama above. In addition to the serpentinite you see, if this is indeed a plate boundary where chunks of seafloor are being recycled, what other rock types might you expect to find contained as sheared-out chunks within the mélange?

Plate Boundary 1: Subducting Seafloor

? Exercise 62.12.1

If oceanic lithosphere is subducted beneath continental lithosphere, and mélange deposits are a part of the result. What other bits of rock would you expect to see included as the mélange is produced?

a. Evaporite deposits

b. Metamorphically altered blocks of basalt and seafloor sediments

c. Igneous materials resulting from local volcanism

Answer

b. Metamorphically altered blocks of basalt and seafloor sediments

You can learn more about California's Coast Ranges and their accretionary wedge rocks in this case study.

Oceanic-Oceanic Convergence: The Semail Ophiolite Complex

For our second stop, we will examine another convergent boundary situation, but this time it will be only oceanic lithosphere that was involved. We now journey to Oman, where we can examine the rocks within the Murri Anticline to help us understand an ancient oceanic-oceanic subduction complex. Do you remember what that layers of rock composed an ophiolite... or chunk of ocean floor? You can learn about the nearby Semail Ophiolite Complex in another VFE.



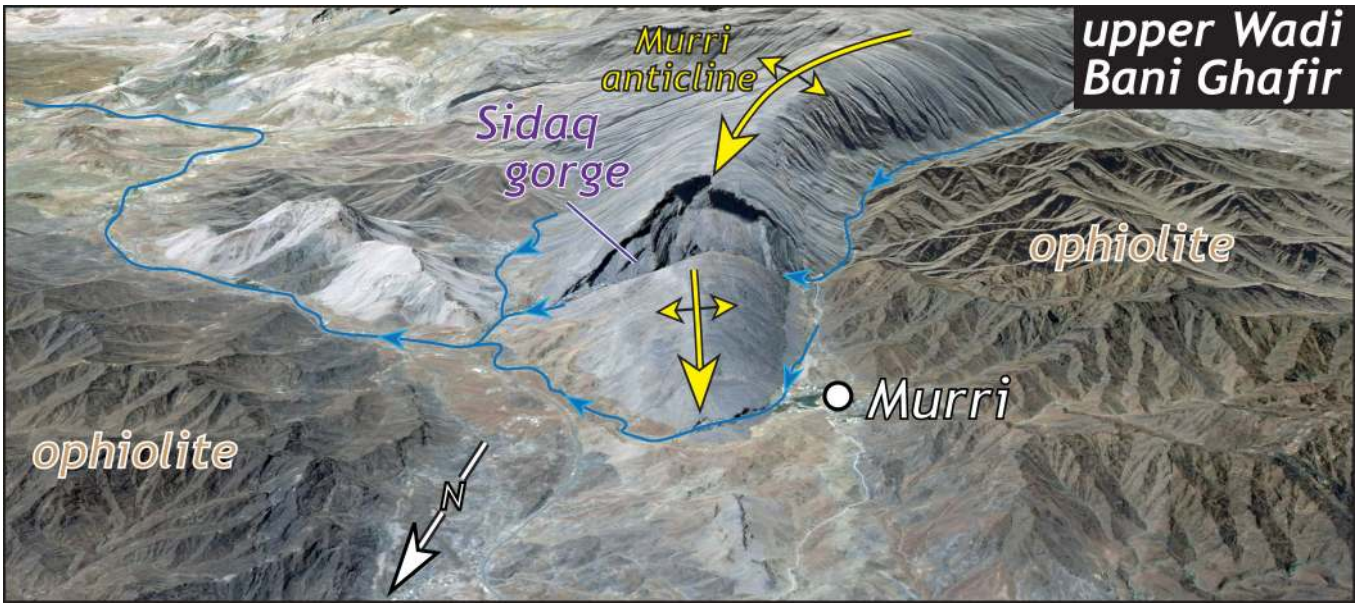


Figure 62.12.3: Sidaq Gorge, Murri Anticline, Oman (Image courtesy of Chuck Bailey, College of William & Mary Geology. Used with permission.)

This site is upper Wadi Bani Ghafir. First, note that a "wadi" is Arabic for a dry stream bed. Note how Wadi Ghafir is cutting right through the uplifted anticline. How did an intermittent stream like this manage to cut through a folded hillside? Wadi Ghafir behaves like any other stream above its base level, downcutting as it flows. You might think that the fold (the Murri Anticline) must be younger than the stream, warping upward in more recent times as the stream downcut into it. However, was not the case, as the anticline was formed as the two blocks of oceanic lithosphere converged over 65 million years ago!

In the aerial image above, you can see how ophiolite blocks, or chunks of oceanic lithosphere, surround the limestone layers of the Murri Anticline. The ophiolites were obducted onto the contemporaneous Cretaceous ocean floor limestones (the strata exposed in the Murri Anticline). The rocks in the anticline above are *autochthonous* (in place), where the ophiolite blocks are *allochthonous*, having been thrust here from elsewhere during convergence. (Put another way, autochthonous means it's in the same place where it started, while allochthonous means it's arrived where we find it today from an original place that was somewhere else.) Given the age of the limestones and ophiolite blocks, the ability of this stream to cut across the anticline and create the dramatic Sidaq Gorge is puzzling.

When we think of convergent boundaries, we often discuss the subduction of oceanic lithosphere beneath continental lithosphere. But this situation is different. The Sidaq Gorge cuts through layers of Cretaceous carbonates deposited on the ocean floor, creating a "water gap", but then the anticline you see above (an upwarped fold created as a result of this convergent boundary), dives ***beneath*** the ophiolites! If all of the structure you see here was formed as the result of the collision of two pieces of oceanic lithosphere, some of which obducted atop oceanic carbonates, then it is likely that Wadi Ghafir represents a stream that has downcut in its current path as erosion has over time exposed all of this. What we see occurring today and that has exposed these structures does not represent the landscape this stream (or perhaps its predecessors) existed in. As streams in this area downcut, these structures were exposed and the current landscape was created as differential weathering did its work. Less resistant rock around the edge of the ophiolite and through the Murri Anticline gave way to the water. Over more time and downcutting, the current landscape emerged.

The rocks that allow the wadi to create the water gap were less resistant to erosion. Ultimately, the work of streams like Wadi Ghafir over time allow us windows into the geologic past. In this case, it shows what happens when oceanic lithosphere meets more oceanic lithosphere.

Plate Boundary 2: Obducting Seafloor

? Exercise 62.12.1

During the oceanic-oceanic convergence that occurred during the Creteaceous Period at the present site of the Sidaq Gorge, Oman, which chunk of rock obducted over the other?

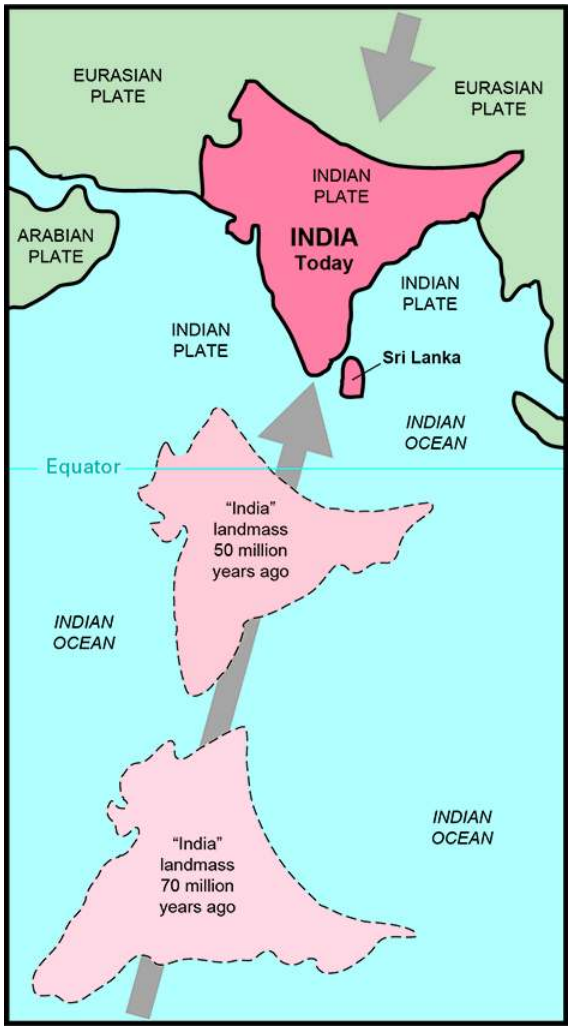
a. Seafloor rock over carbonate seafloor sediments

b. Carbonate seafloor sediments over seafloor rock

Answer

a. Seafloor rock over carbonate seafloor sediments

Continental-Continental Convergence: The Himalayas and Mt. Everest (Chomolungma)

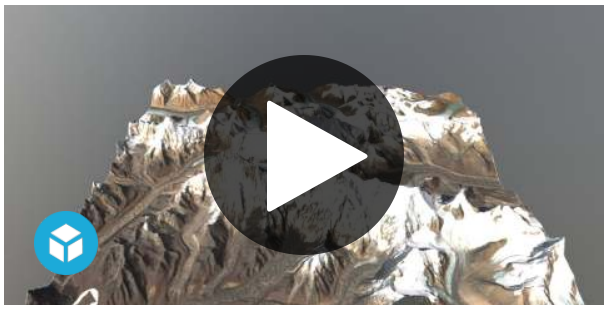


Grab your hiking gear and hire a Sherpa guide, because our third stop on this VFE is the Mt. Everest region!

Here are some key aspects of the current view of the formation of the Himalaya. First, India was a small separate continent, moving northward on the Indo-Australian Plate's western edge. It was one of many fragments of continental crust produced from the breakup of the southern supercontinent Gondwana. About 45 Ma, India had moved far enough northward that it began colliding with southern Asia. This action was so powerful that it led to a change in direction for the Pacific Plate to the east, causing the bend in the Hawaiian Island/Emperor Seamount Chain!

The collision is not over, but continues today, shoving the highest peaks ever higher even as rivers, glaciers, and landslides attempt to tear them down.

Check out the region around Mt. Everest in this 3D model:



Here is a geologic map of the Himalaya region (Searle et al., 2008)

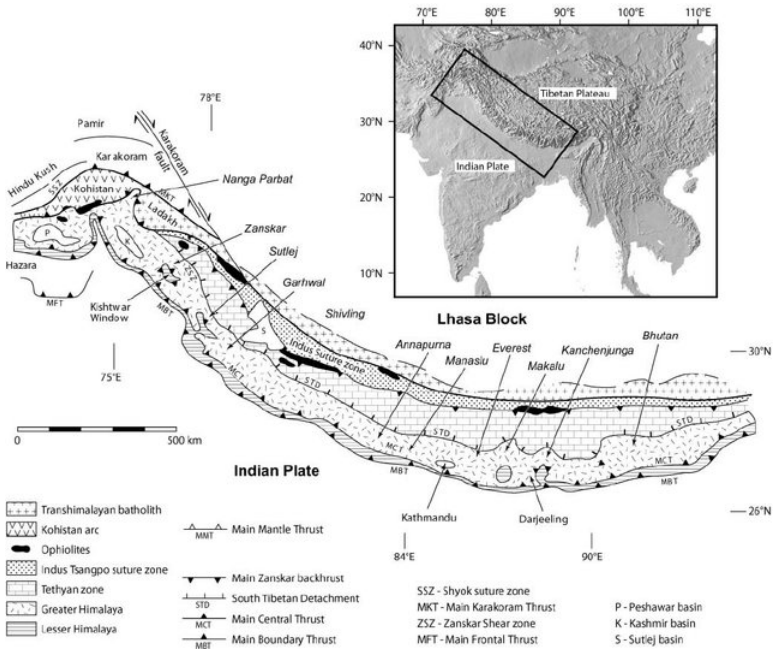


Figure 62.12.4: Geologic map of the Lhasa Block region of the southwestern Himalaya. Note the major peaks highlighted. Mt. Everest sits just within the northern edge of the Greater Himalaya unit just to the south of the limestones of the Tethyan zone.

Study the image above. Consider the location of Mt. Everest and the kinds of geologic units that are very close by.

The image below (also from Searle et al., 2008) provides additional detail. You can find Mt. Everest in the northwest corner of this map. Again, located just to the south of the Tethyan limestones, the detail on this map notes that the peak region of Mt. Everest and the peak to its north, Cho Oyu, also contain these same rocks. Such limestones are normally deposited in marine basins. But, here they are on top of the tallest mountain in the world! Below these are metasedimentary rocks the rest atop the Miocene leucogranites (plus sign unit). Leucogranites are commonly found in continental collisional zones. How do these rocks together on Mt. Everest explain the history of the region?

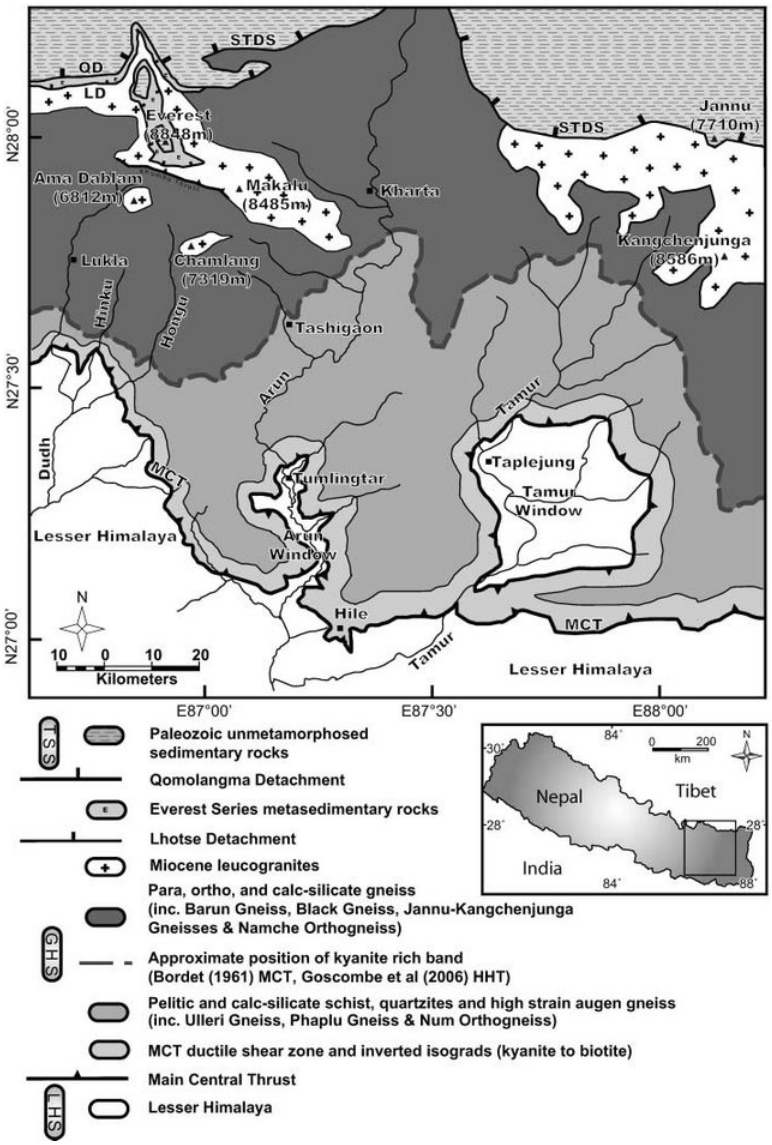


Figure 62.12.5: Geologic map of the Mount Everest Region, Nepal. (From: Searle et al., 2008)

Now, consider again the question of what kinds of geological units surround Mt. Everest. What kinds of rocks exist...

1) ...above and below the Lhotse Detachment Fault (LD)?

- Ordovician-Mesozoic sedimentary rocks above and Everest Series metasedimentary rocks below.

2) ...above and below the Qomolangma Detachment Fault (QD)?

- Everest Series metasedimentary rocks above, Everest-Makalu leucogranites below.

Here is a cross-section of the Everest region (Searle et al., 2008):

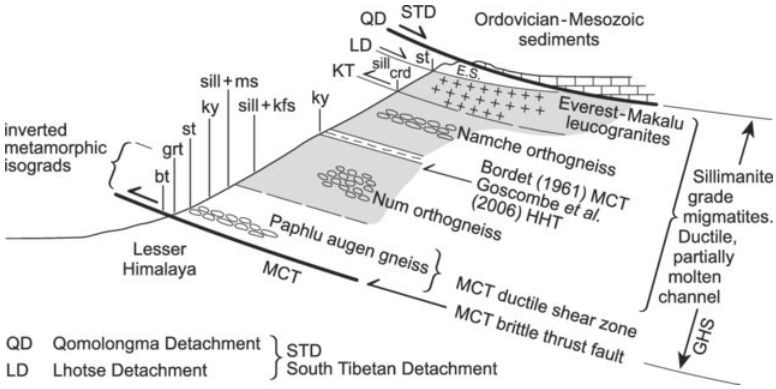


Figure 62.12.6: Simplified cross-section of Everest-Makalu Himalaya region. Our focus is on the transition near the top of the mountains from igneous continental rocks to marine sedimentary rocks across the Qomolongma and Lhotse Detachment Faults.

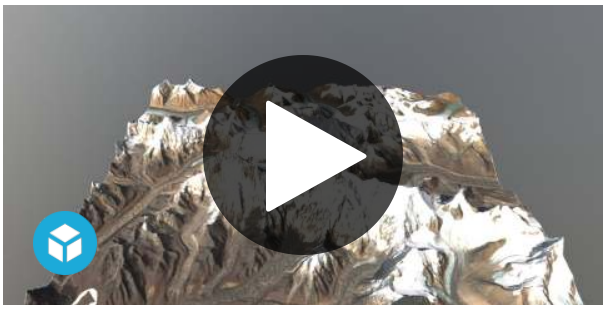
There are three major rock units that are visible on Mt. Everest, including in the 3D model below. See if you can find them!

1) Qomolongma Formation (Summit to the top of the yellow band – ***you can see this*** if you look carefully in the model). These fossiliferous limestones represent an ancient marginal marine environment.

2) Everest Series North Col Formation (Yellow band is the upper part, made up of meta-sedimentary rocks that weather a characteristic yellow-brown. The Everest Series ("ES," above) and associated metasedimentary rocks are flysch deposits that span shallow to deep water deposition in a marine basin.

3) Rongbuk formation (Bulk of the base of Mt. Everest, containing a progression of crystalline metamorphic rocks and intruded leucogranite dikes and sills. Leucogranites are common in intrusions resulting from orogenies due to continental collisions. They contain relatively few dark-colored minerals).

Now, take another look at Mt. Everest in the 3D model. Rotate the mountain until you can see the yellow banded rock at the very top of the mountain. Use the image below the model for further guidance in your quest:



Here is the particular view you should find in the model and a side to side comparison with what how it appears in actuality (from Corthouts et al., 2018):

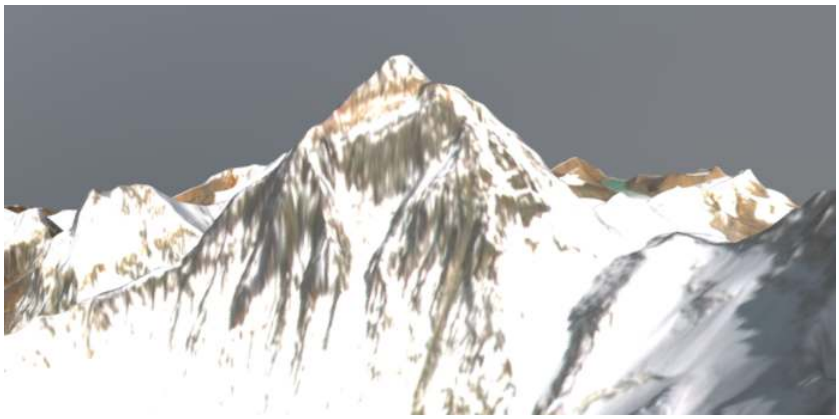


Figure 62.12.7: Image of the peak of Mt. Everest. Two of the stratigraphic units are evidence. The yellow/brown North Col formation of the Everest Series lies just below the Limestones above the Qomolongma Detachment.

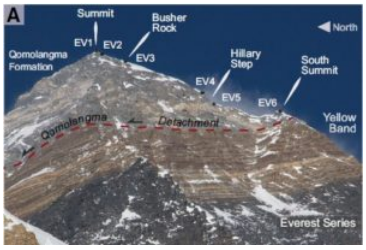


Figure 62.12.8: From: Corthouts et al., 2018. The Qomolongma Detachment separates the yellow banded North Col formation below from the Qomolongma Formation limestones above. Can you see these units in the 3D model?

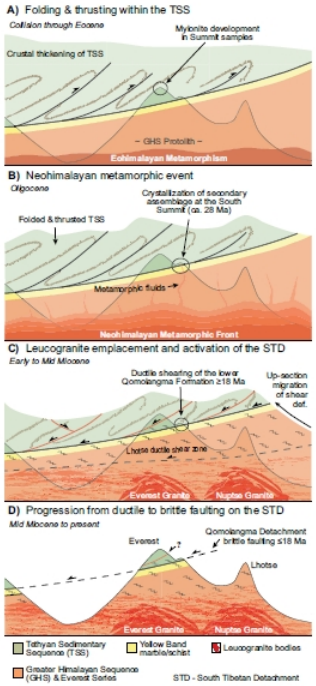


Figure 62.12.9: From: Corthouts et al., 2018. In this series of images, a model for the stratigraphic development of Mt. Everest and surrounding peaks is illustrated. Image A illustrates initial folding during the collision of India and Asia. Over time (2 and 3) erosion and leucogranitic intrusions developed, leading finally to scene 4. This last scene represents what we see today.

Given the stratigraphy of Mt. Everest, how can you tell from the sequence of rocks that this mountain was created by a convergent boundary where India and Asia collided and sutured together?

The rocks at the peak of Mt. Everest represent sediments that accumulated in the marine basin for millions of years prior to the collision of India with Asia. The Leucogranites below represent intrusions resulting from thickening crust, something that would happen as the continental collision was underway. Sitting atop Mt Everest are remnants of the marine basin that existed between the continents.

Himalaya: Whence Came These Limestones?

? Exercise 62.12.1

Limestones, formed in marine basins, atop a mountain may seem unusual. What might explain why such rocks exist atop Mt. Everest?

a. The limestones were formed in their current location well before the Himalayan Orogeny began.

b. The limestones were originally on the seafloor and ended up on the top of the mountain due to faulting during the Himalayan Orogeny.

Answer

b. The limestones were originally on the seafloor and ended up on the top of the mountain due to faulting during the Himalayan Orogeny.

Divergent Boundaries

Oceanic Divergence: Iceland and the Mid-Atlantic Ridge

Below is an image taken of the Þingvellir Rift Valley of Iceland. At this point, it's quite narrow. North America is on the left and Europe is on the right. So different from one another, right? In some ways, this narrow rift is very reminiscent of the early days of the breakup of Pangaea, well before this very long crack in that supercontinent filled with water to form the Atlantic Ocean.

Here, in Iceland, 250 million years later, the two plates continue to stare at one another.

Pan around – View the environs. Make a list of observations about the rocks, the valley between, etc.



Compare what you see above with what you see in the diagram below:

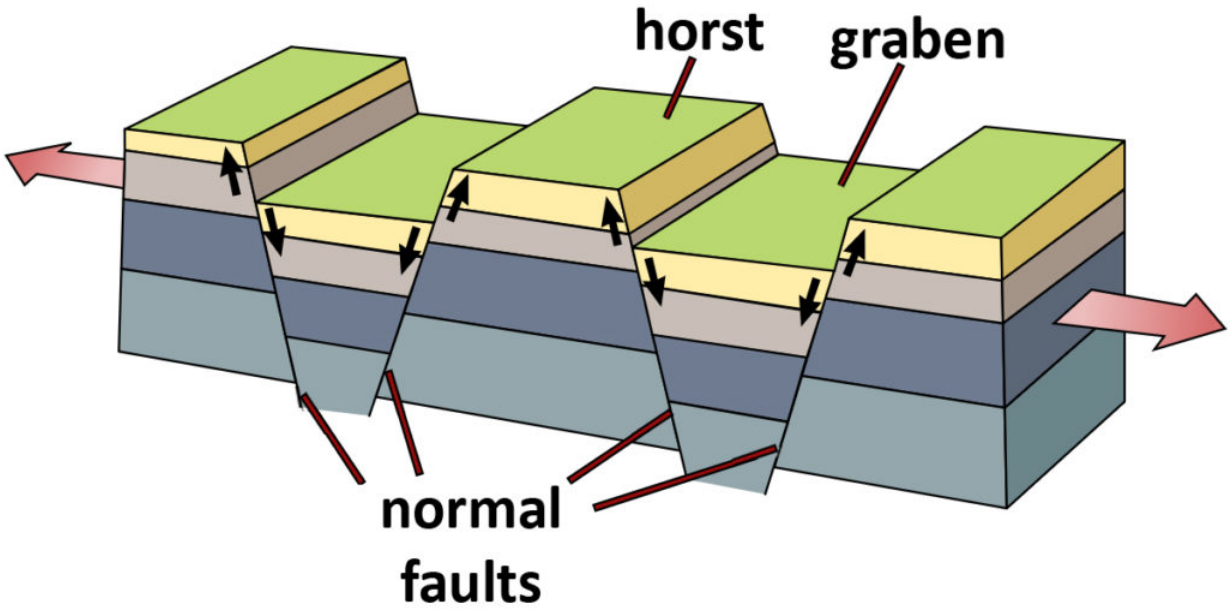


Figure 62.12.10: Horsts and grabens separated along normal faults.

What you are seeing is a traditional horst/graben structure. These are bounded by normal faults. What kind of forces are involved in creating this tectonic situation?

Divergent Boundaries: Forces

? Exercise 62.12.1

What is the primary force applying stress at divergent boundaries?

a. Tension

b. Compression

c. Shear

Answer

a. Tension

Iceland exists because of two important geological situations. First, it sits on the Mid-Atlantic Ridge, which provides the country with its stunning topography while also tearing it apart slowly. But, without also sitting atop a hotspot (mantle plume), it would not protrude above the ocean surface as it does. This means that Iceland is a volcanically active place. The ground there is so hot from the fresh magma supply that Iceland produces a surplus of geothermal energy – which is very cool (or, rather, hot!).

[image]

The map above gives you a simple expression of Icelandic geology. Specifically, you can see the direction of the spreading center. You can also see the large number of volcanoes across the island.

The gigapixel panorama below is taken looking out over a very broad portion of the valley between the scarps on both sides. Pan around – the far left, where the people are hiking, is one scarp. In the distance straight out and to the right is another. One side is the North American plate, and the other is the Eurasian Plate!



What do we call the down-dropped block in the middle of the valley?

Divergent Boundaries: Horst and Graben

? Exercise 62.12.1

What do we call the down-dropped block in the middle of the valley within a rift zone?

a. Graben

b. Horst

Answer

a. Graben

Below are two images of pillow basalts. The top image is from Iceland, from pillow basalts that used to be on the ocean floor but now, due to the uplift by the hotspot, are above the waves.

The bottom image is of pillow basalts under the water near the Juan de Fuca ridge, another active divergent boundary, under the Pacific Ocean, west of Washington and Oregon. Why do you think we call these structures “pillow basalts”?





Figure 62.12.11: Pillow basalts near the Juan de Fuca Ridge, Pacific Northwest, USA.

Divergent Boundaries: Pillow Basalts

? Exercise 62.12.1

How are pillow basalts evidence of the present of a divergent boundary within a marine basin?

a. Divergent boundaries are where new crust forms, which involves volcanism. When lava erupts underwater, it cools slowly, creating smooth "pillows".

b. Divergent boundaries are where new crust forms, which involves volcanism. When lava erupts underwater, it cools rapidly, creating smooth "pillows".

Answer

b. Divergent boundaries are where new crust forms, which involves volcanism. When lava erupts underwater, it cools rapidly, creating smooth "pillows".

One characteristic of either convergent or divergent plate boundaries are volcanoes.

Here, in both shots below, you see the volcano Eyjafjallajökull. If you scroll back up to the map of Iceland, you'll see that this volcano is smack in the middle of the rift.

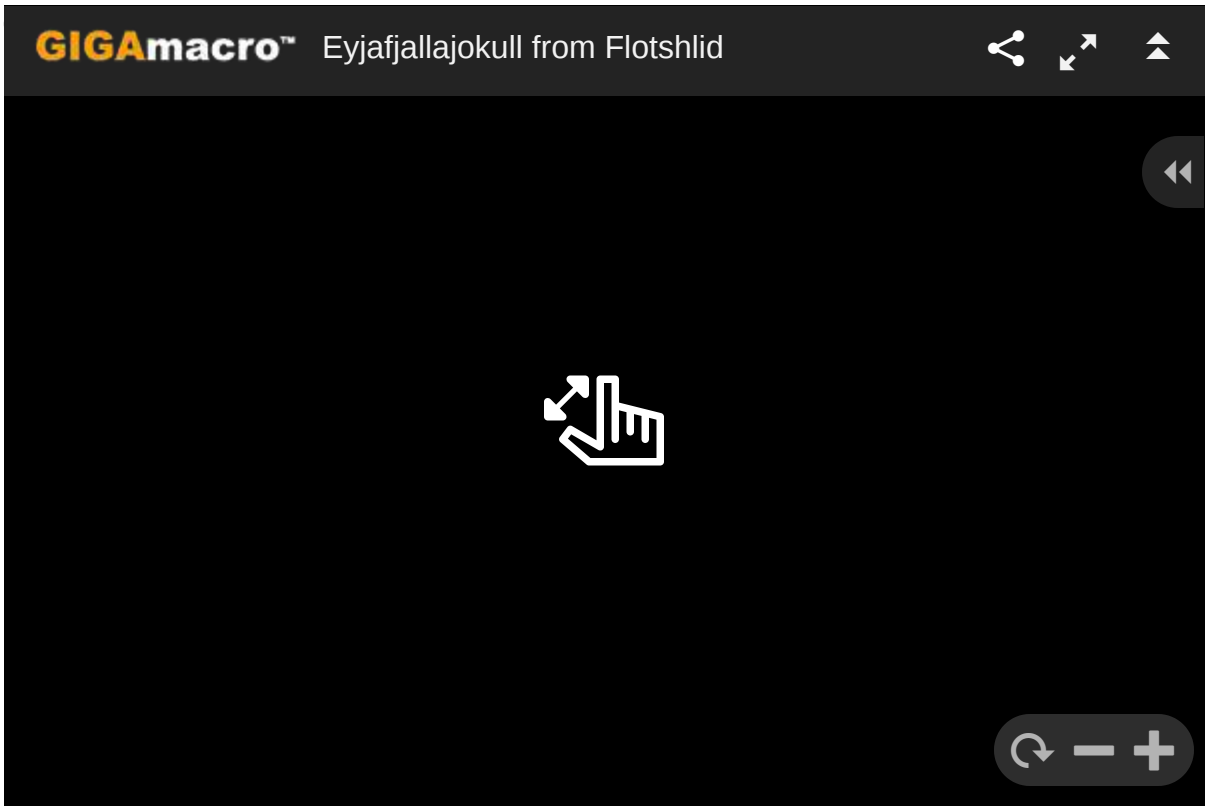




Figure 62.12.12: Eruption of the Icelandic volcano Eyjafjallajökull

The last major time this volcano erupted was in 2010 and it snarled European air traffic for quite some time.

Exploring the Mid-Atlantic Ridge in Iceland provides an opportunity to not only explore a modern active rift zone, but to also catalog a variety of rocks and structures to look for when identifying similar ancient rift zones.

Divergent Boundaries: Key Features

? Exercise 62.12.1

Which of the following features might be diagnostic rocks representing a divergent boundary?

a. Pillow Basalt

b. Basalt

c. Marine Limestones

d. Normal Faults

Answer

a. Pillow Basalt

Continental Divergence: The East African Rift

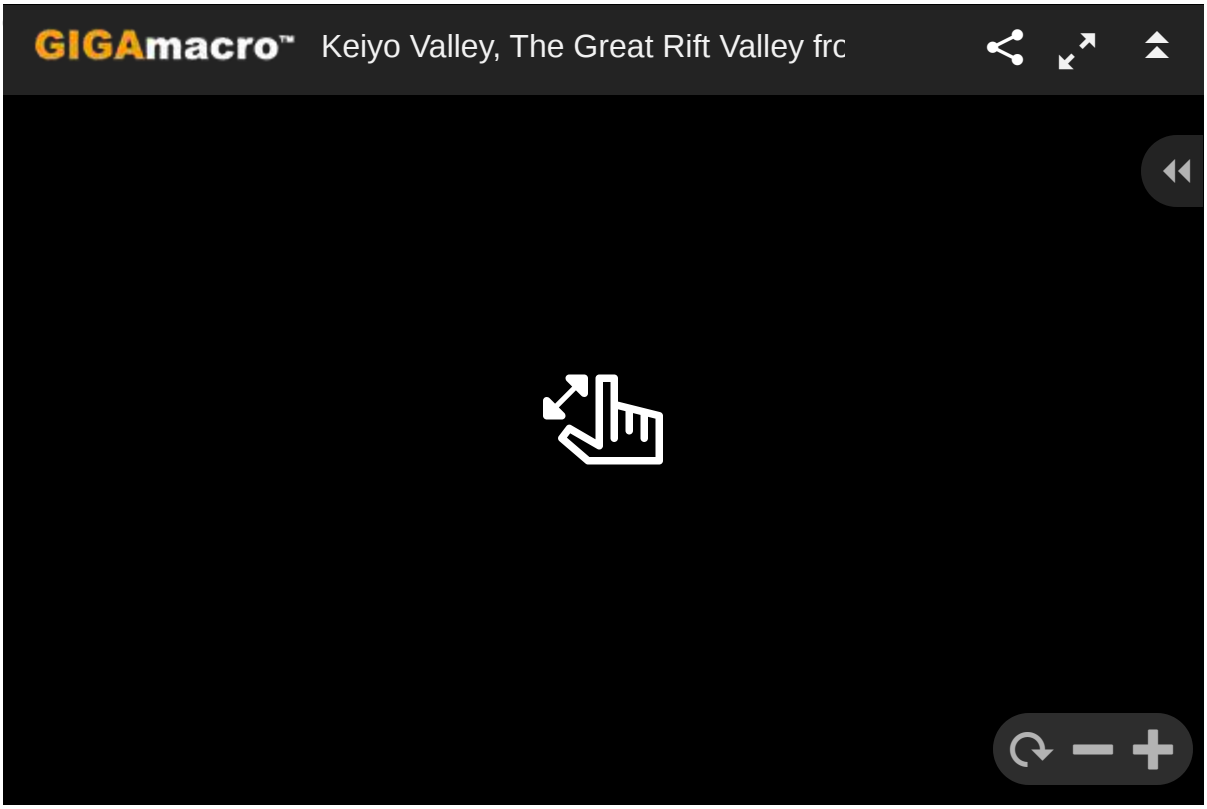
For our fifth stop on this VFE, it's time to go on safari...to Kenya and Tanzania's portion of the East African Rift! The East African Rift is also an active zone of divergence. Unlike Iceland's basalts though, here the divergence is between blocks of continental crust. In this elevated and arid land, the rift is mostly dry, with no water aside from a few lakes. Still, what kinds of features might you expect in a divergent boundary?

The East African Rift Zone makes up the southern portion of a larger zone of rifting, culminating in a triple-junction to the north of the East African Rift in the Afar region of Ethiopia. From there, one rift of the junction forms the Gulf of Aden and the other forms the Red Sea. This latter one connects tectonically to a series of faults that run north through the Sinai region of Egypt into Israel, creating the Dead Sea and Sea of Galilee. The East African Rift portion of this triple-junction of divergent boundaries is an aulocogen, or failed rift. However, some parts of it are re-activated.

[image]

Here are two fantastic GigaPan images from this portion of the rift valley (The Gregory rift). Use the volcano Oldoin'yo Lengai (on the map above) for a sense of location:





The two images above both show important features of a divergent continental rift valley.

The bottom image is taken from near Iten, Kenya, a location famous for producing fabulous marathon runners. You are looking to the east from the western end of the Gregory Rift. Using your new knowledge, can you imagine the structure across which you are looking? Are there similarities with Iceland?

Oldoin'yo Lengai is a volcano located near Lake Natron in the region of Tanzania historically occupied by the Maasai people. Unique among volcanoes, it erupts a natrocarbonotite lave, which is very rare. Thus, it's a volcano that actually erupts carbonate rock! But, this carbonate is very different from what is precipitated in marine environments. Rather than being rich in silica, the lava is rich in sodium and potassium carbonate minerals (not calcium or calcium magnesium carbonate, as in marine environments). Here is a video of a lava flow from Oldoin'yo Lengai (which, in Maa, means "Mountain of God").

Here is a video showing how this unique carbonatite lava erupts:



There are many other volcanoes, to be sure, most of which produce some still unusual, but also more common trachyte (rich in potassium feldspar, high in silica) and phonolite (very high potassium feldspar, intermediate in silica) products.

East African Rift: Diagnostic Features

? Exercise 62.12.1

Based upon your experiences at both divergent boundaries, which of the following is the best diagnostic property of a divergent boundary?

a. Basalt

b. Horst and Graben Structure

c. Carbonatite Lava

d. Volcanoes

Answer

b. Horst and Graben Structure

Transform Boundaries

Transform: San Andreas Fault, Southern California

Transform boundaries are plate boundaries where motion is primarily horizontal. Most examples of these occur along divergent boundaries at the bottom of the ocean and act as displacement boundaries due to the effects of the Earth's curvature on the alternating divergent boundary's motion. These boundaries offset sections of these very long ridges from one another. However, some good terrestrial examples do exist for us to examine.

The San Andreas Fault is a classic example of a transform boundary strike slip fault. Strike slip faults, created by shear, twisting forces, have lateral movement across their trace and are generally vertical in orientation. Of course, no strike slip fault is perfectly vertical – all fault scarps have some amount of dip along which the fault slips. thus, there is often some vertical dip-slip motion in addition to the much more pronounced horizontal strike slip movement.

This 3D model of the trace of the San Andreas Fault was taken from a drone above the Carrizo Valley near Wallace Creek, California.



Review strike slip faults here. Can you identify the fault zone? Find and follow Wallace Creek. For what would you look? Is the fault left lateral or right lateral? List the characteristics you see in the image that help you answer the above questions. Then, answer the quiz question below.

Right or Left? The San Andreas Fault

? Exercise 62.12.1

Is the San Andreas Fault at Wallace Creek right or left lateral in its motion?

a. Left lateral

b. Right lateral

Answer

b. Right lateral

Leaving the Field

Modern plate tectonics utilizes three main kinds of plate boundaries to make it possible for plates to move around. Compression, tension, and shear are forces of stress that are constantly at work at different locations. They dictate the type of plate boundary that forms. In this VFE, you have had the chance to explore all three types of plate boundaries. Some of them were in continental crust, and others in oceanic crust. Some of these were still active. Some of them existed in the past. Regardless of timing, the processes occurring at each different boundary produce a suite of unique rocks and structures that can be diagnostic.

Test Your Knowledge of Plate Boundaries!

? Exercise 62.12.1

Here is a lovely pillow basalt. This one is from the Columbia River Province and is not a part of a plate boundary. However, we do know of at least one plate boundary where such pillows form! Why do you get this particular structure at divergent boundaries like the Mid-Atlantic Ridge or Juan de Fuca Ridge?



a. They erupt on land inside a lava tube.

b. They erupt underwater and quickly cool into this shape.

Answer

b. They erupt underwater and quickly cool into this shape.

Exercise 62.12.2

Here is a gigapan image of a belemnite fossil from the Swiss Alps. These fossils, often in the shape of a bullet, come from marine cephalopods. Again - marine limestones in the tops of mountains!



What does the presence of this fossil on a mountain top tell us about the Alpine Orogeny?

a. The Alps were formed by a continental-continental collision.

b. The Alps were formed at a transform boundary.

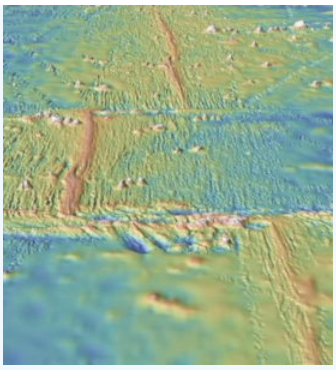
c. The Alps were formed at a divergent boundary.

Answer

a. The Alps were formed by a continental-continental collision.

Exercise 62.12.3

This is a false-color image from the East Pacific Rise created using the program GeoMapApp. The raised red areas are the divergent boundaries. But, do you see how they are offset and by what? There is more than one type of plate boundary here! What do we call this other boundary type?



a. Continental-Continental Convergent (Collisional) Boundary

b. Oceanic-Continental Convergent Boundary

c. Divergent Boundary

d. Transform Boundary

Answer

d. Transform Boundary

For Further Reading:

Bailey, Chuck (2014). Wadi Bani Ghafir at Sadiq Gorge: Water Versus Rock in Oman. College of William and Mary Blogs.

Corthouts, Travis L., Lageson, David R., and Colin A. Shaw (2015). Polyphase deformation, dynamic metamorphism, and metasomatism of Mount Everest's summer limestone, east central Himalaya, Nepal/Tibet. In *Lithosphere* v. 8(1):38-57. DOI:10.1130/L473.1

Kirkpatrick, Jamie; Rowe, Christie; Bentley, Callan; Blisniuk, Kim, and Wakabayashi, John (2019). *Streetcar 2 Subduction*, digital field trip guide the geology of the San Francisco Bay Area. American Geophysical Union.

Searle, M.P., Law, Richard D., Godin, Laurent, and Kyle Larson (2008). Defining the Main Central Thrust in Nepal. *Journal of the Geological Society* v. 165:523-534. DOI: 10.1144/0016-76492007-081

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62.13: Siccar Point, Scotland

Siccar Point: where the enormity of geologic time was first recognized

Introduction

Today, we visit Siccar Point, Scotland. You've probably already seen photos of this place in geology class – they usually look something like this:



To those who aren't familiar, here's what going on:

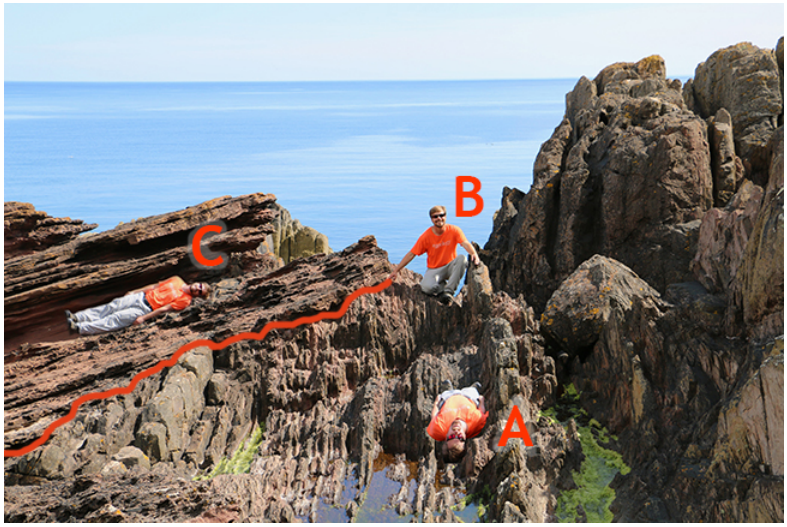
There are two sets of strata here – and the contact between them is an ancient erosional surface. It's an angular unconformity, the very first one identified — **anywhere, *ever*.**

When James Hutton brought his friend John Playfair here in 1788, he explained the significance of such a structure: It implied that (1) Silurian graywackes and shales were deposited horizontally, (2) they were subsequently folded and tilted to more or less vertical orientations, (3) a period of erosion commenced, chewing away at their upper ends, (4) erosion ceased and deposition resumed, but now the environment of deposition had totally changed – instead of deep marine strata, oxidized fluvial redbeds were laid down (the Old Red Sandstone) — again in an original horizontal orientation, then (5) these too were tilted to their present angle gently dipping off to the left of the frame, and (6) modern erosion went to work on the whole mess, carving it up into the three-dimensionally complex exposure we find today.

Hutton pointed out that this whole sequence would probably take a little while. Playfair recounted the sense of temporal vertigo Hutton had induced in him: "The mind seemed to grow giddy by looking so far into the abyss of time."

Siccar Point, then, was the place where the immense depth of geologic time was first realized. It was the first time it was *felt!*

I got into the spirit of the place by putting myself in the place of (A) the turbidites, (B) the ancient erosional surface, and (C) the red beds:



Photos

But there's much more to Siccar Point than just this one oft-photographed scene.

Just walking in to the site is a treat – looking both ways along the coast, you can see the relevant units – with Old Red Sandstone dominating to the northwest, and Silurian turbidites to the southeast:





The first view of Siccar Point itself comes from a great height:



I paused there to capture the view on GigaPan (see the resulting image later in this VFE):



Then, we descended. As you can see, we had awesome weather – but this slope was so steep, next to a knee/chest level strand of rusty barbed wire, that I shudder to imagine descending it on a rainy day:



Because the modern waves have sculpted Siccar Point into a complex shape, there are a dozen ways to view the unconformity, dipping down below it in places (entering the upended Silurian) and popping back up again, like a time-traveling gopher (daylighting into the Devonian).





The unconformity surface is laced with boulders and cobbles – the rubble of the eroded Silurian graywackes that became a basal conglomerate for the overlying red beds.



There is some tafoni to be seen, too – always a treat!



Another thing you can see in the previous image is orangey-yellow lichen, of which there is quite a lot at Siccar Point. I was surprised at this – even though the two rock units here are very distinct in color, both actually appear similar because of all the lichen encrusting them!

In fact, there is very little “clean” rock exposed here – if it’s too high up, it’s covered with lichen, and if it’s too low in elevation, the tides bring nutritious waters up twice daily, meaning it’s prime for colonization by barnacles. In between, you might find some decent views into fresh rock. But don’t hammer! This is as close to sacred ground as geologists get! Here’s a view showing the vertical zonation:



In the thin zone where the rocks are clean, you can see features such as graded bedding. Here, a single turbidite is imaged, with the coarse base on the left, and then fining (younging) toward the right, where the lens cap lies atop some thin laminations:



Here's another, with the opposite geopetal implication:



Visitors can walk out to the hinge of a plunging syncline:



Here's a view of the same fold, from the other side of that little bight — You can see the nested ">>" shape of the barnacle-encrusted turbidite layers, in this case plunging to the lower-left:



Fractures in the upturned Silurian graywackes give the clean rocks a tesselated appearance:



And not all the fractures are joints – some are faults!





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GigaPans

So there's a lot of fun stuff to contemplate here! To provide a user-directed facsimile of a real-life visit to Siccar Point, I gathered some visuals for you. Here are four GigaPans of the site:









360° spherical photos

Next up, here are three 360° spherical photos:

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Siccar Point unconformity, Scotland #theta360 - Spherical Image - RICOH THETA

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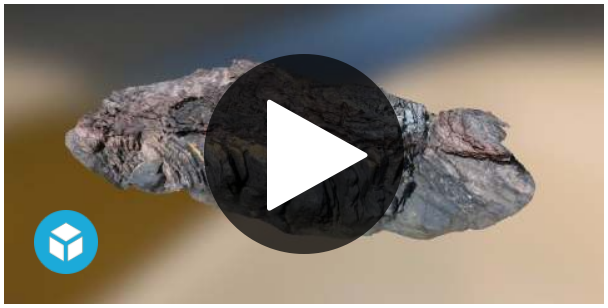
Planet Hutton! Siccar Point unconformity, Scotland #theta360 - Spherical Image - RICOH THETA

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Siccar Point unconformity, Scotland #theta360 - Spherical Image - RICOH THETA

3D models

And then —our *coup de grâce*— we have two awesome 3D models of the unconformity surface – one zoomed out a bit:



...And closer in on a key area:



Conclusion

Because Siccar Point is a wave-blasted headland, the outcrop surface is a complicated shape, and that makes photographing it complex. Essentially, there's one view that easily captures the relationship between the two rock units, and that's the one that everyone takes – the view that started off this VFE. But Siccar Point is bigger than that one view, and I think it takes something like these 3D models to convey a real understanding of the complicated geometry of the place – an important step for inducing the "giddyness" that John Playfair and James Hutton enjoyed while visiting the site.

I feel extraordinarily fortunate to have been able to visit Siccar Point with my family on such a beautiful day. I hope that you are able to access some of the giddy joy I felt there with this virtual field trip!



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62.14: Wind River Canyon, Wyoming

ROCK TYPES

Here are three samples of rock collected at Wind River Canyon, Wyoming. Examine each sample in detail by zooming in and exploring its composition and texture.

Sample A:

Sample B:

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Sample C:

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Now answer these interpretive questions:

Exercise 62.14.1

Identify the rock type that makes up each sample:

A) _____

B) _____

C) _____

Exercise 62.14.2

Which of these three rock samples is **sedimentary**? (answer A, B, or C) _____

Exercise 62.14.3

Which of these three rock samples is **metamorphic**? (answer A, B, or C) _____

Exercise 62.14.4

Which of these three rock samples is **igneous**? (answer A, B, or C) _____

Exercise 62.14.5

Is the igneous sample plutonic or volcanic? _____

Explain how you know: _____

OUTCROPS

Now, let's explore the larger context in which these rocks were collected. We will do this by exploring some outcrops.

The samples were collected at Wind River Canyon, Wyoming, at these sites:

Site 1:

Site 2:

Now answer these interpretive questions:

Exercise 62.14.6

Site 2 shows all three rock types, but Site 1 only shows two. Which two are present at Site 1? (answer 'A and B,' 'B and C,' or 'C and A') _____

Exercise 62.14.7

Based on the principles of relative dating, which of these three rocks is the oldest? _____

Cite your rationale (which principle of relative dating you used for your previous answer): _____

Exercise 62.14.8

Based on the principles of relative dating, which of these three rocks is the youngest? _____

Cite your rationale (which principle of relative dating you used for your previous answer): _____

? Exercise 62.14.9

Therefore, what sort of surface divides samples B and C from sample A? _____

? Exercise 62.14.10

Sample B includes zircon crystals. Imagine you collected some of these zircons and ran them through a mass spectrometer. The analysis produced a ratio of ^{238}U to ^{206}Pb of 3:1. How many half-lives have elapsed since the formation of the rock? _____

? Exercise 62.14.11

Considering that the half-life of ^{238}U is 4.468 billion years, calculate the crystallization age of sample B: _____



? Exercise 62.14.12

Sample A is found to include these fossil specimens: *Lingulepis* is an inarticulate brachiopod. According to the Paleobiology Database, it has a geologic range from 516.0 to 449.5 Ma. What period(s) is/are included in that range? _____ (Feel free to consult a geologic timescale.)

? Exercise 62.14.13

Which of the major North American cratonic sequences was Sample A deposited as part of? _____

? Exercise 62.14.14

Constrain the amount of time ***missing*** along the surface you noted in your answer to question #8 above. Record your answer as a minimum and a maximum number of millions of years missing:

"Between _____ minimum and _____ maximum are missing."

Please submit your answers to the instructor.

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62.15: Stromatolite diversity in the Belcher Islands



Figure 62.15.1: South-plunging anticlines along the east side of Belcher Islands, Hudson Bay. The islands are skinny, separated by narrow, frequently stormy inlets. There are no trees. View approximately to the south.

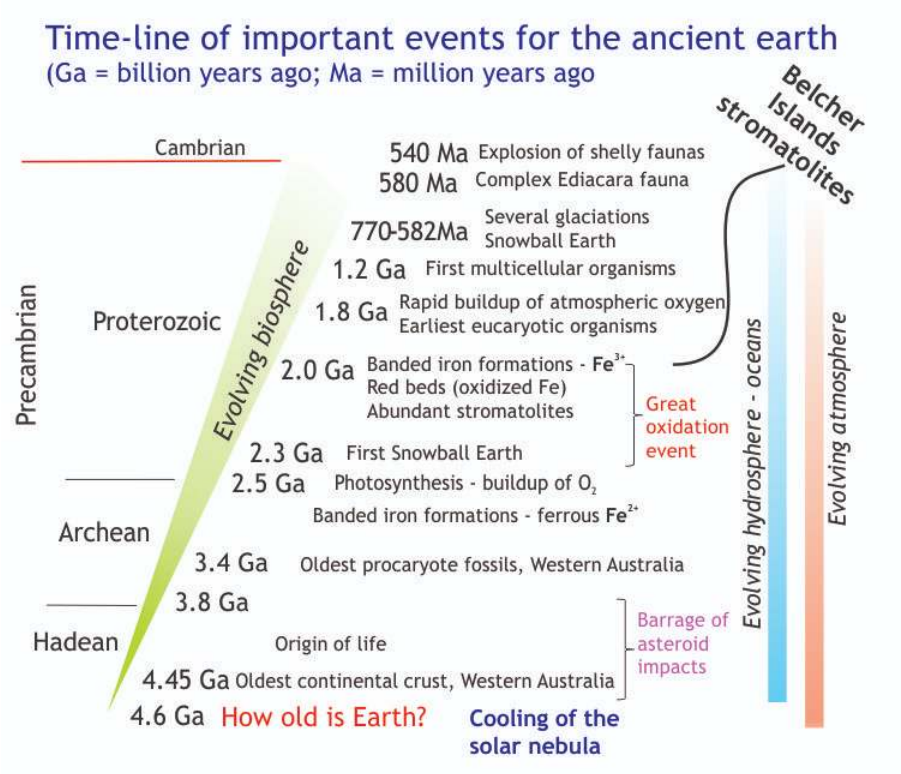
Field work in Hudson Bay (Canada)

A walk through some of the most ancient life forms on Earth

I like the expression "*rock record*." For me, these two words encapsulate Earth's readable history, page by page. Pick up a rock, turn the page, delve into its mysteries, solve the puzzles. There is so much to discover – such a journey.

The journey here will walk you through the world of our most ancient organisms, not the oldest, but pretty old at 2 billion years. This was a world where life was in full swing, where the biomass was huge but occupied solely by bacteria and algae, or more specifically **cyanobacteria** (sometimes called blue-green algae). These organisms thrived in vast carpet-like communities, now preserved in the rock record as delicately laminated mats (or **cryptalgal laminates**), molded into simple domes or complexly branched columns (**stromatolites**), and even vast, reef-like build-ups. The environments in which they grew ranged from geothermal pools to the shallow confines of freshwater lakes, and marine settings that extended from the most landward extent of ocean tides (supratidal) to deeper shelf and platform.

Stromatolites are known from sedimentary rocks as old as 3.4 billion years (**North Pole deposits**, Western Australia – this 1980 paper was one of the first to fully describe these). They dominated Earth's biosphere for the next 3 billion years; clearly, life was taking its time to get over that first rush of DNA replication. They relinquished this foremost position during the Cambrian invertebrate explosion because of competition for their niche and from predation – lots of critters like to eat algae. Today, cryptalgal laminates eke out a living in a few specialized environments, Shark Bay on the Indian Ocean coast of Western Australia being one of the better-known examples.



We should be immensely grateful for this Precambrian flourish. Cyanobacteria are **photosynthetic prokaryotes**, and the general consensus is that they were responsible for the production of nearly all our atmospheric oxygen. Most (but not all) modern cyanobacteria thrive in sunlight which means growth is limited to the **photic zone** (the part of the water column that sunlight penetrates, at depths less than 200 m). This condition appears to be borne out in the rock record where fossil cryptalgal structures are commonly associated with other sedimentary features also developed in shallow water under the influence of tidal currents, the effects of waves, and storms. We will see some of these sedimentary facies on this trip.

This field excursion will take you through a **Paleoproterozoic** succession in **Belcher Islands**, a small group of islands and islets in eastern Hudson Bay. Exposure here is superb – look again at the image at the top of the page. But the weather can be atrocious. I spent 20 weeks in total in 1976-77. For a Kiwi having just arrived in Canada, meeting the local Inuit, and absorbing the scenery and geology were life-affirming experiences.

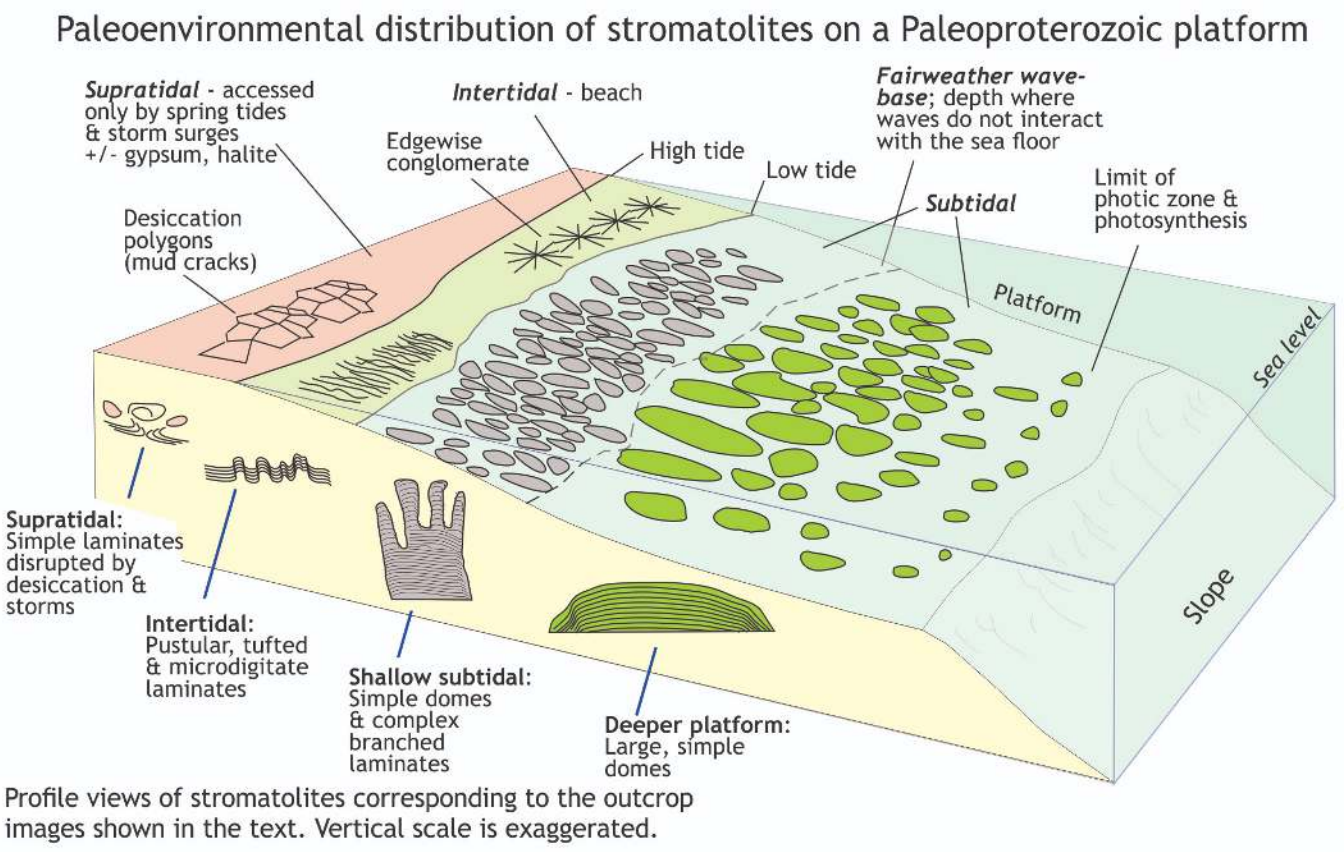


Figure 62.15.2: A relatively sheltered cove on Tukarak Island, the site of an abandoned Hudson's Bay Post, was our base camp for the 1976-77 field seasons. We shared this locale with Inuit who spend summers here. The Inuit lodgings are the large. roomy white tents, that have oil heaters and stoves. The two orange blobs behind were our less-salubrious digs.



Figure 62.15.3: Summer is the only time to do field work. We used two zodiacs to move fly camps around the numerous islands. They are open to the elements, and the weather was not always kind. Hence the two boats – if one sank, the other could return to tell the tale. Here, we are negotiating remnants of sea ice en route to our first destination.

We begin this field excursion at the **high tide** mark, making our way across the **intertidal zone**, through the shallow **subtidal zone** where waves break, and thence to **deeper offshore** waters beyond the immediate influence of anything other than storm waves. You can check your 'paleo-location' at each step using the diagram below.



Profile views of stromatolites corresponding to the outcrop images shown in the text. Vertical scale is exaggerated.

At each location we will compare modern examples with the ancient rock record. We make these comparisons because they are an essential part of how we interpret ancient environments. This way of thinking (sometimes stated as '*The Present is the Key to the Past')* helps us in two ways: first, it provides us with a warrant to reason inductively, to interpret or to predict; and second, it allows us to use the actual processes and products we observe in modern environments to help us solve the puzzles of the past.

Supratidal zone

We begin our geological transect on a broad tidal flat. Its landward extent may include a zone accessed only by spring tides and storm surges – the **supratidal zone**. In arid regions evaporite minerals like gypsum and halite may accumulate (the best modern analogues for these are sabkhas along the Persian Gulf). Any cryptalgal mats that grow will be subjected to frequent desiccation; they are susceptible to reworking by strong winds and storm tides where they are mixed with other sedimentary fragments ripped from the tidal flat.

In the next two images, compare the ancient example with some modern supratidal microbial-cryptalgal mats.



Figure 62.15.4: Compare the profile view of structures in this image with the modern example from Texas. The laminated carbonate mudstones and thin cryptalgal mats accumulated on a two billion year-old supratidal to intertidal flat. The muds periodically dried, resulting in the upturned edges of desiccation polygons. Subsequent erosion by the occasional storm produced mud-pebble conglomerates, or rip-ups.



Figure 62.15.5: These algal mats are forming today in a tidal marsh south of Galveston, Texas. They are only covered during spring, or king tides. Here they have partly dried. The desiccated edges tend to curl upwards. The dried fragments are susceptible to reworking by the next tide. Compare this bedding view with fossil examples opposite.

Intertidal zone

Working seawards we pass through the **intertidal zone**, a region washed twice daily by tides. Desiccation of mats is less intense here, but they are more susceptible to erosion by waves, particularly storm waves. The example here shows many erosional discordances where mats were regularly disrupted and regrown.

Again, we can compare modern analogues with the 2 billion year-old examples. The modern structures below are from an Abu Dhabi sabkha.



Figure 62.15.6: Detail of pustular algal (microbial) mats on an Abu Dhabi tidal flat. Sediment between the pustules is mainly carbonate mud. Coin (circled) for scale, gives a sense of the synoptic, or growth relief of each microbial structure. Compare the scale of these structures with the images of fossil mats. (Photo by Stephen Lokier, University of Bangor, Wales.)



Figure 62.15.7: This photo of a modern tidal flat in Abu Dhabi shows how extensive these environments can be. The tidal flats are covered by algal, or microbial mats, all showing signs of desiccation. They are covered more frequently by shallow tides. Evaporation here is intense. (Photo by Stephen Lokier, University of Bangor, Wales.)



Figure 62.15.8: Laminated intertidal mats overlain by some very nice but small, finger-like (digitate) columns. The digitate mats were subsequently eroded, possibly during a storm surge and covered by coarse sand. Coin is 20 mm diameters.

In the images that follow, fan-shaped clusters mid-image are edgewise conglomerates that consist of semilithified slabs of carbonate mud and cryptalgal laminates, ripped up and stacked one against the other (note this is a profile view). This kind of structure commonly forms on beaches – the modern example (right image) consists of shale slabs that when stacked, present a crude radial pattern (this is a bedding view).



Figure 62.15.9: These intertidal cryptalgal (microbial) laminates show many stages of growth, disruption of mats by erosion during storms or desiccation, and regrowth. Zoom in on different parts of the image to see the detail. Chert replacement of carbonate in some mats causes them to be more resistant to erosion at the outcrop. Oncoids represent mats that grow as they are rolled around the sea floor – a bit like oversized ooids. A modern example of edgewise conglomerate accompanies this image.



Figure 62.15.10: A modern edgewise conglomerate pavement (sometimes called stone rosettes) formed by wave action on a beach. The slabs are stacked end-on in crude radial patterns. In the ancient example shown above, slab composition is different (cemented carbonate mud), but the stacking arrangement is identical.

Ripples are extremely common on sandy tidal flats. In the rock record, layers containing ripple crossbeds may alternate with layers containing mats. The examples below are interference ripples where two sets of ripples almost at right angles, appear to interfere. These structures are excellent indicators of ancient tides where one ripple set formed on the incoming (flood) tide, the other on the outgoing (ebb) tide. If we can determine the orientation of the ancient shoreline, then we can hazard a good guess as to which set records the tidal flux.

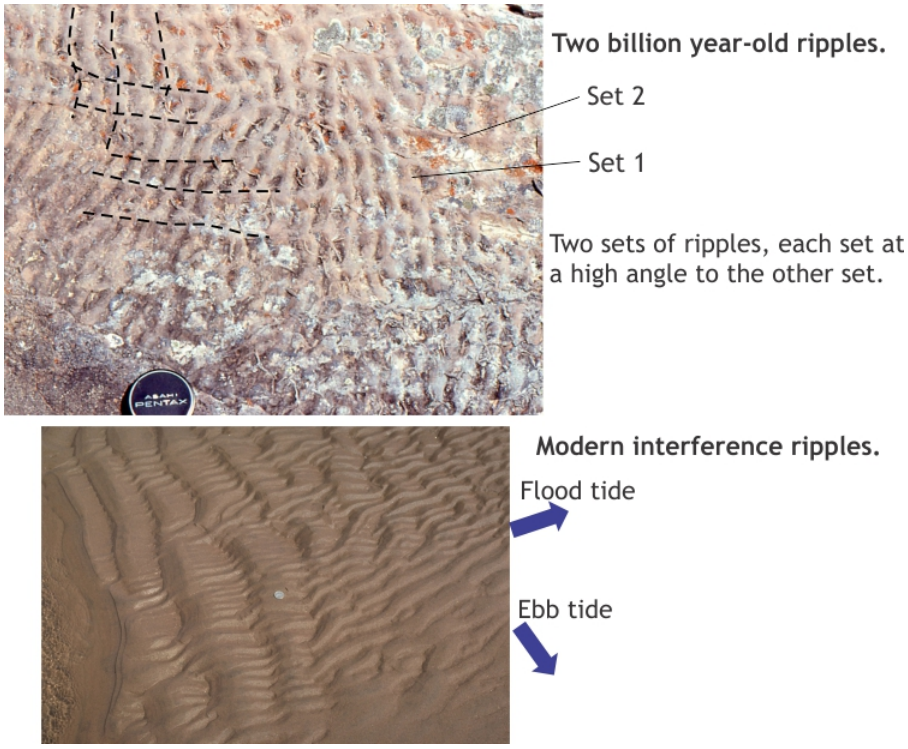


Figure 62.15.11: Compare these two images of interference ripples that formed during alternating flood and ebb tidal currents across a tidal flat.

Shallow subtidal

The subtidal zone below low tide is always submerged. This is the region of constant water movement; it includes the surf zone. Here, cryptalgal buildups tend to be larger and more complex; they may form as simple domes or complex branched columns. In outcrop they can be spectacular, and quite beautiful. Finding modern analogues for this kind of stromatolite is more difficult – there are very few locations where they form today. One notable exception is Shark Bay in Western Australia, an iconic location where modern, actively growing domes bear an uncanny resemblance to ancient structures, as illustrated below.

Even more spectacular are stromatolites that developed **branched columns** of many shapes and sizes. An excellent example is shown below.



Figure 62.15.12: Bedding exposure of simple stromatolite domes, remarkably similar in outward appearance to the modern structures at Shark Bay (W. Australia). One could imagine stepping gingerly between the domes on the 2 billion year-old beach. The domes stand out because rock between each dome has been removed by erosion. Hammer for scale (circled).



Figure 62.15.13: The modern microbial domes growing in Shark Bay, Western Australia are iconic analogues for ancient stromatolites. Comparison with the ancient domes shown above is uncanny. Sediment between the domes is fine-grained carbonate, plus fragments of dome broken off during storms. Penknife for scale bottom centre. (Image credit: Alicejmichel 8 June 2007, https://commons.wikimedia.org/wiki/F...omatolites.jpg)



Figure 62.15.14: Coarse-grained, crossbedded sandstone that was probably deposited on subtidal sand bars, in areas where shifting sand prevented growth of microbial mats and domes.



Figure 62.15.15: This example (one of my favourites) show some broad simple domes in the lower half of the image, changing to branched, digitate (finger-like) columns – the transition occurs just below the hammer. Each millimetre-thick curved line represents a former cryptalgal mat, now converted to dolomite. Zoom in and see how far you can trace a single lamination from one column or dome to another. This will give you some insight into the relief of the growing mat on the seafloor. These structures formed in shallow, agitated, subtidal water.

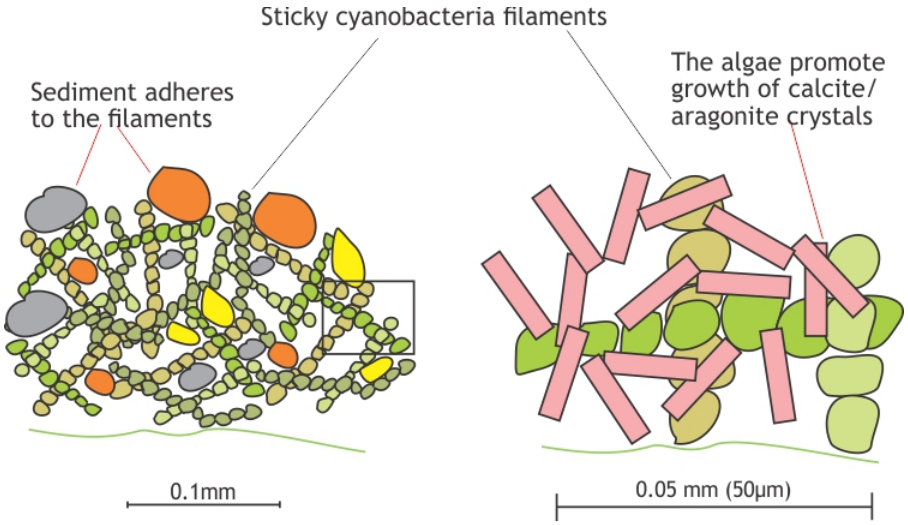


Figure 62.15.16: Fine sediment adheres to sticky algal filaments, or the microbes themselves promote precipitation of calcite or aragonite crystals

Each of the skinny layers, commonly less than a mm thick, represents a once-thriving mat of cyanobacteria, growing on the seafloor. The laminates consist of very fine carbonate mud that adhered to the algal filaments; the filaments may also have promoted precipitation of fine carbonate minerals, such as aragonite. Most of the original organic matter was removed by chemical processes as the sediments were buried.

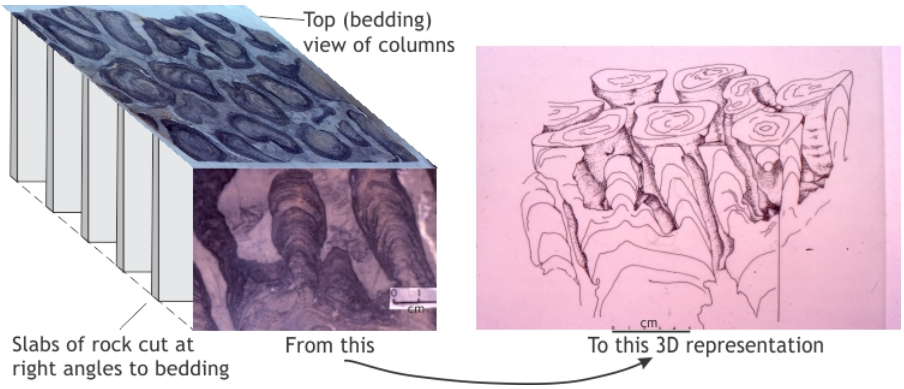


Figure 62.15.17: This 3D reconstruction of branching stromatolites was made from a large block, collected in the field. The block was slabbed (with a diamond saw), each slab was polished, the fine details traced onto paper, and the 3D representation of branches and laminae drawn as an isometric image. (Note, this was done before the advent of computer graphics).

The stromatolites shown above consist of columns that are about 50 cm high – in fact the columns in some stromatolites can reach several metres. However, the column height is a bit deceiving. Each lamination represents a period of growth on the seafloor (days, months). If you trace a lamination, or perhaps a set of laminae from one column to its nearest neighbours, you will see how much relief the stromatolite had during growth. We call this **synoptic relief**. In many stromatolites synoptic relief was only a few mm or cm. In other words, if you were walking across the surface you would only see low mounds and ridges rather than tall columns. Large columns and domes therefore represent relatively continuous growth over long periods (perhaps hundreds of years) where columns gradually build, keeping pace with sediment accumulation on the seafloor. Sometimes growth was interrupted by storms or long periods of desiccation during exposure; when growth re-established, a new branch or dome would develop over the earlier-formed structures – hence the branched stromatolite columns we now see in cross-section views in outcrop.

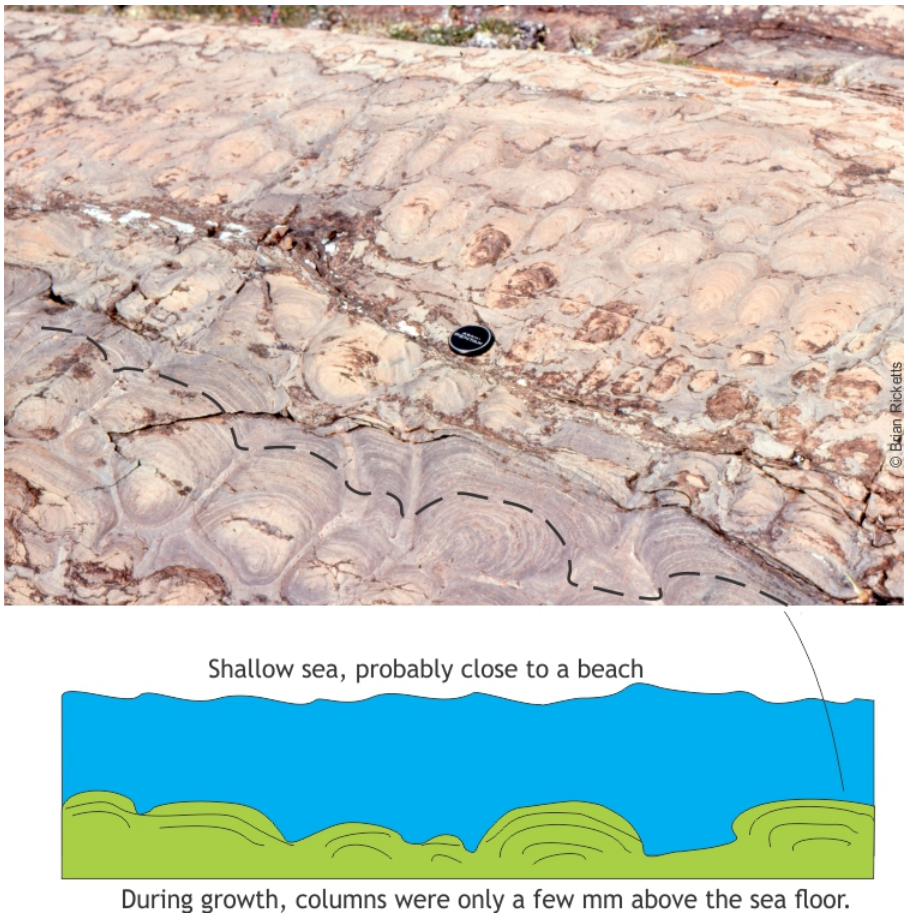


Figure 62.15.18: A nicely polished outcrop containing cross-sections of bulbous stromatolites. We can get a sense of how much these stromatolites were exposed at the sea floor by tracing individual laminae from one column to the next, as in the cartoon below. The traced line represents the sea floor at a point in time when the mats were growing.

Deeper water platform, but within the photic zone

Our transect takes us to deeper water where normal or fairweather waves no longer interact with the seafloor. Here, the cryptalgal laminates grow in quieter waters within the photic zone (perhaps a few tens of metres deep), washed by tidal currents. There is much less disruption of growth by storms (and no desiccation), such that domes extend much farther across the sea floor than in shallower waters. The surrounding sediments are also finer grained, perhaps with ripples migrating across the seabed. However, the synoptic relief is greater; in the example shown, the domes are 2 m to 4 m wide with synoptic relief of 30-60 cm. Branching is less common because there were fewer disruptions to mat growth.

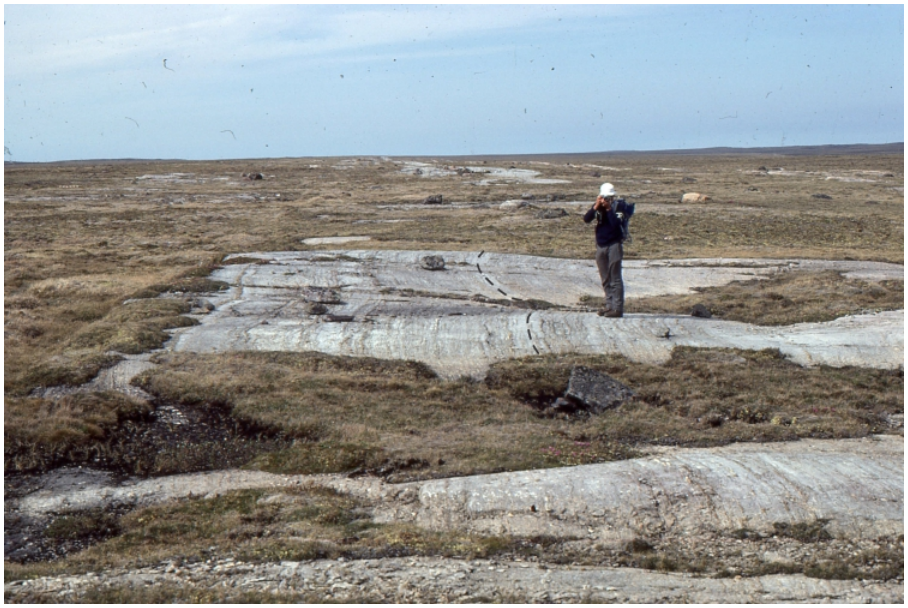


Figure 62.15.19: A cross-section view of cryptalgal domes, some up to 4m wide, interpreted to have formed on a broad carbonate platform at depths below the influence of normal, fair-weather waves. Most are simple structures lacking the complex branching more common in shallower, agitated waters. The dashed black line in front of this keen geologist traces laminae across two domes. indicating synoptic relief up to 60 cm. The carbonate rocks here are 100% dolomite. Bedding dips to the right.

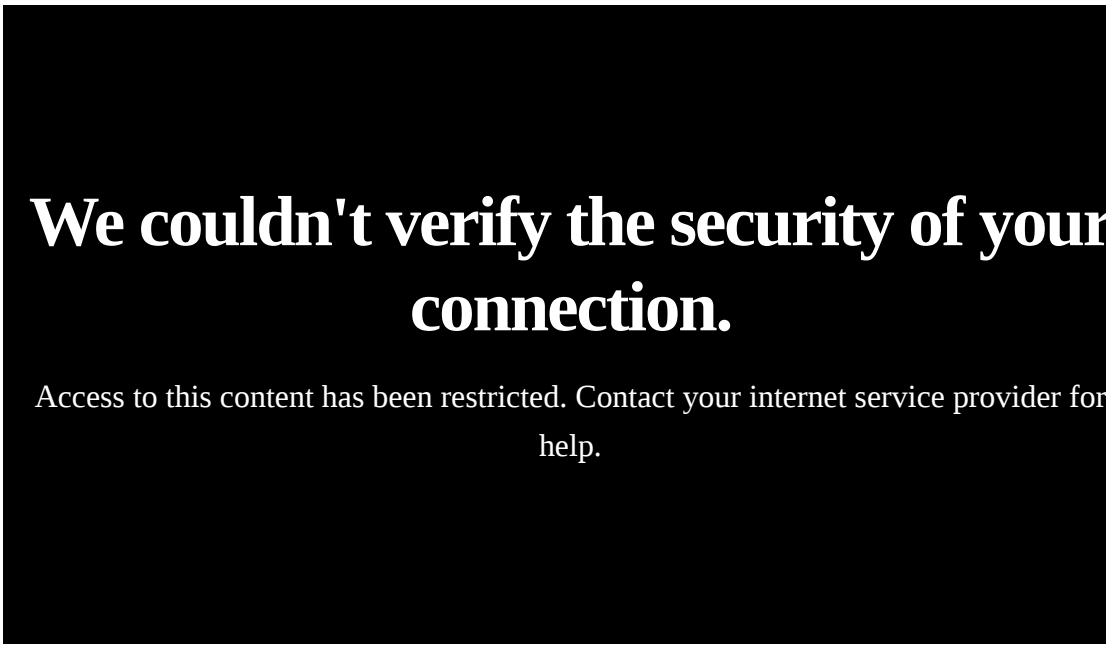
The field excursion ends beneath the waves offshore, figuratively in our interpretations of the two billion year-old rock record, and practically as we made our way across the modern sea floor. Surfacing will allow us to breathe, and reflect on what we have observed.

Unless indicated otherwise, all the images used in this chapter are from geological-digressions.com

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62.16: Modern and ancient microbialites in Western Australia

On this video-based virtual field experience, join paleontologist **Heidi Allen** to explore primary sedimentary structures built by microbes. Heidi works at the Geological Survey of Western Australia, and here presents a look at that region's wealth of stromatolites, thrombolites, and related structures – both from the fossil record and growing alive and well in the modern world!



GSWA Webinar Series: Heidi Allen – Western Australia's wealth of microbialites from DMIRS on Vimeo.

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62.17: Corridor H, West Virginia

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CHAPTER OVERVIEW

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63.1: Minerals

Here are 16 samples (or sample sets) of minerals that you will use to answer lab questions. Examine each sample in detail by zooming in and exploring its physical characteristics.

Sample Set A:



Sample A (all four) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	3	white	Excellent; 3 directions, NOT at 90° to each other	No	fizzes when HCl is applied

Sample B:



Sample B (both) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
metallic	2.5	gray to dark gray	3, at 90°	No	Very dense (specific gravity = 7.5); stinks of hydrogen sulfide (rotten egg smell) if HCl is applied.

Sample set C:



Sample C1 (left) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	1 – 6	brick red	No	No	(none)

Sample C2 (right) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
metallic	1 – 6	brick red	No	No	(none)

Sample D:



Sample D has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	7	(too hard to streak)	No	Brittle; may show parting	Dodecahedral crystals are distinctive.

Sample E:



Sample E (both) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2 – 2.5	white	1 (perfect)	No	can be split into thin, flexible sheets

Sample F:



Sample F has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	6.5	white	2, at 90°	No	lamellae

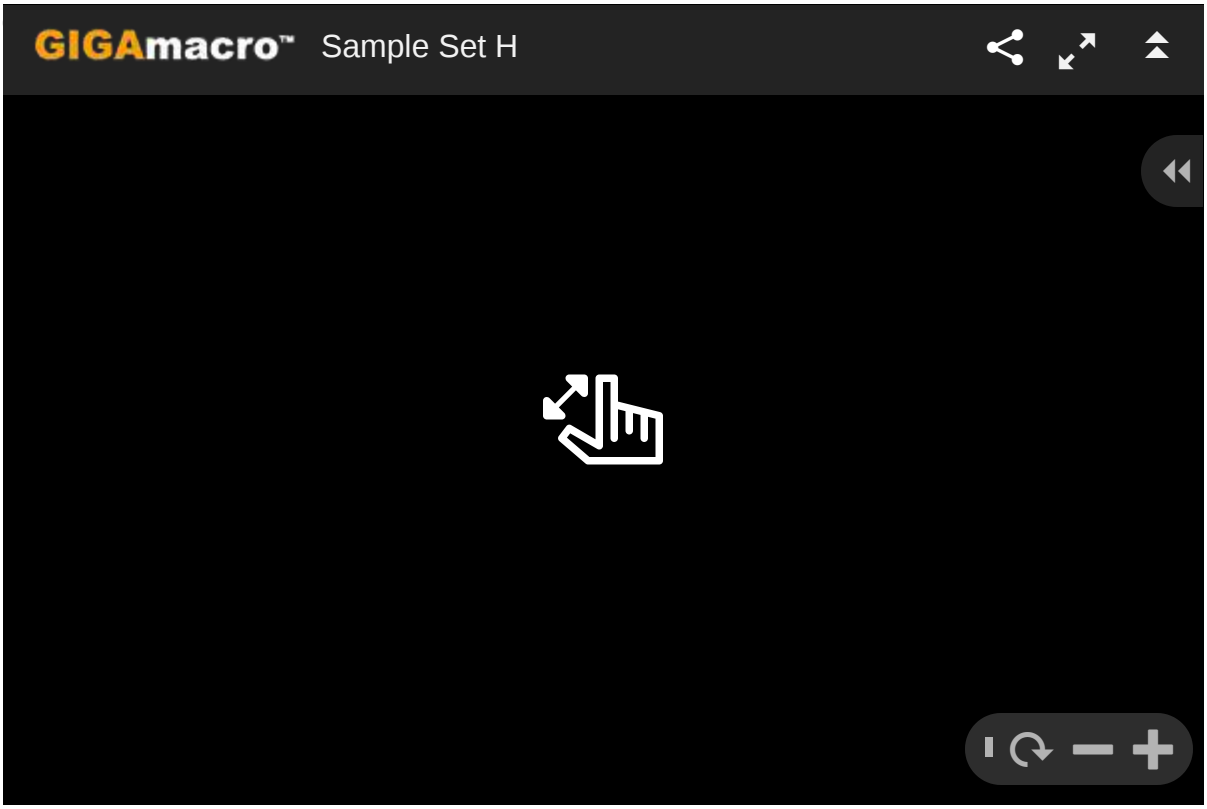
Sample G:



Sample G (both) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2.5 – 3	gray-brown to white	1 (excellent)	No	can be split into thin, flexible sheets

Sample set H:



Sample H1 (left) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	6-6.5	dark gray	Absent	No	magnetic

Sample H2 (middle) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	6	white	2, at 90°	No	lamellae

Sample H3 (right) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
metallic	2.5	gray to dark gray	3, at 90°	No	Very dense (specific gravity = 7.5); stinks of hydrogen sulfide (rotten egg smell) if HCl is applied.

Sample I:



Sample I has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
metallic	6-6.5	dark gray	absent	No	often forms cubes, sometimes with striations

Sample set J:



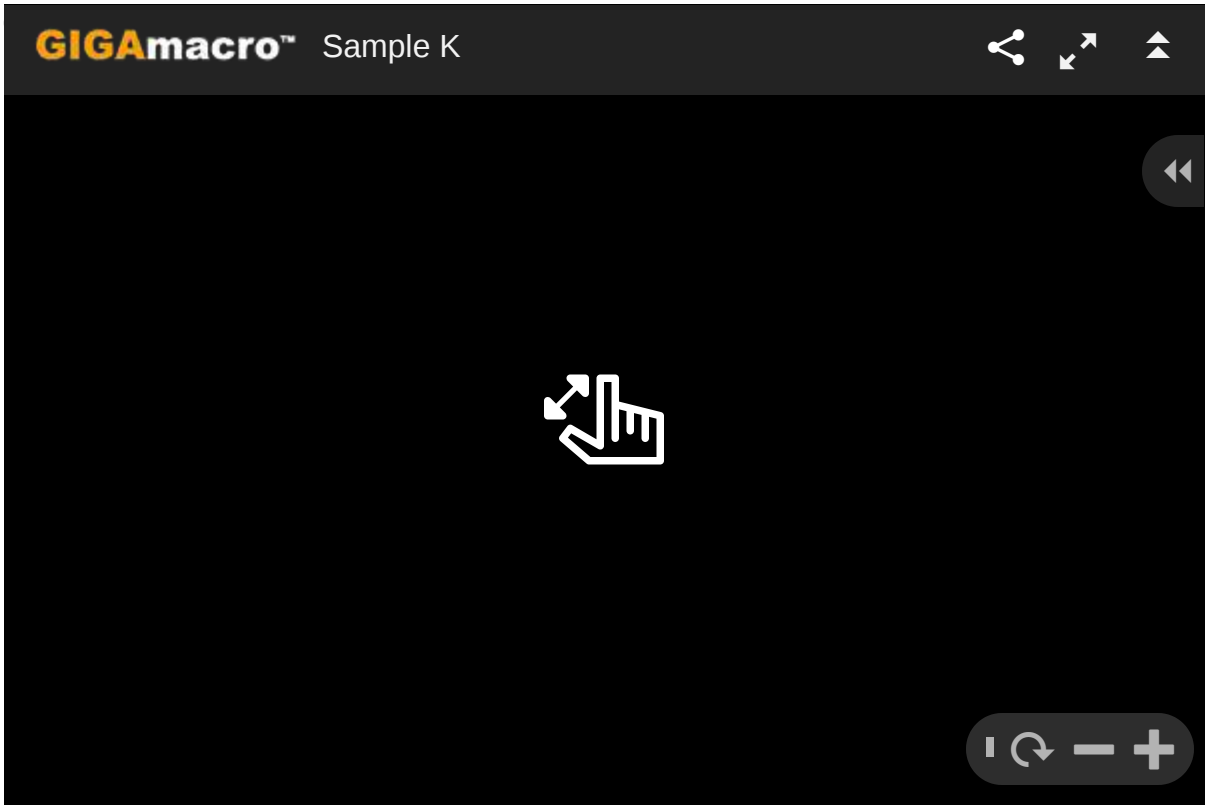
Sample J1 (left) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	5.5-6	white to pale gray	Good. 2 planes, at about 60°/120° to each other	No	color dark gray to black; forms prisms.

Sample J2 (right) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	5.5-6	white to pale gray	Moderate. 2 planes, at about 90° to each other	No	color dark green to brown or black; forms short, 8-sided prisms

Sample K:



Sample K has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2-5	white	absent	No	often forms dull or silky masses and asbestos (fibrous) forms.

Sample set L:



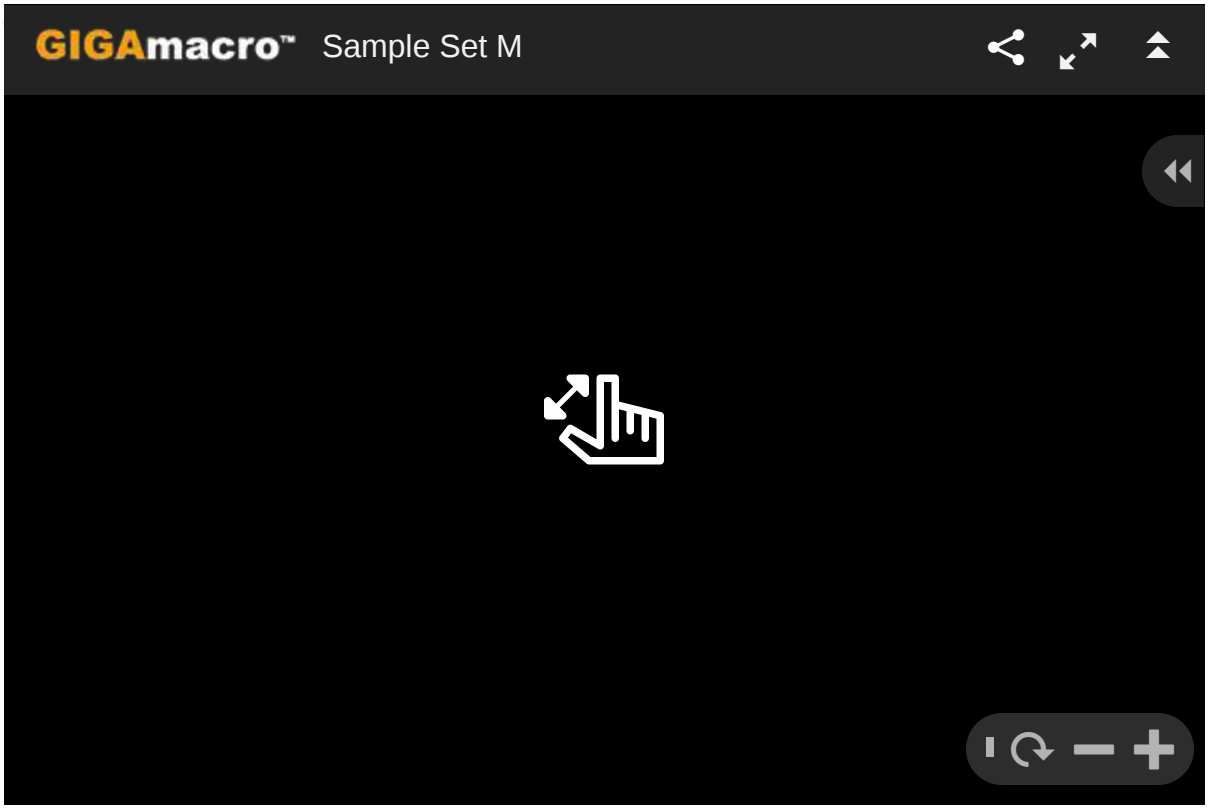
Sample L1 (left) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2	white	Good in 1 direction; poor in 2 more directions.	No	Forms tabular crystals, prisms, or blades. Transparent when pure; translucent when impurities are present.

Sample L2 (right) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	1	white	Absent	No	Very soft. Feels soapy to the touch.

Sample M:



Sample M (all four) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	7	None (too hard to streak)	Absent	Conchoidal	No reaction to HCl. Comes in many colors.

Sample set N:



Sample N1 (left) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2.5	white	Excellent: 3 planes at 90° to each other (so it breaks into cubes)	No	Tastes salty.

Sample N2 (right) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	4	white	Excellent: 4 planes	No	Often colorless (so it would appear black on a black background, like you see here), or translucent purple or green.

Sample O:



Sample O (both) has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	3.5-4	White	Excellent; 3 directions, NOT at 90° to each other	none	Reaction to HCl occurs only if the mineral is powdered first.

Sample P:



Sample P has the following properties:

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
metallic	1	Dark gray	Excellent in 1 direction	none	Greasy feel. Exceptionally soft.

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63.2: 3D mineral samples, set 1

Here are 10 samples of minerals that you will use to answer lab questions. Examine each sample in detail by rotating it and exploring its shape and physical characteristics. Properties of each mineral are listed in the table below the 3D model.

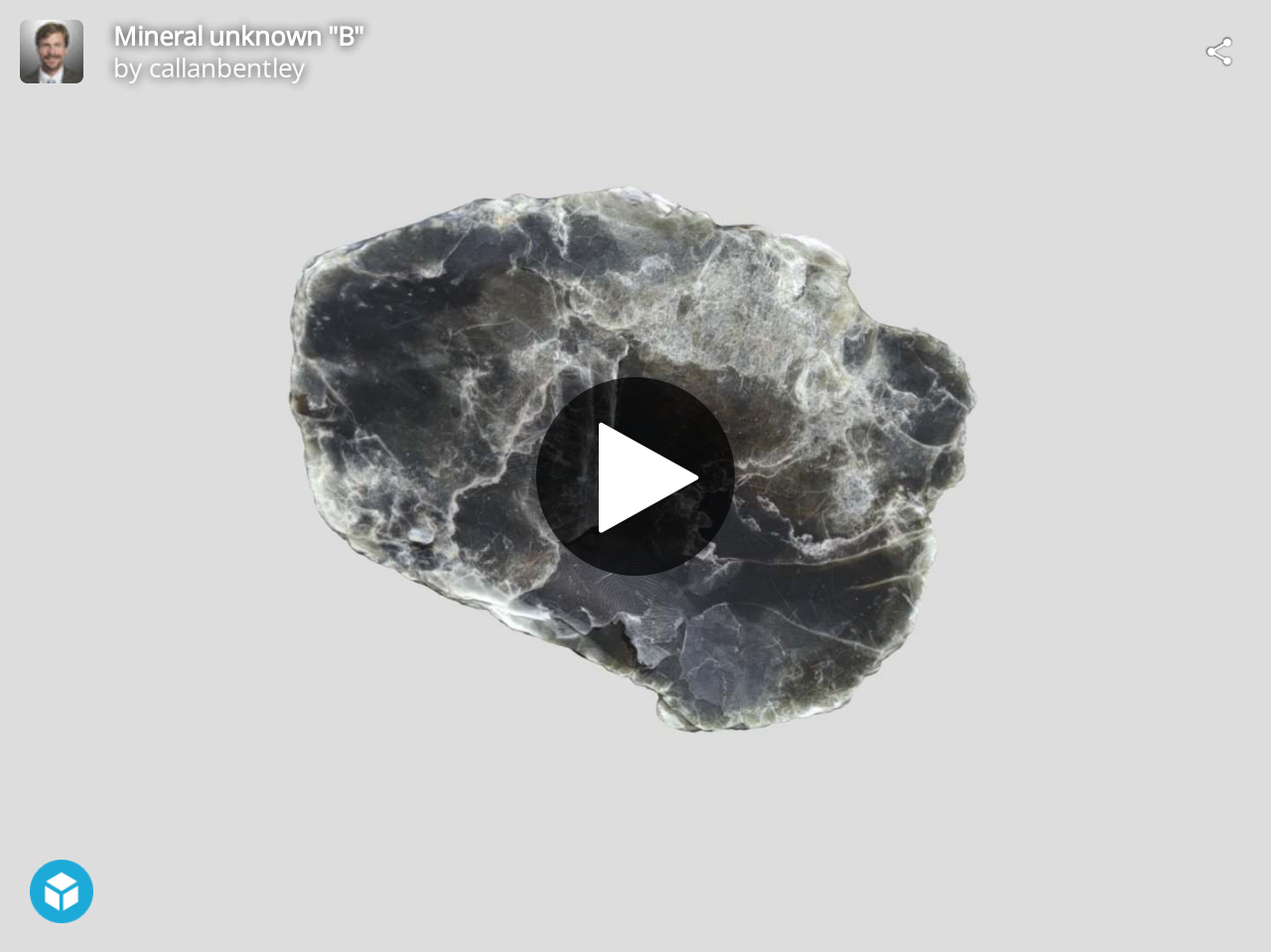
Mineral A



Mineral unknown "A" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	1	white	Absent	No	Very soft. Feels soapy to the touch.

Mineral B



Mineral unknown "B" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2.5 – 3	gray-brown to white	1 (excellent)	No	can be split into thin, flexible sheets

Mineral C



Mineral unknown "C" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	3	white	Excellent; 3 directions, NOT at 90° to each other	No	fizzes when HCl is applied

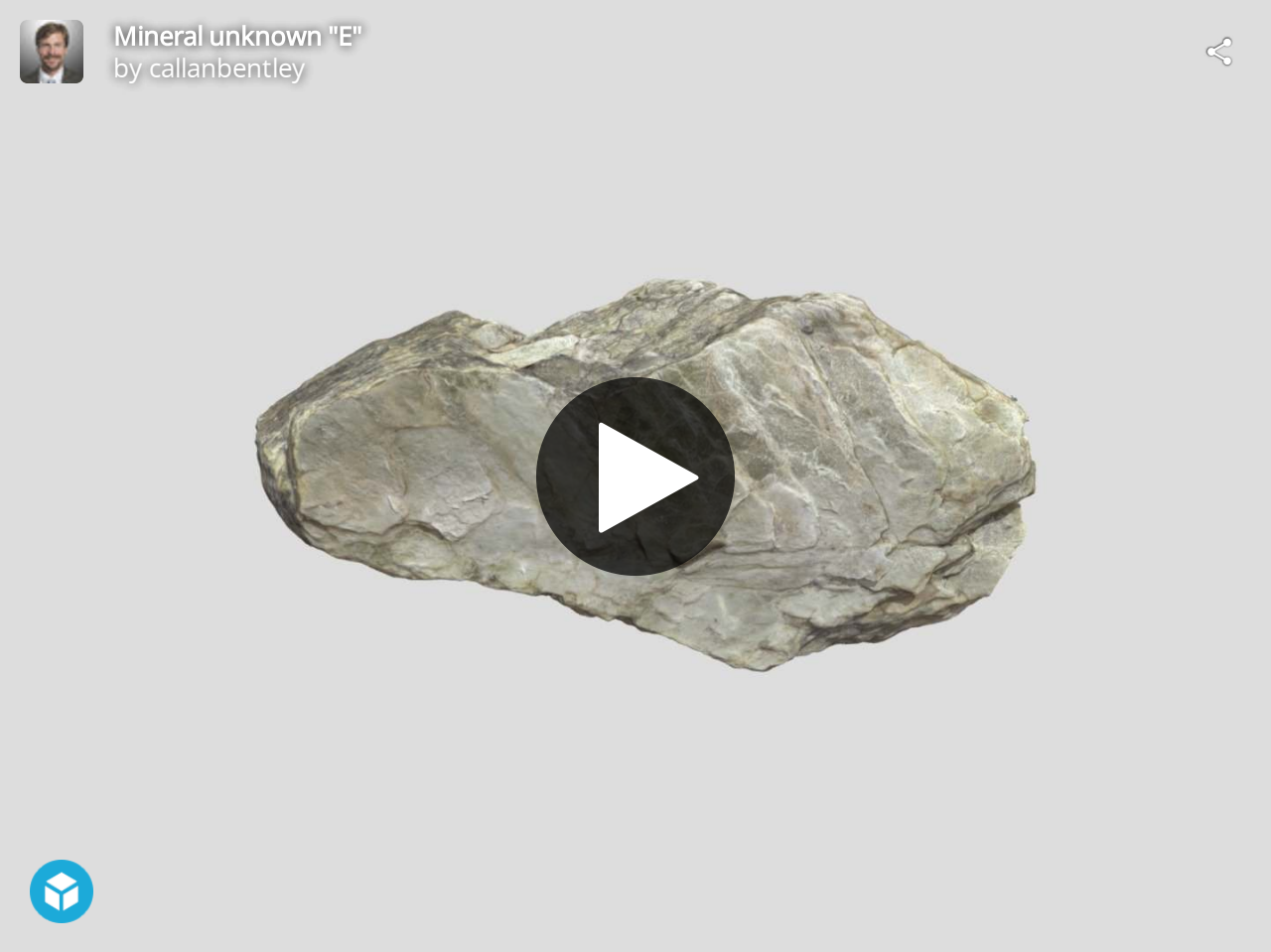
Mineral D



Mineral unknown "D" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	5.5-6	white to pale gray	Moderate. 2 planes, at about 90° to each other	No	color dark green to brown or black; forms short, 8-sided prisms

Mineral E



Mineral unknown "E" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2 – 2.5	white	1 (perfect)	No	can be split into thin, flexible sheets

Mineral unknown "F" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	6.5	white	2, at 90°	No	lamellae

Mineral unknown "G" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	7	None (too hard to streak)	Absent	Conchoidal	No reaction to HCl. Comes in many colors.

Mineral unknown "H" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	7	(too hard to streak)	No	Brittle; may show parting	Dodecahedral crystals are distinctive.

Mineral unknown "H" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	2.5	white	Excellent: 3 planes at 90° to each other (so it breaks into cubes)	No	Tastes salty.

Mineral unknown "J" by callanbentley on Sketchfab

Luster	Hardness	Color of streak	Cleavage?	Fracture?	Other properties
nonmetallic	5.5-6	white to pale gray	Good. 2 planes, at about 60°/120° to each other	No	color dark gray to black; forms prisms.

Submit your answers to your instructor.

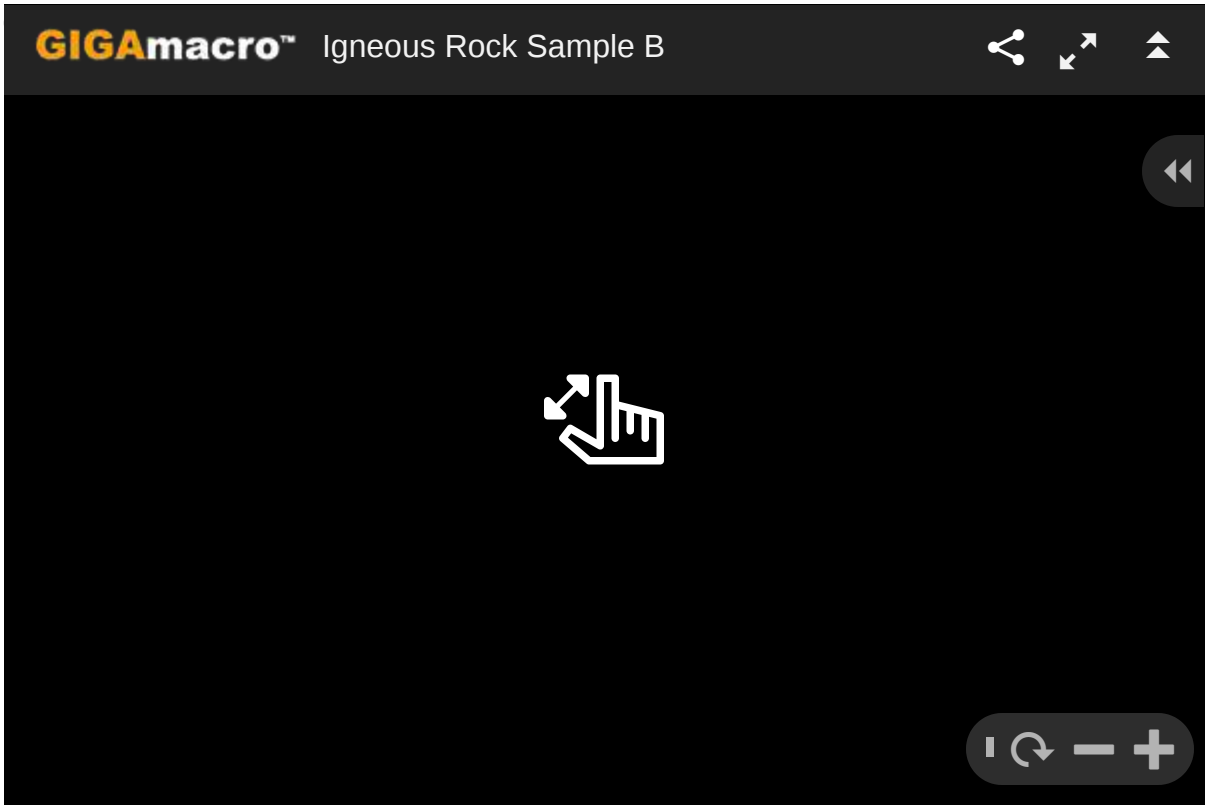
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63.3: Igneous rocks

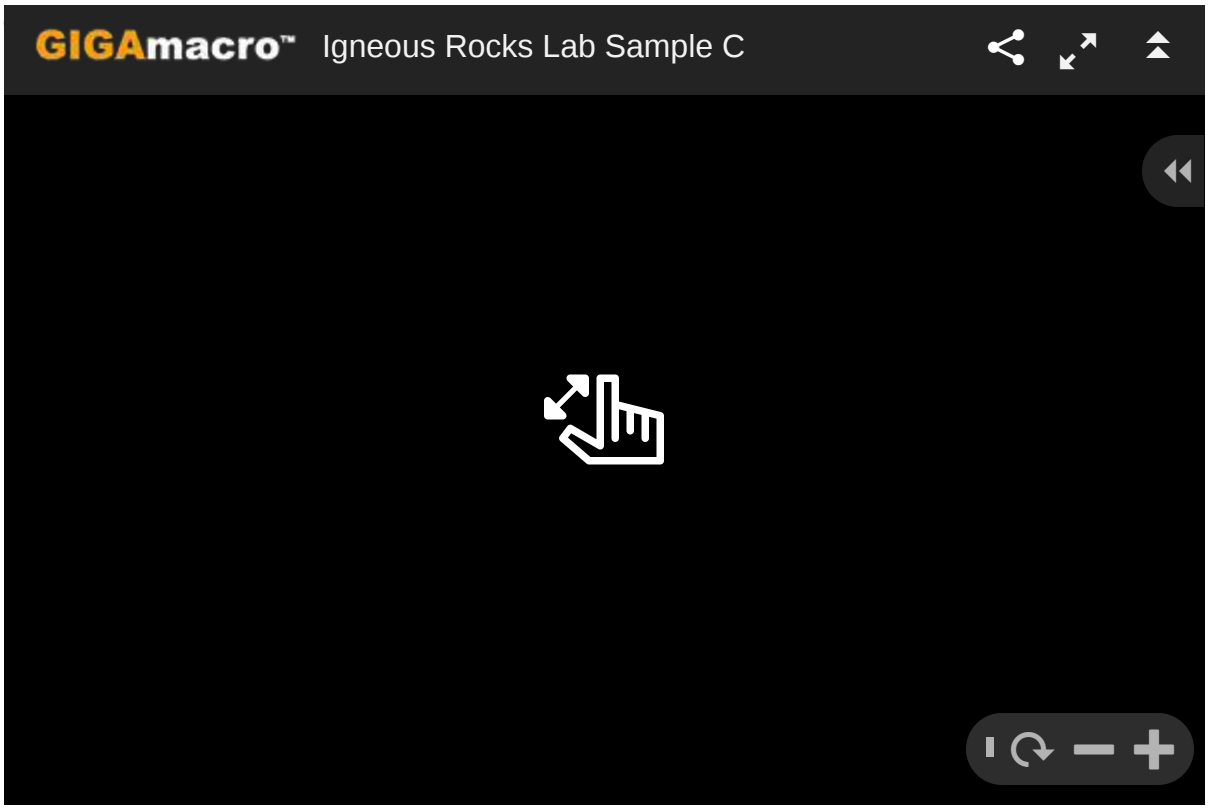
Sample A



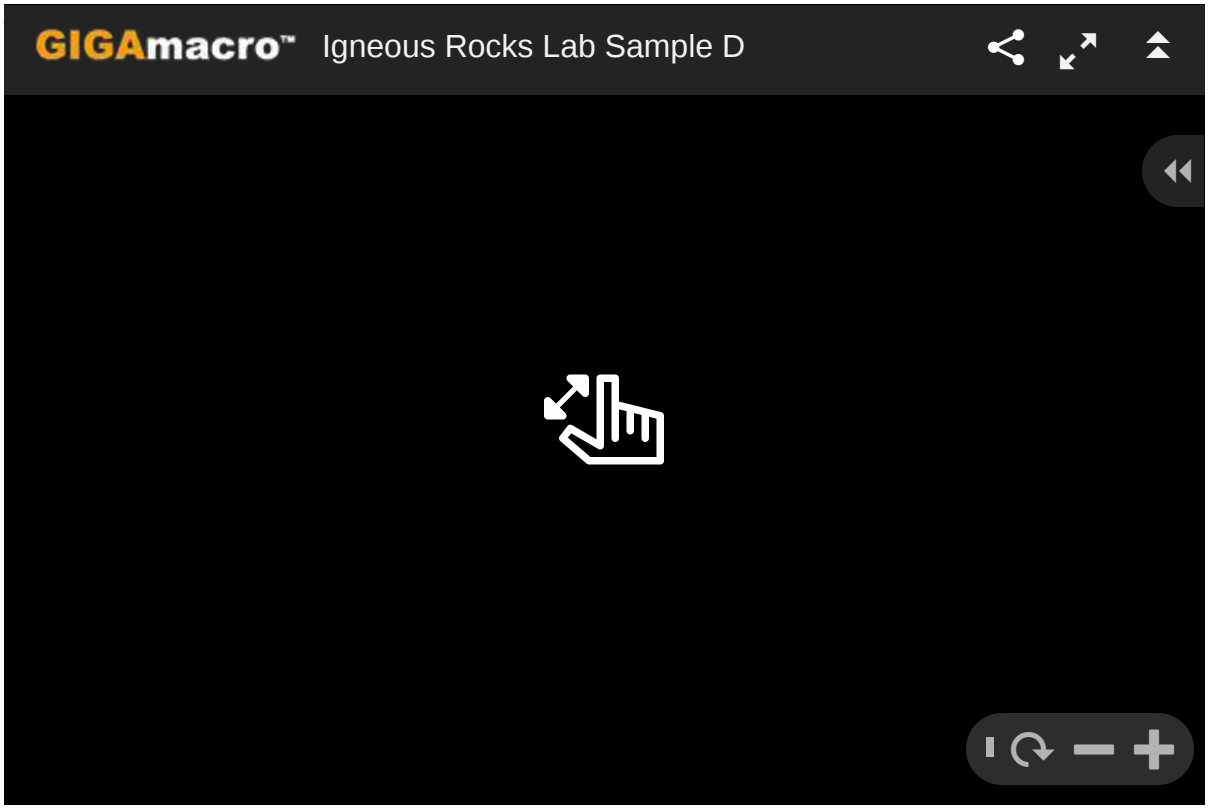
Sample B



Sample C



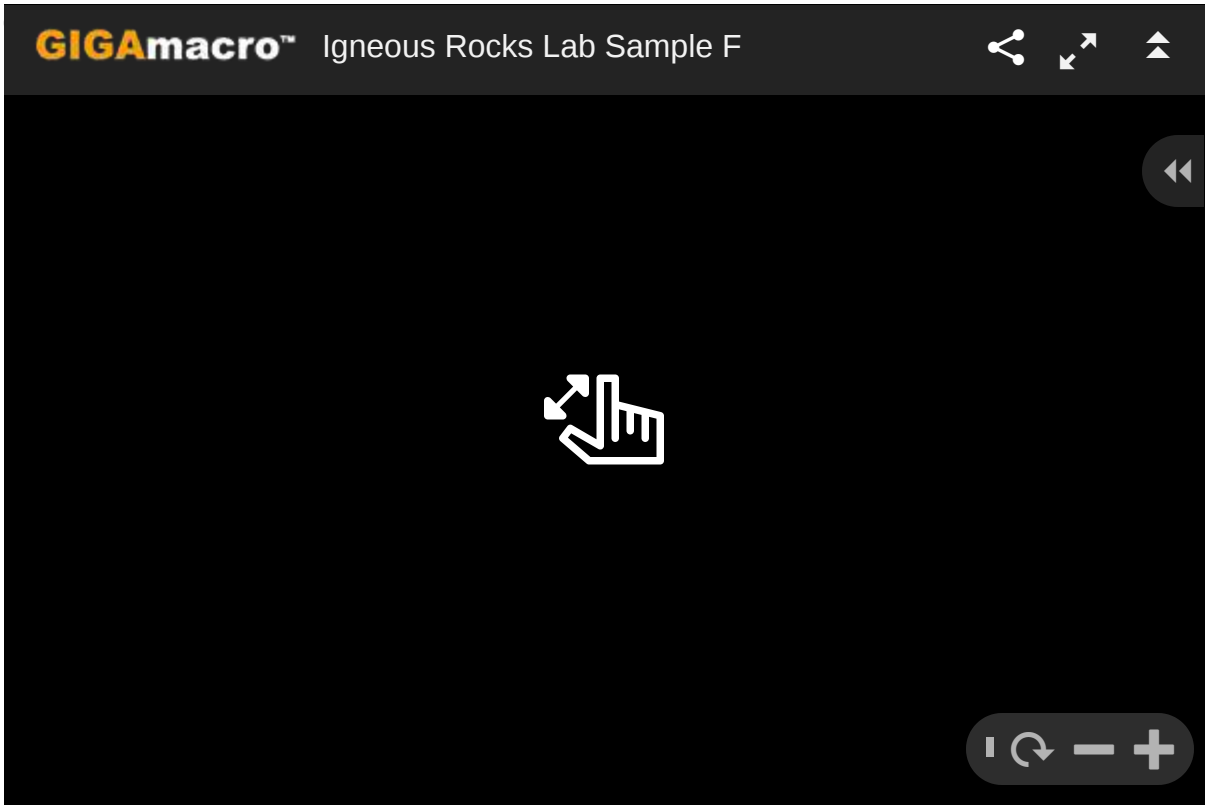
Sample D



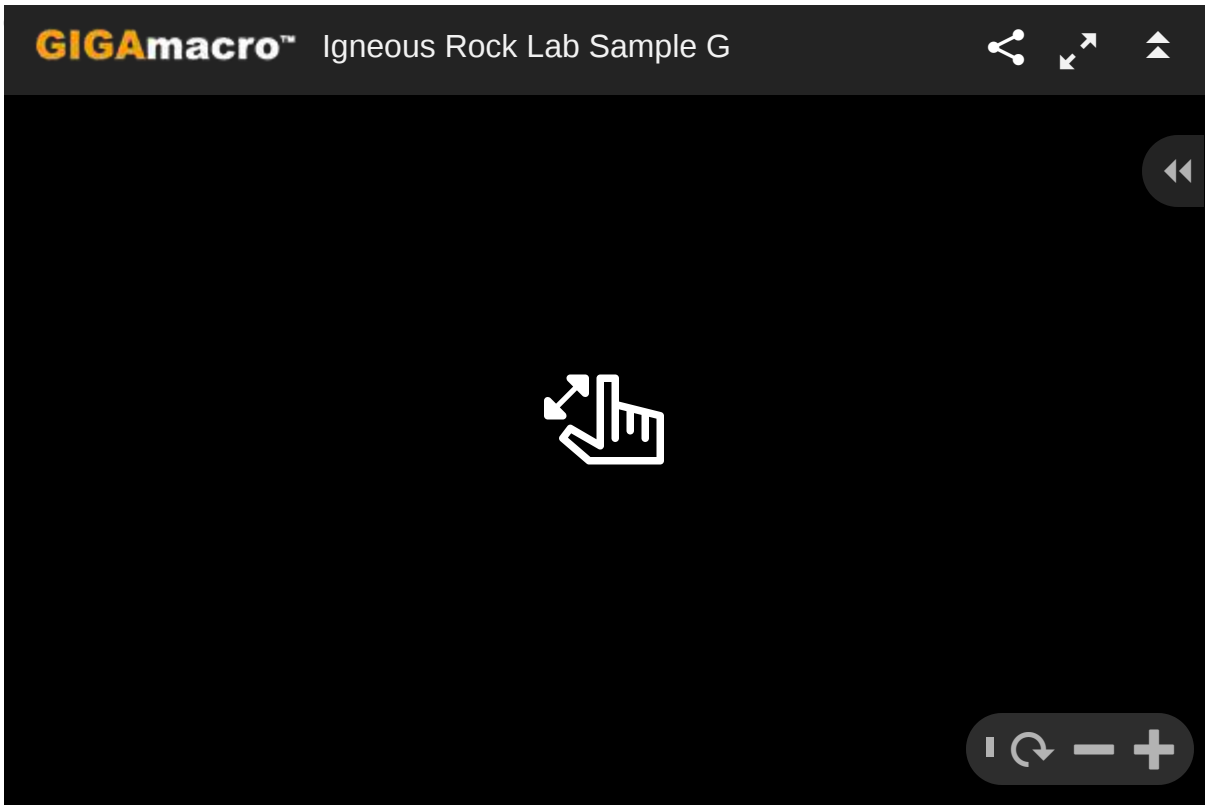
Sample E



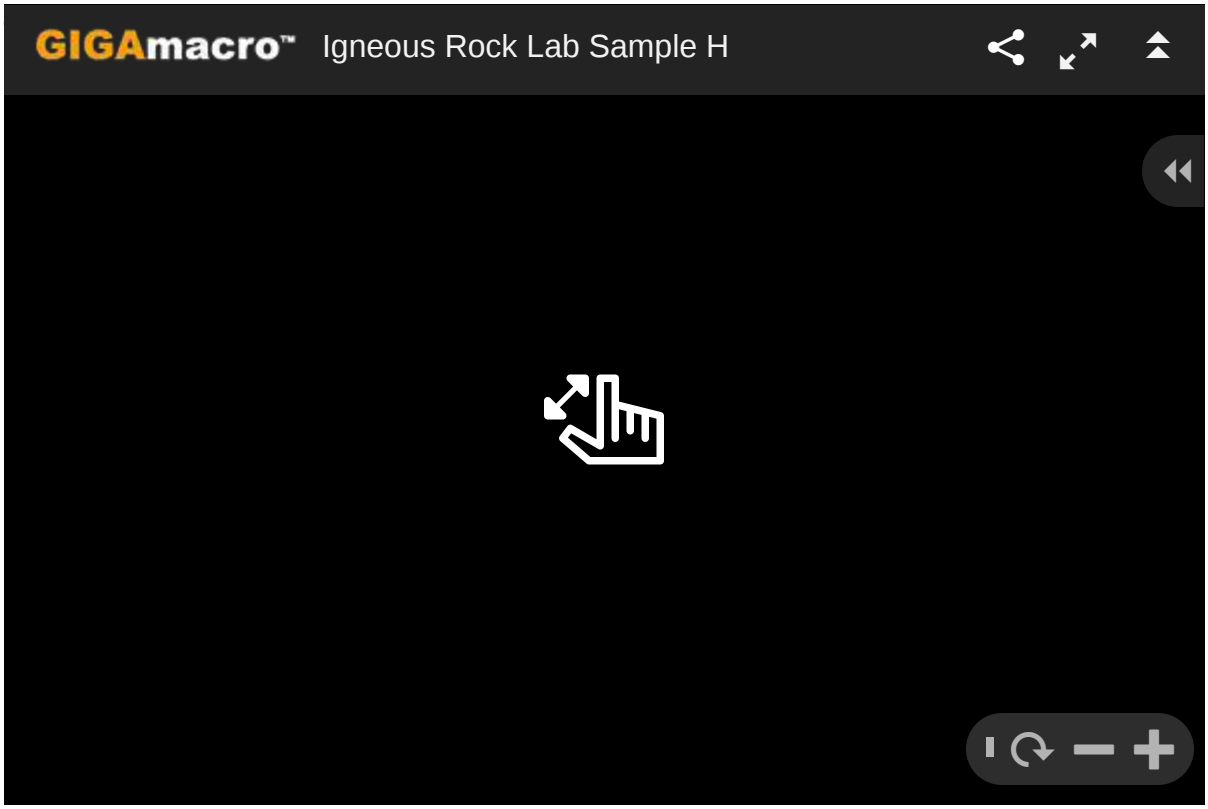
Sample F



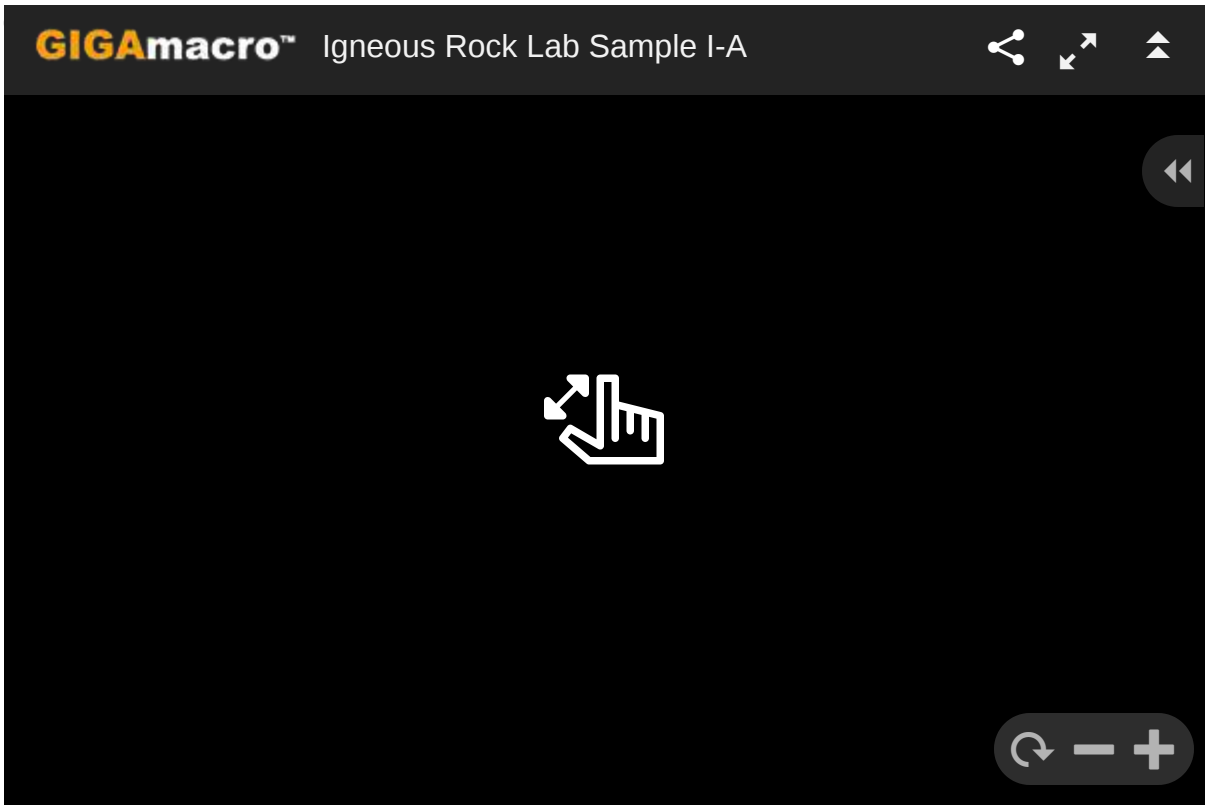
Sample G



Sample H


GIGAmacro™ Igneous Rock Lab Sample H

Sample I

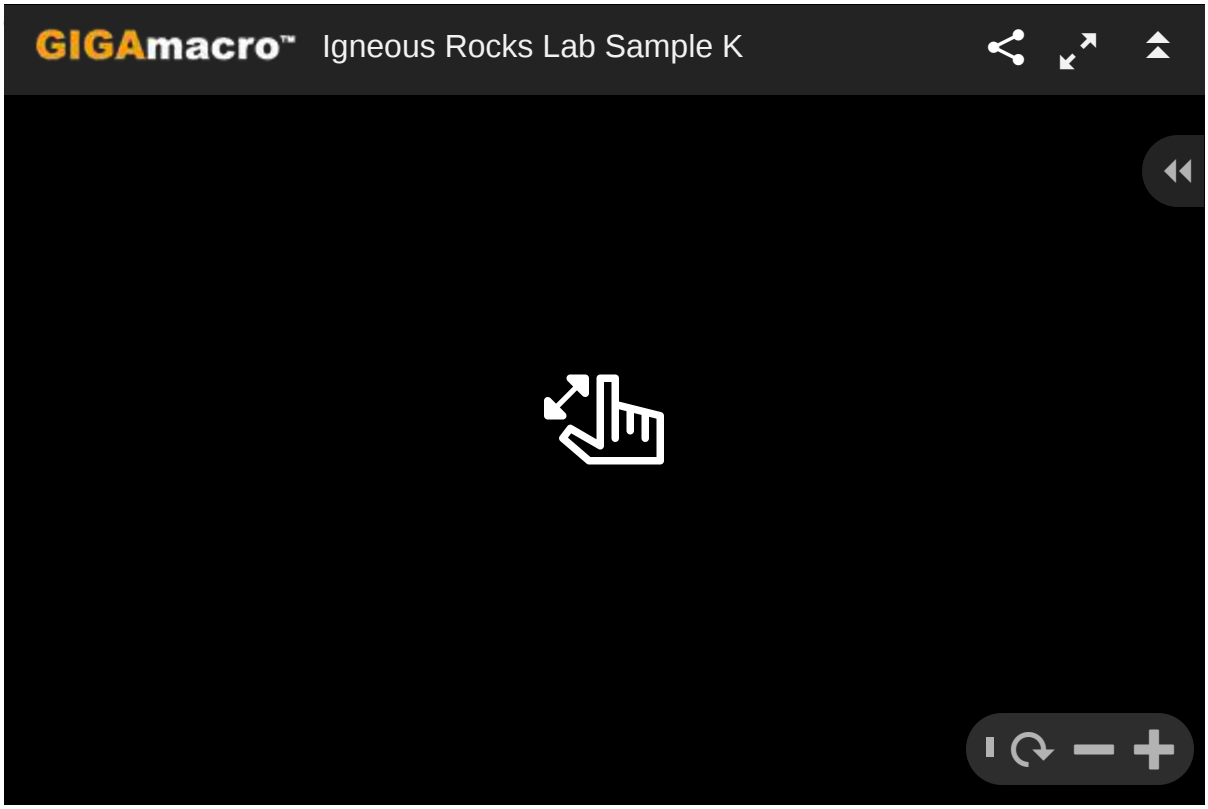

GIGAmacro™ Igneous Rock Lab Sample I-A



Sample J



Sample K



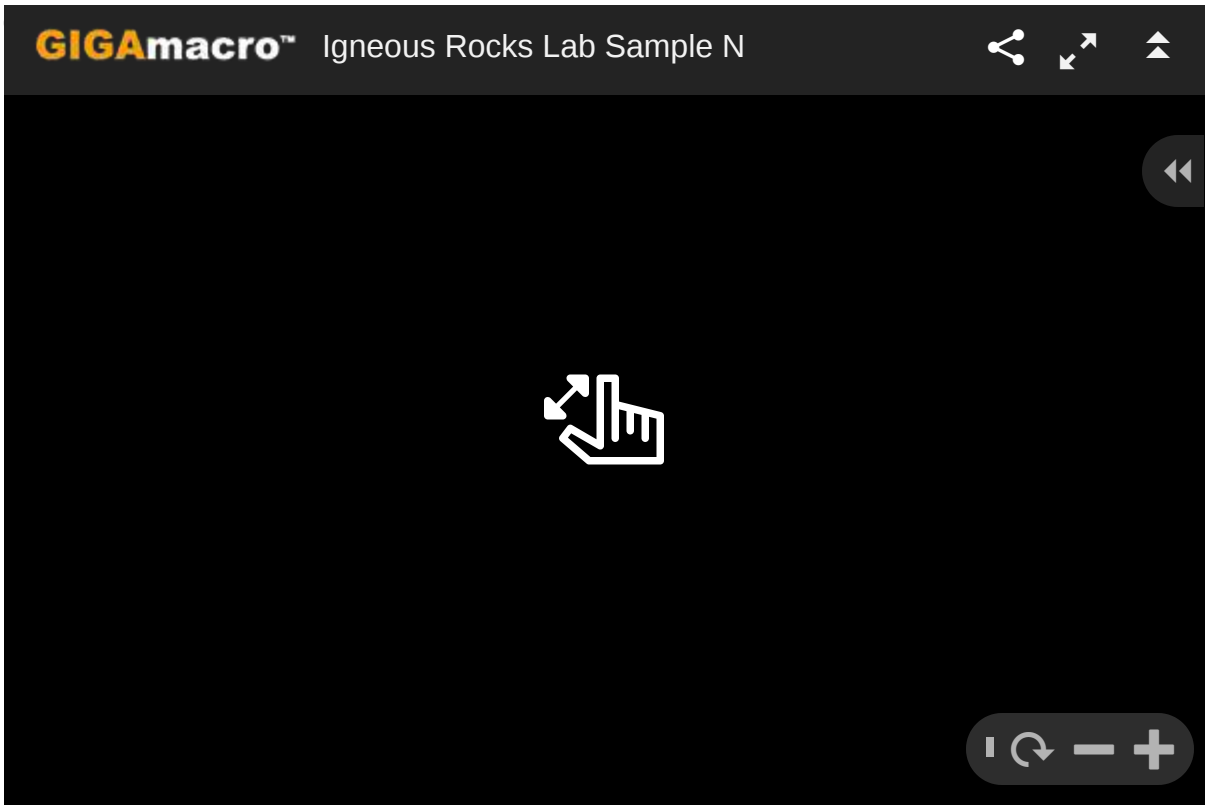
Sample L1 (Left) Sample L2 (Right)



Sample M



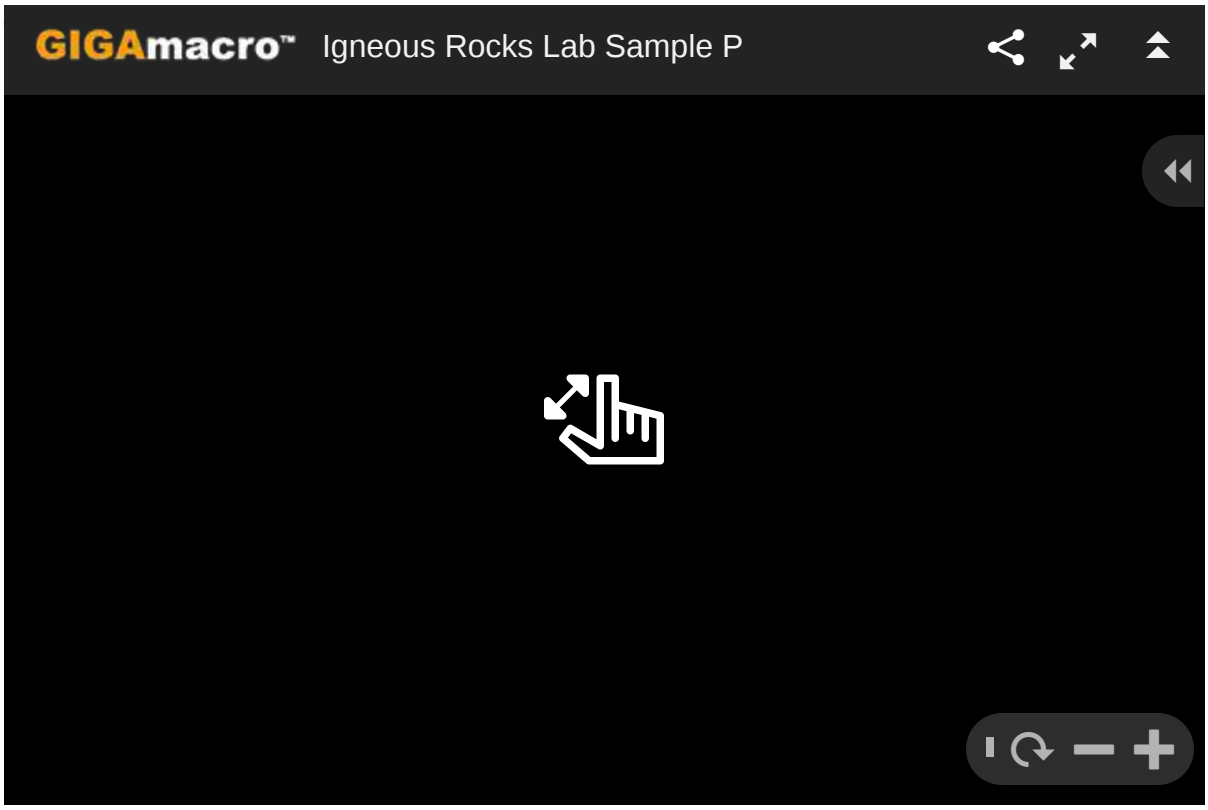
Sample N



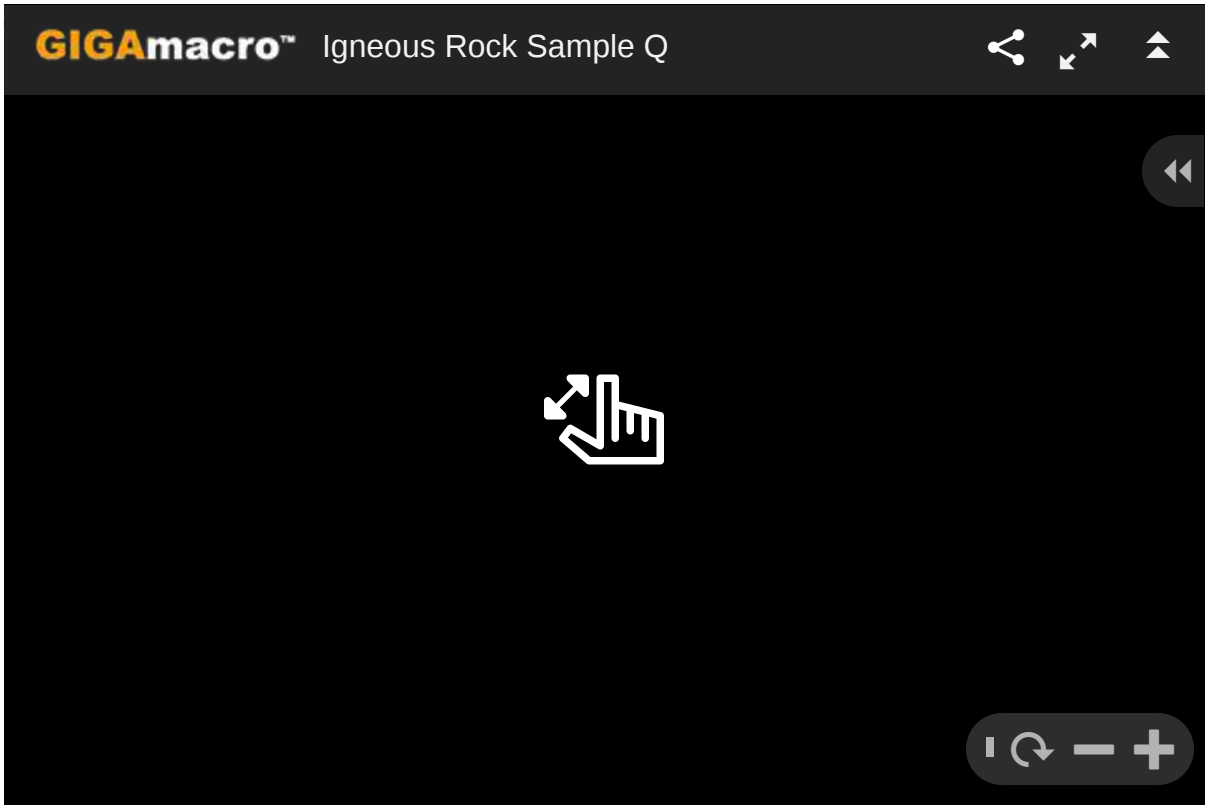
Sample O



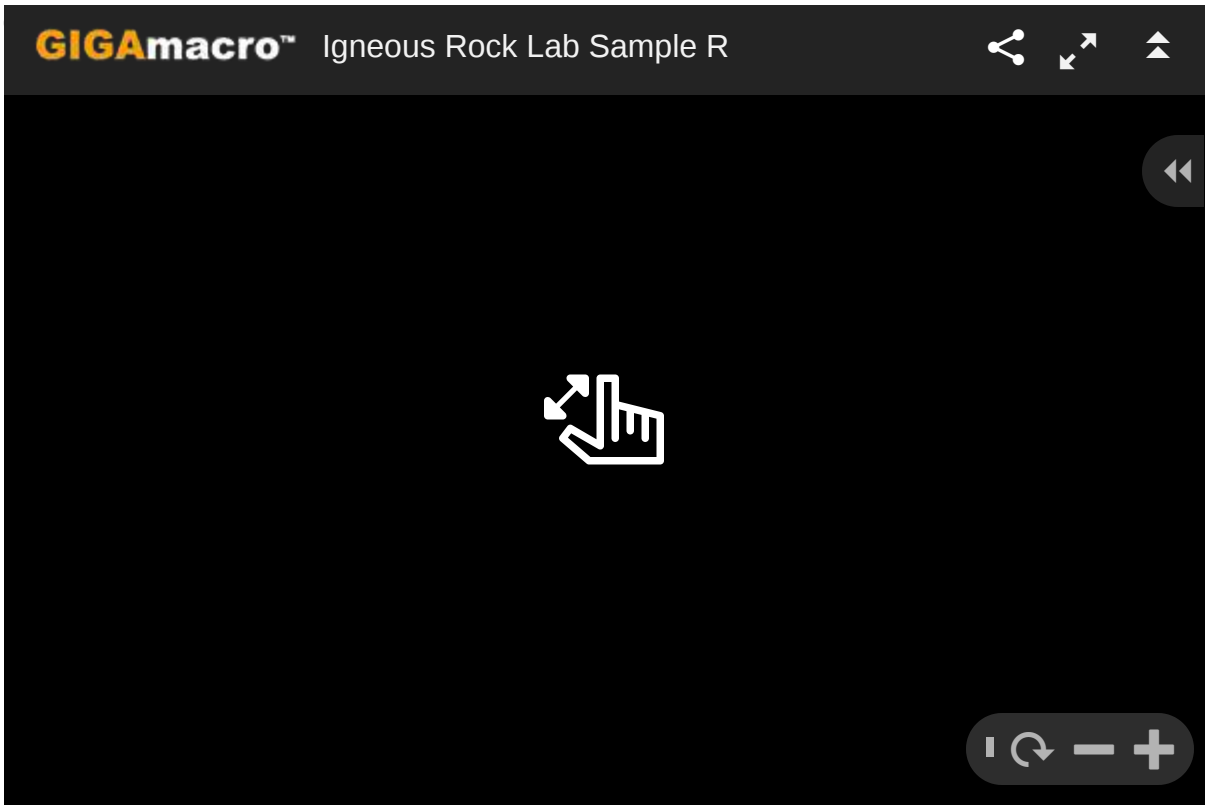
Sample P



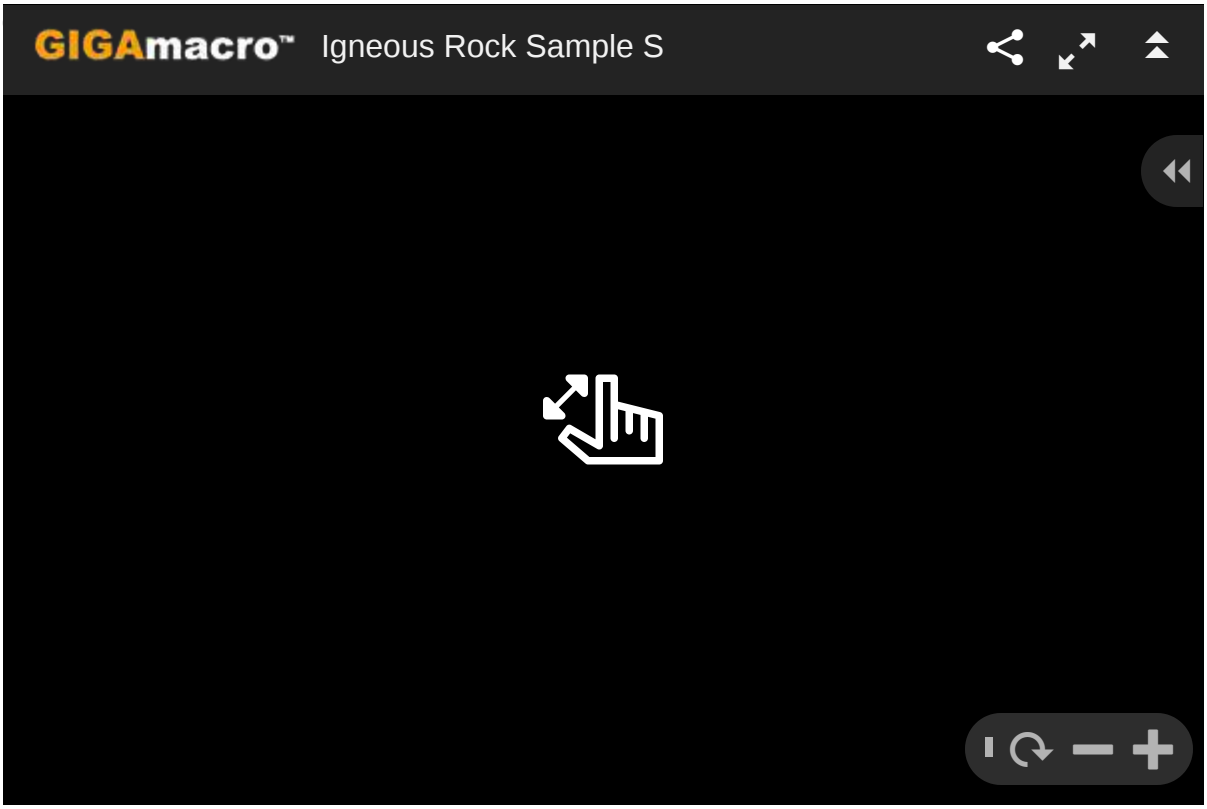
Sample Q



Sample R



Sample S

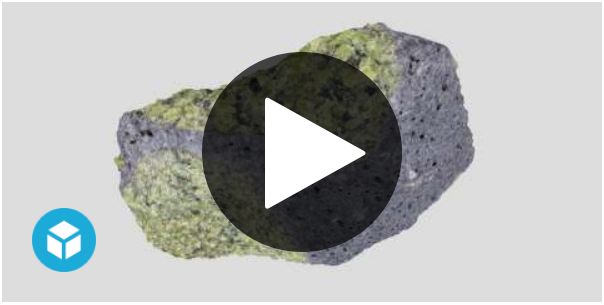


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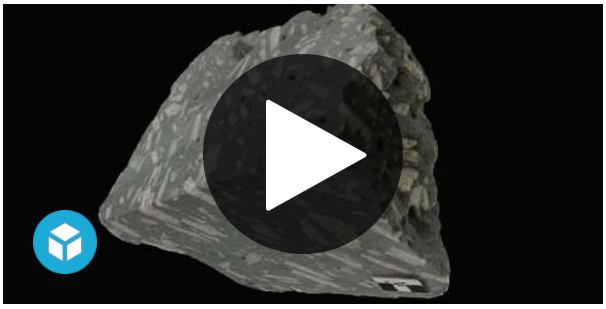
63.4: Igneous supplementary set

A collection of virtual igneous samples for the use of students who wish to practice identification and interpretation.

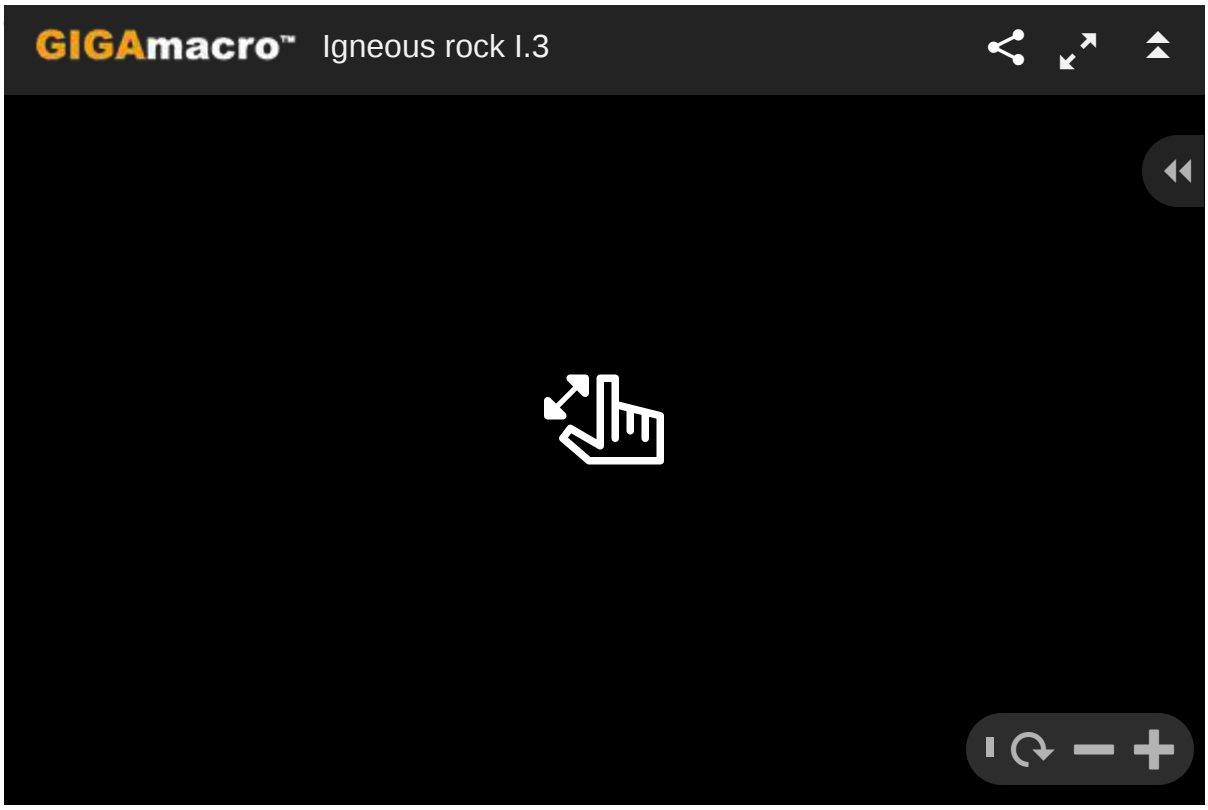
Sample I.1:



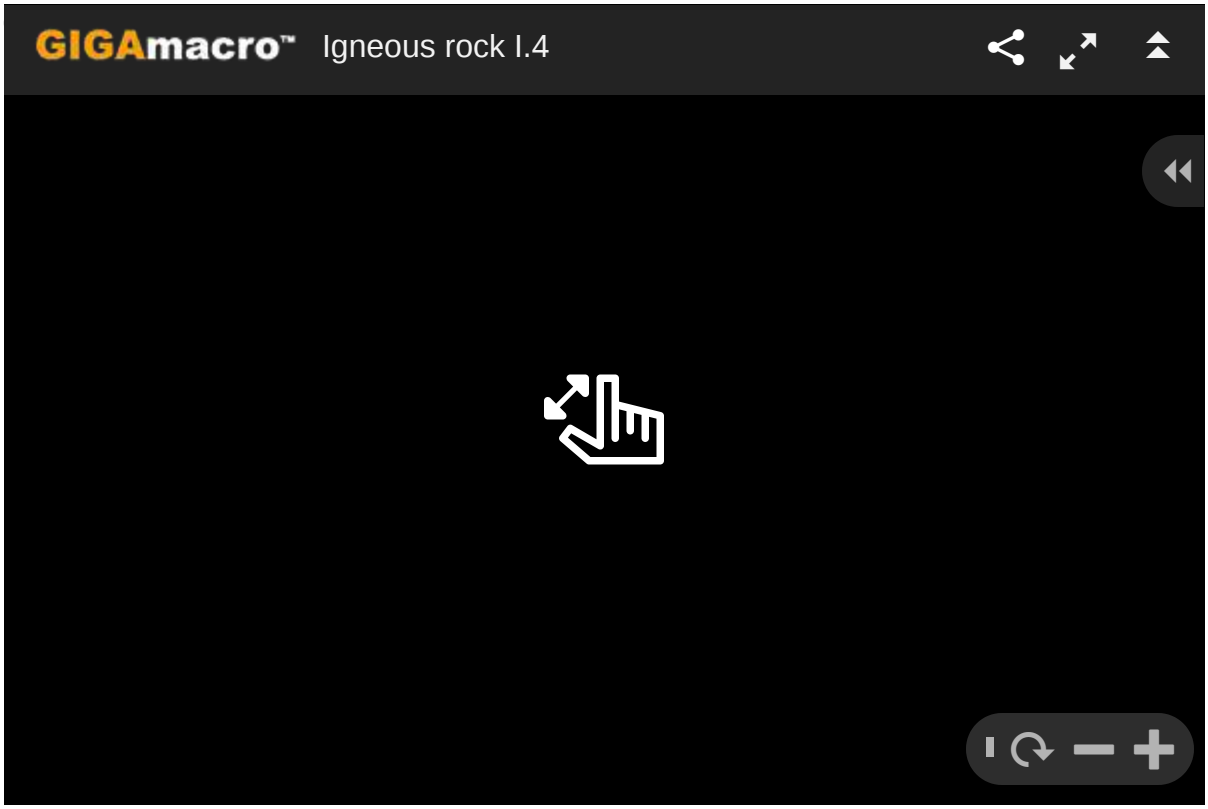
Sample I.2:



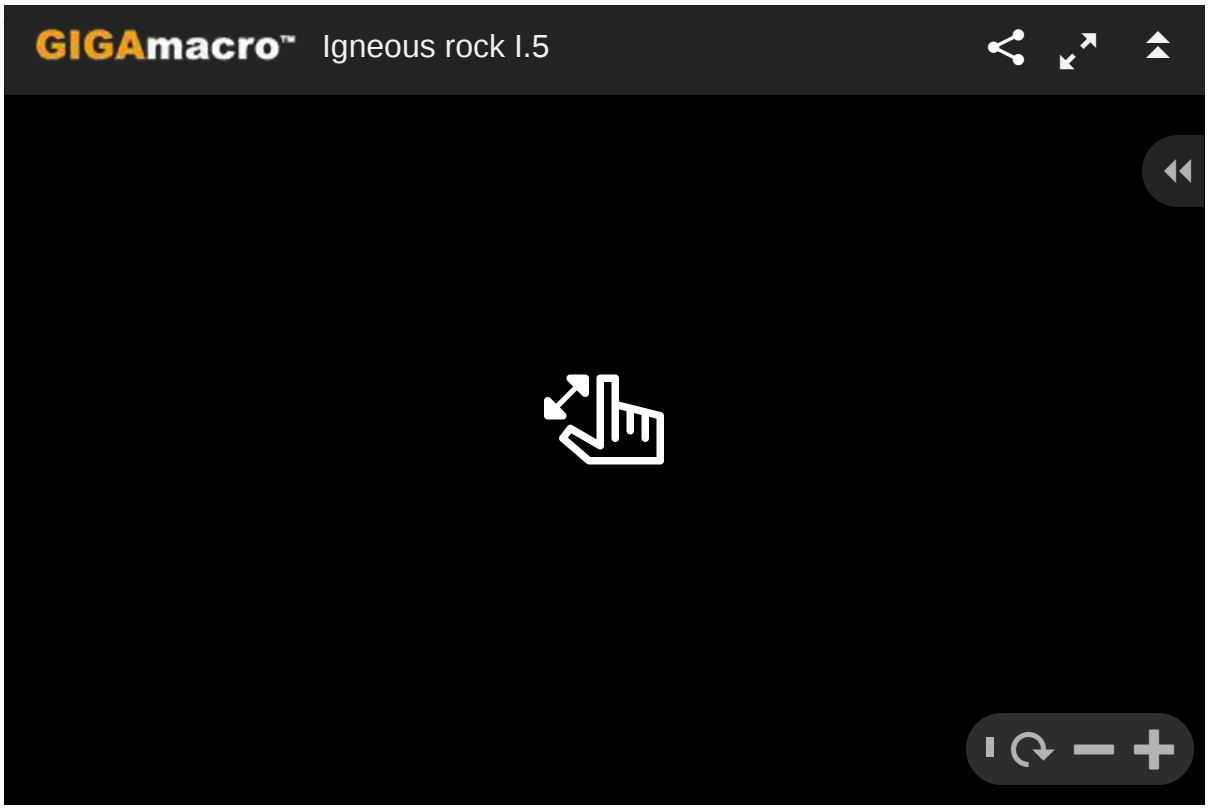
Sample I.3:



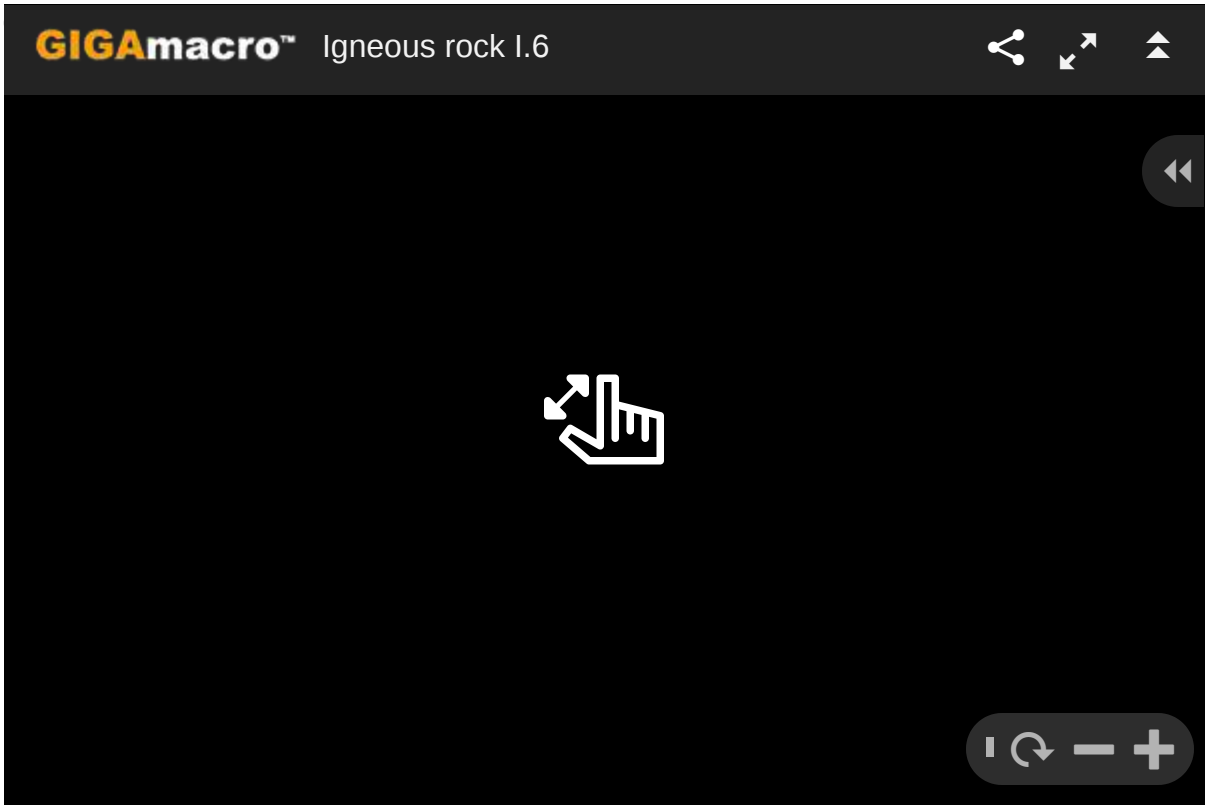
Sample I.4:



Sample I.5:



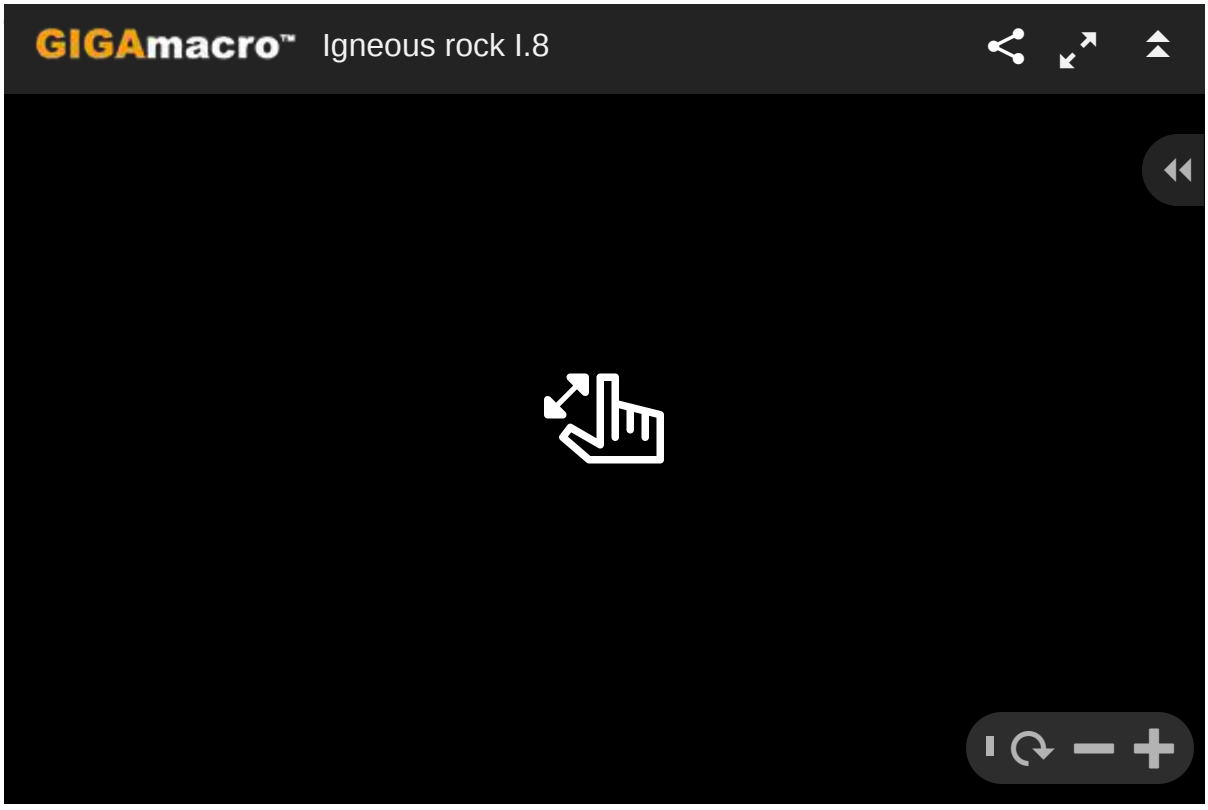
Igneous sample I.6:



Igneous sample I.7

[This Sketchfab 3D model has been disabled.]

Igneous sample I.8:



Igneous sample I.9:



Igneous outcrop I.10:



Igneous outcrop I.11:



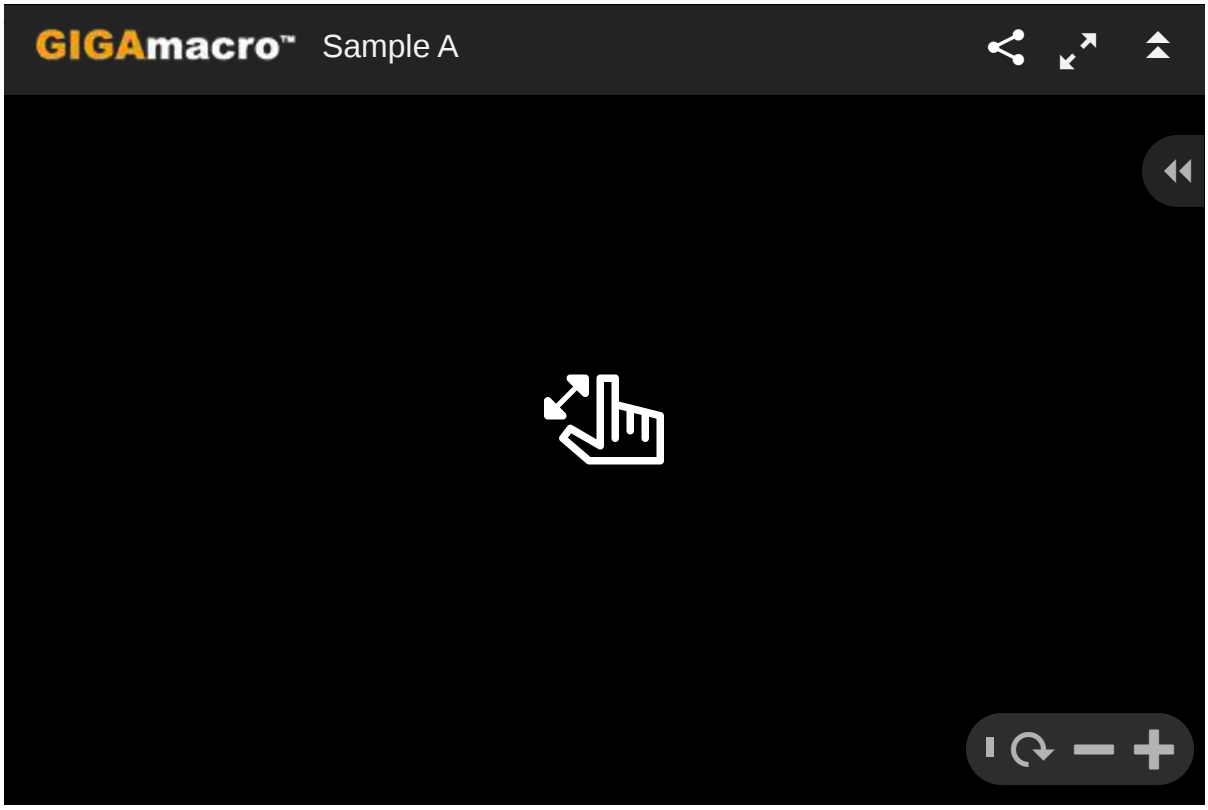
This page titled 63.4: Igneous supplementary set is shared under a CC BY-NC 4.0 license and was authored, remixed, and/or curated by Callan Bentley, Karen Layou, Russ Kohrs, Shelley Jaye, Matt Affolter, and Brian Ricketts (OpenGeology) via source content that was edited to the style and standards of the LibreTexts platform.

63.5: Sedimentary rocks

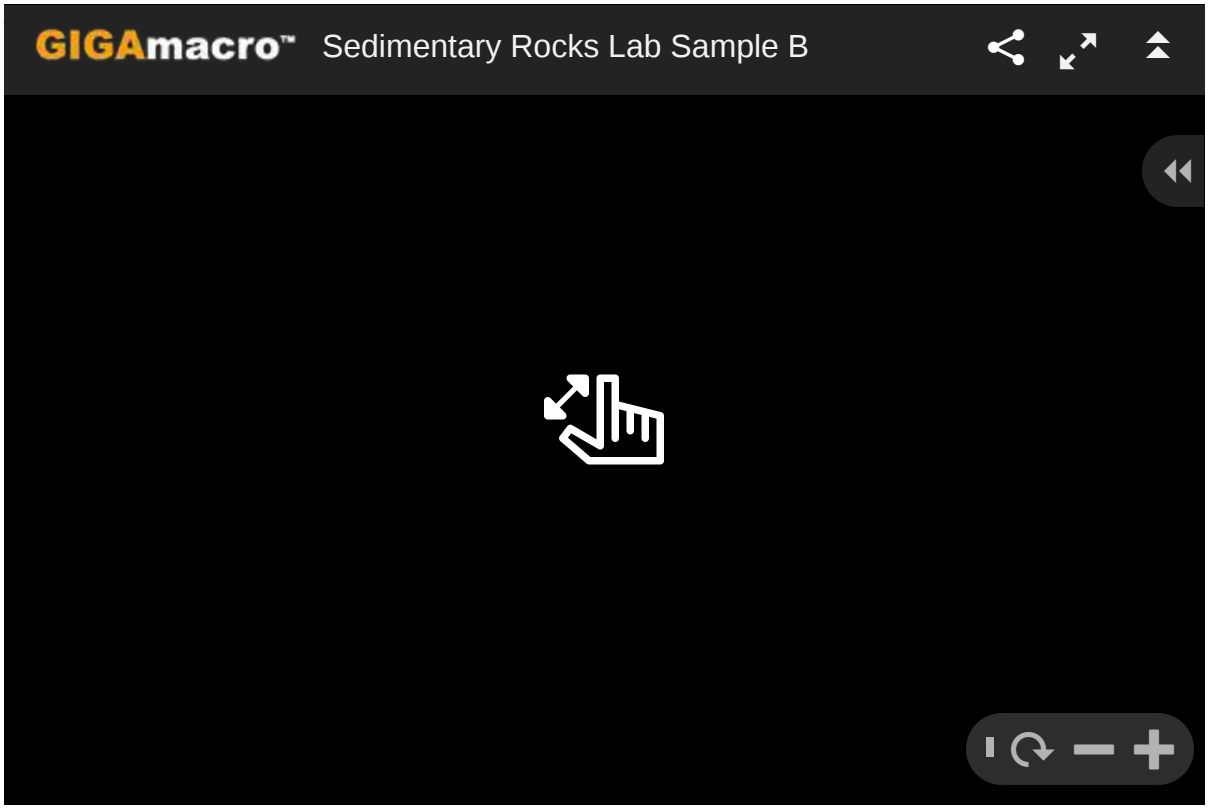
Sedimentary rocks: virtual sample set

Here are 26 samples of sediments and sedimentary rock that you will use to answer lab questions. Examine each sample in detail by zooming in and exploring its composition and texture.

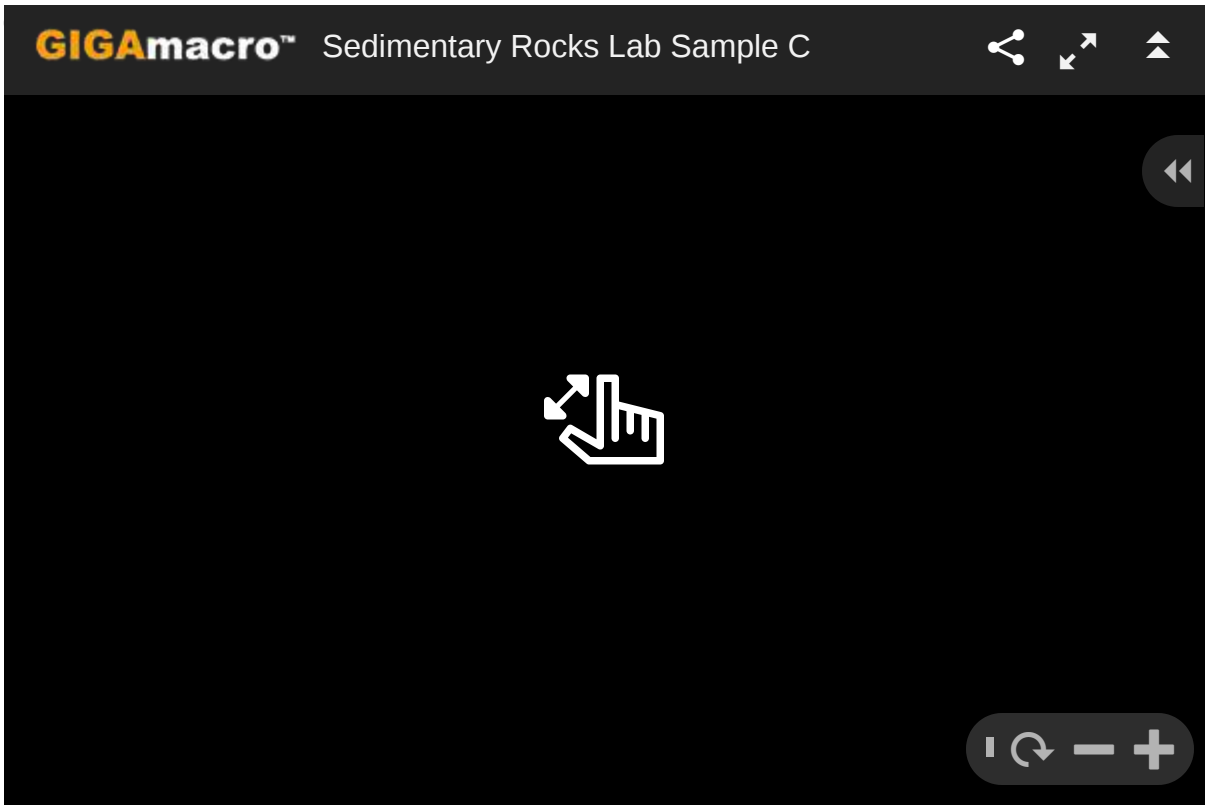
Sample A:



Sample B:

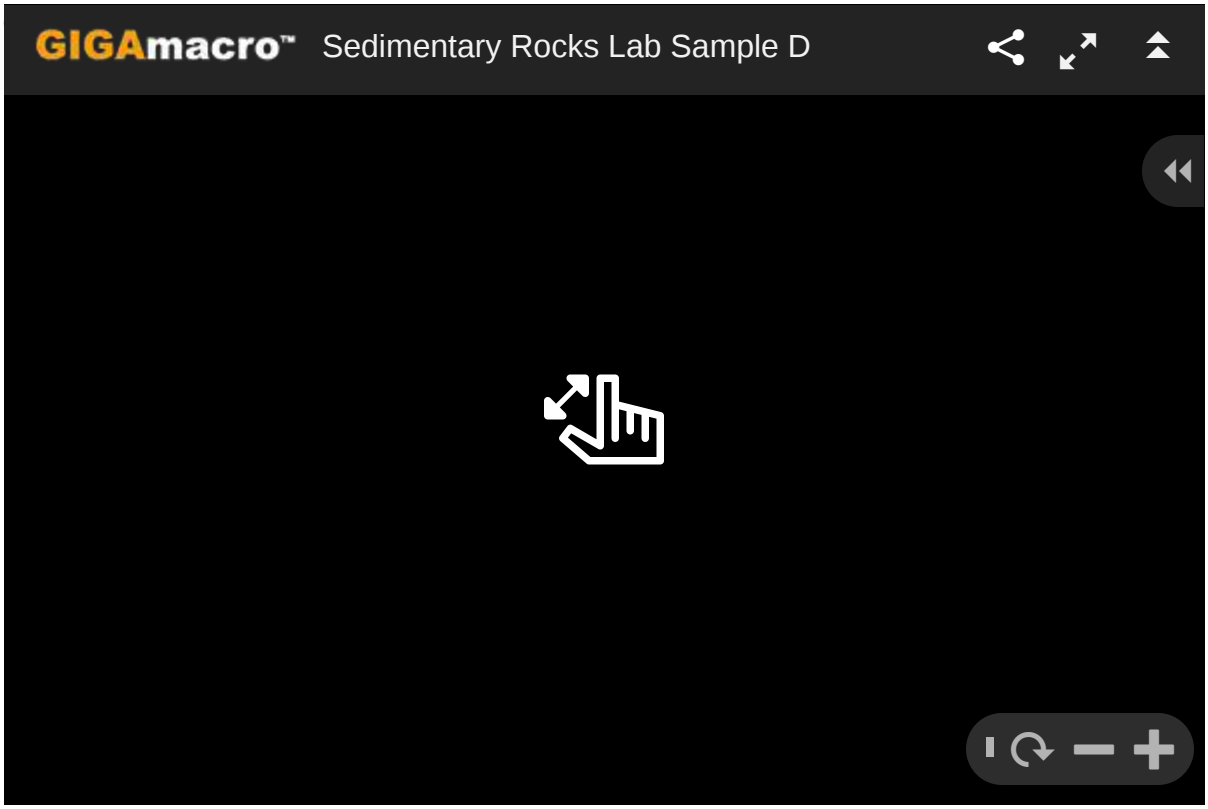


Sample C:

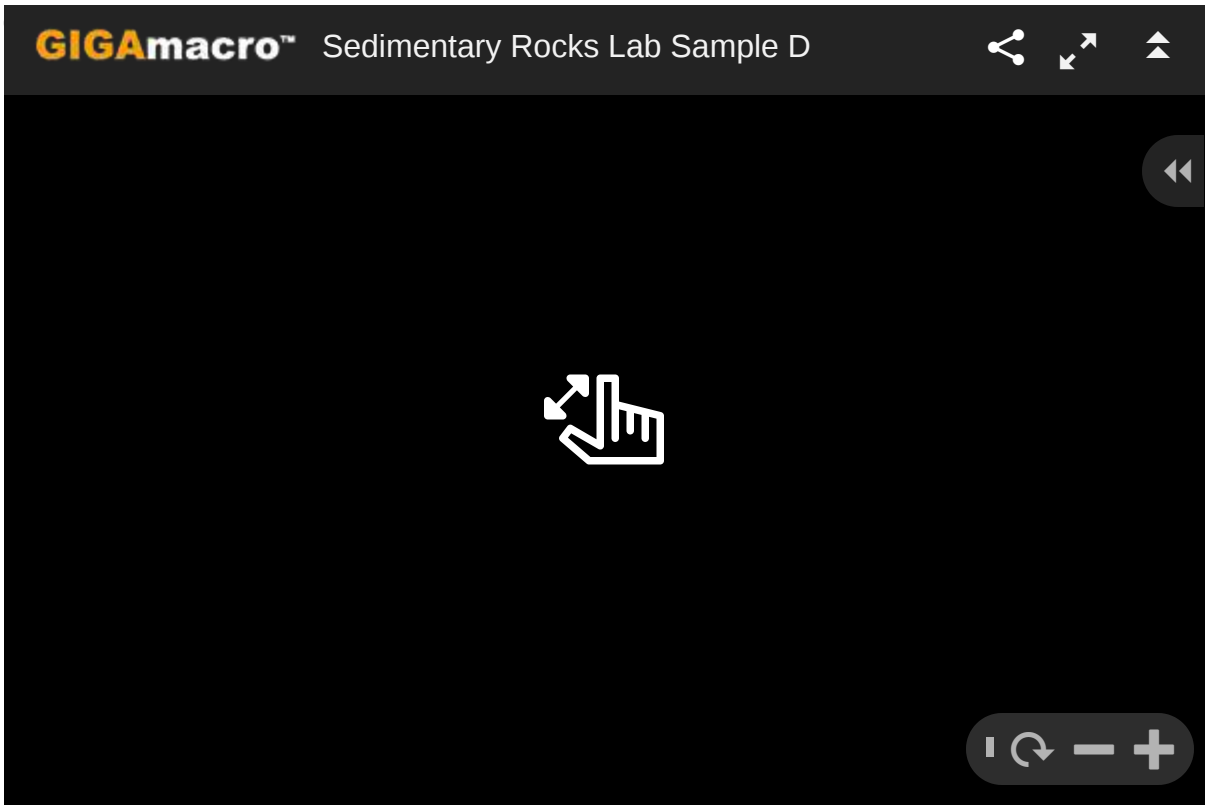


Sample D:

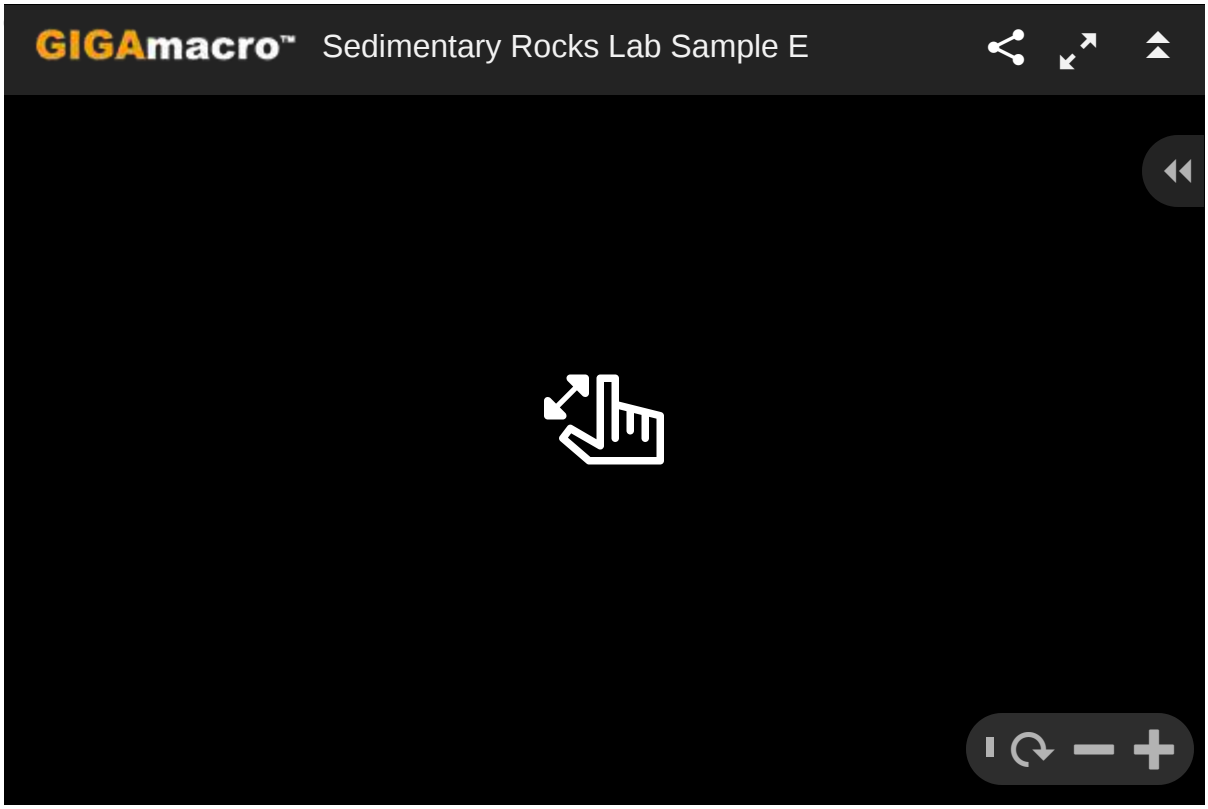
(Uncut original)



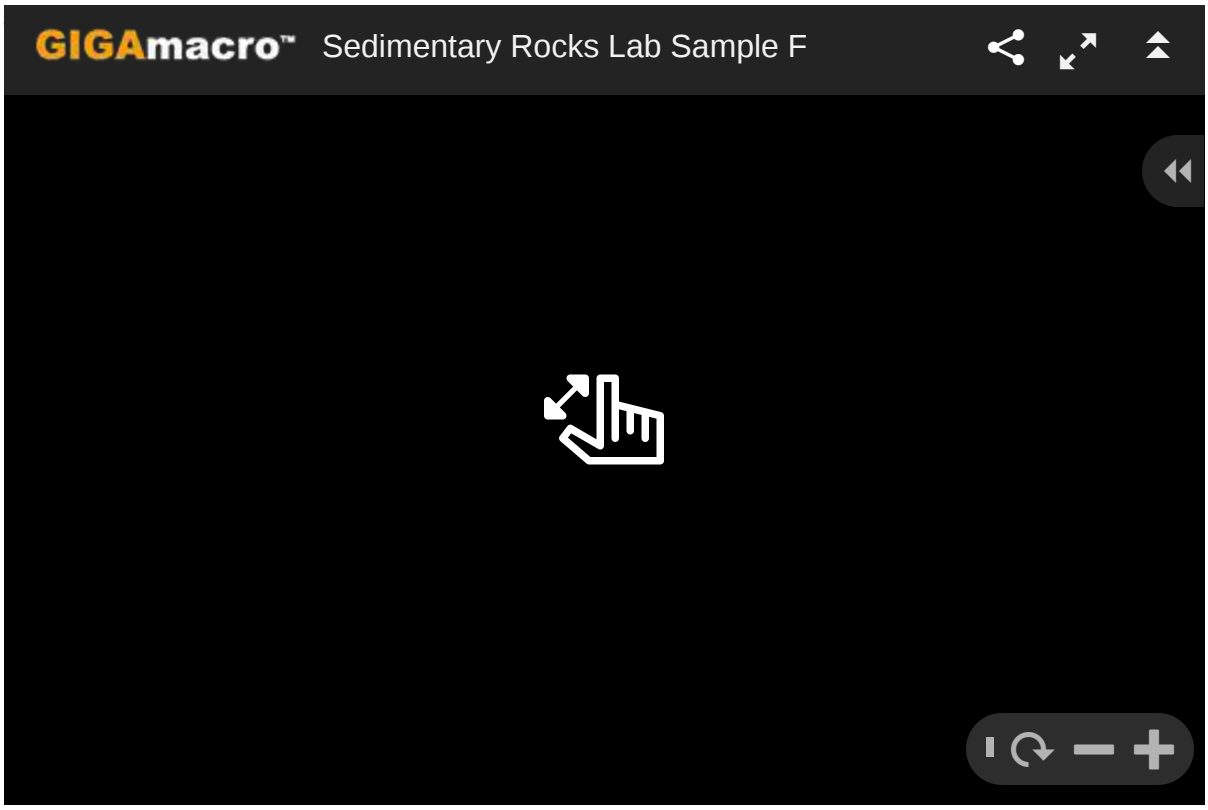
(Flipped and cut into 4 pieces)



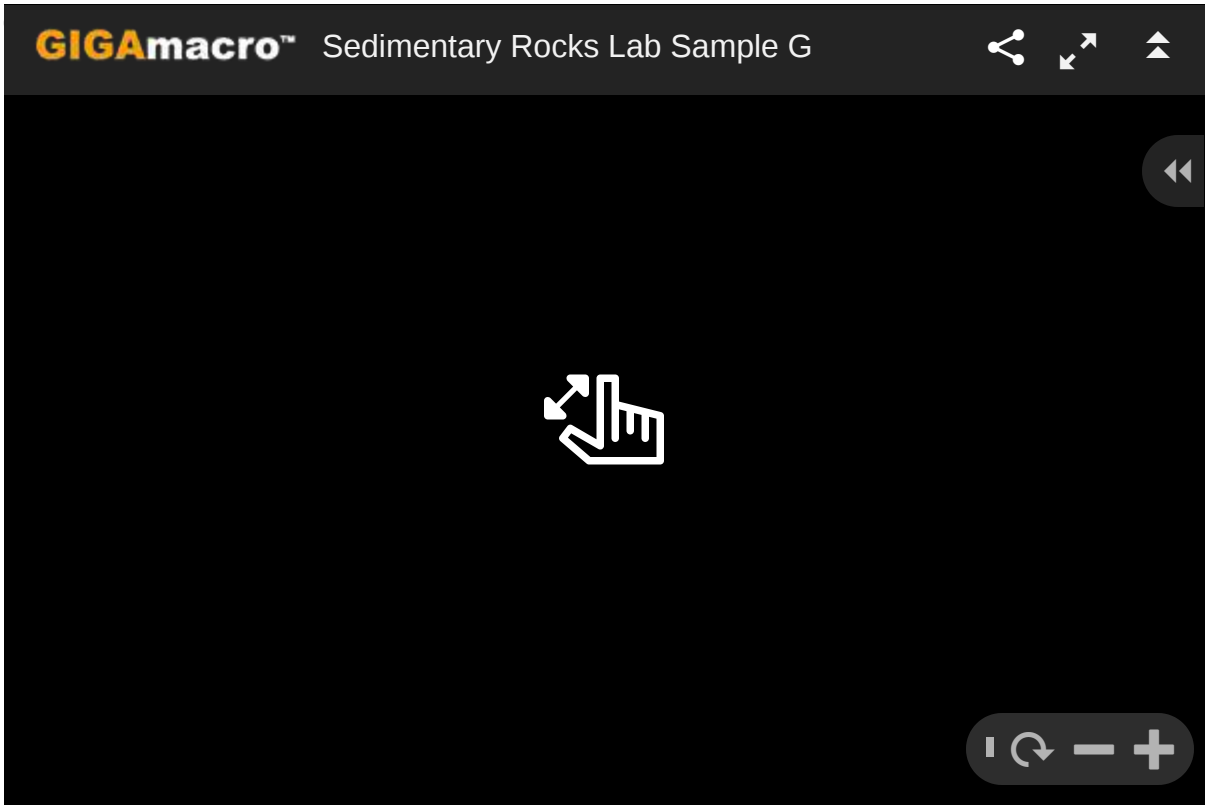
Sample E:



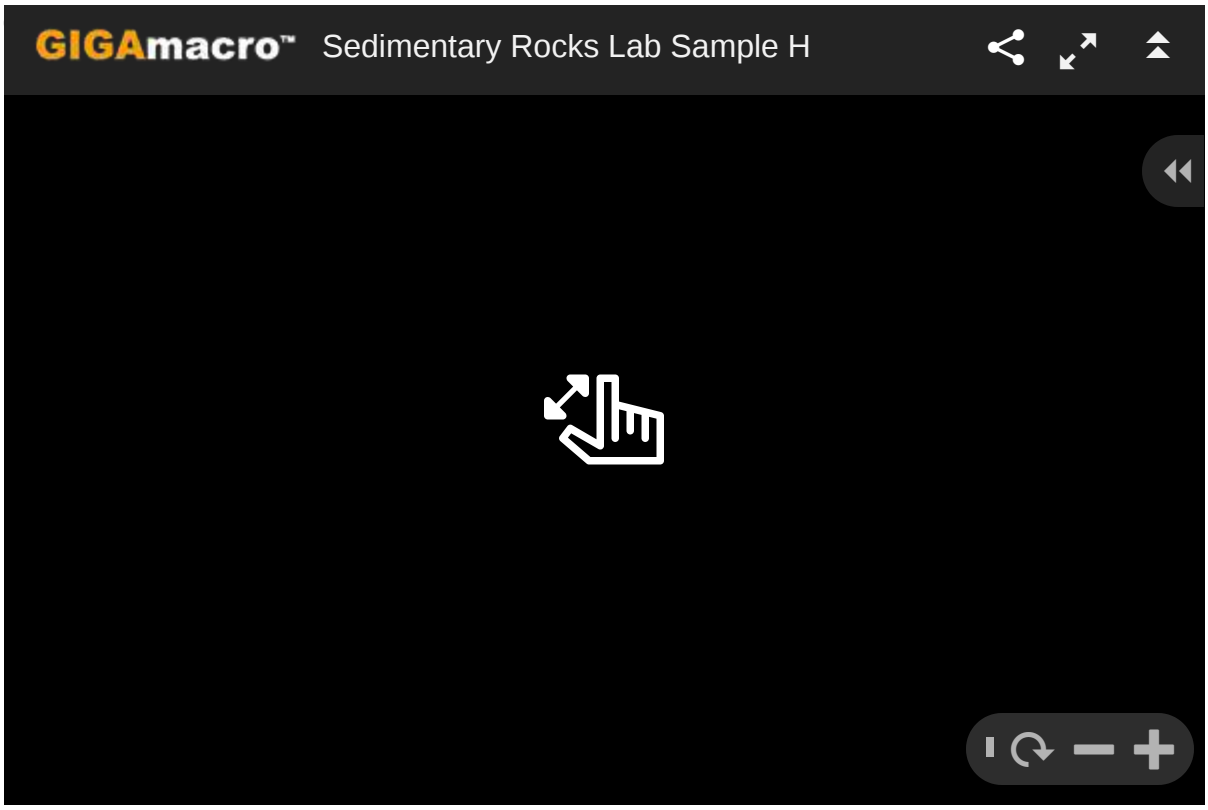
Sample F:



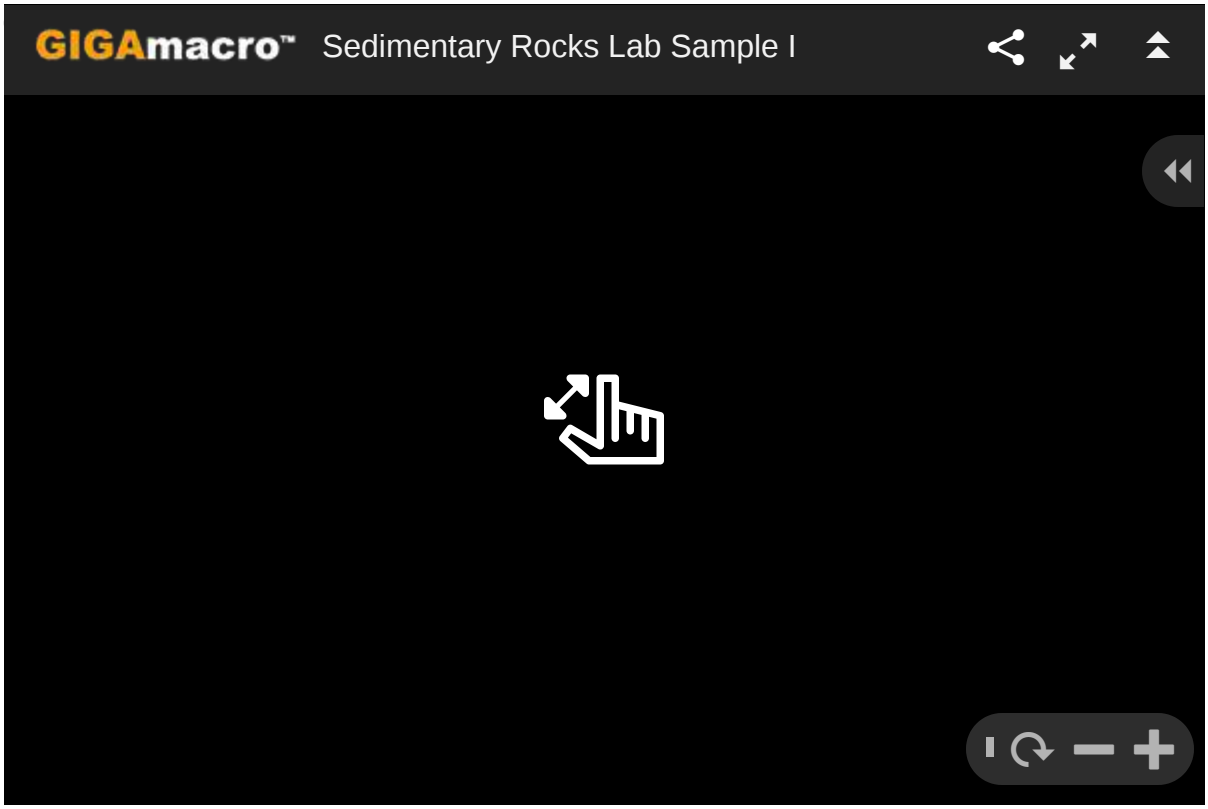
Sample G:



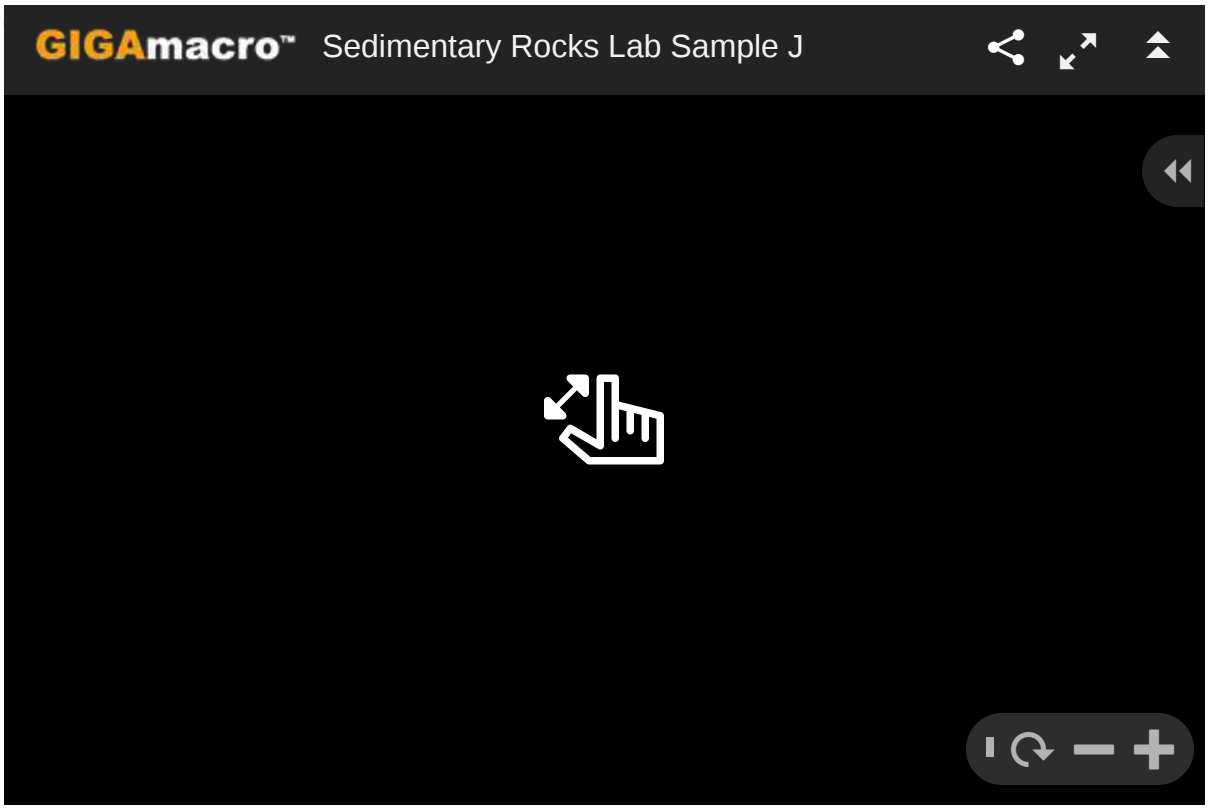
Sample H:



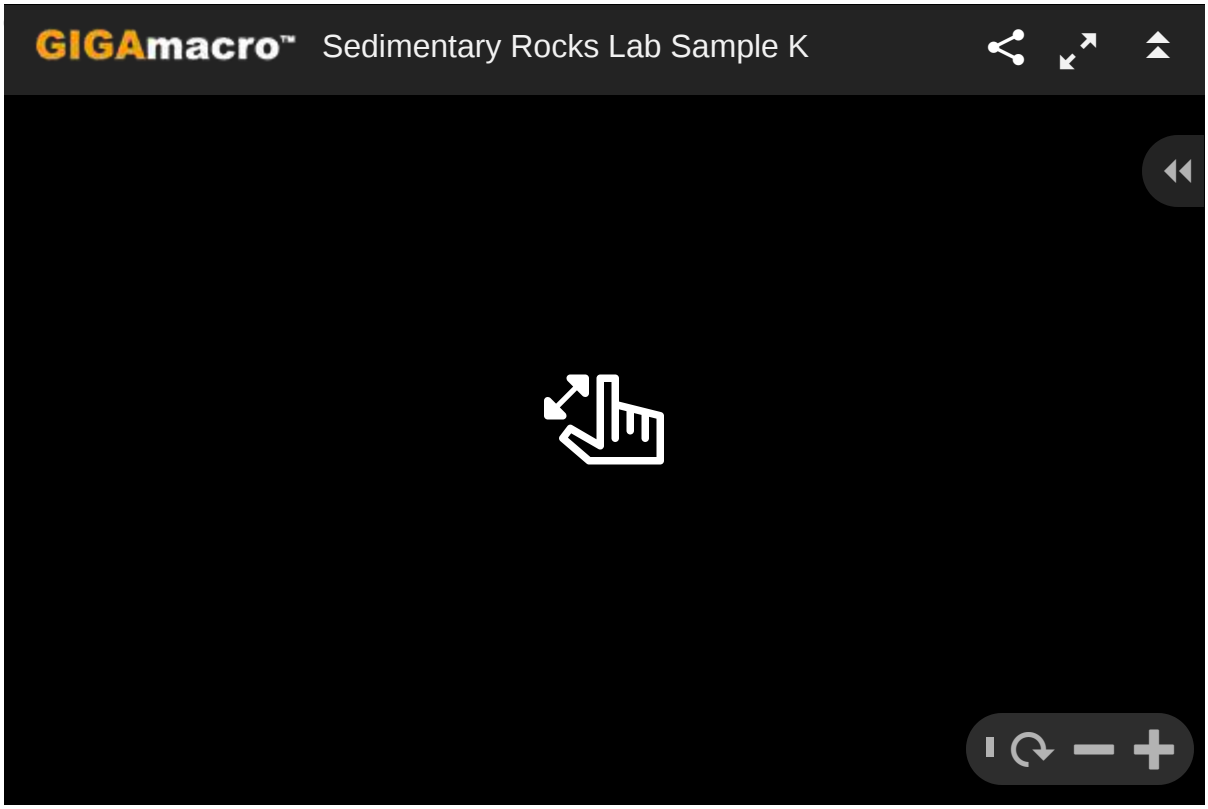
Sample I:



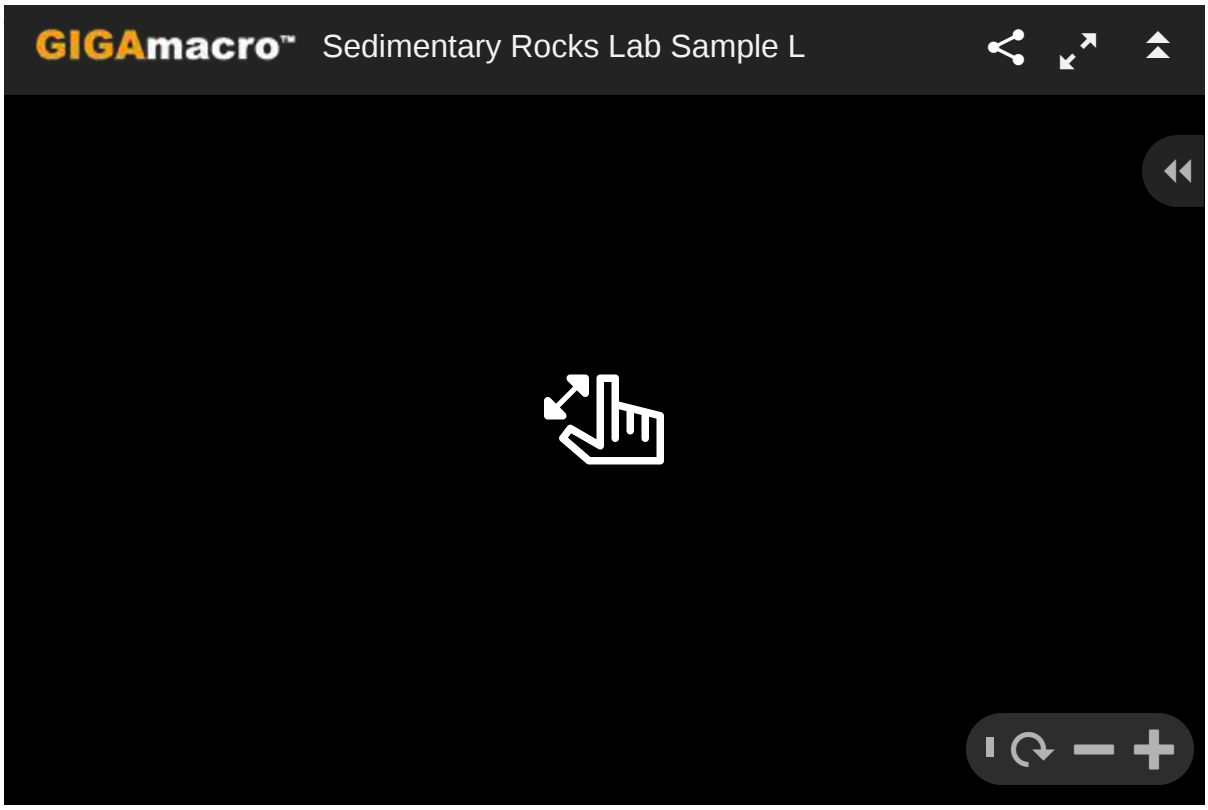
Sample J:



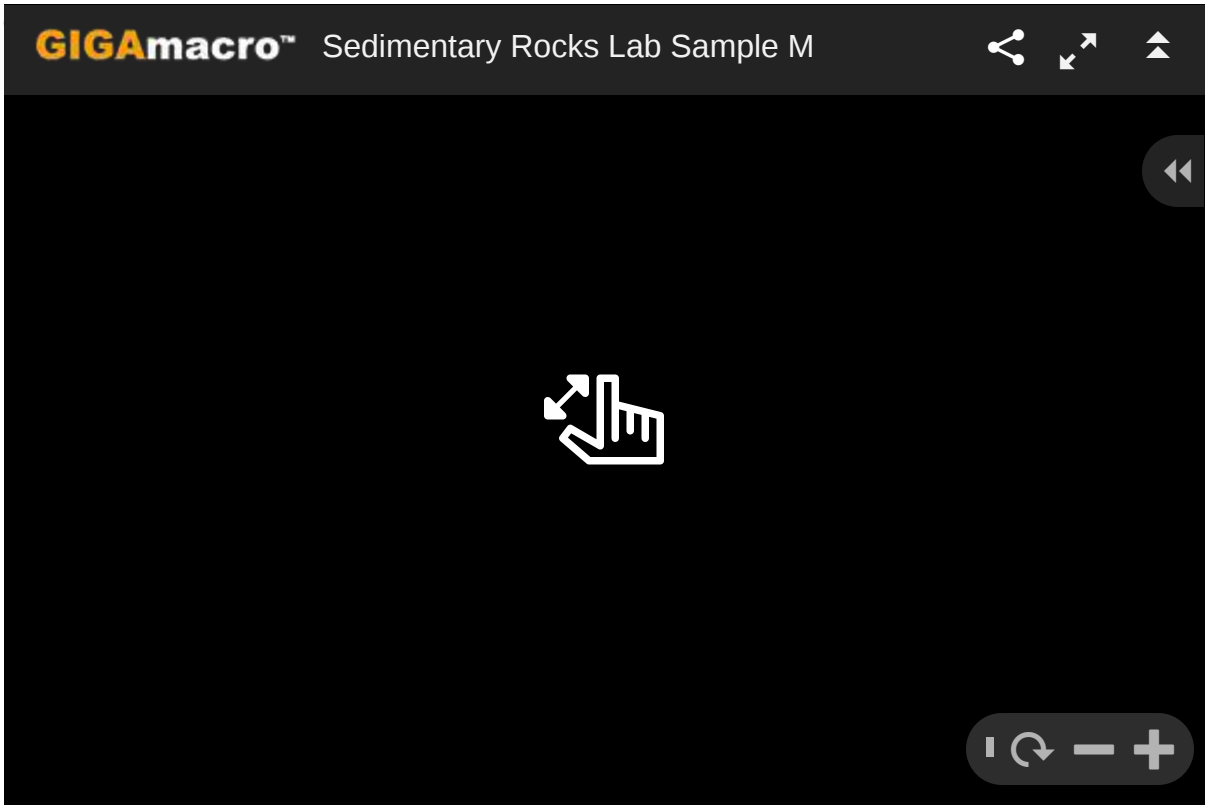
Sample K:



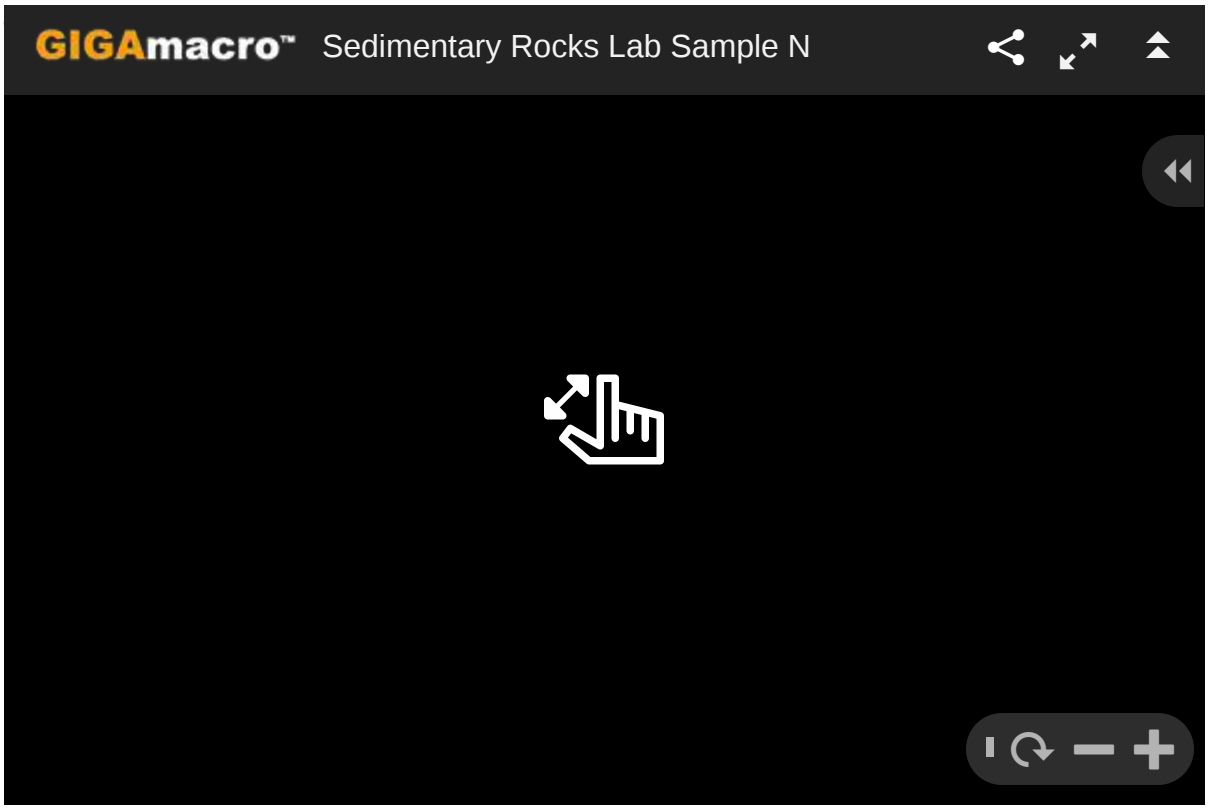
Sample L:



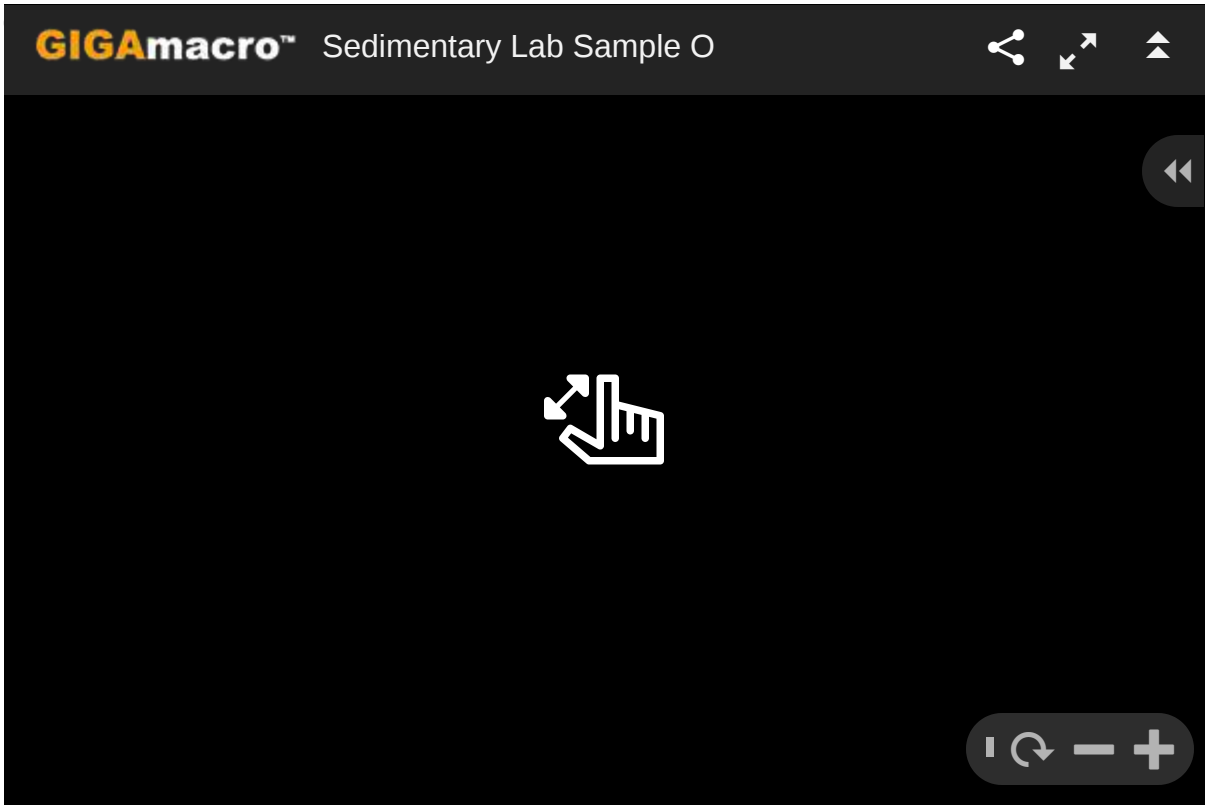
Sample M:



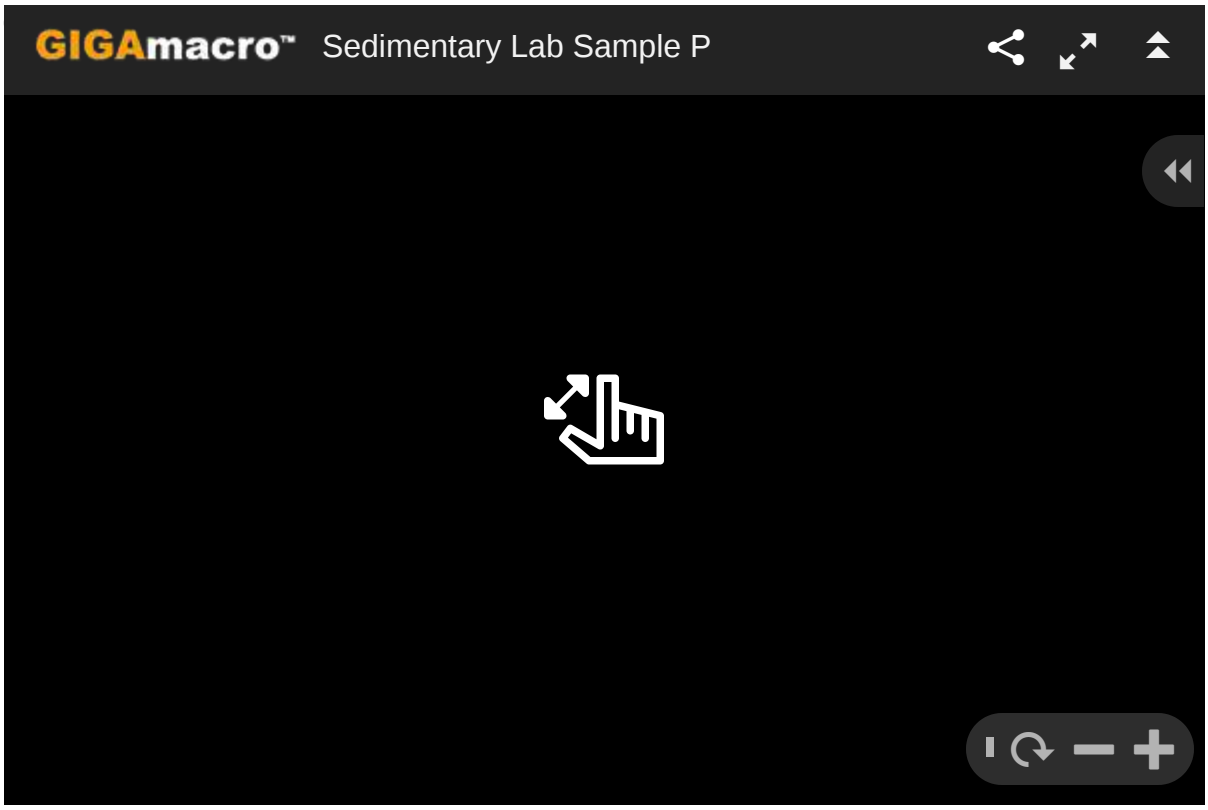
Sample N:



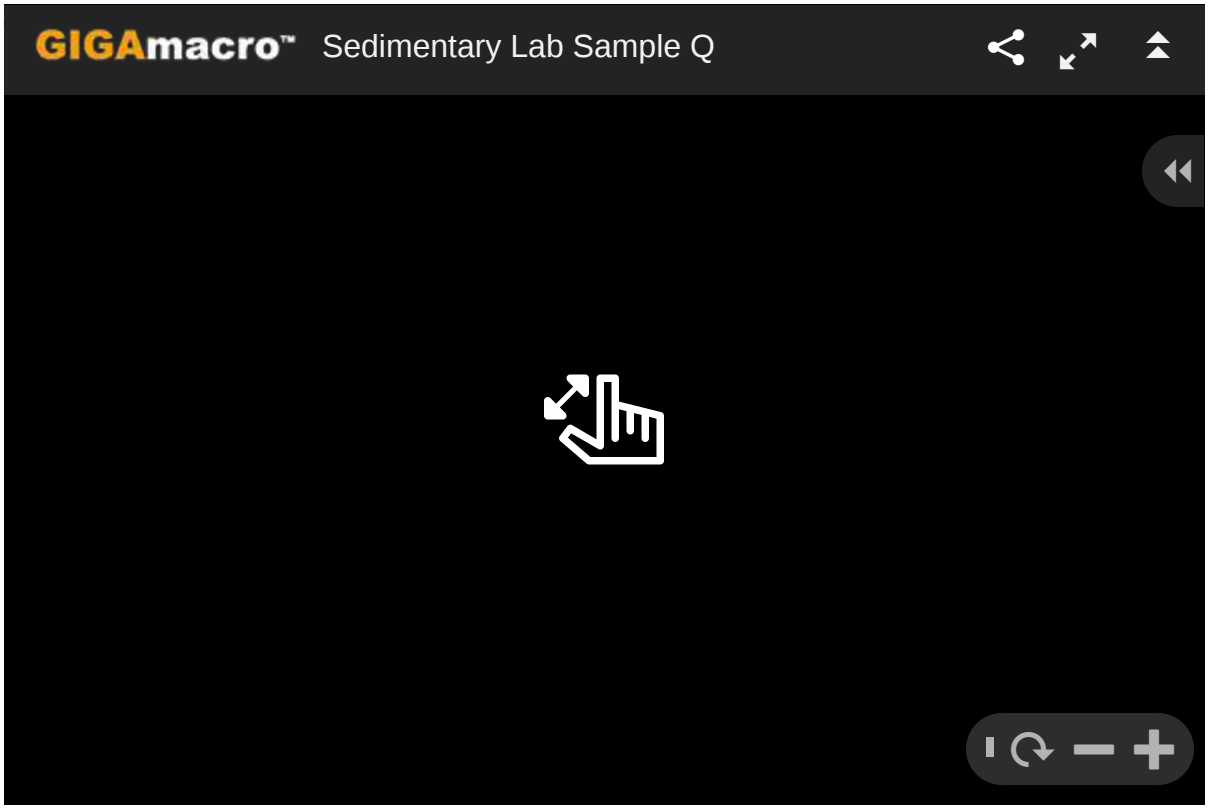
Sample O:



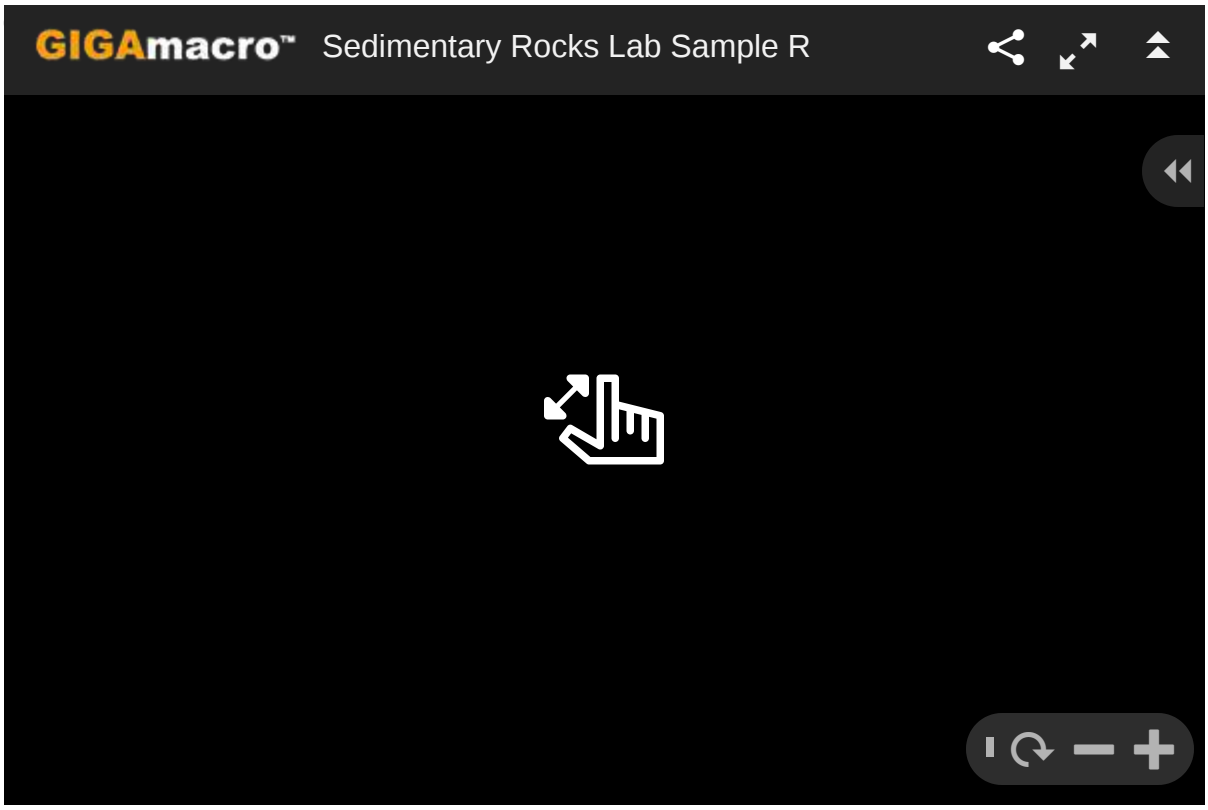
Sample P:



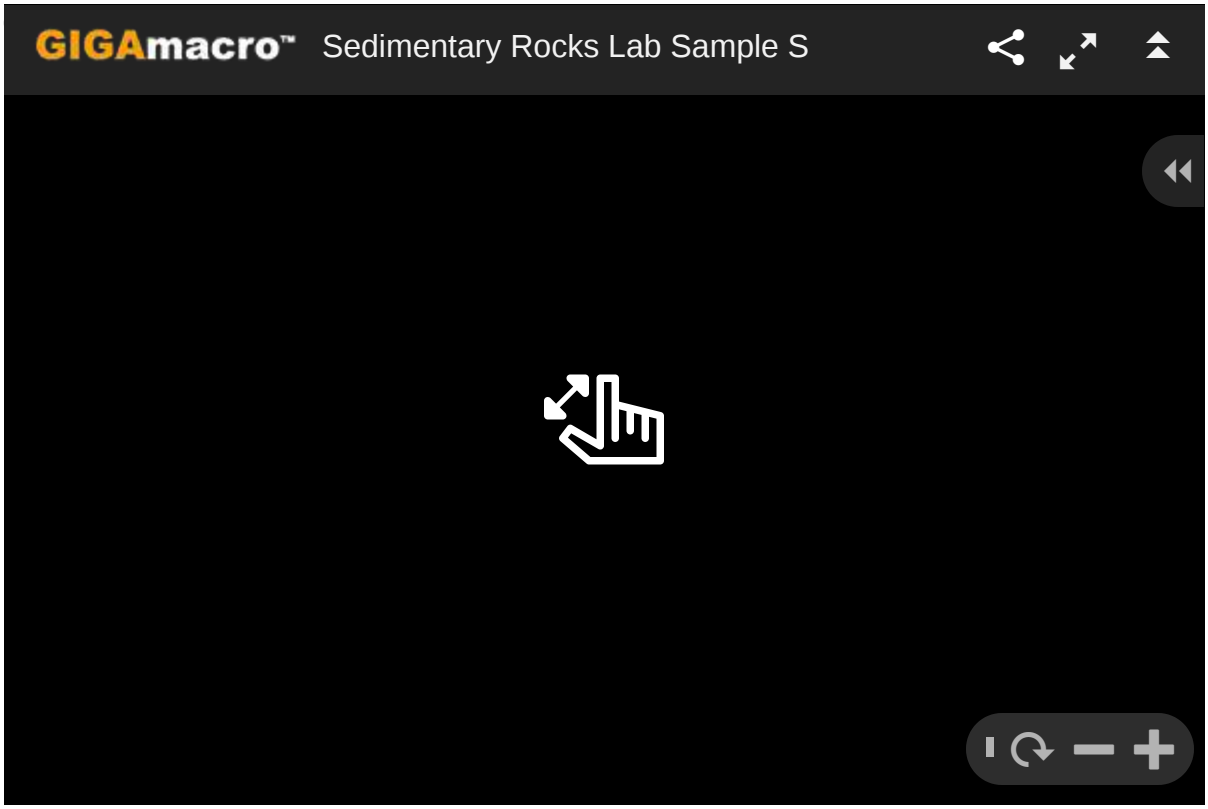
Sample Q:



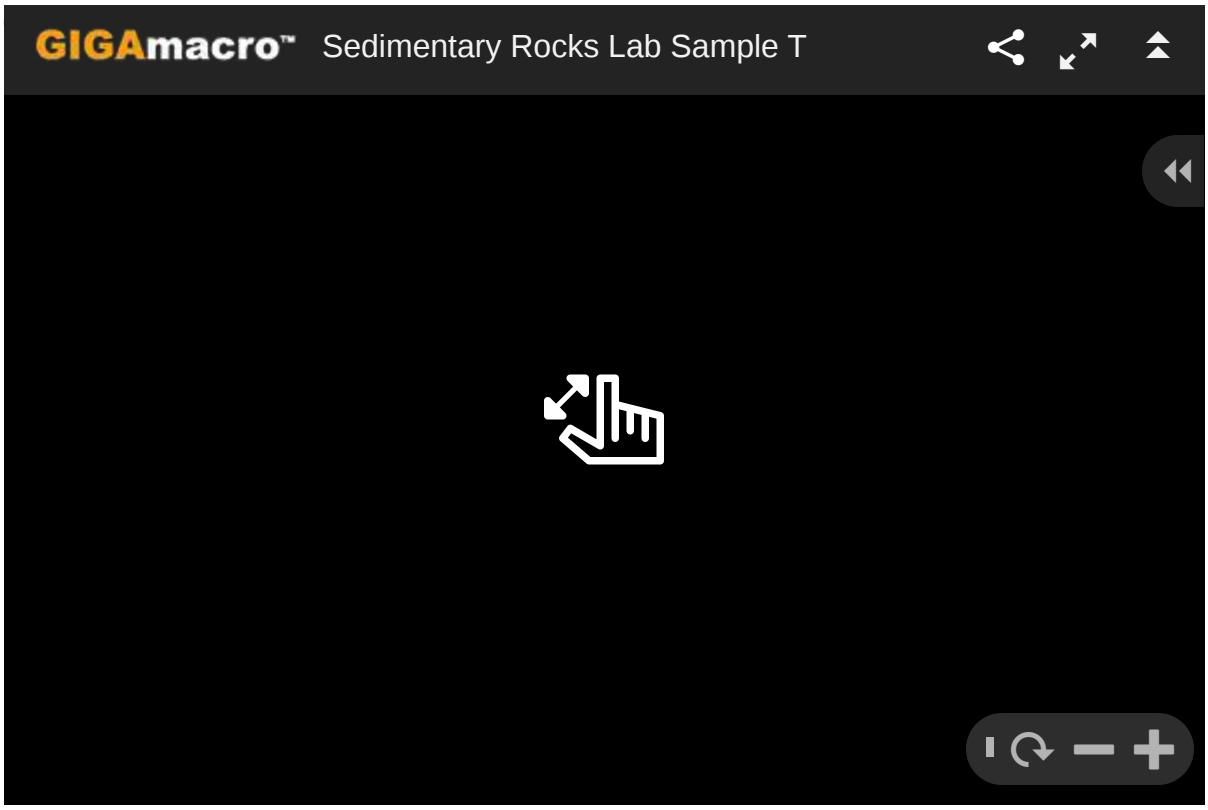
Sample R:



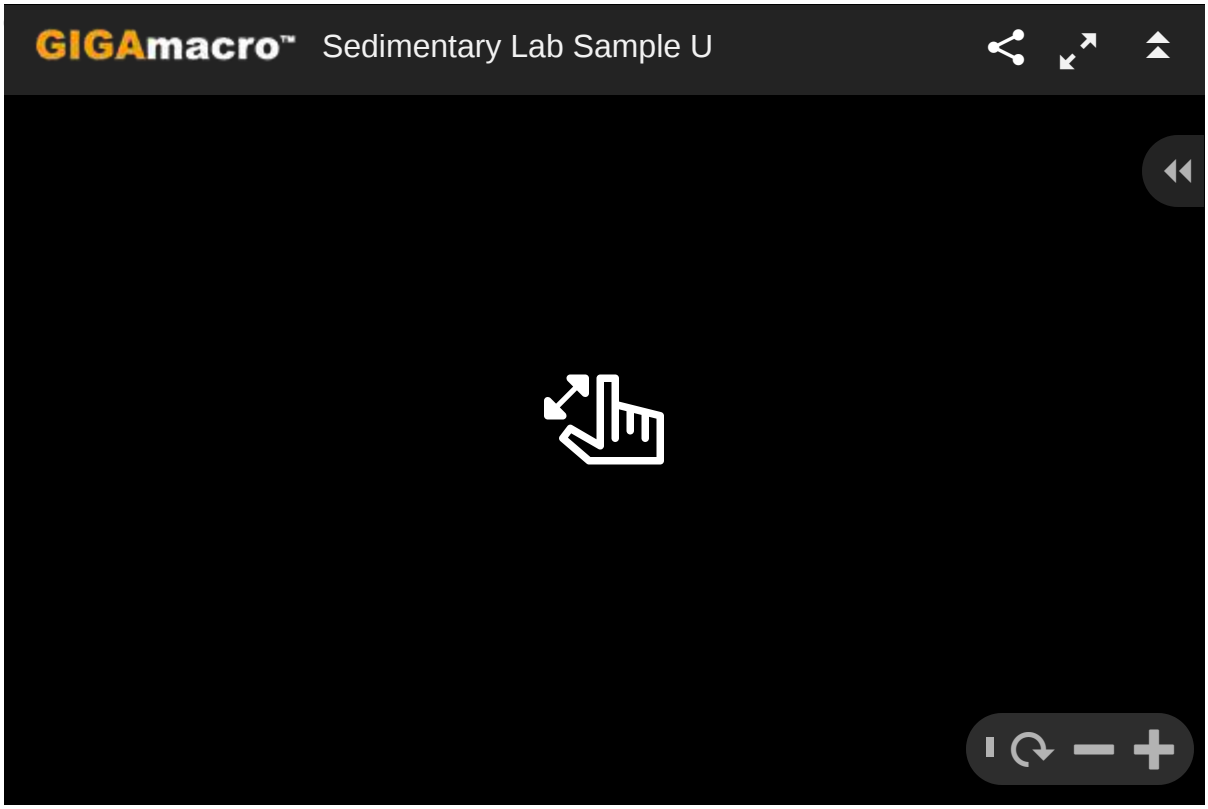
Sample S:



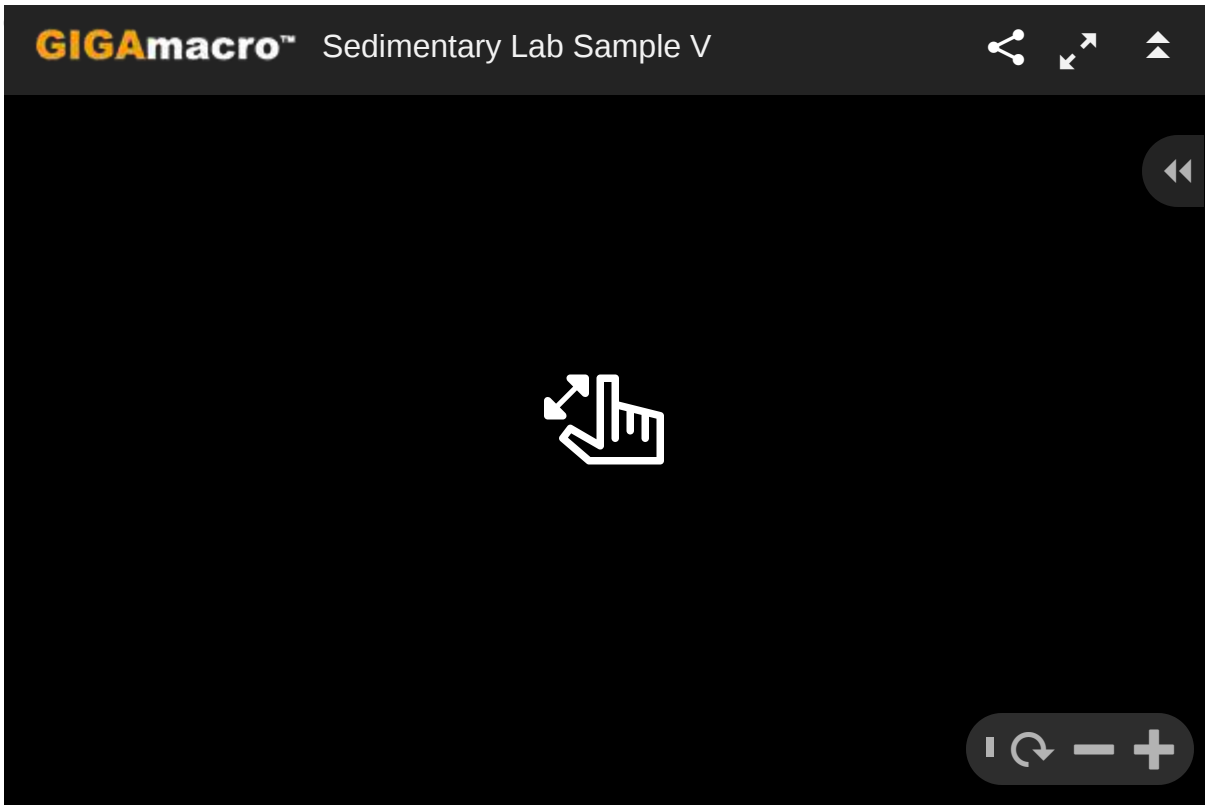
Sample T:



Sample U:



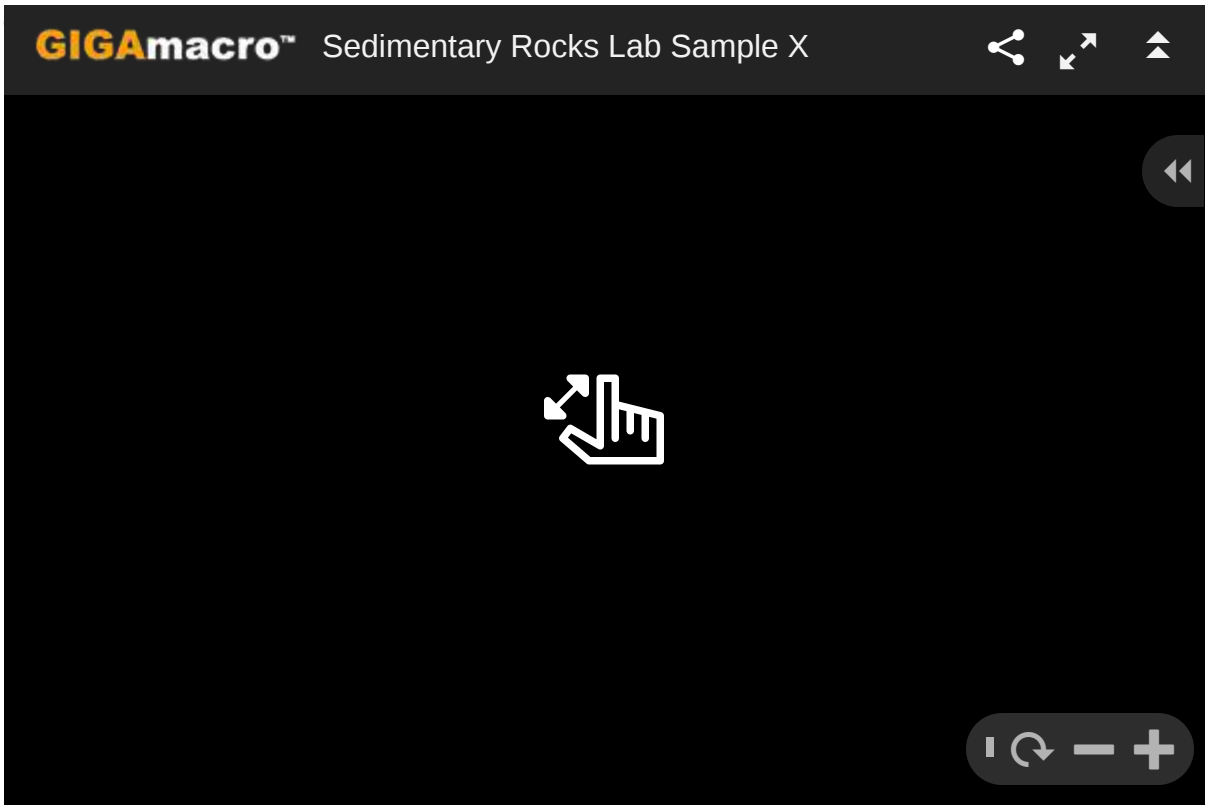
Sample V:



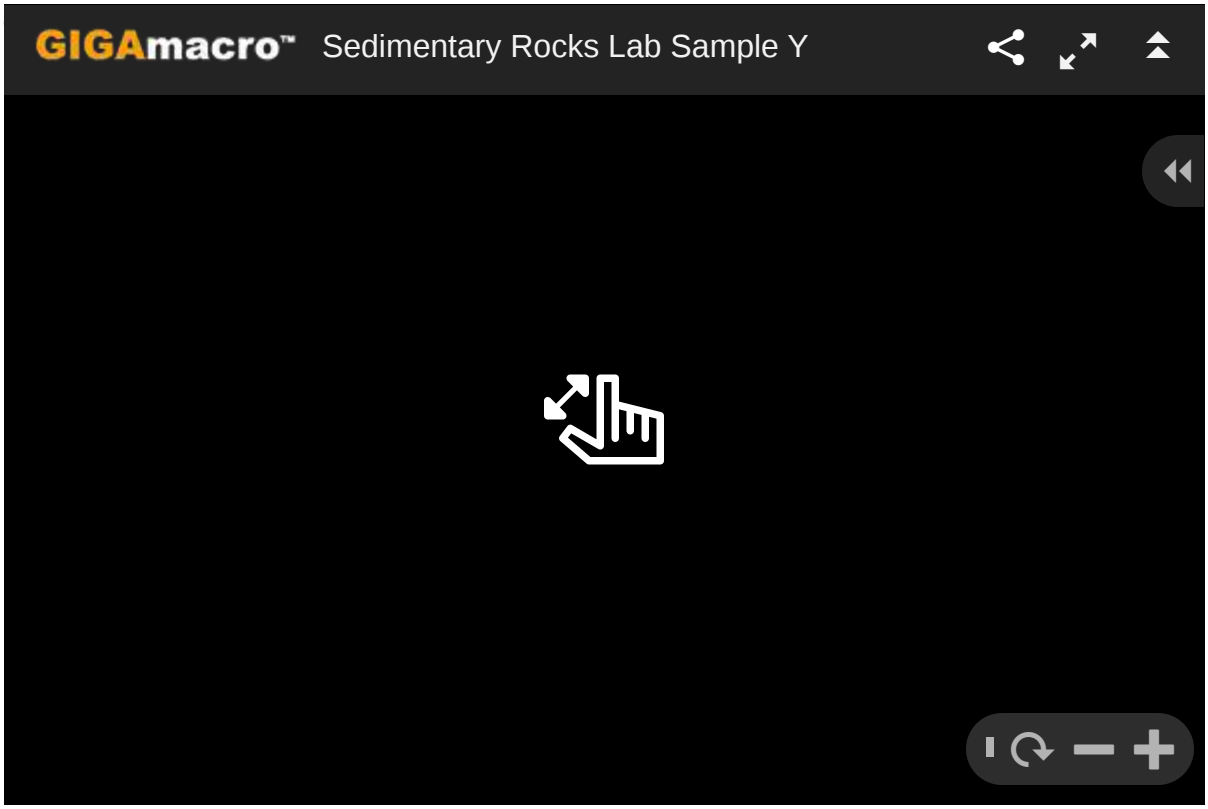
Sample W:



Sample X:



Sample Y:



Sample Z:



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standards of the LibreTexts platform.

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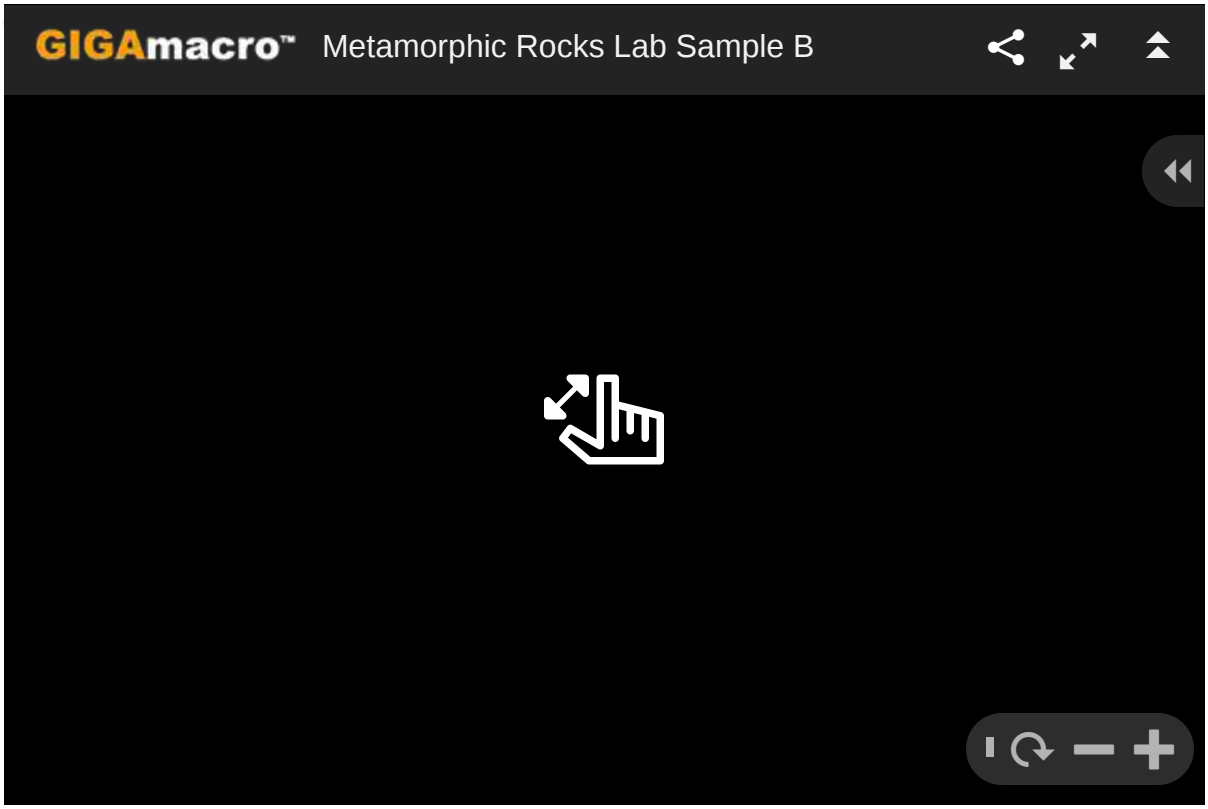
63.6: Metamorphic rocks

Here are 20 samples (or sample sets) of metamorphic rock that you will use to answer lab questions. Examine each sample in detail by zooming in and exploring its composition and texture.

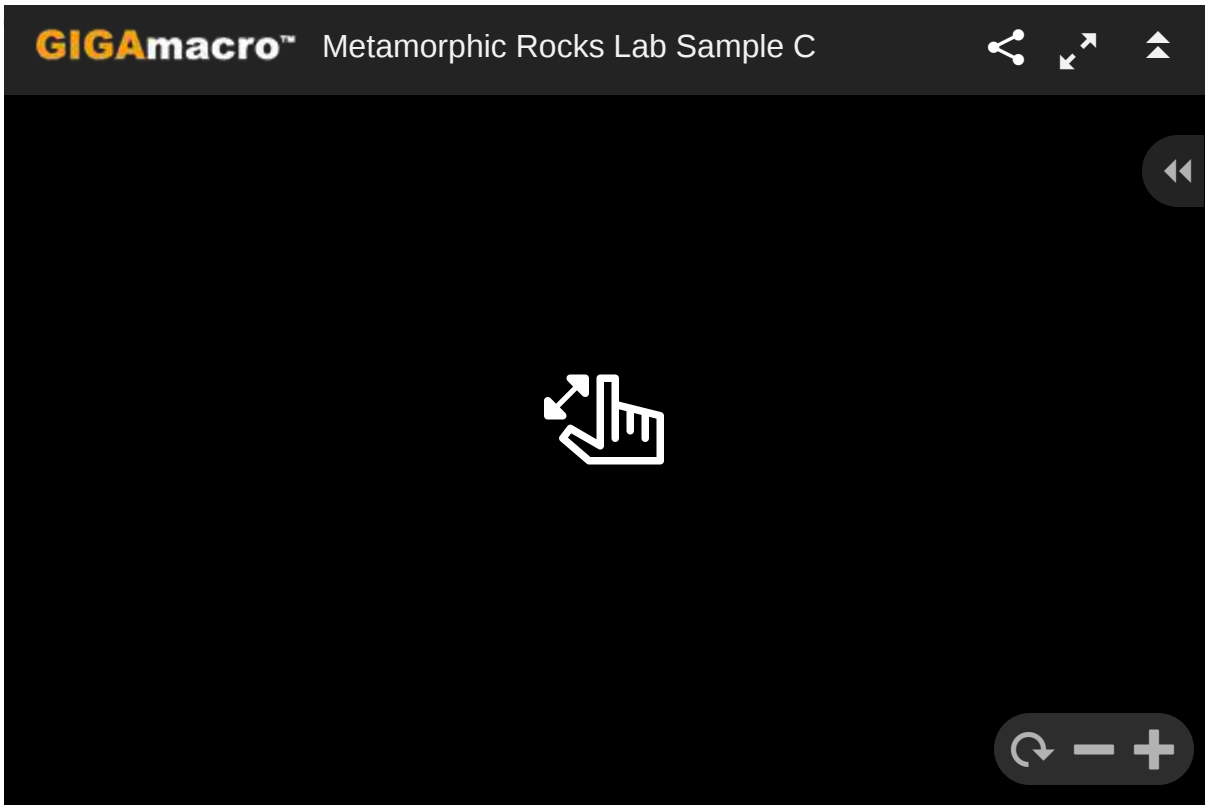
Sample A:



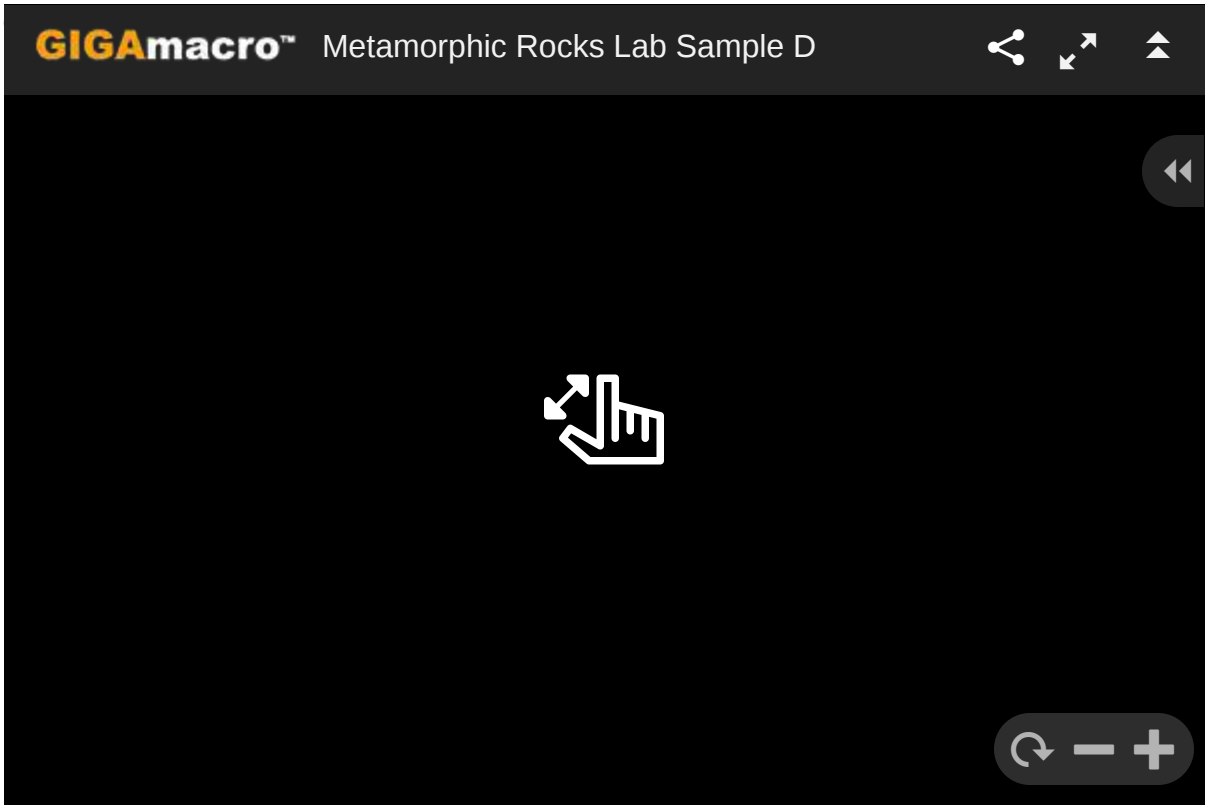
Sample B:



Sample C:



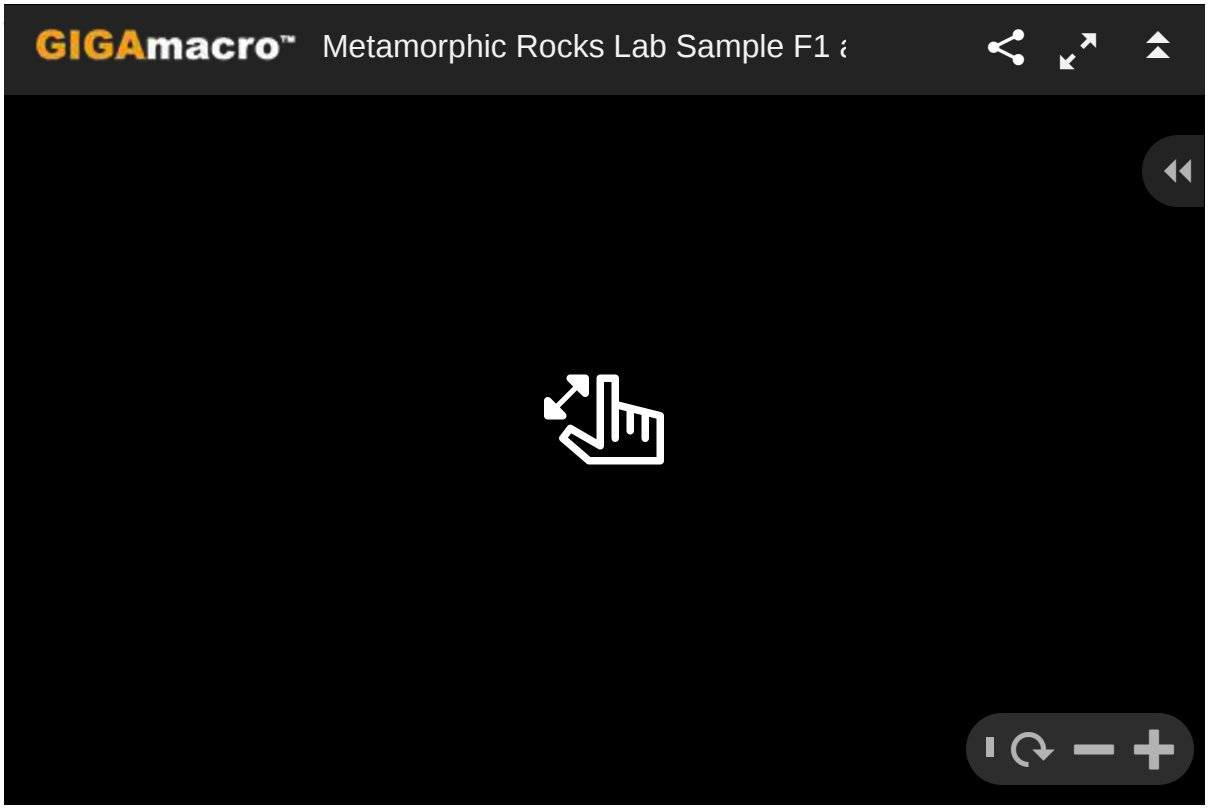
Sample D:



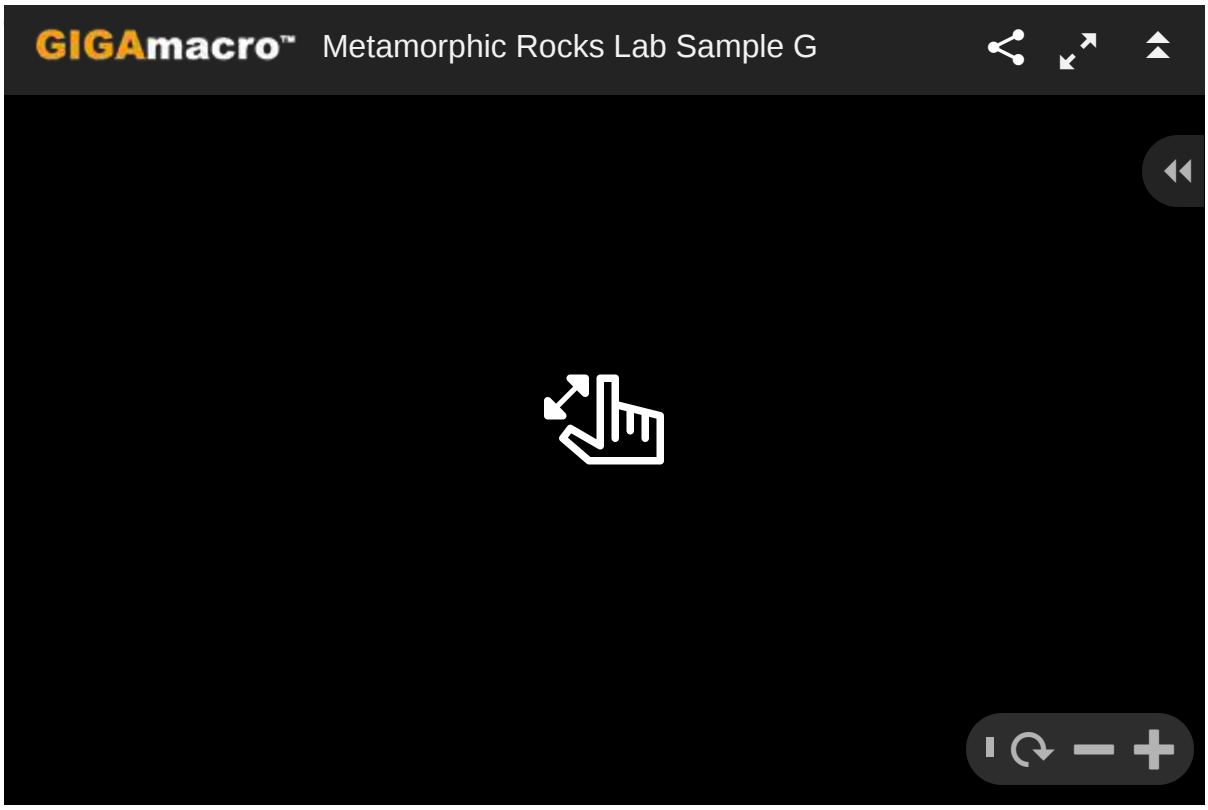
Sample E:



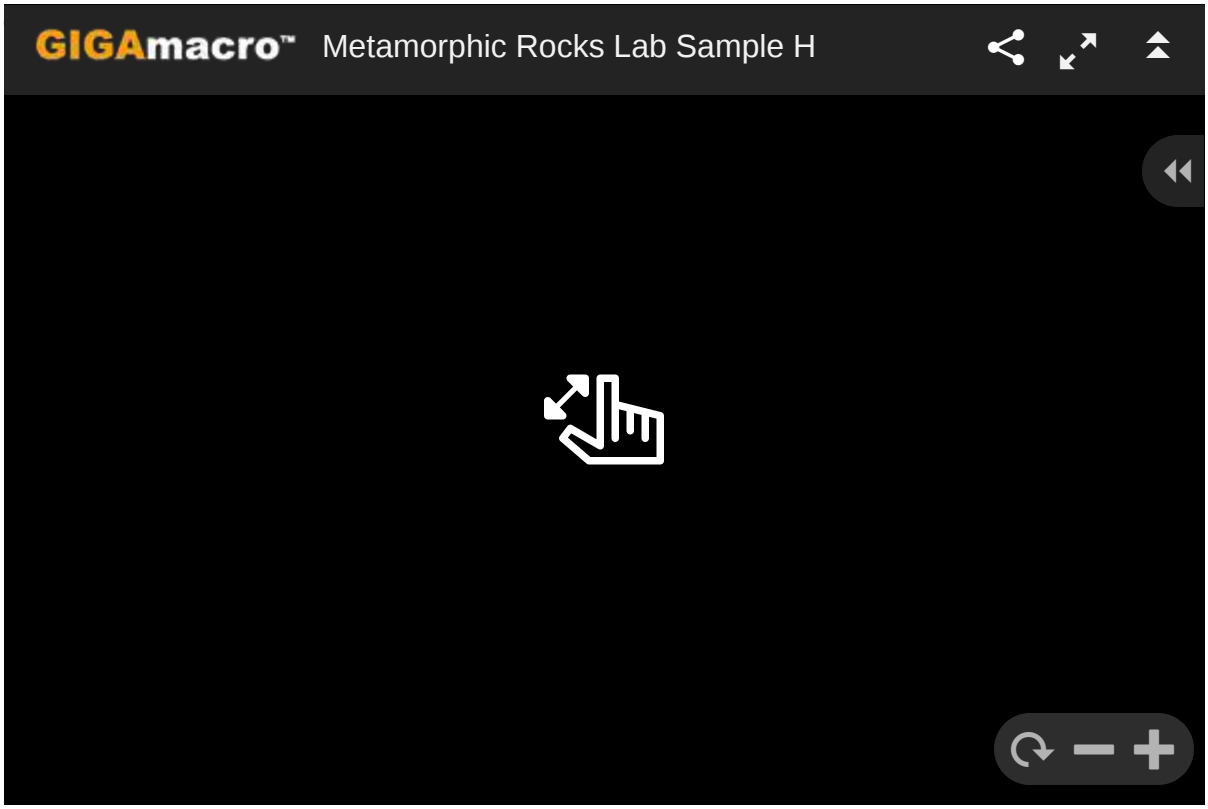
Samples F1 (top) and F2 (bottom):



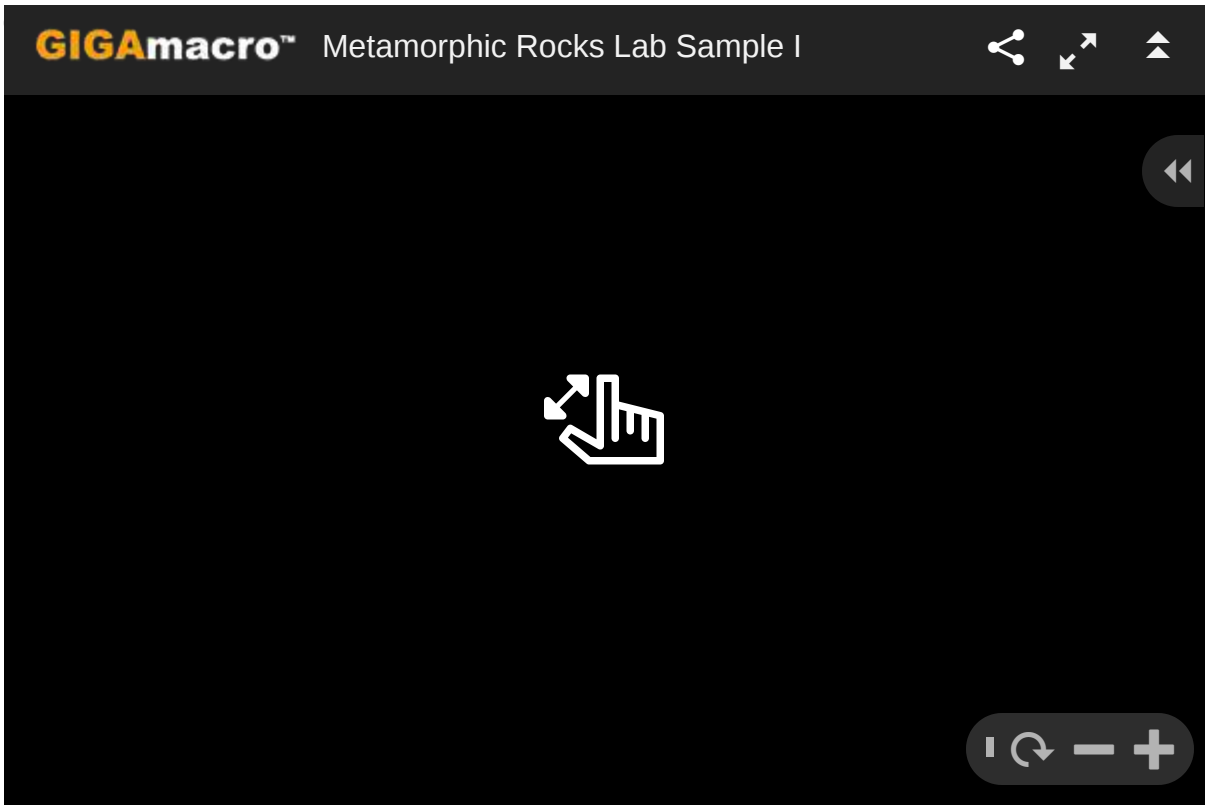
Sample G:



Sample H:



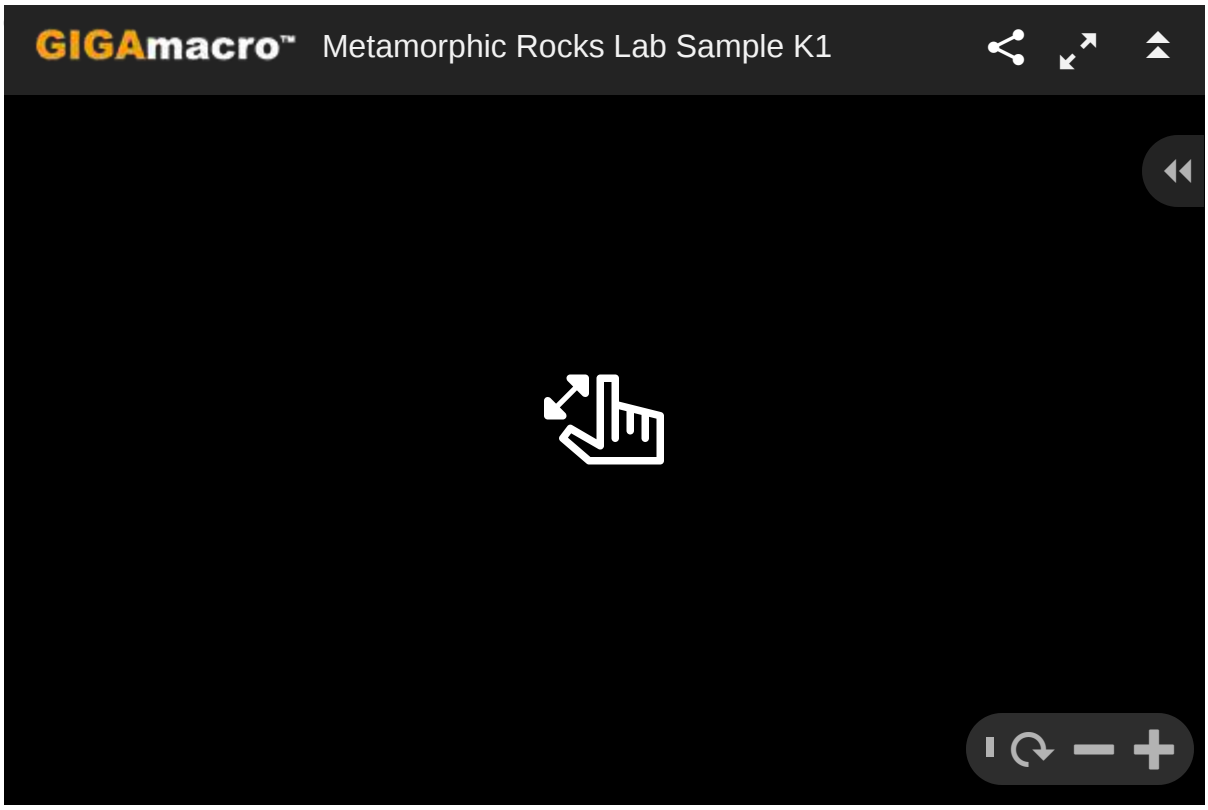
Sample I:



Sample J:



Sample K1, contact with gray wall rock:



Sample K2, side view (left); Sample K2, bottom view (right):



Sample L, opposite side views:



Sample M:



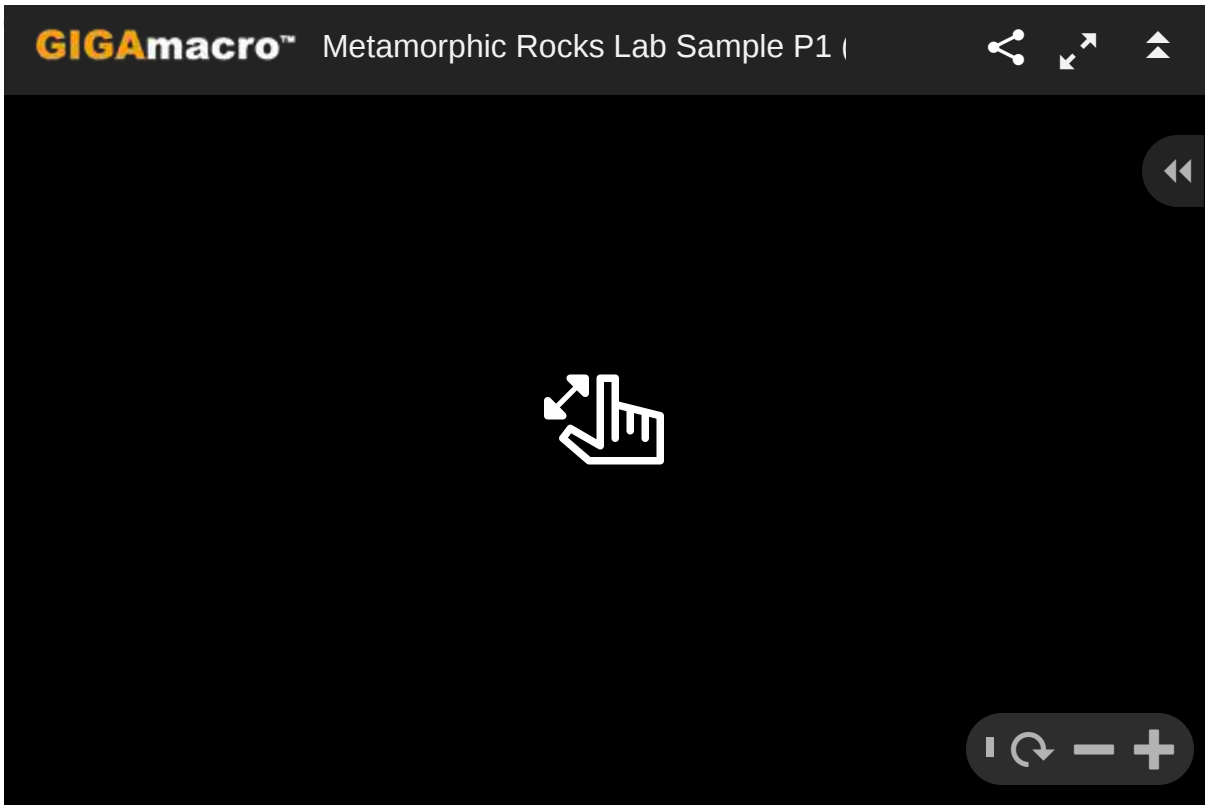
Sample N:



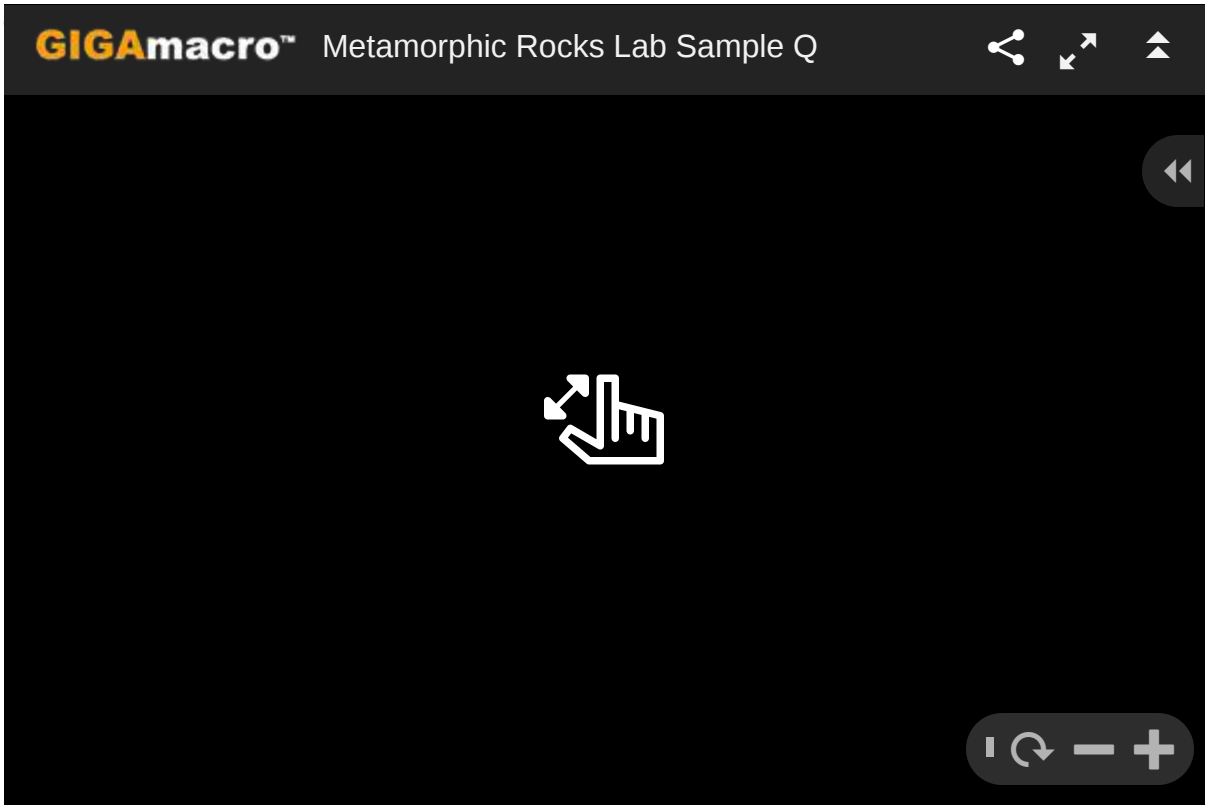
Sample O:



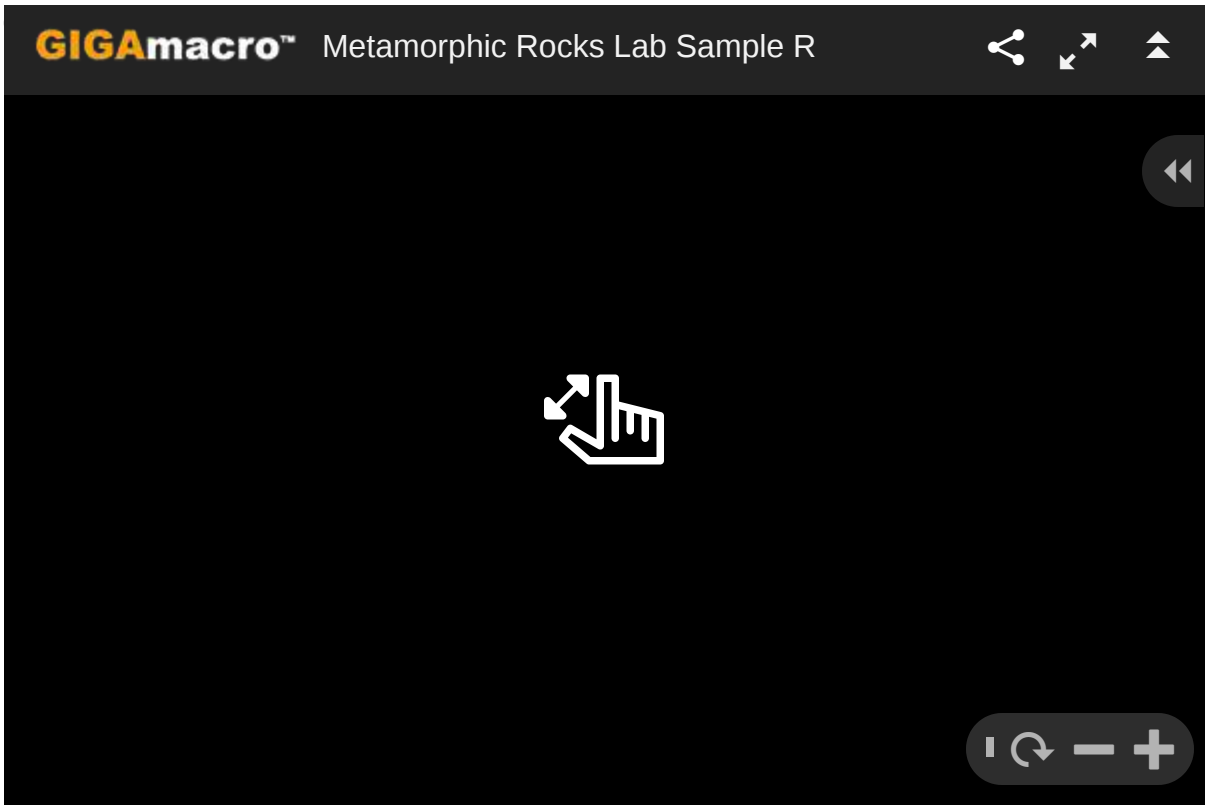
Samples P1 (left) and P2 (right):



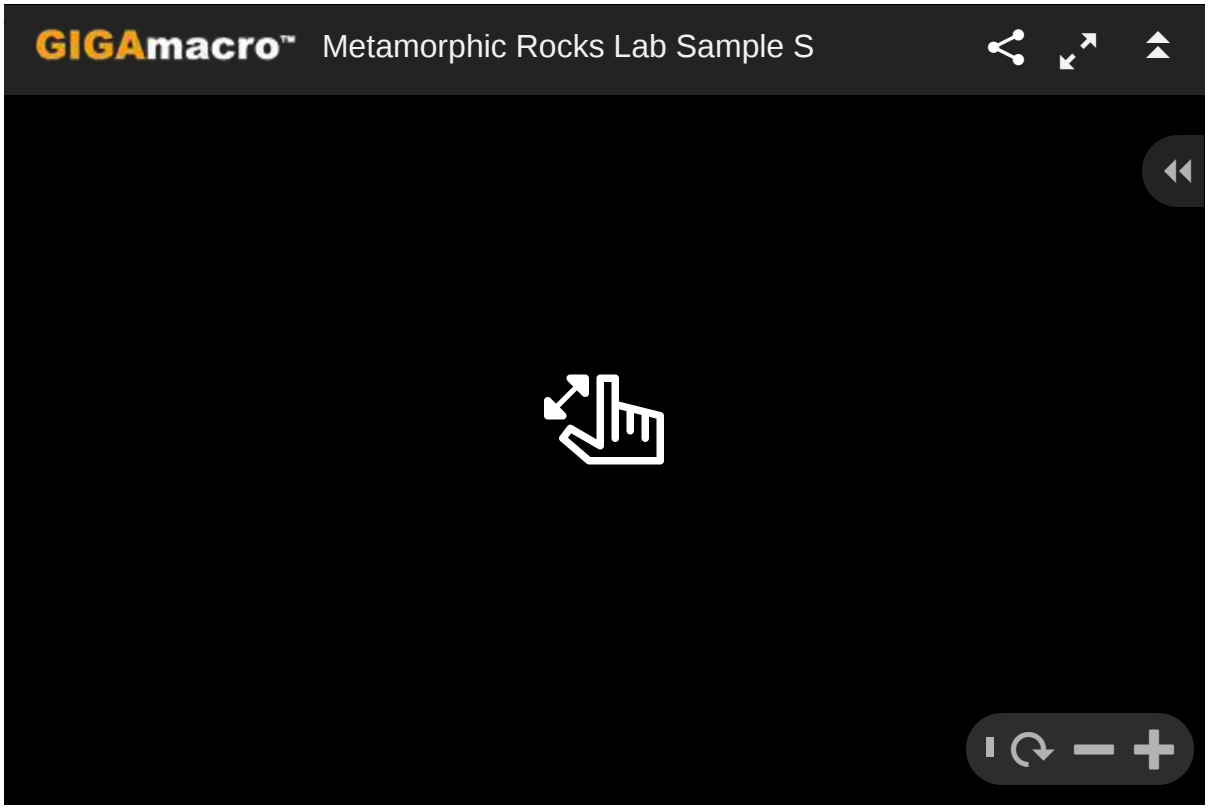
Sample Q:



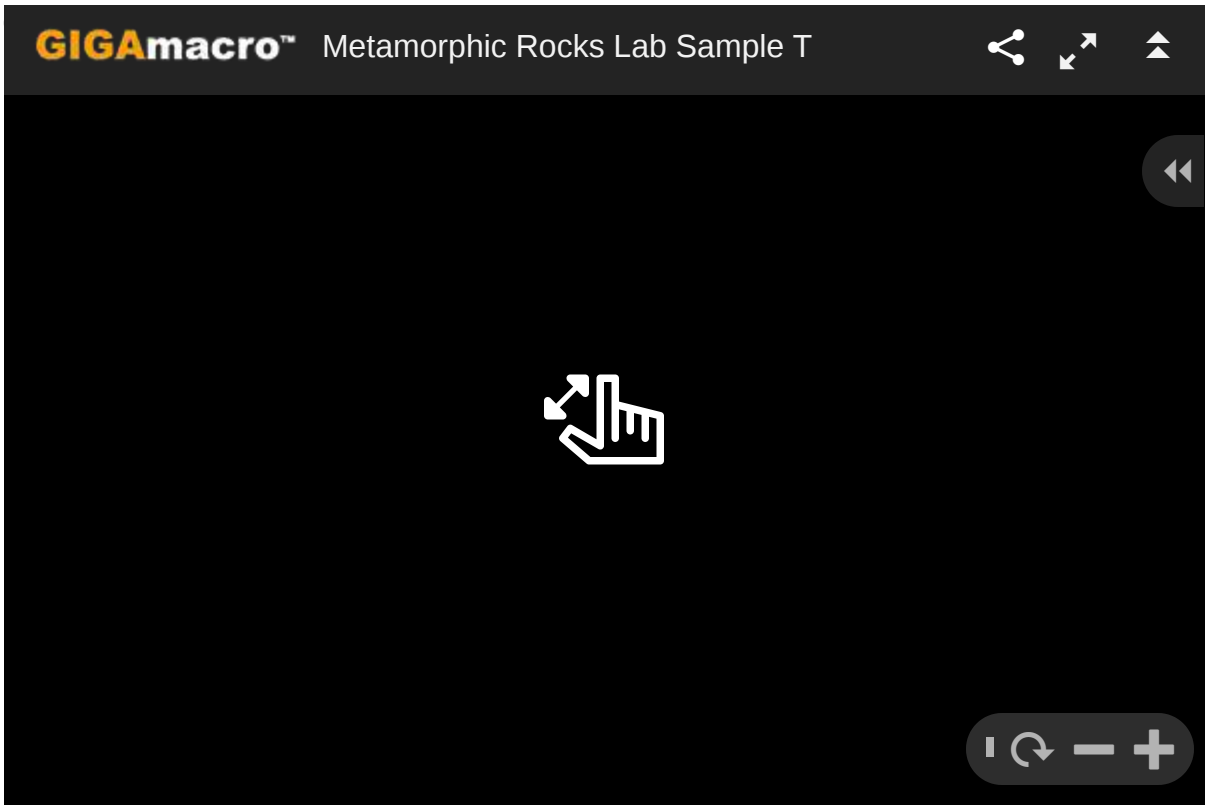
Sample R:



Sample S:



Sample T:



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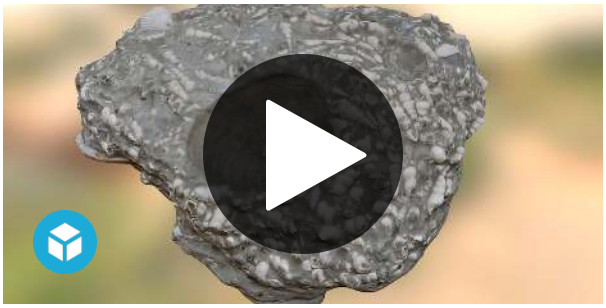
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63.7: 3D rock samples, set 1

Here are 10 outcrops and/or samples of rocks that you can attempt to identify for the sake of review and practice. Examine each sample in detail by spinning it around and exploring its composition and texture.

Sample 1:



Sample 2:



Sample 3:



Sample 4:



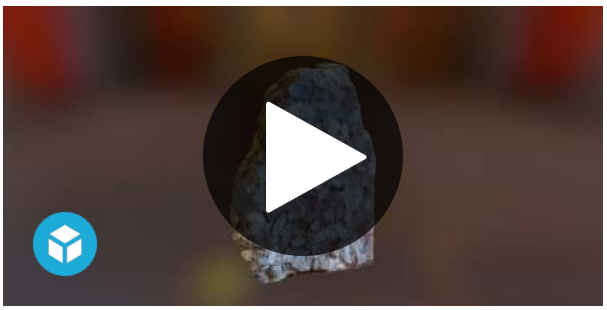
Sample 5:



Sample 6:



Sample 7:



Sample 8:



Sample 9:



https://geo.libretexts.org/@go/page/22484

Sample 10:



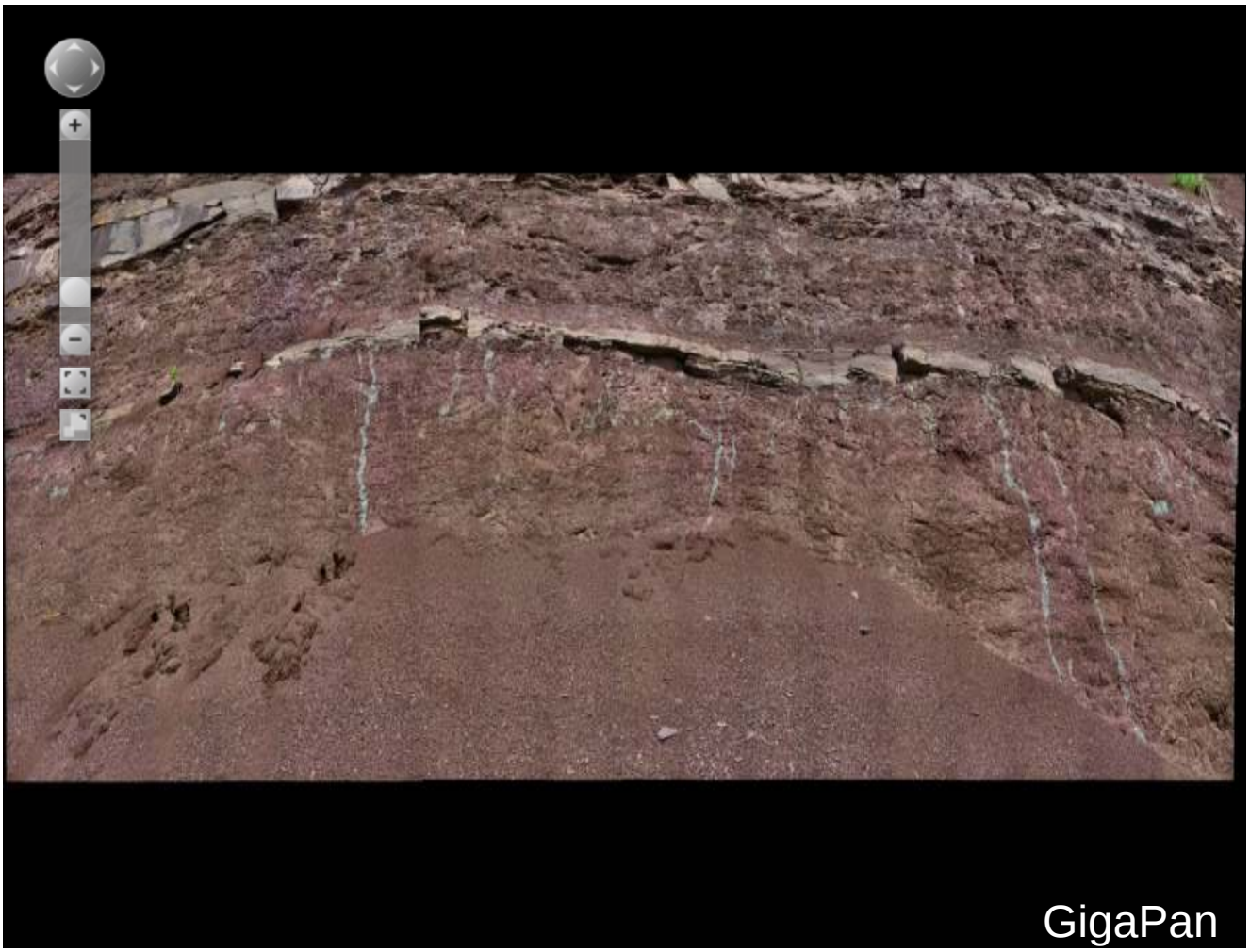
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63.8: Primary sedimentary structures

PART I: OUTCROPS

Here are 9 outcrops of rocks that show sedimentary structures of some kind. Thoroughly explore each scene to determine which sedimentary structures are present. (1) Identify them, and then (2) interpret them [i.e., (a) how they formed, (b) what they reveal about geopetal “up” direction or current direction].

Outcrop 1:

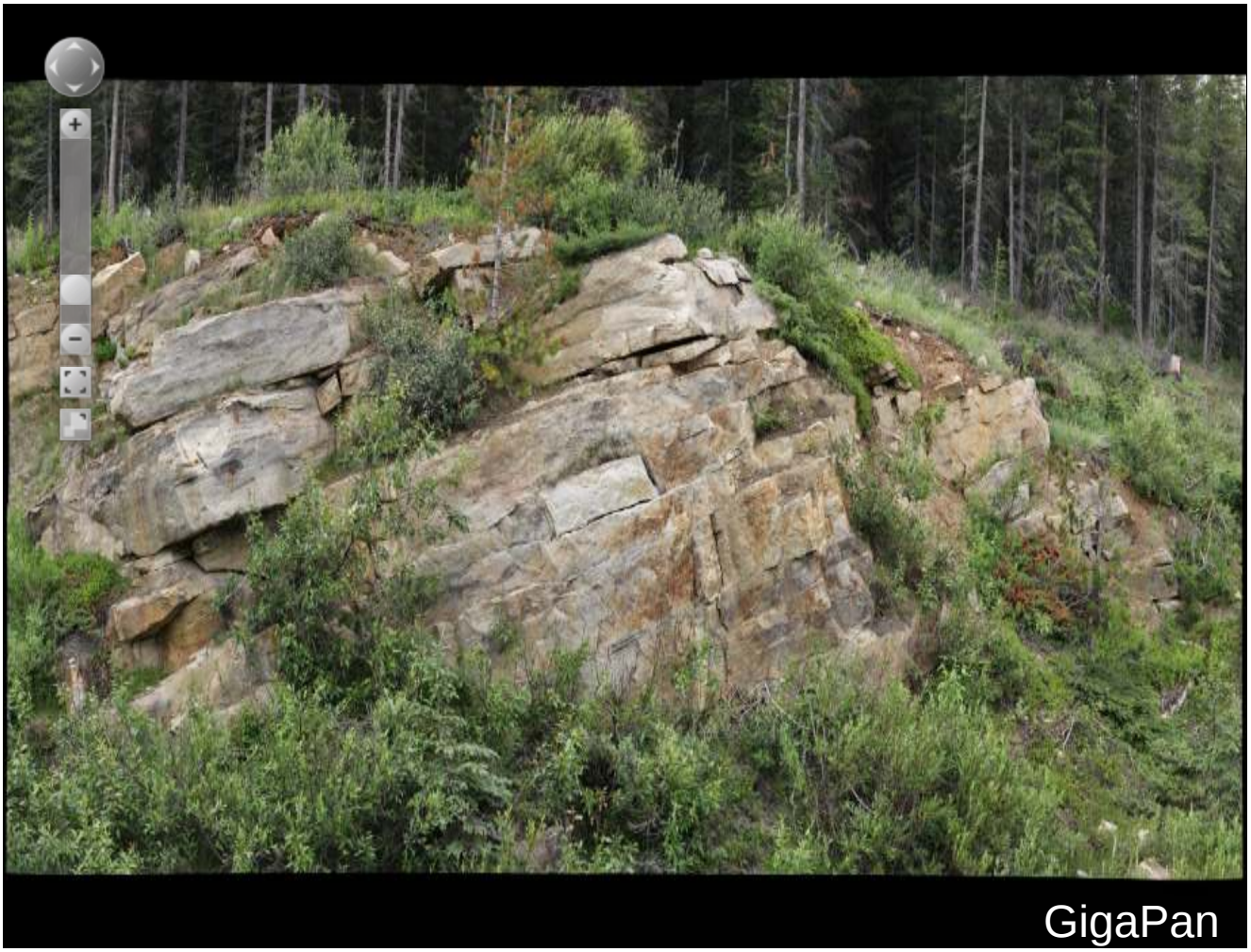


Outcrop 2:

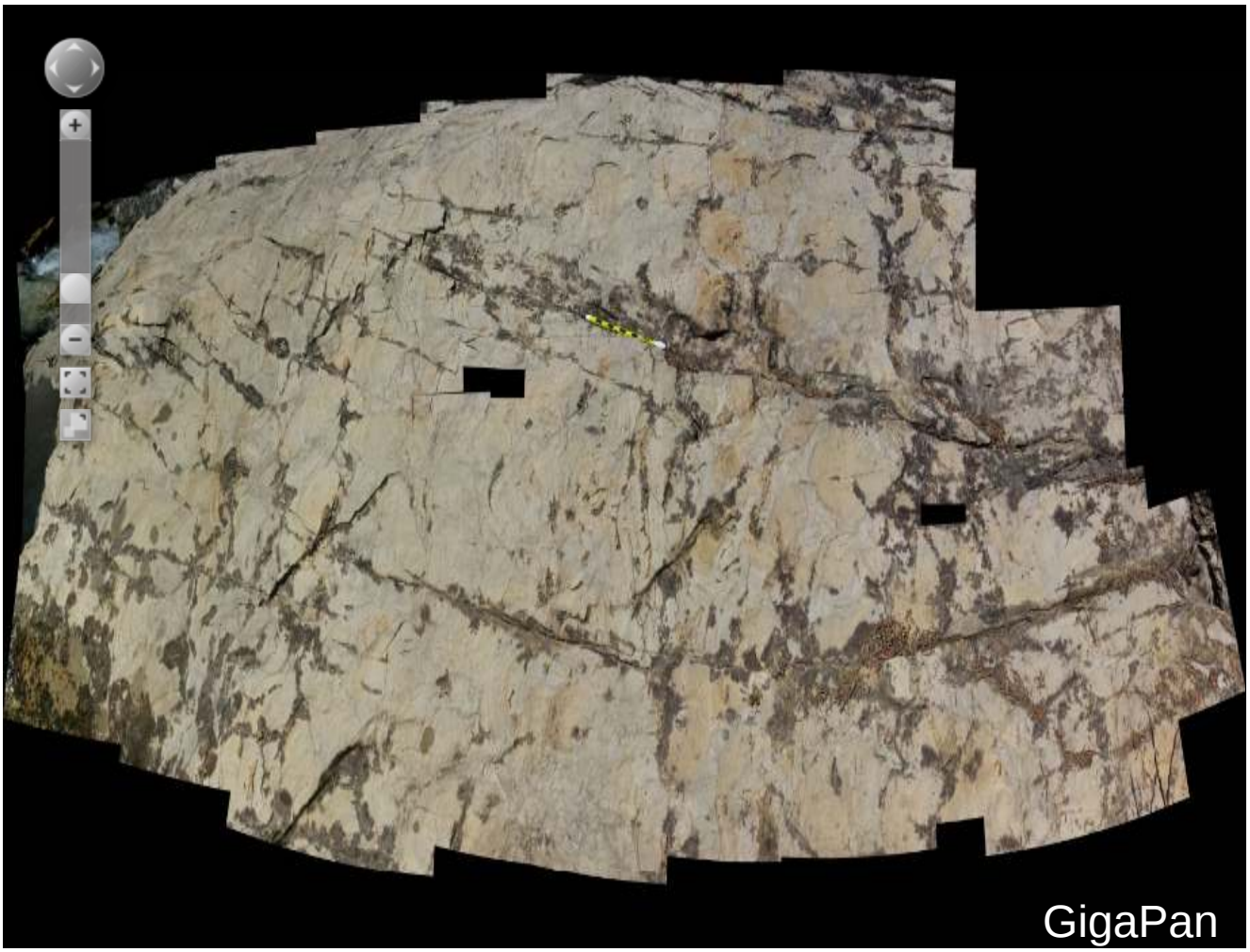


Outcrop 3:

https://geo.libretexts.org/@go/page/22485



Outcrop 4:



Outcrop 5:

https://geo.libretexts.org/@go/page/22485



Outcrop 6:



Outcrop 7:

Outcrop 8:

https://geo.libretexts.org/@go/page/22485

Outcrop 9:

https://geo.libretexts.org/@go/page/22485

PART II: HAND SAMPLES

Examine these 19 samples of rock showing various sedimentary structures and: (a) identify them and (b) interpret them.

Sample 1:

https://geo.libretexts.org/@go/page/22485

Sample 2:

https://geo.libretexts.org/@go/page/22485

Sample 3:

(Note: The little black curled lines here are the shells of small arthopods –shrimplike creatures– called ostracodes. Consider these hemispherical structures as sedimentary particles, as you would grains of sand.)

https://geo.libretexts.org/@go/page/22485

Sample 4:

https://geo.libretexts.org/@go/page/22485

Sample 5:

https://geo.libretexts.org/@go/page/22485

Sample 6:

https://geo.libretexts.org/@go/page/22485

Sample 7:

https://geo.libretexts.org/@go/page/22485

Sample 8:

https://geo.libretexts.org/@go/page/22485

Sample 9:

https://geo.libretexts.org/@go/page/22485

Sample 10:

https://geo.libretexts.org/@go/page/22485

Sample 11: (FYI, the little dash-shaped features that mainly run from left to right are gypsum casts)

https://geo.libretexts.org/@go/page/22485

Sample 12:

https://geo.libretexts.org/@go/page/22485

Sample 13:

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Sample 16:

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Sample 17:

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Sample 18:

https://geo.libretexts.org/@go/page/22485

Sample 19:

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Please submit your answers to the instructor.

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63.9: 63.9 Sediment samples

Follow this link!

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63.10: 63.10 Relative dating

Here are links to several pages of collected ***Relative Dating*** gigapixel panoramas. Each page below contains 10 images at the outcrop or hand sample scale, illustrating principles such as superposition, cross-cutting relationships, original horizontality, etc.

We envision instructors using these images to challenge their students in synchronous activities or asynchronous assignments, individually or in groups, to identify the relevant rock types and structures, and attempt to interpret the basic geologic history of each outcrop or sample.

- Relative dating collection 1
- Relative dating collection 2
- Relative dating collection 3
- Relative dating collection 4

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63.11: 63.11 Relative dating collection 1

Outcrop 1



Sample 2



Outcrop 3



Sample 4



Outcrop 5



Sample 6



Outcrop 7



Sample 8



Outcrop 9



Sample 10



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63.12: Relative dating collection 2

Outcrop 11



Sample 12



Outcrop 13



Sample 14



Outcrop 15



Sample 16



Outcrop 17



Sample 18



Outcrop 19



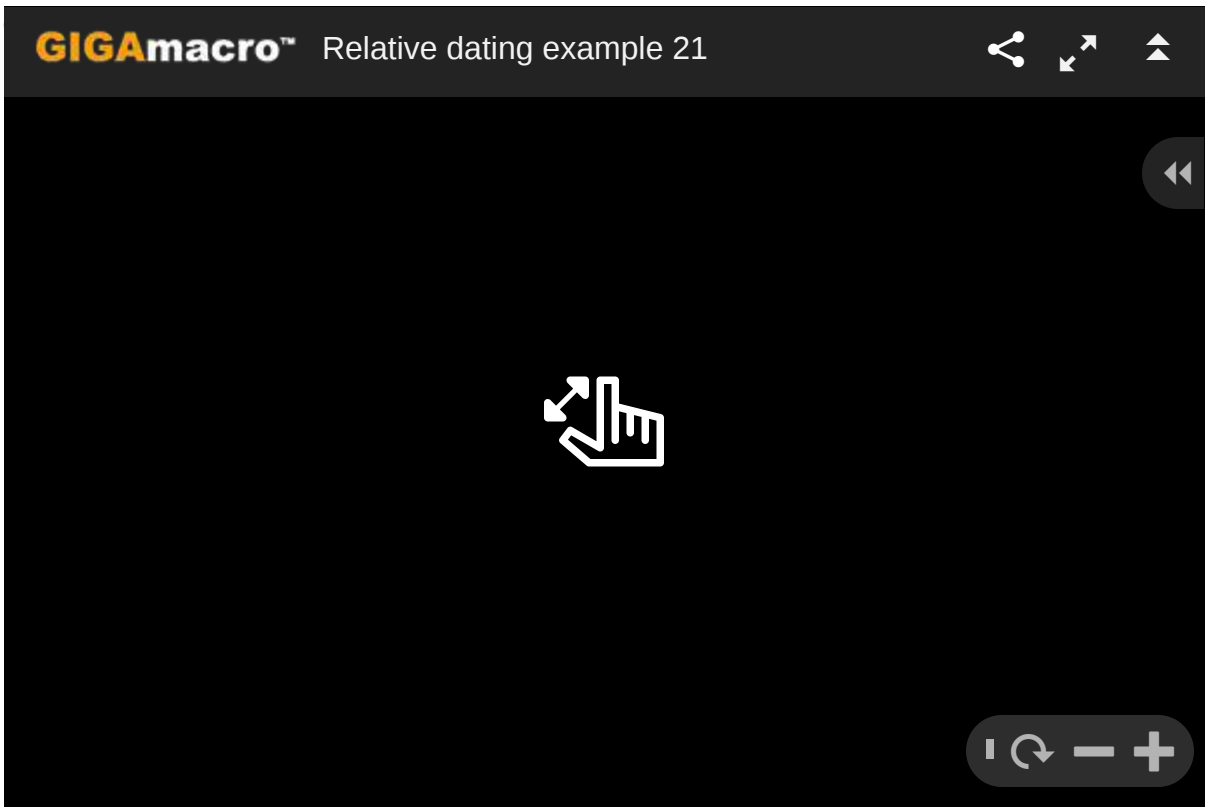
Sample 20



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63.13: Relative dating collection 3

Outcrop 21



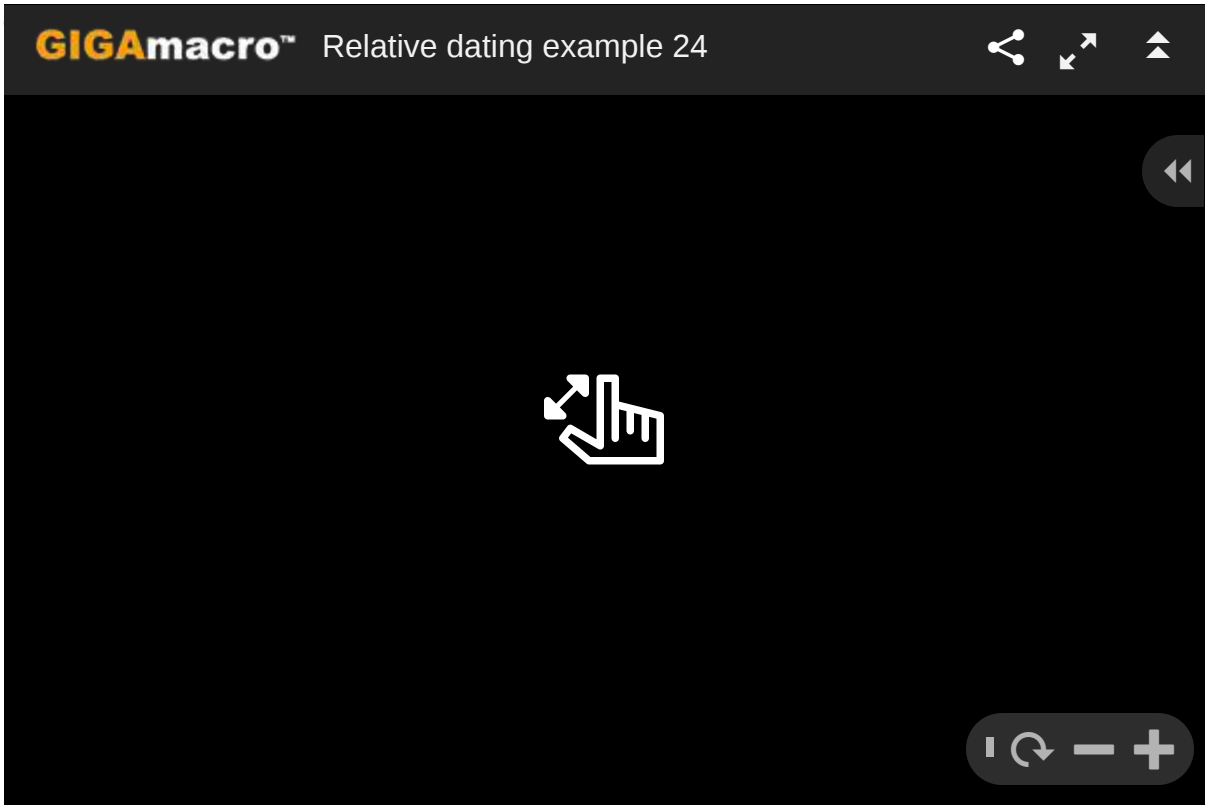
Sample 22



Outcrop 23



Sample 24



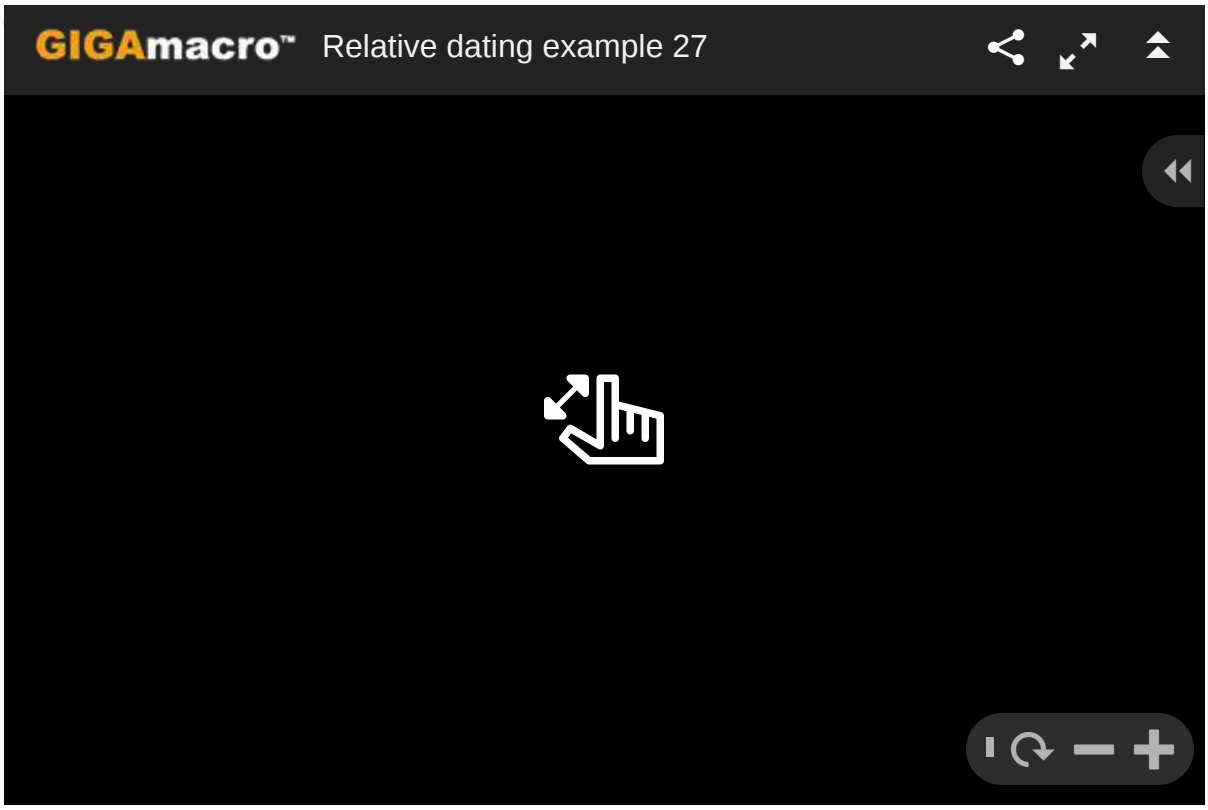
Outcrop 25



Sample 26



Outcrop 27



Sample 28



Outcrop 29



Sample 30



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63.14: Relative dating collection 4

Outcrop 31



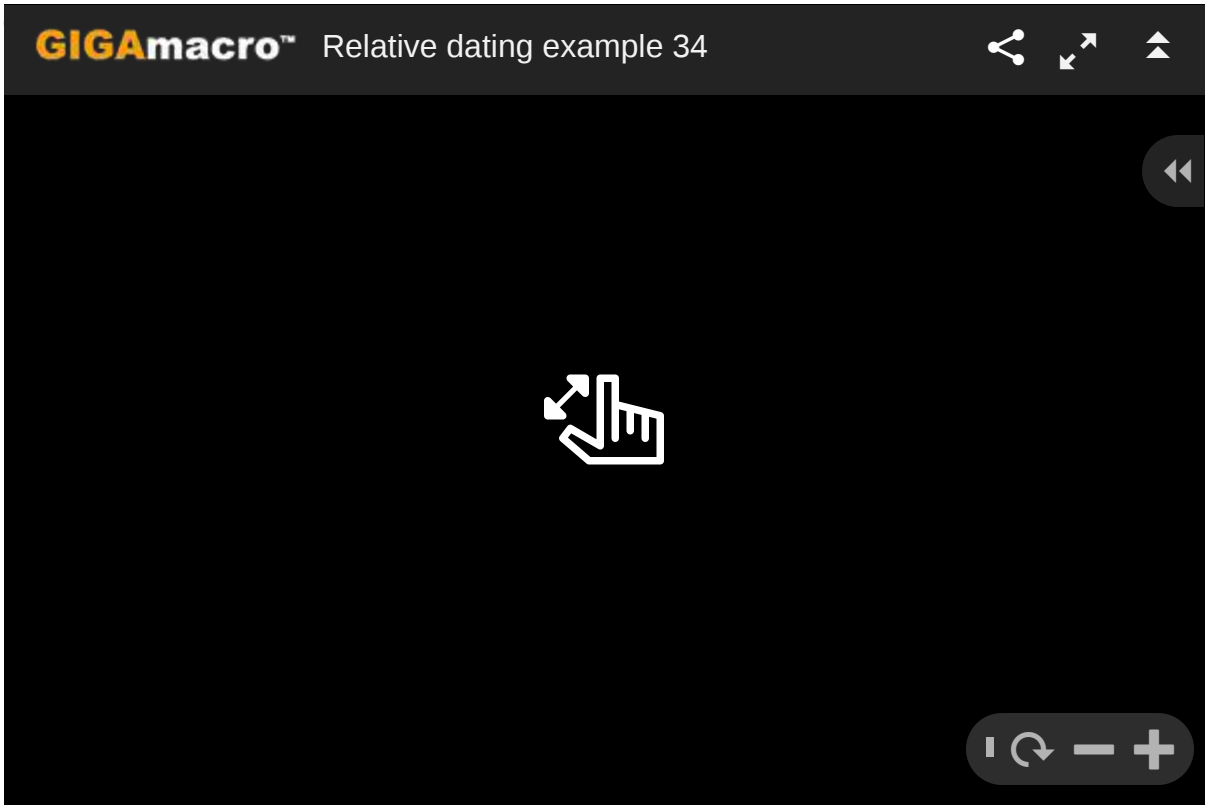
Sample 32



Outcrop 33



Sample 34



Outcrop 35



Sample 36



Outcrop 37



Sample 38



Outcrop 39



Sample 40



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63.15: Tectonic structures

A virtual sample set of structures that formed due to deformation

Examples in this collection include folds, faults, and strained samples and outcrops (bulk strain, cleavage, stylolites, mylonites, and shear zones).

Instructors can assign the identification and interpretation of some or all these structures as an in-class activity, a virtual laboratory experience, or an asynchronous homework assignment.

Structure 1



Structure 2



Structure 3



Structure 4



Structure 5



Structure 6



Structure 7



Structure 8



Structure 9



Structure 10



Structure 11



Structure 12



Structure 13



Structure 14



Structure 15



Structure 16



Structure 17



Structure 18



Structure 19



Structure 20



Structure 21



Structure 22



Structure 23



Structure 24



Structure 25



Structure 26



Structure 27



Structure 28



Structure 29



Structure 30



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stromatolites

11.3: Other 'Mass' Extinctions

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